

## CHAPTER 239

### The Influence of Long Waves on Macrotidal Beach Morphology

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#### Abstract

The aim of the following paper is to demonstrate that standing long waves can generate intertidal morphological features in a dissipative macrotidal setting. In this environment the tidal excursion can exceed the length scales associated with long waves. This extends the work of Simmonds et al (1995) in which it was shown that monochromatic long waves could generate sediment convergence patterns that bore a good similarity to observed topographic features, namely ridges and runnels from a field site on the Belgian coast. The simulation is modified to include a band of long waves and it is found that a good correspondence with real topography can still be achieved on a tidally averaged basis by allowing for a degree of uncertainty in the breaking criterion. It is also argued why the observed scales are more likely to correspond with the shorter end of the long wave spectrum. The simulation helps to shed light on the conditions in which such features are known to form.

#### Introduction

This paper examines the question “can a spectrum of long waves be capable of creating intertidal bedforms on a macrotidal beach?”

Despite the abundance of macrotidal coastlines Short (1991) has been led to remark that the “study of macro-tidal beaches has lagged considerably” in relation to the published material on micro- and meso-tidal coasts. Whilst numerical modellers are beginning to recognise the importance of long wave processes in the modeling of coastal morphology (Roelvink & Broker, 1993) very few of these models, even though applied to macrotidal coastal regions, appear to consider the further complication of a tidally modulated boundary. Although the concept of an equilibrium cross-shore profile is useful in a microtidal setting, the interpretation of equilibrium profile is questionable in the macrotidal setting when considered with

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respect to timescales of the order of days, even when tidally averaged quantities are considered. The cross shore tidal excursion on a dissipative macrotidal beach is measured in hundreds of meters and is further modulated by the Spring–Neap cycle. The boundary is neither steady nor varying by the same amount from day to day. It seems more appropriate to speak of a dynamic equilibrium whereby the profile is constantly adjusting from minute to minute within limits determined by the tidal level and variations in the wave climate.

Standing long wave motion at the shoreline is often associated with the formation of submerged bars (Short, 1991, Sallenger & Holman, 1987) and support for this has been provided by field observations (e.g. Bauer & Greenwood, 1990). But what of the intertidal bedforms such as the bar-like features, referred to as ridges and runnels, that are evident on some macrotidal coastlines?

Wright et al (1982) suggested that macrotidal beach profiles are due to the combination of swash, surf and deeper water processes that operate over different regions of the beach face for different periods of time. It was also suggested that some wave-lain features may survive swash & surf zone excursion in low energy conditions — but that in high energy conditions any bedforms are erased by the more energetic processes.

These studies have been predominantly based on high energy exposed shorelines. In the European context, many dissipative macrotidal shorelines are to be found sheltering in fetch limited environments well away from the continental shelf breaks. In these situations, more subtle, lower energy mechanisms might be more evident.

It is tempting to suggest that standing long waves might be somehow responsible for the development of intertidal features such as ridges and runnels as the length scales of both are comparable. In addition it is widely recognised by field workers that long wave energy is present, to some degree, on all shorelines. In the simplest analysis, topographic features are believed to be generated under a stationary standing long wave. However on a macrotidal dissipative coast the reflection point of the long waves (whether considered at the shoreline or in the surfzone) can move cross-shore by several long wave wavelengths.

Simmonds et al (1995) showed how standing long waves might form bar-like features in the intertidal zone of a linear beach dominated by cross-shore long wave processes. They based their analysis upon a simple transport mechanism resulting from the correlation of the long waves with the incoming wave groups to demonstrate how intertidal bedform development might be initiated through a "tidal bar hypothesis". Using a monochromatic long wave envelope, a pattern for the potential for erosion across the beach face was shown to correspond well with observed scales of ridge and runnel morphology even around the mid-tide region, despite the inclusion of tidal modulation of the transport envelope across the beach face.

However, long waves observed in the field are generally broad-banded in nature (Huntley et al, 1993) and it might be predicted that the result of including of a realistic bandwidth of long waves would smear over the effect of the individual waves. This effect combined with tidal modulation of the reflection point would surely destroy any "memory" that the beach retains of the long wave structure – although Bowen and Huntley (1983) suggested that feedback, in the form of erosion of the bed and subsequent modification of long wave structure might result in a narrowing of the effective spectrum that shapes the beach through resonance.

A further criticism of the hypothesis is that given that there is a band of long waves why does there appear to be a stronger correlation at the shorter wavelength end of the measured band than at the longer scales?

We will investigate these criticisms by reference to an improved version of the original simulation outlined in Simmonds et al. (1995). It will be shown that under conditions recognised as precursory to ridge and runnel formation, rhythmic intertidal features are indeed predicted. We draw upon the geometry and field measurements of a particular example of ridge and runnel morphology. The aim is not to present a full sediment and hydrodynamic coupled model, but instead to show that the mechanism for forming intertidal bedforms involving standing long waves is feasible.

#### A Field Example of Rhythmic Intertidal Morphology

First of all it is useful to reintroduce the field example of ridge and runnel features (Simmonds et al, 1995). The macrotidal beach profile is shown in Figure 1 and is taken from a field site at Nieuwpoort on the Belgian coastline which was the

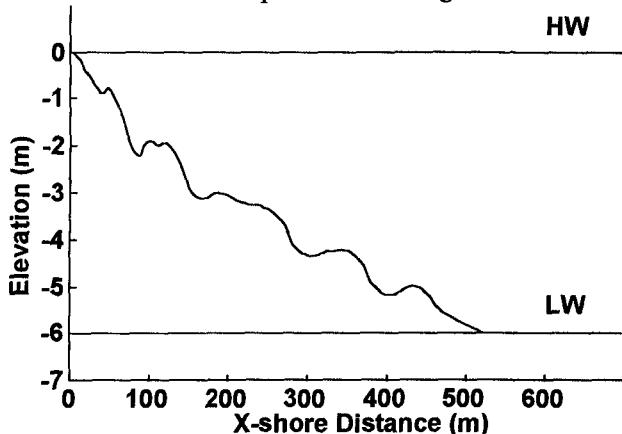


Figure 1. Ridge & Runnel profile at Nieuwpoort, Belgium

focus of the recent MAST project *Circulation and Sediment Transport Around Banks* (CSTAB —O'Connor et al, 1996). This illustrates the kind of intertidal features that are expected to be associated with long waves by virtue of their length scales. These ridges exist as low lying, shore-parallel bodies of sand with cross-shore scales of the

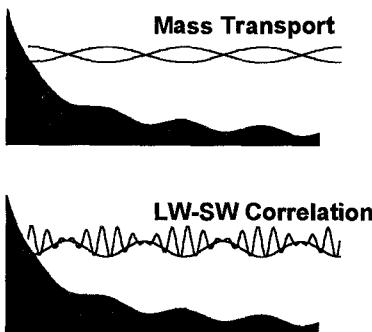
order of 50m separated by runnels. Their longshore extent is interrupted by shore-normal channels spaced every 500m or so, down which water is drained from the runnels on the ebb of the tide. They are exclusive to dissipative macrotidal beaches of medium sand sized material and fetch limited wave climate.

It is hard to justify that these particular features are merely swash bars formed at low and high still stands at Spring and Neap tide levels (King & Williams, 1949). Five bars are clearly visible. Other documented cases of ridge and runnel beaches exhibit even greater numbers of ridges (Mulrennan et al, 1992). Indeed the number and spacing of these particular features on low energy beaches is reported to be strongly related to the beach slope and this is more reminiscent of long wave structure where the node-anti-node spacings are also dictated by the beach slope  $\beta$ . Therefore this swash bar hypothesis should be rejected.

The scales involved are certainly too large for comparison with gravity wavelengths. So the question "Could long waves be responsible?" arises. Simmonds et al (1995) previously reported the identification of standing long wave activity, even during low energy conditions at this particular beach in the frequency band of 0.01-0.04Hz (100s to 25s wave period). The scale of the features was shown to be consistent with shortest of these long wave scales (0.04Hz) measured at the beach which had a slope of around 1%. It therefore seems likely that long waves are indeed responsible.

### **Long Wave Transport Mechanisms**

The mechanism by which a memory of the structure of a standing long wave can come to be imprinted upon a beach is now discussed. The cartoon (Figure 2) illustrates two mechanisms that can be considered.



**Figure 2 Transport mechanisms**

The first involves sediment movement in response to the mass transport induced under a standing wave. This has been investigated in association with subtidal bar formation (e.g. Carter et al. 1973). In this, the sediment is assumed to move in response to the drift velocities induced by the standing long wave structure. Sediment accumulates either at the nodes or anti-nodes according to different

investigators though there is no general agreement as to which, although the answer is thought to depend upon the sediment grain size.

The second is a mechanism based upon the correlation of short-wave groups with a bound long wave. O'Hare & Huntley (1994) have discussed a model that furthers the concept of short waves acting as sediment stirrers and correlated long waves acting as transporters (pseudo-steady currents). In the case of a bound long wave the long wave is generated through the radiation stress variation under wave groups so that the largest short waves occur around the long wave troughs and the smaller at the long wave crests. Such a correlation creates a net offshore transport when integrated over a long wave cycle.

If a planar beach face is considered, in the absence of any other processes, this offshore sediment flux would be predominantly modulated by the simple depth dependence of the short wave orbital velocity. The inclusion of a phase-locked reflected long wave, however, introduces a rhythmic cross-shore spatial variability in the sediment flux. The function which is the gradient of this can be interpreted as regions of net erosion or accretion.

Both of these two mechanisms would produce an erosion pattern that would predict features of the same scale. However, the efficiencies of the two in transporting sediment will have different magnitudes. We argue here that the correlation mechanism is likely to be more important.

In the first case the transport mechanism is related through a power law to the "drift velocity". This second order parameter arises from the non-linearity in orbital motion under the standing waves. In the case of long waves this is likely to be very small in relation to the orbital velocity of the long waves.

In the second, the net transport is the result of averaging the correlation between the square of the heights of the (stirring) short wave and the velocity of the (transporting) long waves. Again, it is difficult to estimate the relative importance of this quantity since the resulting transport over a long wave cycle will depend upon, amongst other factors, the amplitude of the long wave orbital velocity and the groupiness of the short waves — that is, the net effect depends upon the difference between the onshore and offshore directed correlation's.

However, the magnitudes of the latter velocities are much greater than those of the drift velocities, and so the second mechanism has the potential to be significantly greater.

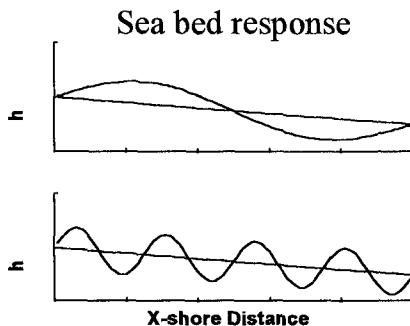
### **Bias to Shorter Scales**

If it is accepted that there is a process for the imprinting of long wave scales on the beach face the next question to address is "why is a stronger correspondence with the shorter end of long wave spectrum observed in the field?" As we stated earlier, the observed correlation between the measured long wave spectrum and the observed morphological scales at Nieuwpoort indicated a good agreement but only at

the shorter (0.04Hz) end of the long wave spectrum. This bias can be understood by considering the following arguments.

First, we should define what we understand by the description "broad banded" in relation to the long wave spectrum. If we take the naïve interpretation of "broad-bandedness" as implying equal wave heights at all frequencies in the spectrum, quite clearly the velocity (transporting) spectrum will be biased towards higher frequencies. This is because the velocity is obtained as the gradient of the time series. Put another way, a "white" amplitude spectrum corresponds to a "blue" velocity spectrum (predominance of higher frequencies).

We can also argue, on a geometric basis, that shorter long wave scales grow more rapidly. This is because the work done on the bed in changing from a flat to an undular bed is less if the undulations are of shorter wavelength (Figure 3). The distance that the centroid of sand has to move in order to create a depression and an adjacent hole is less for the shorter wavelength feature than the longer, although the



**Figure 3 Geometric argument: mass centroid in (a) moves further than in (b)**

amount of material that has moved is the same. It is therefore likely that the shorter scales are quicker to develop upon the beach and can be concluded that there is an inherent bias to short scales forming. This is especially important for a system in a state of dynamic equilibrium where features that require longer timescales in order to establish themselves will be discriminated against.

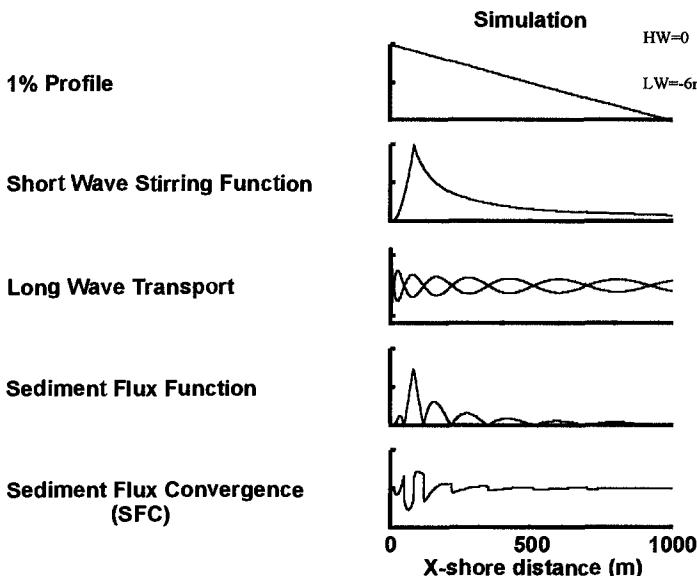
### Simulation

Attention is now turned to the simulation reported in Simmonds et al. (1995). The simulation is used to predict the spatial cross-shore profile development. The basis for this is the assumption that the sediment transport is everywhere offshore directed due to the correlation mechanism outlined above. Obviously this is unrealistic—beaches do not simply disappear off-shore, (at least, in the short term), but this assumption is made so that the potential of this mechanism can be investigated in isolation. As such, other processes such as surf-zone and swash processes are not represented.

A bound long wave is assumed to interfere with a reflected free long wave to give a shallow water standing long wave. The sediment transport rate is assumed to be proportional to the product of the magnitude of the standing long wave orbital velocity and of the square of the short-wave envelope.

It is further assumed that the short wave envelope that stirs the sediment can be described by simple shoaling waves which break with a linear decrease in waveheight to the shoreline. The standing long wave envelope, calculated from the bessel function solutions of Lamb (op cite Kirby et al, 1981) defines the magnitude of the offshore flow. The product of these two represents an expression of the uncalibrated offshore sediment flux whose spatial representation is the sediment flux function. The gradient of this is the quantity that is of interest, and is referred to here as the *sediment flux convergence* (SFC). This is interpreted as an indication of zones of sediment accretion or erosion across the profile.

The bias to higher frequencies is incorporated in the model in terms of a "blue shift" of the velocity spectrum. This is generated by assuming uniform amplitude spectrum and hence a velocity spectrum biased linearly towards the high end of the frequency band.



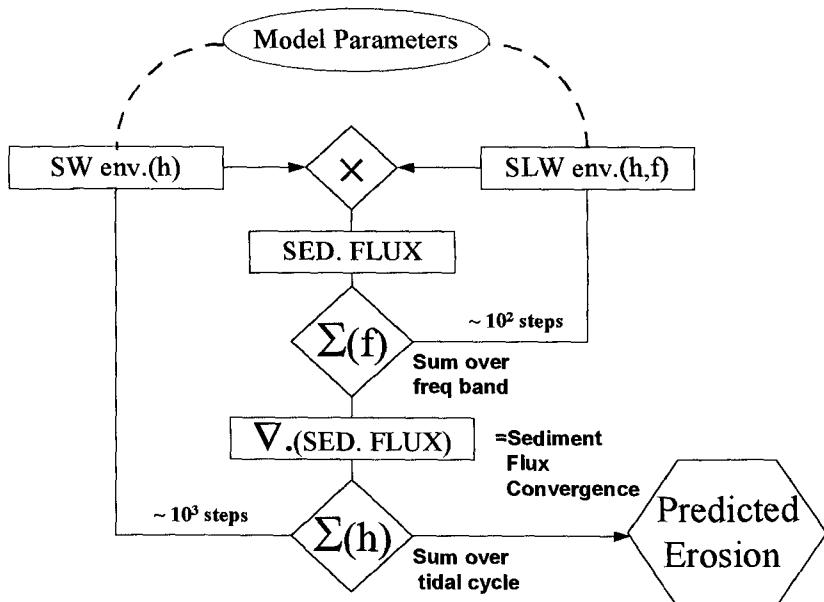
**Figure 4. Components of the simulation**

Taking parameters similar to those observed at the site at Nieuwpoort beach, a Spring tidal modulation of 6m height is introduced over a planar beach of 1% slope. Figure 4 shows how the simulation is constructed. This was implemented in the MATLAB language on a PC.

The simulation is also represented by the flow diagram (Figure 5). The model parameters specified include the short-wave height, the breaking depth, the long wave spectral amplitudes and the beach slope. From this information the standing long wave envelope (SLW env) is calculated and the short wave envelope (SW env). Multiplied together, these give the sediment flux function for a particular long wave frequency and reflection point on the profile.

The simulation has then been adapted to permit the summation of the sediment flux functions over a suitable spectrum of standing long waves at each position of the tide. The summation of the calculated sediment flux convergence functions is then performed to calculate the tidal average.

It is also possible to modify the short-wave envelope to allow for a degree of uncertainty in the short wave height and breaking criterion. This is desirable because, firstly, groupy waves do not all break at the same position, and, secondly, breaking is believed to be influenced by instabilities which determine that the nature of this process is chaotic (Longuet-Higgins, 1994).

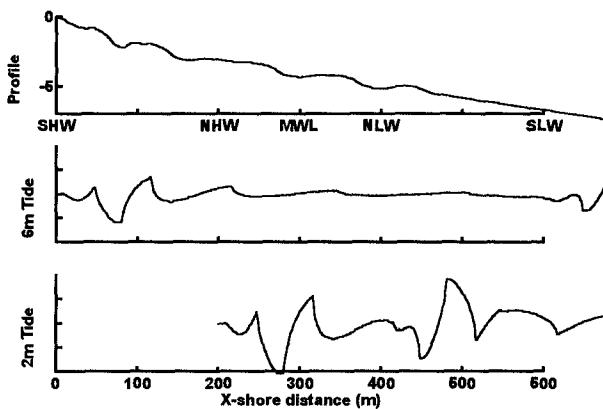


**Figure 5** Flow-chart for improved simulation

As mentioned, the simulation involves two summations. Summation over frequency band and summation over all tidal amplitudes. An appropriate indication of the magnitude of the number of steps for these two summations is indicated in the figure. These have been selected to optimise processing time against numerical "noise".

It should also be pointed out that at each position of the tide the sum of the superimposed velocities is being calculated. It is therefore being assumed that all the different standing long wave modes occur simultaneously and, further, that their respective correlation's with the short waves also all occur together. Whether this is realistic is a matter for debate. Usually the statistical analysis of a wave record is interpreted to mean that the spectrum of frequencies is stationary. It may be that, in reality, there is a "wandering" of the peak band of frequencies from long wave to long wave.

The present simulation is not sensitive to the order in which the two summations are calculated. However to develop the model further, feedback, in the form of bed evolution, is to be introduced before summing over the next tidal



**Figure 6(a)** Ridge & runnel profile. Monochromatic calculations of the SFC for (b) 6m tide & (c) 2m tide.

amplitude after the summation over frequency bandwidth.

Tidally averaged runs for Spring & Neap tides using values for these parameters obtained from the Nieuwpoort beach produces the results in figure 6(b) & (c) for monochromatic long waves (0.04Hz). As can be seen the predicted SFC shows good correspondence with the actual topography, Figure 6(a), in terms of scale.

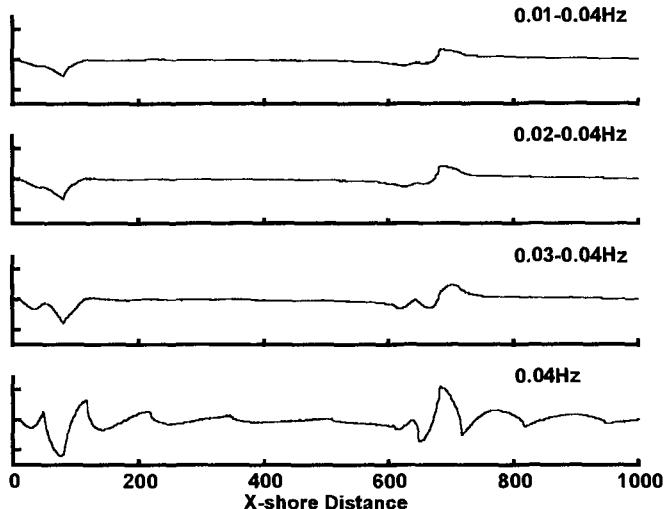
#### **A Realistic Bandwidth of Long waves**

The main query regarding the above simulation is: "what effect does a realistic, broad long wave bandwidth have on these predictions?" This question will now be addressed and the sensitivity of this simple model to other parameters will also be discussed.

At first sight, Figure 7 indicates that calculating the SFC for a band of long waves destroys the intertidal undulations, except in the case of a very narrow bandwidth indeed. However this can be shown to be misleading by studying the

sediment flux functions. Figure 8(a) indicates that the flux function (the offshore transport) is dominated by the position of the breakpoint. This creates two relatively sharp peaks in the SFC near high and low water throwing the remaining details in to the shadows. The convergence here is emphasised by the fact that the breakpoint dwells at these two positions for the longest. As mentioned above, in reality the breaking of the short waves has a statistical variation caused by variations in the waveheight due to groupiness, return flows, interactions with preceding waves and an inherent chaotic instability in the breaking process.

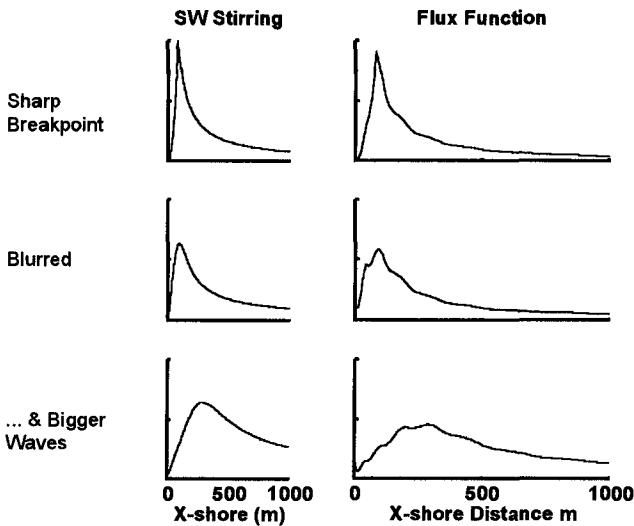
To study the effect of this, an uncertainty in the breakpoint was modeled assuming a gaussian description of wave height and the breaking criterion. This was included in the description of the short wave envelope or "stirring function". The effect in the sediment flux function is evident in Figure 8(b). The breakpoint is rounded off and de-emphasised, allowing detail of the long wave structure to emerge. Figure 9(b) shows the effect of this in the predicted SFC which displays once again some undular detail across the profile, more obviously than in Figure 9(a). Again the scales correspond well with those observed in the field.



**Figure 7** SFC calculated for different long wave bandwidths

The uncertainty in the breakpoint acts to widen the surf zone. Another way this can be achieved is to consider short waves with larger waveheight. As can be seen in Figures 8(c) and 9(c), the effect of this is to smooth off the breakpoint further and to bring out more intertidal profile detail. This remarkable result shows that despite summing over a band of long waves and averaging over a tide, the predicted SFC still shows details that correspond to the observed topography even in the mid tidal region. However, although larger wave heights are helpful in this model it is probably true, as Wright (1982) indicates, that in very high energy events other

processes will dominate and the surf zone "bulldozer" will mask the effects of this subtle mechanism.



**Figure 8** Short wave envelopes and sediment flux functions for (a) sudden breaking, (b) uncertain breaking, (c) uncertain breaking & larger wave height

#### Further Investigations

The simulation was also used to investigate other parameters, including shape of the tidal curve and of the spectrum. These predictably showed that a flatter tidal curve de-emphasised erosion at the high and low water marks, and that a "blue" long wave amplitude spectrum brought out even more detail in the undulations of the sediment flux convergence.

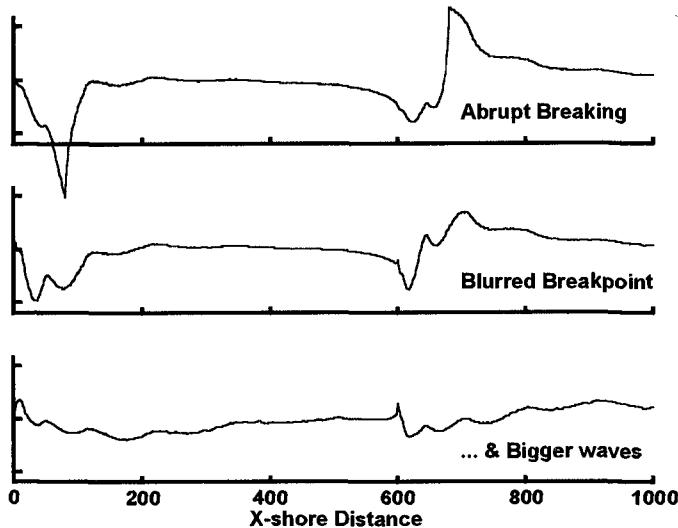
#### Time-scales

The results so far have all been calculated as averages over the tidal cycle. Figure 10 shows the sediment flux convergences calculated for averages over shorter fractions of a tide (1/6ths). As can be seen, the tidally averaged result shows no indication of some of the features that can develop during the middle of the tide. Two features at 200m and 500m are clearly visible in the mid tide SFC but these are cancelled by the SFCs approaching high and low water.

It is likely that the bed will respond over shorter time-scales than that of a tide. If the bed is modified, the long and short wave field are changed accordingly

and the erosion patterns will be changed. It may be that the erosion pattern is capable of "tuning in" with the topography to favour growth of certain scales.

Southgate (1995) has remarked that since the interaction between hydrodynamics and the transport of sediment is non-linear, the chronology or sequencing of wave events is of direct influence. Further, in the macrotidal case, the erosion of the beach face will be dependent upon both the temporal variation of the long wave transport functions and the tidal variation.

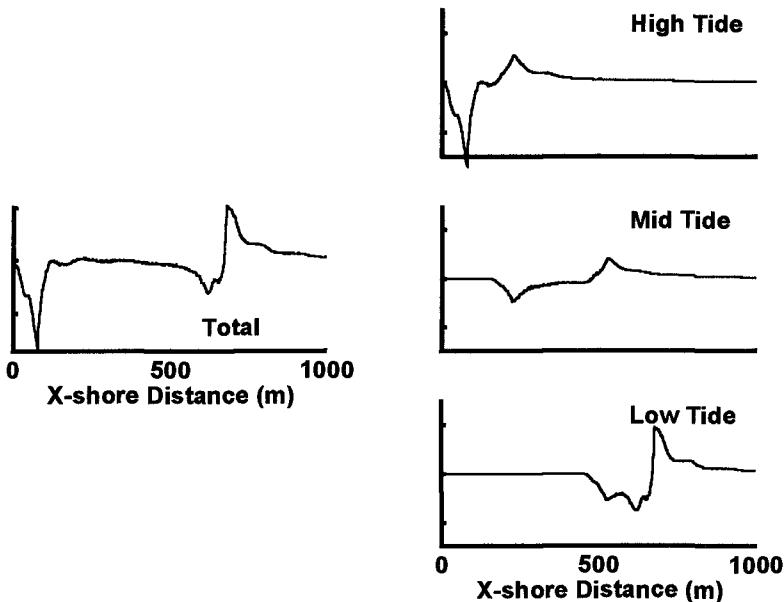


**Figure 9(a)** SFC for a band of long waves (0.01–0.04Hz), (b) inclusion of uncertain breaking & (c) bigger short waves.

### Summary & Discussion

Simmonds et al (1995) demonstrated that a standing long wave mechanism such as the bound long wave-short-wave correlation is capable of generating intertidal bedforms at scales observed in nature. This idea has been extended in this paper to the consideration of a band of long waves. It has been argued that there exists an inherent bias towards the quicker development of shorter scales. It has been shown that a band of longwaves will still generate undular intertidal topography, even in the mid-tide region so long as a smoothing of the breakpoint and hence a widening of the surfzone is introduced. This was necessary to de-emphasize the erosion at the high and low water levels caused by the constancy of the breaking at these places. The smoothing is justified by considering the uncertainty in the breaking process. The resulting predicted regions of sediment convergence agree remarkably well with the observed topography.

It has also been suggested that the current approach to integrating results over tidal timescales needs re-examining since it is that the bed will respond over much shorter timescales.



**Figure 10** SFCs calculated for different sixths of the tide

The model helps us to explain the environment in which ridge and runnel features are found. For the mechanism to work, the cross shore tidal excursion must exceed the scale of the shorter long wave wavelengths otherwise no intertidal structure can be created. Broad, dissipative macrotidal beaches are ideal for this. The mechanism is, however relatively subtle, depending upon the long wave orbital velocities to carry out the transporting of the sediment, and would probably be masked by other processes in a more energetic environment. This might explain why these features are only found on fetch limited macrotidal coastlines.

Future developments of the model are envisaged that allow erosion of the bed as a feedback mechanism which perturbs the structure of the sediment flux function. It is hoped that such a coupled sediment transport and hydrodynamic model can be used to study the parameters that control the behaviour of these intertidal features and the time-scales of their development.

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