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A model of inter-annual variability in beach levels

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ABSTRACT

Beach profile data, collected twice per year at 19 stations over a 25 km length of coastline in Tremadoc Bay, have been analysed to quantify the inter-annual variability in beach levels over a 7 year period and the results compared against the output of a numerical model. Using hourly wind data as forcing, the morphological development of northern Tremadoc Bay was simulated by wave, tidal, longshore transport, total transport and bed level change models. The modelling methodology was efficient and innovative, allowing realistic simulations of long duration with a time step of 1 h, hence capturing the high frequency nature of wind events. The model was run for each of the 7 autumn/winter periods (generally November–April) and the modelled net change in beach levels compared with the data from all 19 stations. The model results had reasonable agreement with the beach profile surveys. However, the observed magnitude of bed level change in the bay lagged the model output by 1 year, indicating that sediment processes acting over a larger area are important in a relatively localised study of inter-annual variability.

1. Introduction

Tremadoc Bay is located at the northern end of Cardigan Bay in the eastern Irish Sea (Fig. 1). During the past 20 years, several sections of the coastline in the region have been reinforced with rock armour to protect dune systems from erosion caused by winter storms. The coastal dunes act to protect the low lying adjacent land from coastal flooding. During the first few years of the 21st century, the dunes suffered damage during severe winter storms and the local authorities were required to make urgent reinforcements to the rock armour. With increased storm frequency (rather than intensity) predicted over the northeast Atlantic due to climate change (Houghton et al., 2001; Schmidt et al., 1998; WASA, 1998), the problem of coastal flooding is likely to be exacerbated in the future. The goal of the present study was therefore to provide insight to the local authorities of the processes involved in the erosion of the dunes by tidal currents and storm generated waves. This study involved an analysis of in situ (beach profile) data and the development of a morphological model.

As part of the monitoring programme of the local authority, beaches are surveyed at specified locations in northern Tremadoc Bay twice per year (once in the spring and once in the autumn). There are 19 stations distributed along the coastline from 5 km west of the Dwyryd Estuary to Abersoch (Fig. 1), a distance of approximately 25 km. The monitoring has been continuous from 1997 to 2005.

The focus of this study is Traeth Crugan, a beach to the west of Pwllheli (Fig. 1). The backshore is characterised by a single, narrow sand dune ridge. The height of the dune is generally 3–5 m and the crest width is approximately 3–4 m. Much of the dune system is poorly vegetated, hence the system is vulnerable to erosion and coastal flooding. This has been an engineering problem since 1967 when erosion was first noted, and timber groynes were placed along the beach in 1974. Since 1976, rock armour has been placed on the beach and frequently extended/strengthened. During February 2002, a large tide and prolonged southerly wind caused significant damage to the dune system to the east of the rock armour. As well as considerably extending the rock armour after this storm, the beach was nourished with $34 \times 10^3\,m^3$ of sand and gravel dredged from the harbour entrance at Pwllheli. The problem at Traeth Crugan continues, especially as climate change research generally predicts more short term storms (Alexandersson et al., 2000; WASA, 1998; Beniston et al., 2007). Since the foreshore of Traeth Crugan is designated a site of special scientific interest (SSSI), the problem, and the impact of possible engineering solutions, is sensitive to the natural environment.

Morphological changes in the nearshore zone occur (and have been studied) over a range of scales. Generally, studies of this nature focus on the detail at a single site, e.g. the detailed topographic surveys of Teignmouth over a time period of 2 months (Van Lancker et al., 2004), or cover a range of independent

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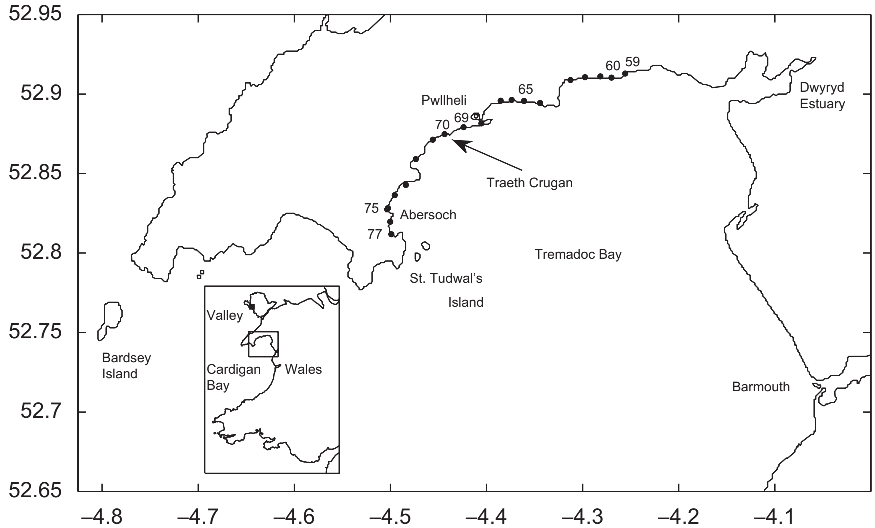

Fig. 1. Location of Tremadoc Bay and Traeth Crugan. The 19 beach profile locations are plotted as filled circles, with reference numbers shown for selected stations. Beach profile data was obtained for all stations twice per year from 1997 to 2004.

coastal locations, e.g. a range of sites along the coastline of western Ireland to study historical shoreline change over a time scale of 200 years (Cooper et al., 2004). However, studies of large scale rhythmic changes (i.e. alternate cells of erosion and deposition along a coastline) are more useful to this current work in Tremadoc Bay, e.g. the longshore transport study of Cape Lookout, North Carolina, where coastal cells of order 1 km were studied (Park and Wells, 2005).

The time scale and repeat interval of each survey is crucial to the sediment transport processes (and subsequent morphological change) being studied. Historical shoreline changes spanning centuries (e.g. Cooper et al., 2004) are pertinent to assessing change of land use or perhaps climate change, but the available data are usually not of good quality or have insufficient temporal resolution to study seasonal or yearly evolution. At the other end of the scale are detailed surveys over very short time periods (e.g. Hill et al., 2004) which suffer from uncertainty in the observed temporal trend in beach profiles (i.e. it is not known whether a short term observation of a seasonal trend is a long term trend or an anomaly). Assuming that storms are the dominant mechanism leading to morphological change along a coastline, one sampling strategy would be a survey immediately before and a survey immediately after a storm event. The data set available to this project is biannual, collected in the spring and autumn. Hence, constraints can be placed on the data by comparing beach profiles immediately before and after the dominant autumn/winter storm season.

There are two main approaches to predicting coastal morphology: data based and process based (Reeve et al., 2004). By correlating past measurements of beach profiles with environmental forcings, it is possible to use statistical methods to predict beach response to future climate forcings. Such methods can be applied through techniques such as empirical orthogonal function (EOF) analysis (Hashimoto and Uda, 1980). Deterministic process-based models can be either relatively simple (e.g. observed offshore wave climate transformed to shallow water and applied to empirical sediment transport and continuity formulae) or can incorporate numerical models to calculate the hydrodynamics and morphological response over relatively large areas. In the latter, a series of two-dimensional (2D) numerical tidal and wave models are applied over a domain. These hydrodynamic forcings are applied (at the appropriate morphological time scale) to sediment transport formulae and sediment continuity to predict bed level change over the desired time scale. This is the modelling approach developed in this paper.

Many studies exist in the literature on the use of wave models to describe the evolution of beaches. Studies at the qualitative end of the spectrum use wave models to infer the paths of sediment transport and subsequent morphological change (e.g. Cooper et al., 2004). These simple wave models are either wave refraction models (Carter et al., 1982) or model monochromatic waves (e.g. Park and Wells, 2005). Of more use to the present study in Tremadoc Bay are spectral wave models such as SWAN (Booij et al., 1999) which include the natural statistical distribution of wave height, period and direction (SWAN is described further in Section 3.1). To make a quantitative morphological study, a model for sediment transport and bed level changes is required. In a few studies (e.g. Ranasinghe et al., 2004), total load sediment transport formulae are used such as Bailard (1981). However, it is more common for a longshore transport formulation to be used to predict sediment transport in the nearshore zone (e.g. Kamphuis, 1991). To capture high frequency events, a relatively short time step (e.g. 1 h) should be used for all stages of modelling, rather than data reduction (e.g. Latteux, 1995; Jones et al., 2007).

In this study, a morphological model is developed, consisting of wave, tidal, longshore transport, total transport and bed level change modules. Efficient methods are applied to the tidal and wave models to allow multiple morphological simulations of time scales ~6 months at high temporal (hourly) and spatial (100 m) resolution over a relatively large geographic area (~220 km^2) without significant data reduction. The aim of the study is to determine whether such a model can successfully reproduce bulk features observed in beach profile data. In addition, we will be determining whether such a model can be used as a tool to study inter-annual variability of beach levels due to inter-annual variability of wind forcing. This has important implications with climate change since storm frequency is likely to increase in the northeast Atlantic (WASA, 1998).

The data sources are described in Section 2, consisting of the hydrography of the study area, wind and beach profile data. A morphological model is developed in Section 3 consisting of wave, tidal, sediment transport (longshore and total transport) and bed level change models. The model is applied and compared with 7 years of beach profile data in Section 4, and sources of model error are discussed in Section 5.

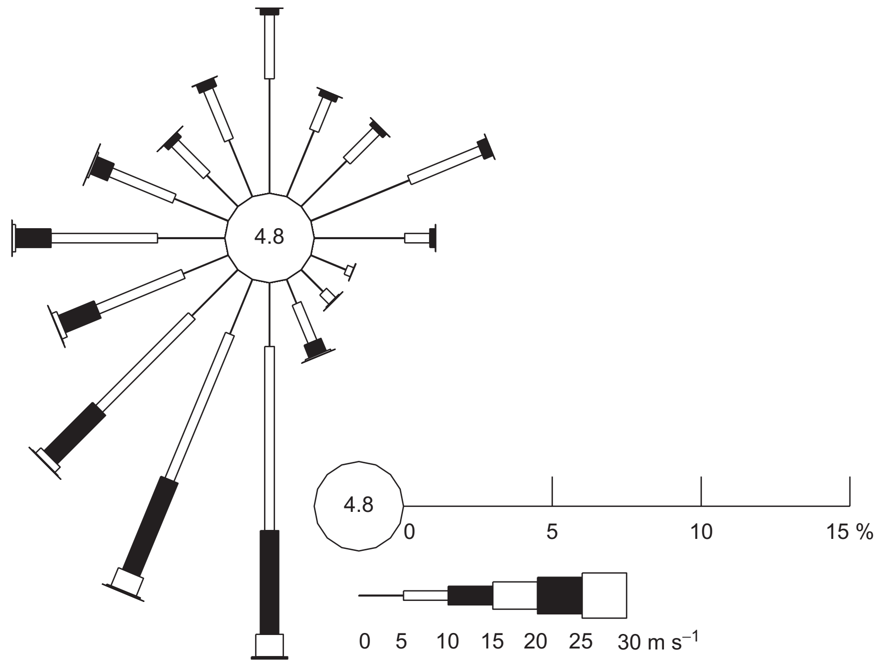

Fig. 2. Wind rose for Valley meteorological station, 1992–2004.

2. Data

2.1. Hydrography of the study area

Tremadoc Bay is a shallow water bay (mean depth of order 10–20 m) with semi-diurnal tides of range 4.5 m (spring) and 1.5 m (neap). Tidal currents in the bay are variable with speeds of order $1-2\,\mathrm{m\,s^{-1}}$ in the region around St. Tudwal's Island and of order $0.1\,\mathrm{m\,s^{-1}}$ in the northeast of the bay (Neill et al., 2007). This latter region stratifies during the summer. Numerous sandy beaches are distributed along Tremadoc Bay, interspersed by rocky promontories. Many of the beaches are popular for tourism and leisure activities, particularly those close to Pwllheli. The town of Pwllheli has a marina, and the entrance to the harbour has to be dredged annually due to sediment accumulation. This dredged material is stockpiled and used for beach nourishment as required. The wave climate in Tremadoc Bay is generally from the southwest, relating to the dominant wind direction which is also southwest (Section 2.2).

2.2. Wind data

Wind data at Valley (Fig. 1) is representative of conditions in Tremadoc Bay as confirmed by correlations at several (coarse temporal resolution) meteorological stations in the bay. Valley was used as the data source for model forcing (Section 3) since a long term data set at high resolution (hourly) was available at this, the closest synoptic meteorological station to Tremadoc Bay. Valley is approximately 40 km north of Tremadoc Bay, but is central to the Irish Sea, making it a reasonable location to represent wind conditions over a wider region. Since the wave modelling methodology assumes a spatially uniform wind field (Section 3.1), such a central location is desired. Jones (1999) demonstrated that such an assumption is valid for a region such as the central Irish Sea which has a length scale of around 300 km. Fig. 2 shows the wind rose for Valley from 1992 to 2004, a time period covering the beach profile data set. The most frequent wind direction is SSW (210°, $6.3\,\mathrm{m\,s^{-1}}$), but the highest magnitude wind events tend to be southwesterly. Since the meteorological station at Valley is exposed to the Irish Sea to the southwest, it is an ideal location in which to measure these major southwesterly wind events with minimal topographic effects.

Table 1
Details of beach profile surveys from autumn in one year to spring in the following year

Autumn survey date	Spring survey date	Length (days)
12/11/1997	08/04/1998	148
01/12/1998	27/05/1999	182
08/10/1999	04/05/2000	211
12/10/2000	05/06/2001	236
16/11/2001	25/04/2002	162
18/11/2002	29/04/2003	162
04/11/2003	04/05/2004	182

2.3. Beach profile data

Beach profile data for years 1997–2004 for 19 stations in Tremadoc Bay have been provided by Gwynedd Council (Fig. 1, Table 1). Each profile was surveyed twice per year. With such a large quantity of raw data, presentation had to be considerably condensed. Hence, two adjacent profiles have been selected for detailed plotting[1]: a profile with a dune system (station 69) and a profile with rock armour (station 70). At each profile location, a fixed land reference point was used as a starting point, and horizontal and vertical measurements were taken along the beach profile to a horizontal distance which extended to mean sea level (MSL). Bearings were strictly defined so that each beach profile was normal to the coastline. The positions of the stations and bearings were accurately duplicated at each repeat interval of the surveys (~6 months). Beach surveys taken during the spring were plotted since this period indicates the annual trend as well as showing a snapshot of the beach profiles immediately after the autumn/winter storm period. Autumn surveys are less useful in

[1] Although all data were used for bulk comparison, see later.

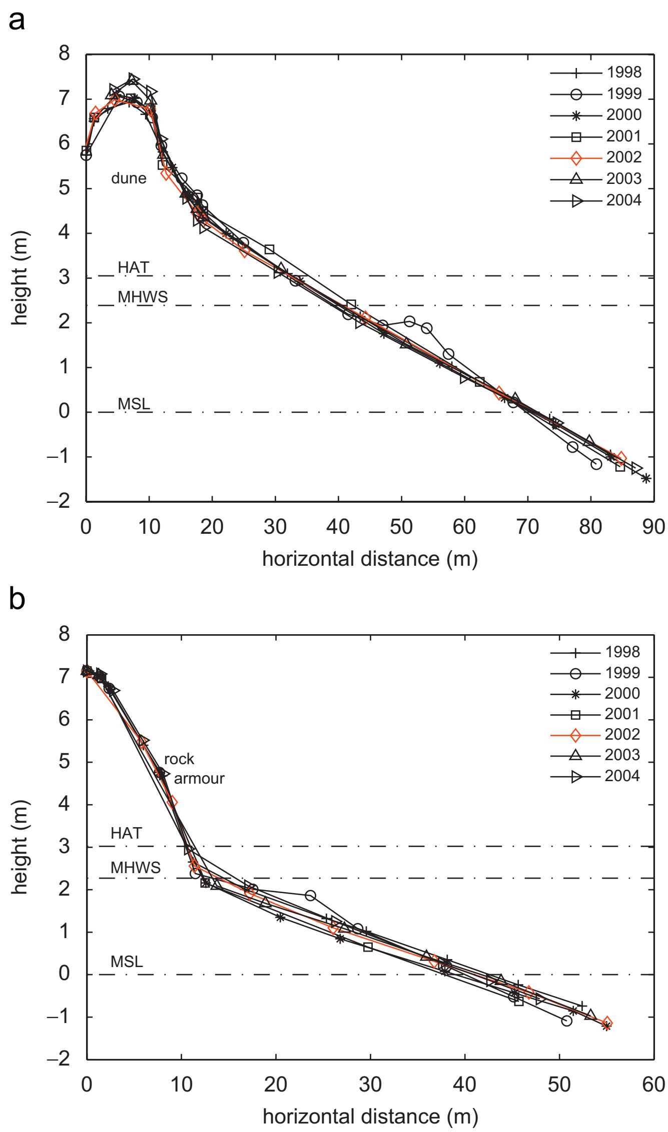

Fig. 3. Annual (spring) beach profile surveys for years 1998–2004. The levels of mean sea level (MSL), mean high water springs (MHWS) and highest astronomical tide (HAT) are plotted as dot-dashed lines: (a) station 69 (dune system) and (b) station 70 (rock armour).

assessing the impact of storms since they include the summer beach recharge due to swell waves over a period when storms are rare. Since the morphological model does not include swell waves (Section 3), inter-annual variability of beach levels is here taken to be the change in bed level occurring over the autumn/winter storm period.

Profiles from 1998 to 2004 of a dune beach (station 69) are plotted in Fig. 3a and profiles of a rock armour beach (station 70) are plotted in Fig. 3b. In the case of the dune beach, the crest of the dune grew in 2003 and continued to grow until 2004. In 2002 the dune was undercut on its beach face. This reflects the storm damage which is known to have occurred in February 2002 (Section 1). Winds were either from the SSW (65%) or south (35%) during this storm with a peak wind speed of $24.2\,\mathrm{m\,s^{-1}}$ (meteorological data at Valley). In the period following this storm, the beach was nourished with $34 \times 10^3\,\mathrm{m}^3$ of dredged material from Pwllheli marina. Note that in the case of both locations (station 69 and station 70), a beach berm has been captured between MSL and mean high water spring (MHWS) due to the timing of the spring 1999 survey in relation to the preceding wave

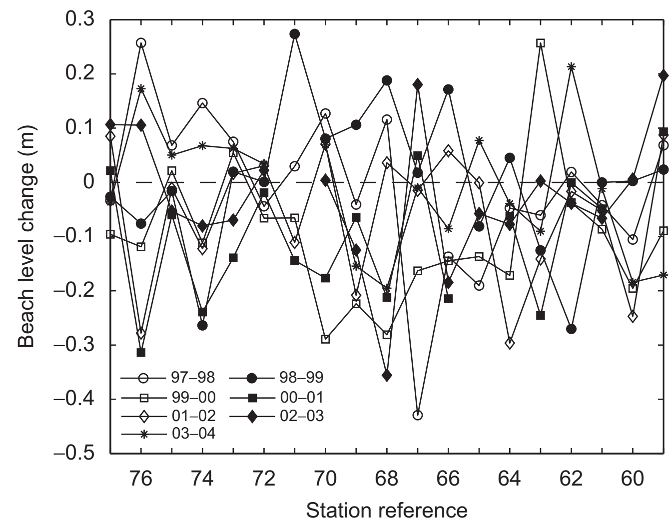

Fig. 4. Difference between autumn and spring beach levels calculated between mean sea level and highest astronomical tide for all beach profile data.

activity. Such features generally form due to swell waves and are destroyed by storm waves. This indicates, therefore, that there was either minimal storm activity or considerable swell activity in the early part of 1999.

For all stations, the mean beach level between MSL and highest astronomical tide (HAT) was calculated for each beach profile survey (autumn and spring). Since the horizontal resolution of raw beach profile data varied between each station and for each survey date, it was necessary to interpolate this raw data to a common sampling interval. A 1 m linear interpolation was found to be a suitable method to capture details over the range of profiles, and to accurately define the intercept of each profile with both MSL and HAT. For each location and survey date, the 1 m re-sampled data was averaged for all data points bounded by these two datums. This produced a single value of beach level for each profile at the time of each survey. By subtracting the autumn beach level in year n from the spring beach level in year $n+1$, this gives an objective measure of the change in beach level due to the autumn/winter storm season at each of the 19 locations. This calculation was made for all 7 seasons, enabling an inter-annual comparison to be made (Fig. 4). Alternative methods such as the momentary coastline (MCL) (Van Koningsveld and Mulder, 2004) could have been used to calculate the inter-annual variability, but the change in beach level provides a measurement which can be compared directly with the model output (Section 3).

3. Morphological model

The morphological model consists of wave, tidal, sediment transport (longshore transport and total transport) and bed level change modules.

3.1. Wave model (SWAN)

A wave model was used to determine the effect of wind speed and direction on wave characteristics, primarily significant wave height, wave period and wave direction. SWAN (Simulating Waves Nearshore) is an Eulerian formulation of the discrete wave action balance equation (Booij et al., 1999). The model is spectral discrete in frequencies and directions and the kinematic behaviour of the waves is described with the linear theory of gravity waves. The deep water physics of SWAN are taken from the WAM model (Komen et al., 1994). SWAN has two modes: stationary and non-stationary. Non-stationary mode is time dependent, hence the evolution of the wave field for a storm can be modelled realistically, using boundary conditions of time-varying wind speed and direction. This is, however, computationally expensive since a time step $\ll$ wind forcing time step is required for stability depending on the spatial cell size (Elliott and Neill, 2007). Since a long time series (>1 year) simulation was required for this study, a more economical method was used. This involved running SWAN in stationary mode.

In stationary mode, the evolution of the action density N is governed by the time-independent wave action balance equation (Booij et al., 1999)

$$\frac{\partial}{\partial x} c_x N + \frac{\partial}{\partial y} c_y N + \frac{\partial}{\partial \sigma} c_\sigma N + \frac{\partial}{\partial \theta} c_\theta N = \frac{S}{\sigma} \tag{1}$$

where c_x and c_y are the propagation velocities in the x and y directions, σ is frequency, θ is wave direction and S represents the source terms, i.e. generation, dissipation and non-linear wave–-wave interactions.

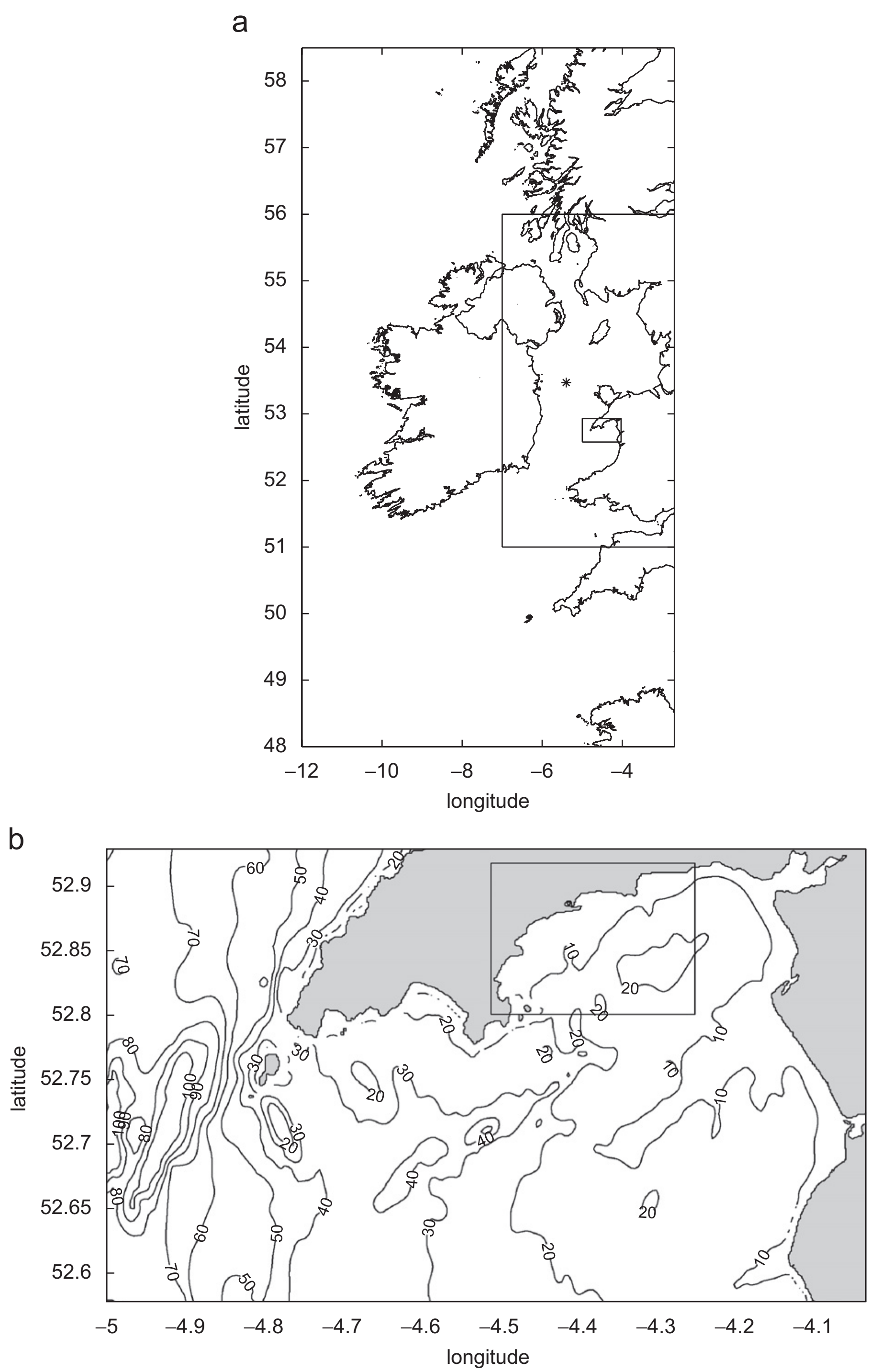

Fig. 5. Wave model nests. The asterisk shows the position of the M2 wave buoy (used for validation of the wave model): (a) Shelf Sea, Irish Sea and Tremadoc Bay nests. (b) Tremadoc Bay and Traeth Crugan nests. Contours are of bathymetry (in metres) relative to mean sea level.

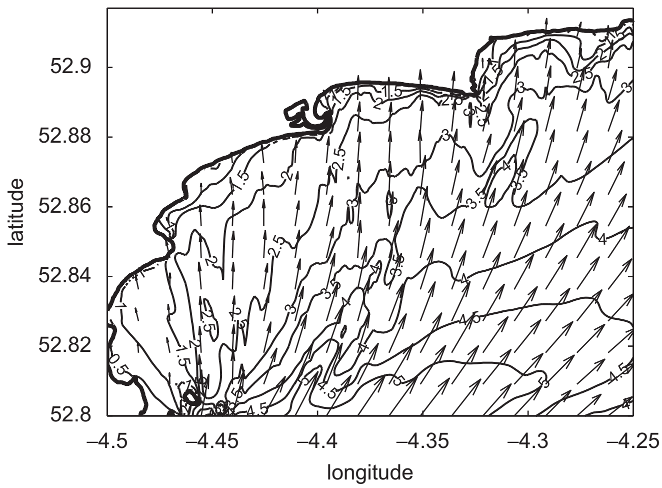

Fig. 6. SWAN output from Traeth Crugan model of H_s contours (m) and vectors of direction θ_p with magnitude H_s for SW wind with speed 24 m s^{-1} (~47 kn). Every tenth modelled vector is shown for clarity.

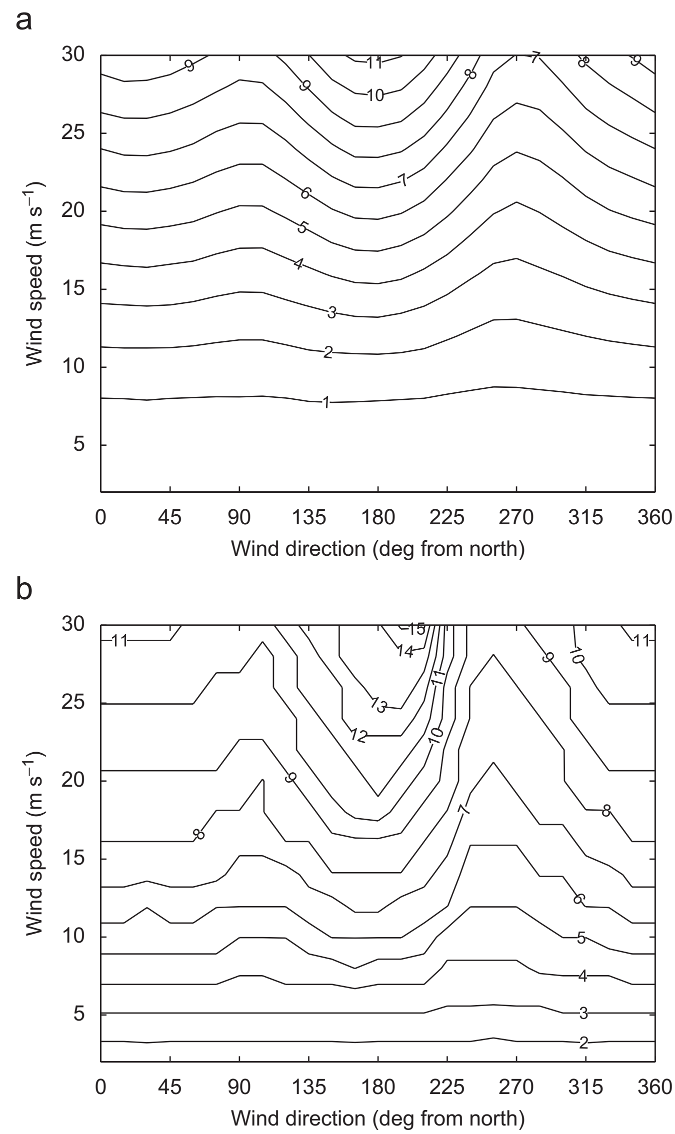

Fig. 7. Graphical representation of significant wave height and peak wave period matrices at M2 buoy (for location, see Fig. 5a) based on SWAN stationary wave model of a range of wind speeds and directions. (a) H_s (m) and (b) T_p (s).

From the 12 year analysis of wind data at Valley on Anglesey (Fig. 2), a suitable range of discrete wind direction and speed bins was selected using $\theta = 0, 15, \ldots, 345^\circ$ and $W_r = 2, 4, \ldots, 30\,\mathrm{m\,s^{-1}}$, respectively (i.e. $24 \times 15 = 360$ simulations). SWAN was run in stationary mode by applying each of these wind vectors as a constant over the entire model domain. An outer shelf model was run initially at a resolution of 12 km (Fig. 5a) with a high resolution (1.85 km) nested model of the Irish Sea run with boundary conditions of the action density spectrum extracted from the outer grid. Bathymetry data for the shelf model was provided by the Proudman Oceanographic Laboratory (Liverpool), and data for the Irish Sea from Brown et al. (1999). Within this Irish Sea grid was nested a 300 m resolution model of Tremadoc Bay, and nested within this was a 100 m resolution model of Traeth Crugan (Fig. 5b). Bathymetry for the Tremadoc Bay and Traeth Crugan models was digitised from Admiralty Charts 1971 and 1512, based on survey data collected between 1961 and 1983, hence a potential source of error. A typical output of H_s and θ_p is shown in Fig. 6 for a SW wind of speed 24 m s^{-1} (~47 kn). For each cell of the Traeth Crugan grid, a matrix of H_s, T_p, θ_p and U_{rms} (output directly from the spectral model rather than calculated using linear theory) was produced for the range of wind speeds and directions. For validation, matrices of H_s and T_p at the position of the M2 buoy (Fig. 5a) are shown in Fig. 7. From this matrix, the dominant wind direction for producing high significant wave heights at the M2 buoy is southerly (180°), relating to the longest fetch. The method has been validated with hourly data of H_s and T_p over a period of 3 months in 2005 (Fig. 8) using hourly wind data at Valley meteorological station applied to the lookup tables. The agreement is excellent for such a simple statistical method, but it should be remembered that a third generation wave model was used at high resolution, hence considerable computational effort was required to compute the lookup tables.

3.2. Tidal model (POLCOMS)

POLCOMS is the Proudman Oceanographic Laboratory Coastal Ocean Modelling System (Holt and James, 2001). POLCOMS is three dimensional (using σ coordinates in the vertical) and is formulated in spherical coordinates. For turbulence closure, the Mellor–Yamada–Galperin level 2.5 scheme is used (Mellor and Yamada, 1974; Galperin et al., 1988). Boundary conditions required for POLCOMS are elevation and the normal component of velocity. POLCOMS was applied in this study by first running a 12 km outer grid of the northwest European continental shelf (12°W–12°E and 48°N–62°N) with astronomical boundary conditions and bathymetry provided by the Proudman Oceanographic Laboratory. An hourly time series of elevation and velocity was stored at the boundary locations of the first inner nested region: the Irish Sea (Fig. 5a) (grid details as in Section 3.1). Harmonic analysis was performed on each time series to create an independent high-resolution Irish Sea model with no feedback to the outer nest. This process was repeated on a second nested region of Tremadoc Bay (Fig. 5b) and finally on an inner nested region of Traeth Crugan (Fig. 5b), both grids as described in Section 3.1. For this morphological study, the two dominant constituents at Pwllheli were used to force the model: M_2 and S_2. The modelled astronomical tide was validated with data from the UK Tide Gauge Network (Table 2). M_2 current ellipses for Traeth Crugan are plotted in Fig. 9. Currents in the region of Traeth Crugan are generally low (of order 0.2 m s^{-1}). To the south of Traeth Crugan, the currents are rectilinear, with the semimajor axis aligned approximately north/south. Closer to the coastline, the character of the tidal currents is more rotary, particularly to the southeast of Pwllheli harbour. The amplitudes and phases of the astronomical

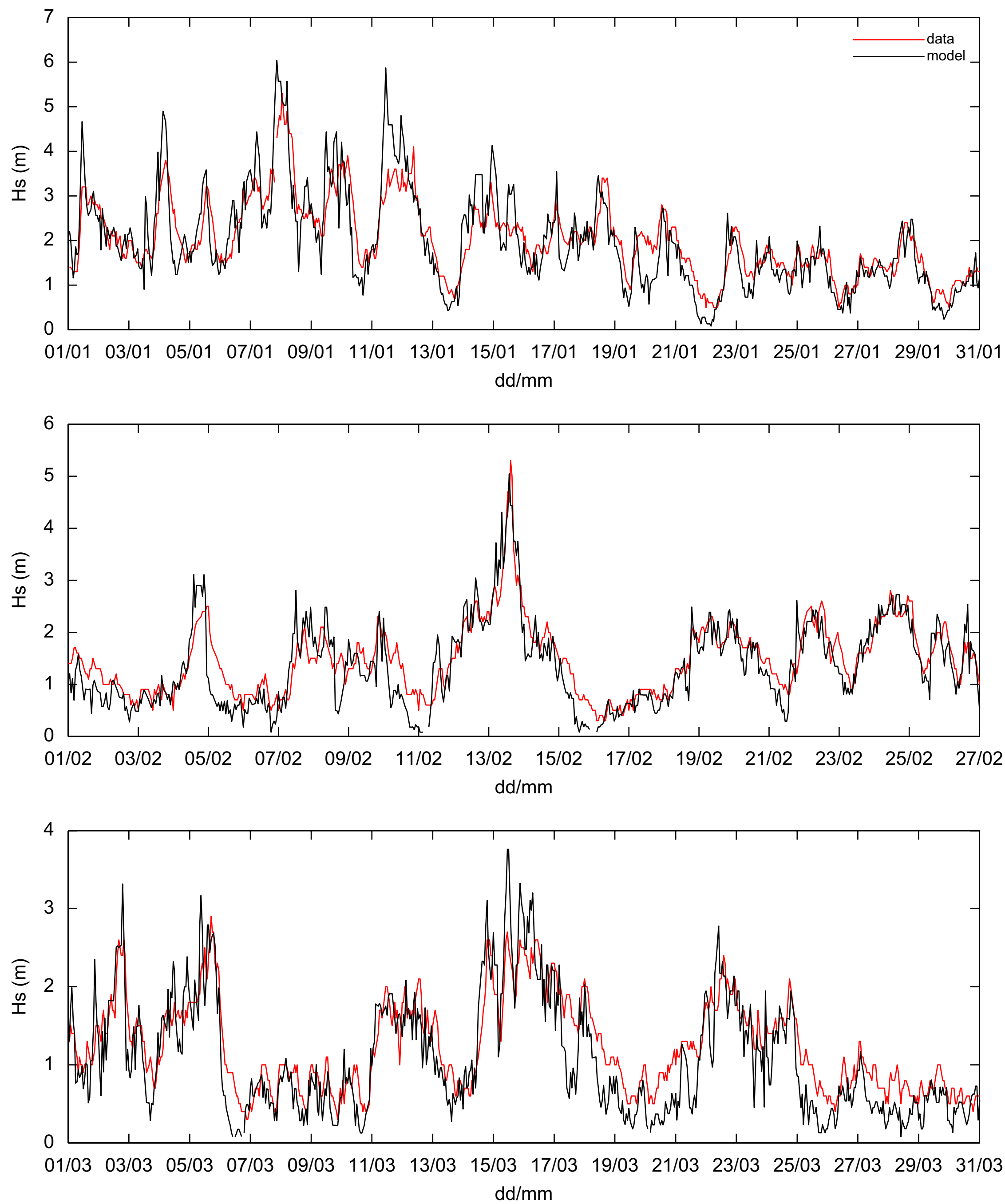

Fig. 8. Agreement between H_s measured at M2 wave buoy and H_s modelled by SWAN, January–March 2005.

Table 2
POLCOMS modelled amplitude ζ (m) and phase g (deg) compared with values at tidal stations around Tremadoc Bay for M_2 and S_2 constituents

Station	M_2				S_2			
	Data		Model		Data		Model	
	ζ	g	ζ	g	ζ	g	ζ	g
Barmouth	1.47	244	1.52	243	0.53	283	0.59	287
Pwllheli	1.47	241	1.49	247	0.58	279	0.57	291

constituents at each cell of the Traeth Crugan grid were used with tidal prediction (of velocity components) to provide an economical method (since it is not restricted by the length of timestep) for tidal input to the total transport model (Section 3.4).

Wind-driven flow was neglected by the model, justified as follows. The mean wind speed over a typical simulated autumn/winter period (2001–2002) was $7.2\,\mathrm{m\,s^{-1}}$. Assuming that the surface current (U_s) has a speed of approximately 3% of the wind speed (Bowden, 1983), $U_s = 0.2\,\mathrm{m\,s^{-1}}$. A roughness length z_0 can be taken as a function of wind speed, and the depth at which wind-driven flow is assumed to be zero (z_c) can be scaled on

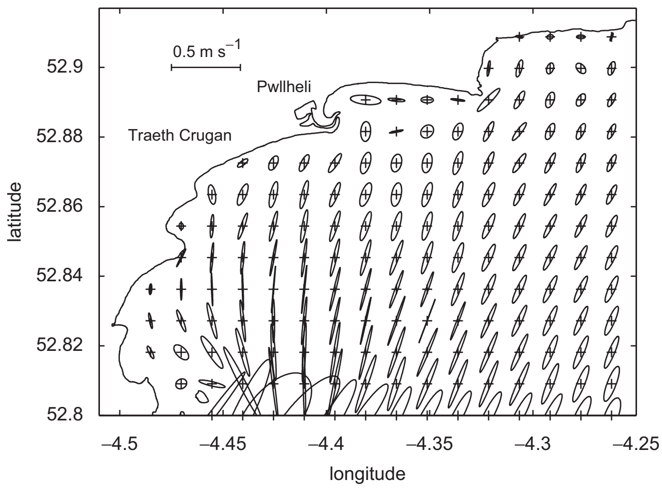

Fig. 9. POLCOMS modelled M_2 current ellipses for Traeth Crugan. Every tenth modelled ellipse is plotted for clarity.

wavelength. Therefore, the vertical profile of wind-driven currents can be calculated using

$$u(z) = U_s\left(1 - \frac{\log(z/z_0)}{\log(z_c/z_0)}\right) \tag{2}$$

The mean water depth in the model domain is 12.8 m. The mean wavelength over the domain during the modelled period was 17 m, hence the wind-driven vertical profile can be calculated and reveals that u reduces to of order $0.01\,\mathrm{m\,s^{-1}}$ in the bottom 1 m of the water column. The mean tidal velocity and root-mean-squared wave orbital velocity over the same modelled period were 0.09 and $0.05\,\mathrm{m\,s^{-1}}$, respectively, averaged over the entire domain. Hence, the wind-induced currents at the bed are an order of magnitude less than either the mean tidal currents or the mean wave orbital velocity. Since bed shear stress is a function of velocity squared, bed stress due to wind-driven currents is correspondingly two orders of magnitude less than either tide- or wave-induced bed stress.

3.3. *Longshore transport model*

The CERC formula (USACE, 2001) is commonly used for the estimation of longshore transport in practical or engineering applications (e.g. Miller, 1999). However, several more complicated longshore transport formulae which include additional parameters have been developed and compared to each other via experimental or field data (Van Wellen et al., 2000; Bayram et al., 2001; Soulsby and Damgaard, 2005). Kamphuis (1991) developed a longshore transport equation based on physical model experiments and dimensional analysis. He found that using controlled model test results in the laboratory may yield more accurate results compared with field studies because of uncertainties associated with field measurements and subjectivity of interpreting field measurement results (USACE, 2004). In the present study, therefore, the method of Kamphuis (1991) was used to estimate the longshore transport. The longshore transport rate can be estimated as

$$Q_{lst} = 2.27H_{sb}^2T_p^{1.5}m_b^{0.75}d_{50}^{-0.25}\sin^{0.6}(2\theta_b) \tag{3}$$

where Q_{lst} is the longshore transport rate (in $\mathrm{kg\,s^{-1}\,m^{-1}}$), H_{sb} is the significant wave height at breaking, T_p is the peak wave period, m_b is the beach slope from the breaker line to the shoreline, d_{50} is the median grain size and θ_b is the wave angle at breaking (the angle which the wave crest forms with the coastline). The application of Eq. (3) is not ideal for our model resolution of 100 m, since we have not fully resolved wave characteristics at breaking. However, it provides a reasonable compromise between the relatively high resolution output (spatial and temporal) from the wave model and the desired accuracy of sediment transport in the surf zone. It is anticipated that future developments of the modelling methodology will incorporate higher resolution of processes in the inner nearshore zone.

Eq. (3) was implemented by applying a time series of hourly wind speed and direction values to the lookup tables described in Section 3.1. Each coastal grid cell at the resolution of the wave model (100 m) was flagged and properties of H_s, T_p and θ_p extracted. Rather than using the discrete cells to derive the angle of the coastline relative to the incoming wave, a high resolution vector (rather than raster) coastline was used to calculate the angle of the coastline (by using the position in the coastline vector nearest to the centre of each discrete coastline cell). This is important in longshore transport modelling since Eq. (3) produces a scalar quantity, and the direction of transport along the coast is sensitive to the orientation of the coastline. Values of sediment size were parameterised from a particle size analysis of shoreline data collected in 1997 and 2003 in Tremadoc Bay (Gwynedd Council, 2004). From this data, the mean value at MSL was $d_{50} = 0.4$ mm.

Using typical measured and modelled values, the width of the surf zone was calculated using the following empirical formulae for breaking waves in shoaling water (USACE, 2001)

$$\frac{H_b}{H_0'} = \frac{0.563}{(H_0'/L_0)^{0.2}}, \quad \frac{d_b}{H_b} = 1.28 \tag{4}$$

where H_b is the breaker height, d_b is the breaking depth, H_0' is the unrefracted deepwater wave height and L_0 is the deepwater wave length. For example, with a beach slope of 0.08 (Fig. 3) and typical storm values of $H_s = 3$ m and $T_p = 7$ s, the width of the surf zone is ~50 m. Since the model is fixed grid (resolution 100 m) at MSL this equates to the centre of the coastal grid cells. Hence, application of Eq. (3) at grid cells adjacent to the coastline is justified.

3.4. *Total transport model*

Numerous non-cohesive sediment transport models exist in the literature and these are often compared against each another (e.g. Davies and Villaret, 2002). In this study, sediment transport is calculated as a total load transport by waves plus currents using the Soulsby-Van Rijn formula (1997). It is based on the model of Van Rijn (1989) with curve fitting over a range of wave and current conditions by Soulsby (1997). This formulation contains a large enhancement of transport rate due to wave action. The wave action has an important contribution to the suspended load when considering total transport in shallow waters. The formula is valid for non-cohesive sediments in the range of 0.1–2.0 mm. Total sediment transport rate (in $\mathrm{kg\,s^{-1}\,m^{-1}}$) is

$$Q_t = \rho_s A_s \overline{U}\left[\left(\overline{U}^2 + \frac{0.018}{C_D}U_{rms}^2\right)^{1/2} - \bar{U}_{cr}\right]^{2.4}(1 - 1.6\tan\beta_b) \tag{5}$$

where $A_s = A_{sb} + A_{ss}$ and

$$A_{sb} = \frac{0.005h(d_{50}/h)^{1.2}}{[(s-1)gd_{50}]^{1.2}} \tag{6}$$

$$A_{ss} = \frac{0.012 d_{50} D_*^{-0.6}}{[(s-1) g d_{50}]^{1.2}} \tag{7}$$

and

$$\overline{U}_{cr} = 0.19 (d_{50})^{0.1} \log \frac{4h}{d_{90}} \quad \text{for} \quad 100 \leqslant d_{50} \leqslant 500\,\mu\text{m} \tag{8}$$

or

$$\overline{U}_{cr} = 8.50 (d_{50})^{0.6} \log \frac{4h}{d_{90}} \quad \text{for} \quad 500 \leqslant d_{50} \leqslant 2000\,\mu\text{m} \tag{9}$$

where $\overline{U}$ is the depth-averaged current velocity, U_{rms} is the root-mean-square wave orbital velocity, C_D is the drag coefficient due to current alone, $\overline{U}_{cr}$ is the threshold current velocity, ρ_s is the density of sediment, β_b is the bed slope, h is the water depth, s is the relative density of sediment and D_* is the dimensionless grain size. For the typical water depths in Tremadoc Bay (12.8 m), the threshold current velocity was calculated as $\sim 0.4\,\text{m}\,\text{s}^{-1}$ using $d_{50} = 0.4$ mm. Importantly, the direction of total sediment transport is determined by the tidal flow and not the wave direction. The relatively simple and easy to apply Soulsby-Van Rijn formula enables reasonable predictions to be made of sediment transport in combined wave–current conditions during storm and mean events as demonstrated by in situ studies (e.g. Williams and Rose, 2001). Eq. (5) was implemented in this study by using a lookup table for U_{rms} (Section 3.1) and tidal current prediction (Section 3.2) at each grid cell in the model except for the grid cells adjacent to the coastline where the sediment flux was assumed to be dominated by longshore transport (Section 3.3). This assumption is justified since tidal currents in the cells adjacent the coastline were small (Fig. 9). In addition, the modelled magnitude of longshore transport was found to be $\gg$ the magnitude of total transport close to the coastline (Section 4).

3.5. Bed level change model

Nicholson et al. (1997) have stated that, when considering long-term morphodynamics, it is important to include the interaction between the hydrodynamic and the morphodynamic components of the scheme. In the present paper, the time scale is of order 6 months, but the expected ratio of bed change to mean water depth was found to be small (of order 0.01), hence bathymetry changes will be negligible and this feedback has not been included.

Assuming that the sediment content of the water column does not change significantly over time, morphological development can be modelled in two dimensions using (e.g. Van der Molen et al., 2004)

$$\frac{\partial z}{\partial t} = -\frac{1}{1-p}\left\{\frac{\partial q_x}{\partial x} + \frac{\partial q_y}{\partial y}\right\} \tag{10}$$

where z is the bed level, p is the bed porosity and q_i is transport of sediment in the i direction (from Eqs. (3) and (5)). This equation, known as the Exner equation, was solved using the Lax-Friedrichs finite differencing scheme which has first order accuracy (Chung, 2003).

4. Model results

Hourly wind data at Valley was used in conjunction with the lookup tables generated by the wave model (Section 3.1) to provide hourly hindcasts of wave conditions for each cell of the Traeth Crugan (100 m) grid. Each of the grid cells adjacent to the coastline was flagged and the longshore transport model applied at these locations (Section 3.3). At all other grid cells, the total transport model was applied (Section 3.4) with hydrodynamic input from the wave and tidal models. Finally, the Exner equation was applied to calculate the change in bed level (Section 3.5). This modelling methodology was applied over each of the (~6 month) periods of beach profile surveys, i.e. from autumn in one year to spring in the following year (Table 1). A typical contour plot of the resulting change in bed level from 16/11/01 to 25/04/02 is given in Fig. 10 and typical time series of modelled sediment transport and bed level change at station 70 are given in Fig. 11.

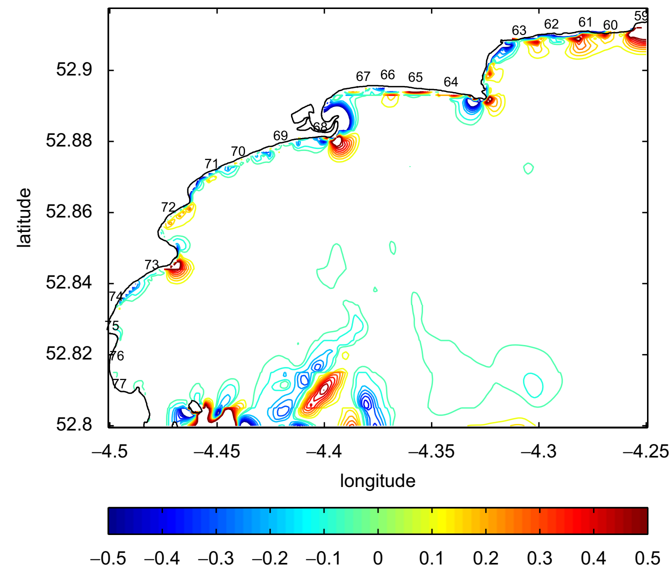

Fig. 10. Contour plot of modelled bed level change from 16/11/01 to 25/04/02. Colour scale is bed level change in metres. Numbers along coastline indicate beach profile station reference.

Along most of the coastline, the simulations generally show an alternating series of sources and sinks, but with localised discrepancies (Fig. 10). The magnitude of bed level change along the coastline was greatest at headlands where polar systems of erosion and deposition developed. Note that the Lax-Friedrichs scheme has introduced a small amount of 2D diffusion into the bed level change model, hence the sources and sinks tend to diffuse away from the coastline.

Significant wave height is closely related to wind speed (Figs. 11a and b). Wind direction is not critical since winds (and hence waves) tend to emanate from the S to SW sector in the study region, and waves tend to be refracted approximately normal to the coastline in shallow water. Longshore transport, however, is critically dependent on wave direction in addition to a threshold wave energy (Fig. 11c). Total transport in the region of Traeth Crugan is $\ll$ longshore transport (Fig. 11d) (however, total transport is important further offshore, e.g. the shallow regions centred around 4.4°W, 52.81°N). The resulting bed level change (Fig. 11e) tends to occur suddenly during a storm event (steep positive gradient) but then to diffuse relatively slowly after a sustained period of storm activity. Bed change can be positive or negative at a particular station during a simulation.

In Fig. 12, the modelled bed level change at the coastal cells closest to each of the beach profile locations is plotted at the end of each ~6 month simulation, noting that the x-axis (station reference) is reversed to provide a similar orientation to Fig. 10. The modelled data is plotted along with the observed change in beach level at each station over the same time period. The observations were processed as described in Section 2.3 and

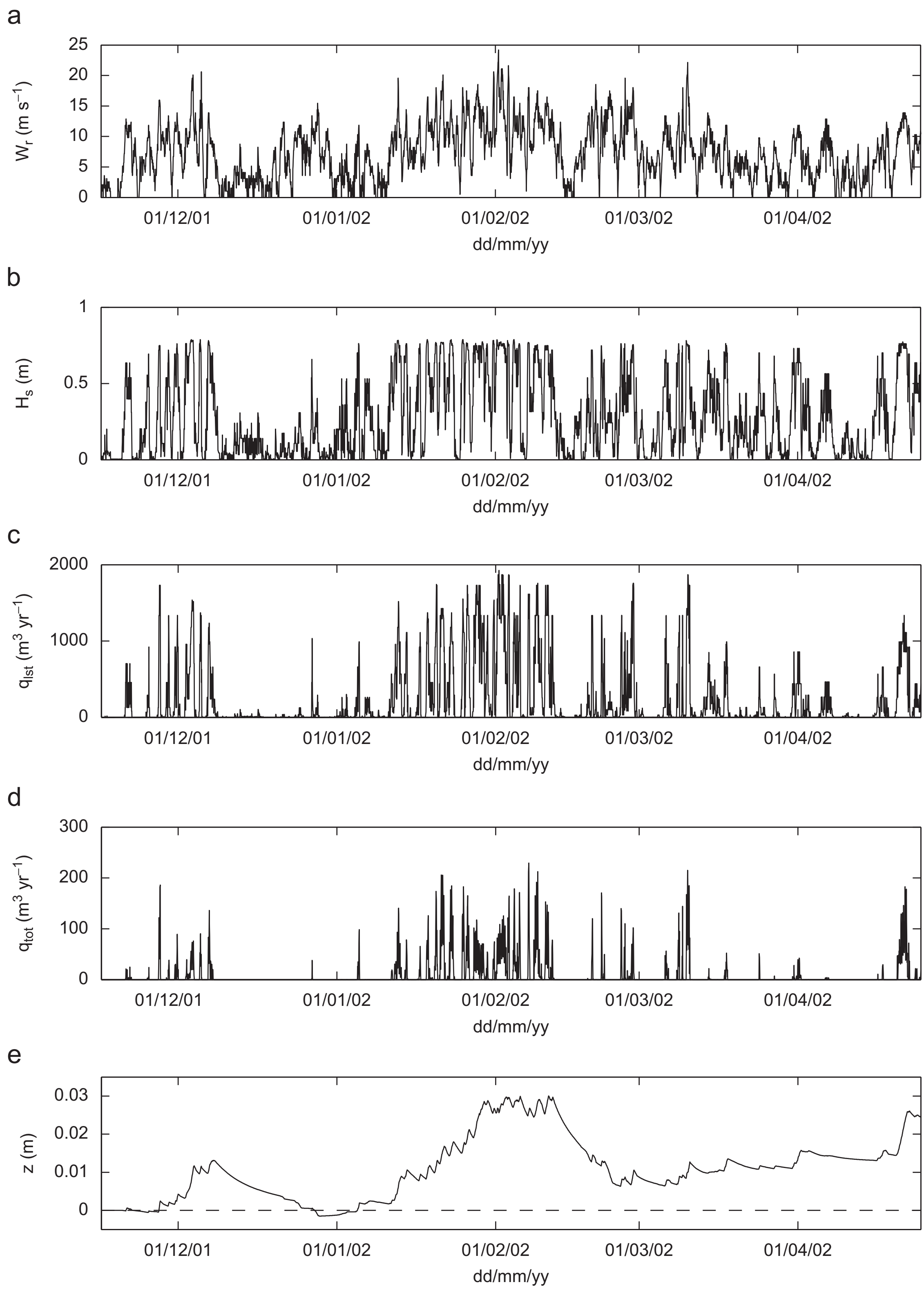

Fig. 11. Time series of wind speed (at Valley), significant wave height, longshore transport, total transport and bed level change at Traeth Crugan (station 70) from 16/11/01 to 25/04/02. Longshore transport, wave height and bed level change are output at model grid cell adjacent to the coastline. Total transport is output at the second model grid cell normal to the coastline: (a) wind speed, (b) significant wave height, (c) longshore transport, (d) total transport and (e) bed level change.

represent the change in beach levels between MSL and HAT. It is acknowledged that this method provides only a proxy for sediment movements due to longshore and total sediment transport, but the agreements in Fig. 12 are reasonable.

5. Discussion

The bed level change in the near coastal zone predicted by a morphological model of longshore and total sediment transport

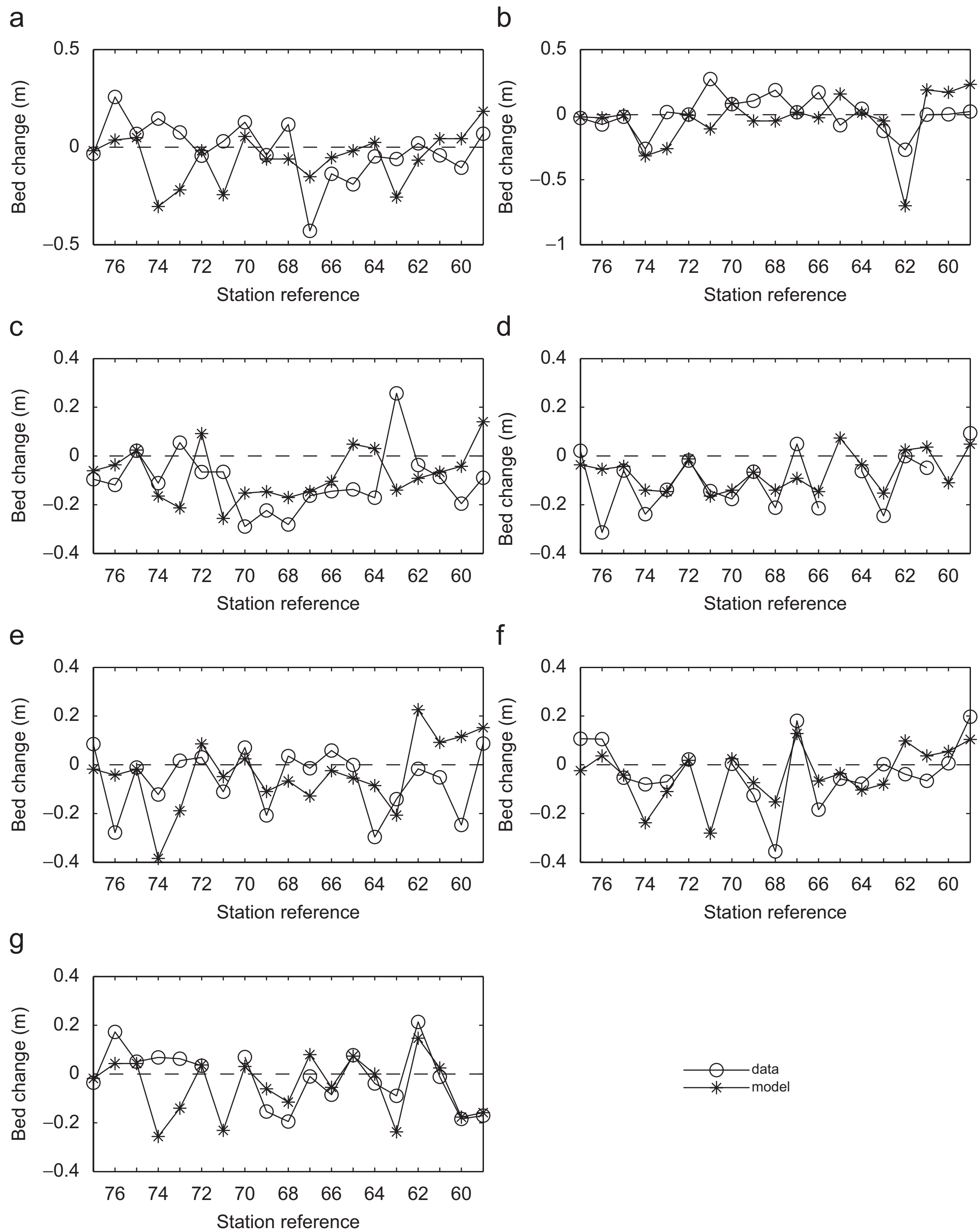

Fig. 12. Comparison between modelled and observed change in beach level for all years for all 19 stations: (a) 1997–1998, (b) 1998–1999, (c) 1999–2000, (d) 2000–2001, (e) 2001–2002, (f) 2002–2003 and (g) 2003–2004.

(with hydrodynamics provided by wave and tidal models) had reasonable agreement with observed beach level changes (Fig. 12). The model successfully captured much of the detail in the along shore erosion/deposition pattern. The overall mean error (for all 19 stations and for all 7 autumn/winter periods) between model and data was calculated to give an indication of the error in magnitude. This error was 0.11 m, which is of the same order as the observed magnitude changes. Hence, caution is required when interpreting this data. However, considering uncertainties in morphological modelling (e.g. empirical sediment transport formulae), this is not considered to be an unreasonable error in magnitude. Sources of this error are discussed further below.

Magnitude errors between model and data are due to (a) processes absent from the modelling methodology and (b) inaccuracies in representing the processes present. In the latter case, the key sources of error are in the empirical sediment transport formulae, model resolution (particularly the lack of resolving waves at breaking) and the accuracy of the lookup table technique for wave prediction. By tuning the empirical formulae (particularly Eq. (3)), it may be possible to reduce the error in magnitude. However, in the absence of local in situ measurements of longshore transport, it is difficult to justify such tuning. Since the wave model used a fixed grid (100 m resolution), it was not possible to accurately transform the offshore waves to the surf zone (a resolution of order 2–5 m would be required to resolve wave breaking). A suggested extension to the methodology is to nest a more detailed near-shore wave model within the (relatively) coarse 100 m grid. This further nested model would transform waves from a location approximately 100 m (i.e. one model grid cell) from the coastline to the location of wave breaking. A one-dimensional model applied at each coastal grid cell would suffice for this application, hence minimising the additional computational cost.

Clearly, the lookup table technique for wave prediction has reasonable accuracy over long time periods but has substantial local errors (Fig. 8). Wave energy for high magnitude wind events is often over-estimated since the technique is based on instantaneous wind speed and hence contains no 'memory' of wave energy. In addition, the assumption was made that wind was spatially uniform over a large geographical region. Finally, the source of wind data for the lookup tables was spatially remote from Tremadoc Bay, hence no account was taken of the influence of local topography on wind. However, the use of wind data from Valley meteorological station was justified in terms of the available temporal resolution (hourly) and its central location in relation to the Irish Sea (Section 2.2). Other sources of model error include the dated bathymetric surveys (Section 3.1) and the assumed sediment particle size distribution, taken as constant over the model domain.

The method presented in this paper for morphological modelling takes no account of swell waves which are suggested to recharge beaches after storm damage (USACE, 2001). There is no equivalent economical method to the wave modelling presented here (Section 3.1) which can include swell waves in (for example) the form of lookup tables since they are generated externally to the computational domain. For this reason, it is only realistic to apply the morphological model over relatively short time periods (e.g. 6 months). In addition, the assumption was made that an infinite supply of sediment was available for the model simulations, since all of the surveyed beaches were sandy beaches (i.e. sediment was redistributed throughout the model domain). This is a realistic assumption in the inner nearshore zone, but is less realistic further offshore. The total transport formula calculates potential and not actual transport since the seabed composition varies throughout Tremadoc Bay (BGS, 1988). Tremadoc Bay is composed of regions of sand, gravel, mud and combinations thereof. This problem could be partially resolved by imposing an initial condition of sediment distribution (i.e. sand), including a specified depth of sand deposits in the form of an erodible layer.

One of the objectives of developing this model was to assess its suitability for studying inter-annual variability of beach level changes over the autumn/winter storm period. This can be quantified for both beach profile data and model output in the following way. The mean absolute change in beach (or bed) level was calculated using

$$\frac{1}{m}\sum_{i=1}^{m}|\Delta Z_i| \tag{11}$$

where m ($=19$) is the number of beach profile locations and ΔZ is the change in bed level from the autumn survey to the spring survey (which also corresponds to the modelled period). This calculation was made for each of the seven ~6 month periods. Hence, the information contained in Fig. 12 was reduced to a single value (per year) for the in situ data and a single value for the model data. These values represent the magnitude of sediment movement over the relevant time scale, but give no indication as to the distribution of sediment. The values are plotted in Fig. 13 as a function of time (defined as the time of spring survey). It is clear that the model results lead the observations by approximately 1 year. Therefore, also plotted is the mean absolute bed level change for the model outputs lagged by 1 year.

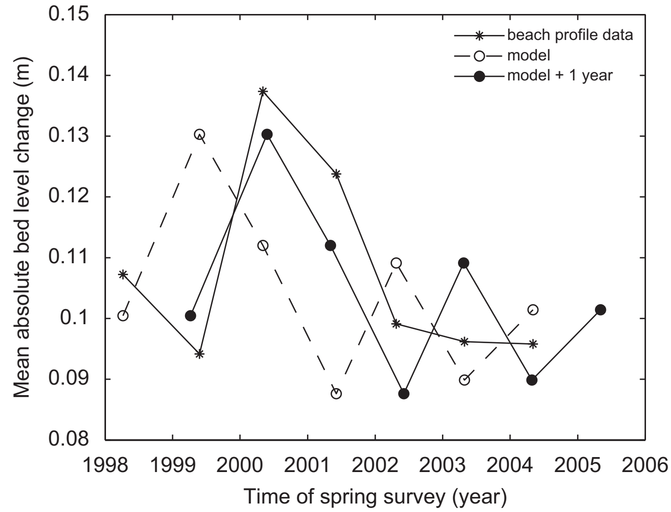

Fig. 13. Mean absolute bed level change for all 19 stations for all 7 years: in situ beach profile data and model data. The model data is also shown lagged by 1 year.

Since the model results lead the beach profile observations by 1 year, this represents, apart from errors in the modelling methodology discussed above, sources of sediment external to the domain. It is suggested that storms from the previous autumn/winter season may have moved sediment from an adjacent coastal sediment sub-cell into the north of Tremadoc Bay. Two such sediment sub-cells exist in the region (Cooper and Pontee, 2006): one extending from Bardsey Island to the Dwyryd Estuary (Fig. 1) and the other extending from the Dwyryd Estuary to the south of Cardigan Bay. This quantity of sediment is placed into storage (e.g. in the extensive sand deposits at the mouth of the bar built Dwyryd Estuary) and then redistributed within Tremadoc Bay in the following year. Therefore, in terms of the magnitude of beach level changes, the observations would be expected to lag a model based on local sediment transport.

The model generally under-predicted the magnitude of beach level change over the model domain by O(0.01 m) (Fig. 13). However, there was one significant exception: the observations for the autumn/winter period 2002–2003 (from the above discussion, this magnitude is proportional to the model output and hence the sediment transport for the period 2001–2002, i.e. $n-1$). The model over-estimated the mean magnitude of beach level change by O(0.01 m) over this period. It was during February 2002 when significant dune damage was recorded and urgent rock armour repairs were made (Section 1), and the sudden change in bed level can be seen in Fig. 11e (the actual date of the storm was 1–2 February and the peak wind speed was 24.2 m s^{-1}). Since the lookup table technique for wave prediction often over-estimates the wave energy for high magnitude wind events (Fig. 8), it is likely that for a major storm such as occurred in February 2002, the model has over-estimated the magnitude of bed level change and the discrepancy is discernable at the end of the simulation period, i.e. 3 months later (Fig. 11e).

The time series of bed level change (Fig. 11e) shows how the timing of the beach profile surveys is critical to the understanding of modelled processes. The storm activity around the beginning of

February 2002 led to a localised increase in bed level of order 0.03 m. However, if the beach were to be surveyed at the beginning of March 2002, little net change would be discernable at this location (according to the model). It is only the modest storm activity towards the end of April 2002 which has contributed directly to the increase in bed level observed at the end of the simulation period.

6. Conclusions

An efficient morphological model was developed consisting of wave, tidal, longshore transport, total transport and bed level change modules. The model is suitable for application to relatively long duration high resolution (hourly) simulations, and hence includes the effect of high frequency wind events. The model had reasonable agreement with beach profile data over ~6 month time periods. With further work, it may be possible to minimise many of the sources of error within the model without significantly compromising the computational efficiency, e.g. the nesting of a high-resolution one-dimensional wave model to predict accurately the wave characteristics at breaking, or extending the morphological computational domain to include sediment processes acting over a larger area. Other errors, such as the inclusion of swell waves, are not easily minimised without resorting to less efficient numerical techniques (i.e. explicit time-stepping models applied over large regions).

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