HOW IMPORTANT IS GLOBAL WARMING FOR COASTAL EROSION?

An Editorial Comment

While there is now ample evidence of an increasing global temperature rise, we still lack convincing global evidence of an increasing rate of eustatic sea-level rise. However, physically such an increase seems very plausible, and anticipating the effects is a wise policy. The IPCC has successfully raised awareness on political and societal levels, which in many nations has resulted in including sea-level rise scenarios in new designs of shore protection works, both hard (structures) and soft (nourishment) or combinations thereof (Hamm et al., 2002). Crucial in this context is the quantification of the relative role that sea level rise plays in low-lying coastal areas. Simple inundation will occur, when these areas are sediment starved. This implies coastal retreat to be determined by the local coastal slope. Coastal slopes in such areas may be as low as 1 in 1000, which results in a retreat three orders of magnitude stronger than the rate of sea level rise.

More interesting and more complicated is the question what the impact will be on dune and barrier coasts. Is this a simple question? No, because the response of dune and barrier coasts to sea-level rise is a complex morphodynamic issue. Then, is the analysis of Zhang et al. (2004) valuable? To the extent that it highlights one, potentially important, effect of a range of effects, certainly. If no other sediment sources or sinks are present or if no other sediment transport gradients in cross-shore and longshore directions occur, the so-named Bruun effect (Bruun, 1962) is the only effect operational. Hence for this idealized case Zhang et al. (2004) confirm the theoretical finding that coasts will retreat two orders of magnitude stronger than the rate of sea level rise. However, first, this idealized situation is rather exception than rule, second, the Bruun effect in near stillstand sea level rise conditions is rather moderate, and third there are other effects of sea-level rise on the coastal sediment budget. We will discuss these issues below.

1. What's the Evidence?

A general point-of-view, triggered by Bird (1985), is that since 70% of the sandy beaches is erosive, sea-level rise is the most probable cause (c.f. Leatherman et al., 2000). However, there are many coastal systems that have been accretive in the Holocene, even though sea level was rising. A few examples are the Australian coast (Short, 2003), deltaic coasts (Mississippi, Ebro, Po, Yangtze, and many other deltas at earlier stages of the Holocene) and composite coasts such as the U.S.

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Northwest Washington coast and the Dutch coast (Cowell et al., 2003a). What then can we conclude from this evidence? Besides the Bruun effect there must be a number of other processes that can override the Bruun effect to such extent that coasts are accretive. Also, many other coasts experience larger erosion then is explained from the Bruun effect. Note for example the coastal sections that Zhang et al. (2004) indicate as inlet influenced (their Figure 3). This triggers the important question what these other processes are and whether these are impacted by accelerated sea-level rise. If this is the case the additional impact of global warming on coastal evolution can not be quantified from the Bruun effect alone.

In an aggregated way these other processes may be collectively indicated as sediment availability. These are implicitly included in earlier kinematic models for first approximation of long-term coastal change (Curray, 1964). Swift (1976) extended Curray's (1964) ideas into a general framework for long-term coastal change entailing *transgression* (landward retreat) and *regression* (seaward advance) of the shoreline due to sea-level rise and fall, with corresponding tendencies toward *retrogradation* and *progradation* due to net sediment losses or gains alongshore.

Cowell et al. (2003a) show how Swift's (1976) concepts can be quantified and related back to what has become known as the Bruun Rule (Bruun, 1962), if we consider the sediment balance of the upper shoreface. Cowell et al. (2003a) adopt the assumption that the upper shoreface to a first approximation is form invariant relative to mean sea-level over time periods ($\gg$1 year) for which profile closure occurs (Nicholls et al., 1998). The upper shoreface is represented by an arbitrary, but usually concave-up, profile $h(x)$ to a depth h_* (a morphologically active depth) and a length L_*, in which x is the distance from the shore (Dean, 1991). Sediment-volume conservation for profile kinematics requires that

$$\frac{\partial h}{\partial t} + c_p \frac{\partial h}{\partial x} = 0\,, \tag{1}$$

where c_p is the horizontal profile displacement, or via $h = MSL\text{-}z_b$, where MSL is Mean Sea Level and z_b is the bottom level:

$$\frac{\partial z_b}{\partial t} + c_p \frac{\partial z_b}{\partial x} = \frac{\partial MSL}{\partial t}\,, \tag{2}$$

where c_p is the horizontal translation rate of the shoreline position. The sediment-transport balance equation for a fixed spatial control volume is

$$\frac{\partial z_b}{\partial t} + \frac{\partial q_x}{\partial x} + \frac{\partial q_y}{\partial y} + s = 0\,, \tag{3}$$

where $q_{x,y}$ are the cross-shore and alongshore sediment transports, and s is a local source or sink. These equations may be combined to yield

$$c_p = -\frac{\partial MSL}{\partial t}\left(\frac{\partial h}{\partial x}\right)^{-1} - \frac{\partial q_x}{\partial h} - \frac{\partial q_y}{\partial y}\left(\frac{\partial h}{\partial x}\right)^{-1} - s\left(\frac{\partial h}{\partial x}\right)^{-1} \tag{4}$$

or, after cross-shore integration over L_*,

$$c_p h_* = \frac{\partial MSL}{\partial t} L_* - (q_{x,sea} - q_{x,dune}) - \frac{\partial Q_y}{\partial y} - s \tag{5}$$

in which Q_y is the alongshore transport integrated over L_*.

In the absence of littoral transport gradients and other sources or sinks (including sand exchanges with the lower shoreface and backbarrier) the above reduces to the standard *Bruun Rule* (Bruun, 1962; alternatively Zhang et al., 2004):

$$c_p = \frac{\partial MSL}{\partial t} \left(\frac{L_*}{h_*} \right) . \tag{6}$$

Equation (5) is similar to the Dean and Maurmeyer (1983) version of the Bruun Rule, an analytical precursor of the coastal-tract concept (see below). The shoreline change rate is determined quantitatively by the balance between the 'sink' term, for accommodation-space generated due to sea-level rise (first term on the right-hand side), and sediment availability (being the sum of sinks and sources, the last three terms on the right-hand side of Equation (5)). The relative sea-level change is a virtual sink/source term since there is no absolute loss, although the response is comparable to the impact of a real source/sink regarding horizontal movements of the upper shoreface.

The source and sink terms in Equation (5) allow the qualitative Curray-Swift model of coastal evolution to be quantified as a time trajectory in sediment source/sink phase space: e.g., evolution of the well-documented central Netherlands coast between Hoek of Holland and Den Helder in Figure 1. The trajectory is based on (a) estimates derived from radiometric data by Beets et al. (1992), for the period 5000–0 years BP; and (b) the results of reconstruction simulations for 7200–5000 BP. The line separating advance and retreat of the coast is fitted for the trajectory in the top-right quadrant, with its mirror image assumed for the bottom-left quadrant in the absence of other data. The trajectory bifurcates after 2000 BP because differences develop in rates of shoreline change averaged alongshore north and south of Haarlem. The shape of the advance/retreat threshold curve demonstrates that coastal evolution is governed mainly by (a) sediment supply (+/-) under near-stillstand sea-level conditions (such as those predominating in the late Holocene), and (b) change in accommodation space when sea-level changes rapidly (such as during global glaciation and deglaciation).

What do we learn from the evidence? In periods of near stillstand sea-level conditions the Bruun effect is operational, but is commonly overridden by the sediment availability terms in Equation (5), i.e., cross-shore gradients, longshore gradients and/or sources/sinks. An attempt to make a general quantification of these other terms is presented next.

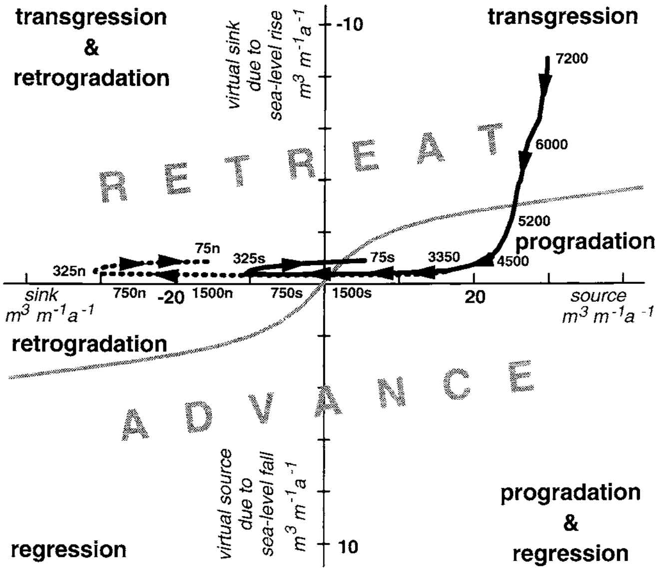

Figure 1. Evolution of the central Netherlands coast (Hoek of Holland to Den Helder) as a time trajectory in sediment-supply/accommodation phase space (abscissa and ordinate respectively, scaled in cubic meters per year per meter of shoreline). Numbers along the trajectory indicate time (years BP); suffixes n and s denote north and south of Haarlem respectively (after Cowell et al., 2003a).

2. Quantifying Cross-shore and Longshore Processes

Based on geological reconstructions and associated sediment balances we make a general estimate of cross-shore and longshore processes, i.e., of the effect of transport gradients due to these processes and the resulting shoreline changes. We will first consider cross-shore induced shoreline changes. Indicative numbers are given of both gross and net cross-shore sediment transports (terms 2 and 4 in Equation (5)), which can be converted to shoreline changes, c_p, by estimating the morphologically active depth, h_*.

Geological reconstructions of the Australian (Short, 2003) and the Dutch coast (Cowell et al., 2003a) have strengthened the hypothesis (Cowell et al., 2001) that middle shoreface wave-induced sediment transport is generally onshore on concave shaped shorefaces. This is associated with wave asymmetry and wave boundary layer induced net flow (Bowen, 1980). While this is a contribution that results in a shoreline advance, generally two contributions must be considered that result in shoreline retreat. One contribution is due to aeolian loss, i.e., wind-driven onshore sediment transport that is lost from the active upper-shoreface profile and helps constructing dune ridges. The other contribution is a virtual loss due to the Bruun effect. This virtual loss can be quantified theoretically by the Bruun Rule, and is validated empirically by Mimura and Nobuoka (1995) and Zhang et al. (2004). Assuming an active depth of 10 m it amounts to 500 to a 1000 times the sea-level

Table I

Typical cross-shore sediment transport towards the Dutch and Australian active upper shoreface and associated shoreline changes

Integrated gross	O(10^2) m^3/m/year or /extreme event
Integrated net	O(10^1) m^3/m/year
Effective profile height	10 m
Net natural shoreline changes	O(1) m/year
Net gross shoreline changes (extreme events or adverse years)	O(10) m/year or /extreme event

rise rate (in m^3/m/unit time). In present conditions of near stillstand conditions sea level rise rates of typically 10 cm/century or 0.1 cm/year occur on otherwise stable coasts. This leads to a shoreline retreat due to the Bruun effect of 0.5 to 1 m/year, and an associated virtual loss of 5 to 10 m^3/m/year. The loss due to aeolian transport is of similar magnitude (Cowell et al., 2003a). So, on average long-term cumulative losses in cross-shore direction are 10 to 20 m^3/m/year, while Australian and Dutch observations lead to net (positive) cross-shore contributions of the order of 5 to 20 m^3/m/year. Hence, onshore asymmetry- and boundary layer induced onshore transport on the middle shoreface should amount from 15 to 40 m^3/m/year.

In the absence of longshore gradients one might therefore observe net natural and net gross shoreline changes as indicated in Table I. The gross changes are associated with extreme events, which cause dune erosion. These changes will be restored on the longer term, when no upper shoreface losses due to alongshore transport gradients occur (List and Farris, 1999). These figures are validated for the Dutch and Australian coasts, but are expected to be general for moderate (low boundary figures) to high (high boundary figures) energy coasts.

What can we conclude from these cross-shore quantifications? Under present sea-level rise conditions the Bruun effect is not negligible, but of similar magnitude as other effects. Obviously, when sea-level rise rates multiply by five the cross-shore effect is not negligible. It will lead to an overruling of the shoreface feeding and advance will turn into retreat.

Let us now consider shoreline changes due to gradients in longshore sediment transports. We distinguish low- and high-energy coasts in terms of wave energy, and we assume wave-induced surfzone longshore flow to be the driving agent. This is a safe assumption along coasts that are not influenced or interrupted by coastal inlets and associated tidal basins or lagoons.

Tables II (low energy coasts) and III (high energy coasts) summarize typical cross-shore integrated longshore sediment transport rates, both net (average over many years) and gross (average over events or adverse years), natural and human-induced length scale variations and associated gradients and resulting shoreline

Table II

Typical longshore sediment transport in the surfzone of low-energy coasts (e.g., the Mediterranean coast)

Integrated surfzone gross	O(10^5) m^3/year or /extreme event
Integrated surfzone net	O(10^4) m^3/year
Natural length scale of variations	10 km (long term scale)
Human-induced length scale of variations	1–10 km (medium-term scale)
Net natural gradients	1 m^3/m/year
Net human-induced gradients	1–10 m^3/m/year
Effective profile height	10 m
Net natural shoreline changes	O(0.1) m/year
Net human-induced shoreline changes	O(0.1–1) m/year

Table III

Typical longshore sediment transport in the surfzone of high-energy coasts (e.g., Holland coast, Eastern U.S. coast)

Integrated surfzone gross	O(10^6) m^3/year or /extreme event
Integrated surfzone net	O(10^5) m^3/year
Natural length scale of variations	10–100 km (long term scale)
Human-induced length scale of variations	1–10 km (medium-term scale)
Net natural gradients	1–10 m^3/m/year
Net human-induced gradients	10–100 m^3/m/year
Effective profile height	10 m
Net natural shoreline changes	O(0.1–1) m/year
Net human-induced shoreline changes	O(1–10) m/year

changes. The length scale of natural variations, such as coastline curvature, are usually an order-of-magnitude larger than human-induced length scale variations, such as due to harbor moles and shore protection structures.

These results reveal that cross-shore effects dominate the problem of coastal erosion in case of low-energy natural coasts, while on high-energy natural coasts it is a combination of cross-shore and longshore effects. In the case of human-induced changes it is a combination on low-energy coasts, while on high-energy coasts longshore effects are dominant. This may explain why on low-energy coasts cross-shore impacting structures, such as offshore breakwaters and perched beaches, perform better than on high-energy coasts. It is also clear that both on low- and high-energy coasts the Bruun effect is of similar magnitude as other effects, except for human-induced changes on high-energy coasts. Again, when a multiplication of sea-level rise rates of about 5 occurs, increased erosion or decreased advance will be noticeable.

3. Backbarrier Sources and Sinks

In the above quantifications of cross-shore and longshore effects we have not considered the potential role of the source/sink term in Equation (5) (the last term). This term plays a role in case the backbarrier consists of a river, estuary or a tidal lagoon. In case of a river or an estuary there may be a natural supply of sediment to the coastal system that can compensate all above-mentioned cross-shore and longshore losses. As a result a deltaic formation will result, the evolution of which depends on the relative role of shaping forces due to waves, tides and river flow. In the Holocene many deltas have been outbuilding as a result of abundant sediment supply due to erosion of the catchment basin. Over the last decades many deltas, with few exceptions, are disintegrating due to human intervention in the form of dams regulating river discharge, which leads to a cutoff in the supply.

In case of a tidal lagoon or an estuary with little fresh water discharge the backbarrier tidal basin area may act as a source or a sink for the coastal sediment budget. Interesting examples of sink behavior are the Frisian Wadden basins along the Dutch and West German North Sea coast. Dronkers (1998) analyzed the net sediment transport behavior of these basins and shows that these basins are generally flood dominant, i.e., there is a tendency to accumulate sediment as sea level rises restoring a dynamic equilibrium geometry. Stive and Wang (2003) further analyzed this response and show that in this case the Bruun Rule expressing the impact of sea-level rise on inlet-influenced coasts can be extended as follows:

$$c_p = \frac{\partial MSL}{\partial t}\frac{L_*}{h_*} + \frac{\partial MSL}{\partial t}\frac{A_b}{h_* L_{ac}}\,, \tag{7}$$

where A_b is the tidal basin area and L_{ac} is the length of the adjacent coast impacted.

In the above equation the first term on the right-handside expresses the Bruun effect (Bruun, 1962) and the second term expresses the basin accommodation effect. The Bruun effect is exceeded by the basin effect as soon as:

$$A_b > L_* L_{ac}\,. \tag{8}$$

Typical orders of magnitude for L_* and L_{ac} are 1 km and 10 km respectively, so that basin areas larger than O (10 km^2) cause an impact on shoreline recession rates which exceeds the direct impact due to the Bruun effect.

Friedrichs and Aubrey (1988) made an analysis similar to Dronkers (1998) for a large number of schematized eastern U.S.A. tidal basins. They showed that depending on the basin hypsometry basins can be either flood- or ebb-dominant. This

Table IV

Typical longshore sediment transport in the surfzone of a high energy barrier coast, e.g., the Frisian Wadden coast or the Eastern U.S. coast

Integrated surfzone gross	O(10^6) m^3/year or /extreme event
Integrated surfzone net	O($0.5 * 10^6$) m^3/year
Natural length scale of variations	10 km (long term scale)
Human-induced length scale of variations	1–10 km (medium-term scale)
Net natural gradients	50 m^3/m/year
Net human-induced changes	50–500 m^3/m/year
Effective profile height	10 m
Net natural shoreline changes	O(5) m/year
Net human-induced shoreline changes	O(5–50) m/year

implies that sea-level rise may both lead to importing and exporting basins. When studying Figure 3 of Zhang et al. (2004) it appears that large stretches of coast, which they denote as inlet-influenced, experience stronger recession rates as the non-inlet influenced, Bruun Rule affected stretches. This gives rise to the hypothesis that the backbarrier basins along that coast are flood-dominant. However, when the basins are ebb-dominant sea-level rise may cause an export of sediment. This will decrease or even compensate the Bruun effect to such extent that shoreline advance may occur.

Next we will try to quantify the shoreline response to flood-dominant tidal basins. In case sea-level rise forces positive accommodation in flood-dominant basins sediment import is delivered by surfzone generated sediment transport that is carried into the tidal basin by the flood-currents. These sediments will be trapped when basins have basin areas of over 10 km^2. Typical barrier length scales are O (10 km) which allows to indicate typical figures for shoreline changes as given in Table IV.

What do we observe from these figures? Coastal stretches interrupted by inlets are influenced by the associated tidal basins in a strong way. The amplitude of these changes is significantly larger than on non inlet-influenced shoreline (cf. Figure 3 of Zhang et al., 2004).

4. An Innovative Approach: The Coastal Tract

We have discussed a wide range of processes relevant in the context of the evolution of coastal morphology over decades to millennia (low-order coastal change). This type of coastal change involves parts of the coast normally ignored in predictions required for management of coastal morphology: i.e., shoreline evolution linked to behaviour of the continental shelf and coastal plain. How can we deal with this complexity?

Cowell et al. (2003b) therefore introduce a meta-morphology, the *coastal tract*, defined as the morphological composite comprising the lower shoreface, upper shoreface and backbarrier (where present). It is the first order-system within a cascade hierarchy that provides a framework for aggregation of processes in modelling low-order coastal change. This framework is used in defining boundary conditions and internal dynamics to separate low-order from higher-order coastal behaviour for site-specific cases. This procedure involves preparation of a data-model by templating site data into a structure that complies with scale-specific properties of any given predictive models.

Each level of the *coastal-tract cascade* is distinguished as a system that shares sediments internally. This sediment sharing constrains morphological responses of the system on a given scale. The internal dynamics of these responses involve morphological coupling of the upper shoreface to the backbarrier and to the lower shoreface. The coupling mechanisms govern systematic lateral displacements of the shoreface, and therefore determine trends in shoreline advance and retreat. These changes manifest as the most fundamental modes of coastal evolution upon which higher-order (shorter-term, i.e., subdecadal scale) changes are superimposed.

Prediction of shoreline change adopts different approaches, depending on the space and time scale over which predictions are required. For short-term (subdecadal) coastal change (event and synoptic-scale changes occurring over hours through seasons to years), the focus is generally on the local sediment dynamics. These affect the shoreline planform and the across-shore profile (e.g., shoreline and profile models) in response to fluctuations in environmental conditions (i.e., the wave climate, littoral sediment budgets, sea level and the effects of anthropogenic activities). Theoretical and empirical approaches to these sub-decadal time scales generally focus on changes to the upper shoreface (defined loosely as the *active zone*; cf. Stive and De Vriend, 1995), which correlate with shoreline movements. These changes are moderated by littoral sediment budgets and by sediment 'production' via shoreline erosion cutting into onshore sand reserves (e.g., eroding dunes or cliffs), or through artificial nourishment of beaches.

The practical imperative for long-term prediction (decades or longer), requires an expanded scope as included in the coastal tract concept that includes the lower shoreface and the interaction between the shoreface and backshore environments (Figure 2). The upper shoreface has cross-shore length scales that are typically two to three orders of magnitude less than for the lower shoreface (depicted in Figure 2). This scale difference means that changes on the lower shoreface are associated with disproportionately larger changes on the upper shoreface, due to mass continuity for sediment exchanges between the two zones (Roy et al., 1994; Cowell et al., 1999). The upper shoreface is subject to a similar interaction with

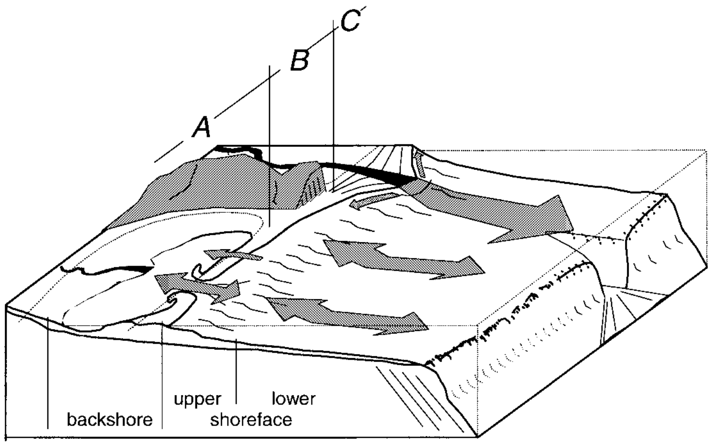

Figure 2. Physical morphology encompassed by the coastal tract (after Cowell et al., 2003b; see text for explanation).

the backshore, which comprises a morphologically active zone located between the upper shoreface (ocean beach) and the mainland. This zone may variously include dunes, washover surfaces, flood-tide deltas, lagoonal basin, tidal flats (Figure 2a), mainland beaches (Figure 2b) and fluvial deltas (Figure 2c). Each of these may be present or absent, depending on local conditions, especially the regional substrate slope (Roy et al., 1994; Cowell et al., 1995).

The sediment exchanges depicted by the arrows in Figure 2 occur in principle during any average year and on all time scales longer than this. These exchanges are summarised schematically in Figure 3, which differentiates sediment fluxes into sand and mud fractions. For coastal change on any scale, antecedent morphology, sea-level change and littoral sediment budgets can be regarded as boundary conditions for the coastal area of interest.

For sub-decadal prediction of horizontal movements in the upper shoreface, sand exchanges with the lower shoreface (Figure 3b) are usually ignored because these fluxes are so small that resulting morphological change is negligible: i.e., the annual closure-depth concept (Hallermeier, 1981; Nicholls et al., 1998). The fluxes of fine sediments (Figures 3c,d) are not directly relevant to the upper-shoreface sediment budget because mud deposition there is negligible. For long-term predictions, like on the scale of climate change, however, none of the internal sediment exchanges depicted in Figure 3 can be ignored. This is because systematic residual fluxes, that are small on the sub-decadal time scale, eventually cumulate through time enough to produce non-negligible (i.e., measurable) morphological changes. Moreover, the changes in morphology of the backbarrier, lower shoreface and upper shoreface cause these three zones to interact dynamically: i.e., the sediment exchanges themselves become influenced by the morphological changes.

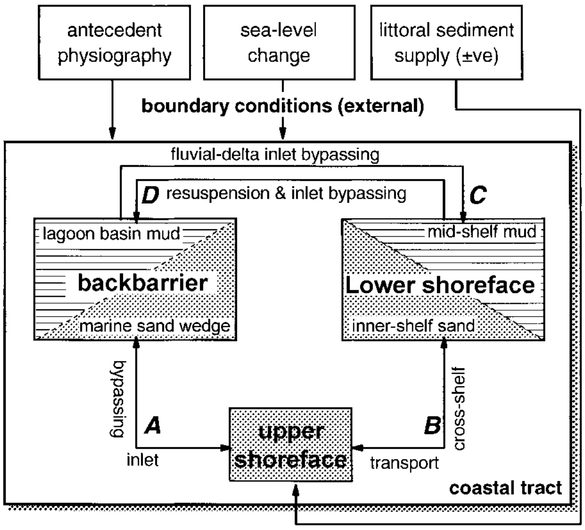

Figure 3. Schematic representation of mechanisms steering the location of the upper shoreface (after Cowell et al., 2003b).

5. Closure

The impact of sea-level rise as suggested by Bruun (1962) seems theoretically sound physically. Under present near still-stand sea level rise conditions the effect can easily be sub-ordinate to a host of other processes. It is therefore difficult to find firm proof for the Bruun Rule, and the study of Zhang et al. (2004) is an admirable contribution in this context. The fact that generally the Bruun effect is sub-ordinate to other effects under present rates of sea-level rise is the probable reason that considerable debate exists in the U.S. of its value (Pilkey et al., 2000). The Bruun Rule validation study of Mimura and Nobuoka (1995) could bypass the problem of hiding by other effects. They studied a part of the Japanese Niigata coast that experienced considerable subsidence, viz. approximately 0.1 m/year during 15 years, which is two orders of magnitude stronger than present eustatic rates. It is therefore fair to conclude that when sea-level rise rates will increase in accordance with our expectations, say a factor of five, the Bruun effect must be observed very clearly along non inlet-influenced shorelines.

Along inlet-influenced shorelines the Bruun effect is expected to be overruled by the response of the tidal basins associated with the inlets. In case of tidal basins with basin areas larger than 10 km^2 the Bruun effect is smaller than the effect due to the response of the basins to sea-level rise. In case of flood-dominant basins the shoreline recession will be much stronger, while in case of ebb-dominant basins shoreline recession may be much less or even revert to shoreline advance. These responses are not well studied and it is therefore advocated that the focus of research regarding response of shorelines to sea level rise is redirected toward inlet-influenced coasts.

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