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Klervi Hamon-Kerivel (1) *, Andrew Cooper (1,2), Derek Jackson (1,2), Mouncef Sedrati (3), Emilia Guisado Pintado (4,1)

(1) School of Geography & Environmental Sciences, Ulster University, Coleraine, UK,

(2) Discipline of Geology, University of KwaZulu-Natal, Westville Campus, Private Bag X54001, Durban 4000, South Africa

(3) Équipe GMGL-Domaines Océaniques UMR 6538, Université de Bretagne Sud, Vannes, France,

(4) Geography Regional Analysis and Physical Geography, University of Seville, Seville, Spain

*Corresponding author: School of Geography & Environmental Sciences, Ulster University, Cromore Road, Coleraine BT52 1SA, UK; Email Address: Hamon_Kerivel-k@ulster.ac.uk

Abstract

At seasonal to century timescales (mesoscale), the shoreface is a critical zone seaward of the surf zone and/or beachface, in which waves interact with the mobile seafloor to cause morphological change. This has important (and often unacknowledged) implications for adjacent shoreline form and behaviour both now and in the near-future. The shoreface has been relatively little studied from a mesoscale morphodynamic (morphological change over time) perspective and various definitions exist regarding its extent and morphodynamic subdivisions. To overcome the diversity and ambiguity of existing definitions we propose a standard terminology involving the external limits and subordinate zones of the shoreface.

In our definition, the landward limit of the shoreface coincides with the seaward limit of the fair weather surf zone, and where no surf zone is present, the base of the beachface. The shoreface itself is subdivided into upper and lower shorefaces, separated by the depth of closure (DoC) as defined by Hallermeier (1981). The seaward limit of the lower shoreface is defined by the limit of significant sediment transport, indicated by bed shear stress according to Valiente et al. (2019). All boundaries are temporally variable according to wave characteristics and timescale of study.

The upper shoreface is dynamic at seasonal to annual timescales and interacts with the adjacent surfzone via wave transformation and two-way sediment exchange. The lower shoreface is dynamic at decadal to millennial timescales and it interacts with the adjacent upper shoreface and inner shelf. The upper shoreface is strongly influenced by wave hydrodynamics whereas the lower shoreface is less dynamic and its shape is more heavily influenced by geological factors (nature and/or abundance of sediment, depth and erodibility of rock outcrop, etc.). Sediment exchange both within the shoreface and between shoreface and adjacent environments is strongly event-driven. Longshore, onshore and offshore transport mechanisms have been documented.

The shoreface profile influences, and is influenced by, wave transformation, however, the widely adopted shoreface equilibrium profile is not universally applicable. Instead, a diversity of shoreface morphologies exists in two and three dimensions. These are likely related to sediment supply and accommodation and we propose a spectrum of shoreface types based on these variables. Recent studies have shown that large-scale 3-D forms (e.g. shoreface-connected ridges and sorted bedforms) strongly influence shoreline behaviour, however, the dynamics of these shoreface bedforms requires further investigation. Each type of shoreface likely exhibits distinctive behaviour at the mesoscale (time scale of 10^1 to 10^2 years and a spatial scale of 10^1 to 10^2km). This is proposed as a unifying model with which to integrate studies of shoreface dynamics at different spatial and temporal scales.

Keywords

Shoreface morphodynamics

Depth of closure

Mesoscale

Shoreline

Shoreface Connected Ridges

1. Introduction

The shoreface is a transitional zone between the continental shelf and the shoreline, in which waves (particularly long period waves) begin to strongly interact with the seabed. These shoreface interactions "condition" the waves, altering their deep water parameters, before they reach the surf zone where more significant wave-seabed interactions (attenuation, breaking and generation of secondary wave motions) occur (Wright and Short, 1984). The contemporary shoreface is difficult to access and direct measurements have been eclipsed until recently by a strong scientific and engineering focus on the surf zone and beach face (swash zone) and areas to landward. Geologically, the shoreface comprises a series of sandy deposits which reflect the influence of waves, currents and organisms on the palaeo-environment (Dashtgard et al., 2009; Hampson and Storms, 2003).

The influence of the shoreface on incoming wave transformation, particularly during storms, is widely recognised (Backstrom et al., 2015; Héquette and Hill, 1993; Niedoroda et al., 1984; Swift et al., 1985) and has been the subject of research attention in recent years (Backstrom et al., 2015). Interest has also been stimulated by recognition of the shoreface's role in coastal response to sea-level rise (Cooper et al., 2018) and its role as a reservoir of sediment accumulated during and since the Holocene transgression (Kinsela et al., 2016). Ancient shoreface sediments have also been studied because they can form hydrocarbon reservoirs (Howell et al., 2008). This paper reviews past research and recent progress in understanding the shoreface in the context of the wider coastal system. It reviews linkages between the shoreface and adjacent surf zone and shoreline and explores the variety of shoreface morphologies and the influences upon it such as geological control and sediment availability.

This paper is largely concerned with shoreface morphology and mesoscale (dynamics which corresponds to a time scale of 10^1 to 10^2 years; and a spatial scale of 10^1 to 10^2 km) which is the typical spatiotemporal scale of morphological change. From a review of existing literature we propose a consistent definition and present a new classification according to shoreface morphologies,. The lack of a uniform definition of "shoreface", and its variation in meaning for different disciplines (coastal

engineering, geology, oceanography…) has led to discrepancies in terminology. While this is addressed in the next section of the review , from this point on we refer to the shoreface as the transitional zone between shelf and beach, delimited on the shoreward limited shore side by the (fairweather) surf zone (or swash zone) and seaward by the threshold beyond which sediment transport is no longer intense enough to contribute to the overall coastal dynamics at a mesoscale (time scale of 10^1 to 10^2 years and a spatial scale of 10^1 to 10^2km). However the shoreface exhibits changes at other spatio-temporal scale, this will be explicated below.

2. Definition of the shoreface

The shoreface can be loosely defined as the transition zone between the beach and the shelf (**Figure 1**). Definitions of its shoreward and seaward limits and internal subdivision are still not fully agreed upon. Some authors (Ortiz and Ashton, 2016; Thieler et al., 2001, 1995) suggest the subaerial coastal zone itself represents the shoreward boundary, thus including the surf zone (Niedoroda et al., 1984) or the lower limit of the swash (Aagaard et al., 2013). Others (Backstrom et al., 2007; Héquette and Hill, 1993; Héquette et al., 2001; Wright et al., 1991, 1991) place the landward boundary between the shoreface and the surf zone. Green et al. (2004) refer specifically to the 'breakpoint' as the landward limit of the shoreface. On beaches without surf zones e.g. Caribbean beaches (Boon and Green, 1988), the shoreface might extend landward to the beach step and swash zone.

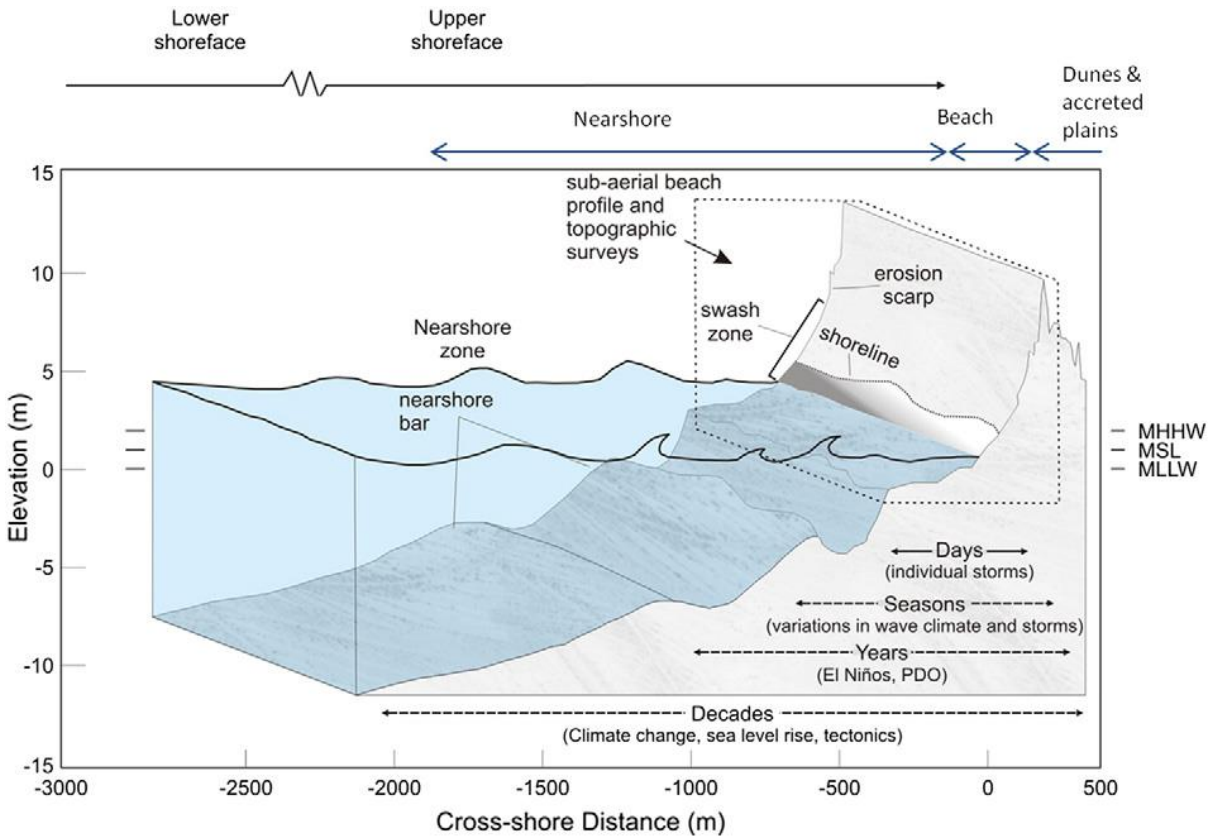


***Figure 1**. (colour) Conceptual diagram of the current definition and limits of the shoreface (from Ruggiero et al. (2016))*

The seaward limit is equally contentious due largely to the fact that it is often hard to identify a definitive morphological feature (e.g. a break in slope) or change in texture between the shoreface and the continental shelf. Without a physical reference location the spatially and temporally variable point at which waves start to mobilise sediment (i.e.in other words the wave base where waves begin to interact with the sea bed sediment) is normally used (Aagaard, 2014; Dashtgard et al., 2012; Stive and De Vriend, 1995). Wave base is temporally variable and hence a definition of the timescale for assessment of wave base is required (see section 4.).

The shoreface can also be defined in terms of the physical processes that occur within it. Héquette et al. (2008a, p. 227) proposed that the shoreface is 'the shallow, friction-dominated zone where the response of the seabed to hydro-meteorological forcing results from the complex interactions between wind, waves and currents'. In this process-based approach, the shoreface comprises two sections: the *upper shoreface* which is influenced by wave-breaking processes and the dissipation of

wave energy (i.e. the surf zone in other definitions) and the *lower shoreface* that is dominated by wave shoaling (Aagaard, 2014; Stive and De Vriend, 1995). The shoreface is a wave-affected zone (Swift et al., 1985; Wright et al., 1991), even at its seaward, deepest point. Incoming waves impact the bottom from a starting point commonly defined as the wave base until they reach the shore. This definition is similar to the profile zonation proposed by Hallermeier (1981) that introduced potentially calculable depths for the boundaries between upper and lower shoreface and between lower shoreface and shelf (Cowell et al., 1999) based on wave parameters. In this definition, the upper shoreface is the zone most likely to experience strong seasonal variability and is therefore equivalent to the shoreface 'active zone' as defined by Stive and De Vriend (1995), with its seaward boundary being the depth of closure as calculated using Hallermeier (1978) formula. In this zonation, the lower shoreface is bounded seaward by the calculated maximum water depth for initiation of motion (wave base) under annual median wave conditions. However, the depth of closure and maximum depth for wave-initiated motion are temporally variable parameters and cannot be considered an absolute calculable boundary for the shoreface.

Other authors have proposed discriminating the shoreface according to slope (Swift et al., 1985) or sediment type and characteristics (Aragonés et al., 2018; Cowell et al., 1999; George and Hill, 2008). Larson et al. (1999) examined the theoretical profile under breaking (surf zone, upper shoreface) and non-breaking (shoaling wave condition, lower shoreface) and determined that the 'equilibrium profile' (see section 3) of the lower shoreface is for that instance distinguished from the upper shoreface by a lower curvature than the one of the upper shoreface. Indeed, the separation between upper and lower shoreface zones is sometime set at a discernible break in slope (Niedoroda et al., 1984).

In general use, the term 'shoreface', when not more fully specified, appears to refer to the lower shoreface, as most authors equate it with an environment dominated by wave shoaling that is active only at a long time scale, i.e decades and longer(Aagaard, 2014; Green et al., 2004; Héquette et al., 2008a; Niedoroda et al., 1984). In current usage, the 'upper shoreface' therefore corresponds to a

seasonally active zone that sometimes, but not always, includes the surf zone. Typically, the upper shoreface is the zone influenced by the offshore migration of the surf zone during storm conditions that results in morphological changes at a seasonal time scale. This inconsistency in terminology regarding the shoreface can be confusing Subsequently, we propose a standard definition.

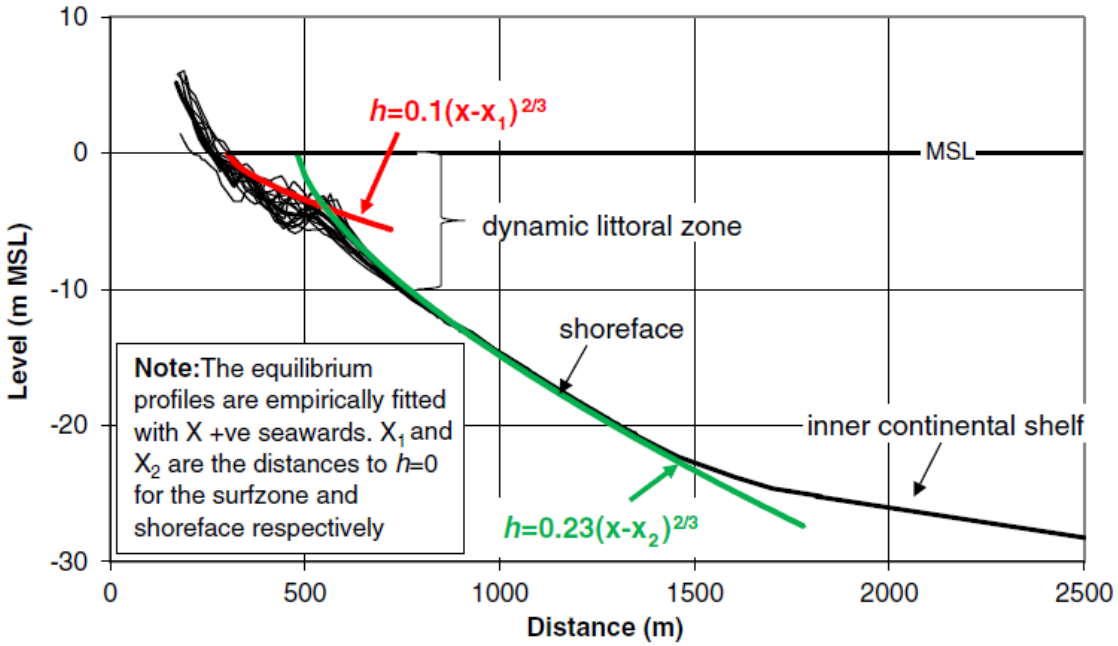
3. Shoreface equilibrium profile

Most models of beach/shoreface profiles are based on the assumption of the existence of an equilibrium profile. The equilibrium profile in the literature has its origins as far back as Fenneman (1902, p. 1) who wrote: “There is a profile of equilibrium which the water would ultimately impart, if allowed to carry its work to completion”. This concept is still widespread in coastal evolution models, especially in engineering applications (Cooper and Pilkey, 2004).

The profile of equilibrium can be defined as the theoretical, two-dimensional shape achieved by a sand-rich, unconstrained coastal system under a given stable wave climate and sediment type. These are the conditions of flume tank experiments that were first used to explore the concept of equilibrium profile (Dean, 1991). In theory, when the system is in equilibrium then no net sand transport occurs and the rate of dissipation is constant. It is the result of an equilibrium between onshore directed (constructive) sediment transport and offshore (destructive) sediment flow and is usually concave upward as described by Dean (1991). Its form is described by an empirical equation of the form:

$$h(y) = Ay^n$$

where: *h(y)* is the water depth *h* at distance offshore *y*, *A* is a scale parameter (dependent on grain size) and ‘*n*’ is a parameter commonly set at 2/3 (the best fit to field data found by (Dean, 1977)) (**Figure 2**). The profile of equilibrium of a coastal system can also be obtained from field data by analysing a time series of cross-shore profiles (Birkemeier, 1985; Hinton and Nicholls, 1999; Nicholls et al., 1998; Różyński et al., 1999).



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159 ***Figure 2****.(colour) Example of equilibrium profiles fitted for the surf zone and shoreface on the Australian Gold Coast (from:*
160 *Patterson and Nielsen 2016). In this example, the dynamic littoral zone equates to the surf zone and upper shoreface.*

161 Pilkey et al. (1993) reviewed the principle of equilibrium profile and demonstrated that the conditions
162 (i.e. a sand-rich system not affected by transient sand bodies or underlying geology and closed by an
163 offshore boundary) that are required for an equilibrium are unrealistic assumptions in a natural
164 environment. Pilkey et al. (1993) also pointed out that the equilibrium equation is flawed because the
165 only variable influencing the shape of the profile is the grain size (expressed as *A*) and the wave climate
166 and inherited morphology have no influence. Similarly, Thieler et al. (1995), pointed out the
167 inconsistent grain size across the shoreface which cannot be accounted for by a single number (*A* in
168 Dean's equation). They also showed that the apparent relationship between '*A*' and grainsize is not
169 evident when only sand grain sizes are considered. Those authors also demonstrated the major control
170 played by underlying geology on shoreface morphology.

171 Backstrom et al. (2007) showed from field measurements that adjacent shorefaces could have
172 markedly different shapes that diverged from potential 'equilibrium' profiles. While one of the sites
173 displayed a concave-up type morphology closer to 'typical equilibrium profile' the other had a linear
174 to convex profile. The presence of other convex morphologies on shorefaces worldwide (Aleman et

al., 2015; Green et al., 2004; McNinch, 2004; Miselis and McNinch, 2006) rules out the concave upward equilibrium profile morphology as a universal principle, as do shorefaces whose morphologies are influenced by rock outcrop and sub-crop (Browder and McNinch, 2006; Riggs et al., 1995; Thieler et al., 1995).

While the principle of equilibrium is a valid concept to describe the evolution of an idealized coastal system, it appears to have limited applicability for "real-world" situations (Athanasiou et al., 2019; Pilkey et al., 1993; Thieler et al., 1995). This is mostly because the forces that affect morphology and drive morphological change on the shoreface appear to be controlled, at least in part, by factors other than waves and grain size. Clifton (2005), states that for a transgressive shoreface profile (which make up the majority of the world's coasts (Bird, 1985)) only the upper part is "wave-tuned" and theoretically able to attain equilibrium, while the lower part of the profile is shaped by differential erosion due to the presence of older deposits.

Extensive research has been conducted on the "active" part of the profile and its influence on shoreline and beach dynamics. This is particularly pertinent when considering seasonal variation, event scale (storm) dynamics and shoreline variability. Most coastal evolution models also focus on this zone because the lower part of the shoreface lies beyond the "depth of closure" and its dynamics are considered to only be significant during storm events when sand is 'lost' (i.e. transported from the upper shoreface to the lower shoreface) (Cowell et al., 1995; Stive and De Vriend, 1995).

While the shoreface has mostly been considered two-dimensionally, it has become more evident that longshore variations often exist and that many shorefaces are strongly three-dimensional. This is related to the presence of various sedimentary features (see section 7) and/or the influence of the geological framework (see section 6).

4. Depth of closure and the shoreface

The depth of closure commonly refers to the seaward limit at which morphological change is not observable for a (usually seasonal) period of time; it also commonly is regarded as the boundary between upper and lower shoreface (Cowell et al., 1999)

The depth of closure can be determined by analysing a time series of profiles at a given location (Różyński et al., 1999) or consecutive bathymetry datasets (Robertson et al., 2007) to identify the depth at which the change in morphology becomes insignificant (not measurable), keeping in mind that the depth of closure is related to the time-scale of the measurements and to the measurement accuracy (Hoekstra et al., 1999). Various statistical methods also exist to identify this threshold (Kraus and Harikai, 1983; Nicholls et al., 1997). However, most commonly, and because datasets necessary to determine the depth of closure are relatively rare, the DoC is usually calculated from wave parameters.

Depth of closure calculations were introduced by Hallermeier (1981, 1978). Hallermeier's DoC calculation is based on the wave period and height and determines the closure depth (*dl*) for an annual wave return period.

$$dI = 2.28Hs - 68.5\left(\frac{Hs^2}{gTs^2}\right)$$

With *H*s the local significant wave height and *Ts* the significant wave period.

Related to depth of closure is the wave base, referred to by Hallermeier (1981) as the outer depth of closure (*di*) (as opposed to the DoC which is referred to as the inner depth of closure (*dl*)) (**Figure 3**). The outer DoC was defined as "maximum depth for motion initiation by median wave conditions" by (Hallermeier, 1981) and corresponds to the limit of wave-initiated cross-shore sediment transport, or the seaward limit of the wave-constructed profile. It is defined as the depth at which the incident wave reaching the shore begins to shoal and can be calculated using:

$$di = (Hs - 0.3\,\sigma)\,Ts\left(\frac{g}{5000D}\right)^{0.5}$$

With Hs the mean significant wave height, σ the standard deviation of significant height, *Ts* the mean significant period, *g* the acceleration of gravity and *D* the sand diameter.

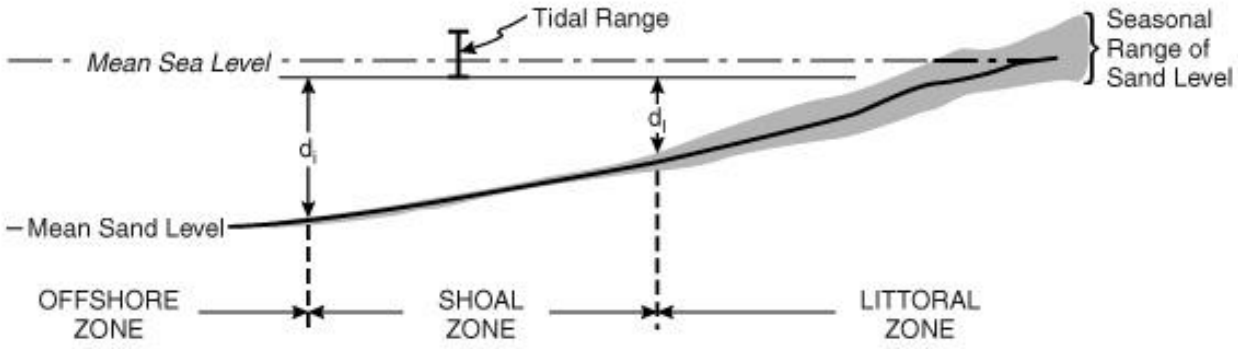


Figure 3. *Definition of inner and outer depth of closure (Hallermeier, 1981) with dl the inner depth of closure (DoC) and di the outer depth of closure or wave base. Here, the littoral zone corresponds to the upper shoreface and the lower shoreface is included in the shoal zone.*

Subsequent studies (Nicholls et al., 1997) building on the work of Hallermeier (1981, 1978) have shown that the depth at which closure is calculated is highly time-variable, and it increases with the time-scale involved (Nicholls et al., 1999, 1998). It is therefore customary to associate a depth of closure with a temporal dimension. The DoC can be calculated for a specific event (Robertson et al., 2008), but more commonly it is calculated for yearly/ seasonal variations or even over a number of years (Valiente et al., 2019) up to decades (Hinton and Nicholls, 1999) and even millennia (Ortiz and Ashton, 2016). Similarly, the wave base depth is usually calculated for a typical year. And as with the DoC, it varies with the return period used for the calculation

The depth of closure is also known to be variable in the longshore (Hinton and Nicholls, 1999; Sabatier et al., 2005) with influences exerted by sedimentology and underlying geology (Robertson et al., 2008), as well as wave exposure.

While most of the literature has focused on the depth of closure, recent studies have proposed new calculations for the lower shoreface/ shelf threshold and advanced the definition of the seaward limit of the shoreface. For example. Ortiz and Ashton (2016) introduced the notion of morphodynamic

depth of closure (MDOC) defined as the depth beyond which 'evolution of the shoreface becomes geologically slow and the bed shape response to environmental change becomes virtually non-existent'. Ortiz and Aston's (2016) calculated MDOC is dependent on wave height and period but is not affected by the sediment grain size; it only influences the shape of the modelled equilibrium profile. This method of calculation, which proposes a calculation of the depth at which morphological variability is geologically slow (millennia), is relevant to studies of shoreface morphodynamics and more generally to the study of macro-scale (time scale $>10^1$ years and spatial scale $>10^2$km) coastal behaviour.

Valiente et al. (2019) used wave bed shear stress to estimate the seaward limit of significant sediment transport (contributing to the overall coastal dynamics). This depth of transport (DoT) outer limit takes into account both wave-induced stress and stress induced by tidal currents. Their approach makes the distinction between DoC which limits the zone of significant morphological change i.e. the upper shoreface (which can be approximated by Hallermeier's formulation) and the DoT that delimits the zone of significant sediment transport. The use of bed shear stress permits identification of the zone where sediment transport is significant in terms of volume but where no morphological changes are observable at the medium term(10^0 to 10^1 years). It is therefore a good potential proxy for the lower shoreface/shelf boundary, as it determines the extent of the active shoreface. In contrast, the wave base (DoC) only encompasses the wave-affected areas of the shelf where significant morphological change is observed.

For the upper/ lower shoreface limit, several methods of DoC identification exist. For example, Aragonés et al. (2018) proposed that DoC could be defined as 'the point where the minimum size of sediment is located'. Aragonés et al. (2019) further proposed a numerical model based on wave parameters and sedimentology to determine the depth of closure.

In the absence of actual measurements, calculated depth of closure is still widely used to characterise coastal systems, especially for engineering purposes. This approach, however, is not universally

applicable especially where rock outcrops or limited sand supply constrain sediment transport on the shoreface. Miselis and McNinch (2006), for example, showed that using calculated depth of closure can cause an over-estimation of the volume of nearshore sediment in geologically (vertically) constrained areas.

5. Shoreface hydrodynamics

Waves have an oscillatory motion that is modified by the interaction with seabed as they cross the shoreface. Incoming waves dissipate energy over the shoreface and this may lead to sediment transport and morphological evolution (Roelvink and Stive, 1991). When the critical bottom friction is reached, seabed sediment starts to be transported either as suspension or as bed load, making incoming waves one of the main driving factors for sediment transport in the coastal zone. Over the shoreface, because of diminishing depths, waves shoal before breaking in the surf zone. Shoaling creates an asymmetry in the wave oscillation, resulting in net sediment transport directed toward the shore (Austin et al., 2009; Huisman et al., 2019; Niedoroda and Swift, 1981). This is especially true under fair-weather conditions when wave oscillation asymmetry is the main process driving onshore sediment transport (Niedoroda and Swift, 1981; Niedoroda et al., 1984; Swift et al., 1985). Incoming waves can also create steady currents such as undertow and rip currents, particularly in the surf zone. These are strongly related to morphology and beach slope (Castelle et al., 2016). These flows are directed offshore and tend to transport sediment from the surf zone to the shoreface, particularly during high-energy storm conditions (Loureiro et al., 2012a). The importance of infragravity waves on cross-shore sediment transport is still unclear (Bakker et al., 2016; Bertin et al., 2018) but it has been hypothesised that they can contribute to offshore transport during storms (Russell, 1993).

Another factor influencing sediment transport in the coastal zone is wind. Currents generated by winds can be drivers of sediment transport, especially down-welling currents that, during storms, have the capacity to move sediment offshore (Héquette and Hill, 1993; Héquette et al., 2001; Niedoroda et

al., 1984). Swift et al. (1985) pointed out the importance of wind-driven coastal flow for the sediment budget, especially when intensified by storms, stating (p. 332) that “Because of the offshore component of bottom flow, sand is swept down the lower shoreface and onto the adjacent shelf. Fair weather processes may be unable to return storm-deposited sand to the beach from such an offshore position”.

Wind forcing may also affect the tidal current velocity by reinforcing or limiting it depending on the concordance of wind and current direction (Héquette et al., 2008a). It also causes near-bottom currents which affect sand ridge migration (Guerrero et al., 2018) (see below in section 7).

Tidal currents can also be a significant component in sediment transport on shorefaces. On microtidal shorefaces, tidal currents are generally overlooked. Moreover, since tidal currents are often symmetrical their significance can be limited when averaged over a tidal cycle (Niedoroda et al., 1984). On shorefaces where morphology creates tidal asymmetry, however, sediment transport from residual tidal currents is non-negligible whereas, in meso- and macrotidal environments, tidal currents are fundamental in driving sediment transport (King et al., 2019; Valiente et al., 2019). In their study of sediment transport on a tide dominated (macrotidal) shoreface environment, Héquette et al. (2008a) found that, although waves were the main forcing parameter in mobilising sediment as they cross the shoreface, sediment transport vectors were linked to average currents directed alongshore, in particular, tidal currents.

Hydrodynamics on the shoreface potentially involve a combination of wave, wind and tidal current action. These processes can interact via positive or negative feedback. For example, winds can play a role in enhancing or inhibiting tidal currents depending on the relative direction of both those forces (Héquette et al., 2008b). During storm wave events, the combined action of waves and winds tends to create offshore-directed currents that move sediment to the lower parts of the shoreface and shelf (Héquette et al., 2001; Niedoroda and Swift, 1981; Wright et al., 1991) whereas, under fair weather conditions, smaller waves tend to move sediment onshore (Wright et al., 1991). As most mean

currents (i.e. rip currents, undertow, down-welling) are directed seaward, onshore sediment transport only occurs under favourable conditions for a limited time. Wright et al. (1991, p. 46) found that at Duck beach (North Carolina, USA), storms were “capable of transporting more sand offshore in an hour than fairweather processes can move onshore in two or more days“.

Over longer time scales (10^0 to 10^1 years), the variability in shoreface morphology may be influenced by climatic factors such as ocean circulation patterns or oscillation. In particular, El Niño/ La Niña periods have been shown to have specific impacts (shift in sediment transport and morphological responses) on the coast and shoreface (Goodwin et al., 2013; Ruggiero et al., 2005; Wright et al., 1991). Other factors such as storm grouping (Lee et al., 1998; Loureiro et al., 2012b) also exert a control on shoreface dynamics. Indeed, a series of storms often has a more extreme effect than a single (even more energetic) storm in creating morphological changes and causing erosion of the beach and upper shoreface and transport of sediment to the lower shoreface (Guisado-Pintado and Jackson, 2019, 2018).

6. Geological influences on the shoreface

It has been shown that underlying geology is a limiting factor for the formation of equilibrium profiles (Pilkey et al., 1993), and various studies have emphasised the role of geology and coastal morphology on mesoscale coastal behaviour(Cooper et al., 2018; Gallop et al., 2020; Jackson et al., 2005). A large proportion of the world’s coast is rocky and/or embayed (laterally constrained) and there is much variability in sediment volume and thickness. The potential for erosion of the underlying lithology to produce shoreface sediment is also highly dependent on lithology and dynamics as well as the rate of sea-level rise.

Coasts can acquire equilibrium profiles if the shoreface is sand-rich and the underlying and offshore geology does not play any part (Pilkey et al., 1993). These conditions are unrealistic for most of the

world's shorefaces. Typically, for the barrier island coast of North Carolina, for example, Pilkey et al. (1993, p. 271) found that 'the shape of the shoreface in sediment-poor areas is determined by a complex interaction between underlying geology, modern sand cover and highly variable incoming wave climate'.

Other studies on North America's Atlantic coast confirm this observation. Riggs et al. (1995) showed that the geological framework determines 3D shoreface morphology, sediment texture and composition as well as shoreline recession rates, concluding (p 231) that 'each barrier beach and shoreface are total products of their geologic heritage; the signature of their history controls and influences the present morphology, shoreface dynamics, and rates of shoreline recession'. Thieler et al. (1995) examined the variation in grain size, underlying geology and sediment transport of Wrightsville Beach (North Carolina) and found a lack of homogeneity in sediment distribution and irregular bathymetry that also indicated a strong control by underlying geology on sediment transport processes on the shoreface.

The geological framework can influence both the volume of sediment available for transport but it can also in some region impede sediment transport reducing the amount of sediment contributing to a coastal system's sediment budget (Menier et al., 2019, 2016). Furthermore, the geological setting of the shoreface and the volume of shoreface sediment readily available for transport seem to have an influence on shoreline evolution pattern (see section 8.3).

For most coasts influenced by their geological setting, conventional theories of shoreface morphodynamics are inadequate. Indeed, in geologically constrained areas it has become evident that commonly accepted theory related to nearshore morphodynamics, such as equilibrium profiles (Pilkey et al., 1993; Riggs et al., 1995; Thieler et al., 1995), depth of closure (Robertson et al., 2008, 2007), or beach state parameters (Jackson et al., 2005; Loureiro et al., 2013) are not directly applicable.

7. Sedimentary Bedforms

Sedimentary bedforms on the shoreface occur at various length scales from ripples (cm) to fields of multi-metric sediment-rich features known as sand ridges (km), they are the result of complex hydrodynamics, including feedbacks between the seabed and tides/currents. Their formational mechanisms and self-organisational processes are still unclear (Durán et al., 2018; Guerrero et al., 2018; Murray et al., 2014b; Nnafie et al., 2015, 2014). In terms of shoreface dynamics, sedimentary bedforms can act as possible sources/sinks of sediment and can also influence sediment characteristics (sorting etc.) and nearshore hydrodynamics (via change bathymetry, bottom roughness etc.) (Latapy et al., 2020; Verwaest et al., 2020).

The main sedimentary features present on shorefaces are sand ridges (Madricardo and Rizzetto, 2018) . On macro-tidal coasts, sand ridges are mostly related to tidal hydrodynamics. They are termed tidal sand ridges or tidal sand banks or waves. They originate from the hydrodynamic climate created by strong tide-related currents, as stated by Anthony (2013, p. 10); “In the course of the Holocene, […] sand has been reworked by the interplay of tidal currents and storm waves into the impressive jumble of tidal sand ridges and banks that have served as sources for coastal accretion.” Tidal sand banks are usually sub-parallel to the shoreline and attuned to the local tidal current climate. They exist on the shoreface and can migrate onshore under storm conditions at the rate of several metres per year (Héquette and Aernouts, 2010; Héquette et al., 2013; Van der Molen and Van Dijck, 2000).

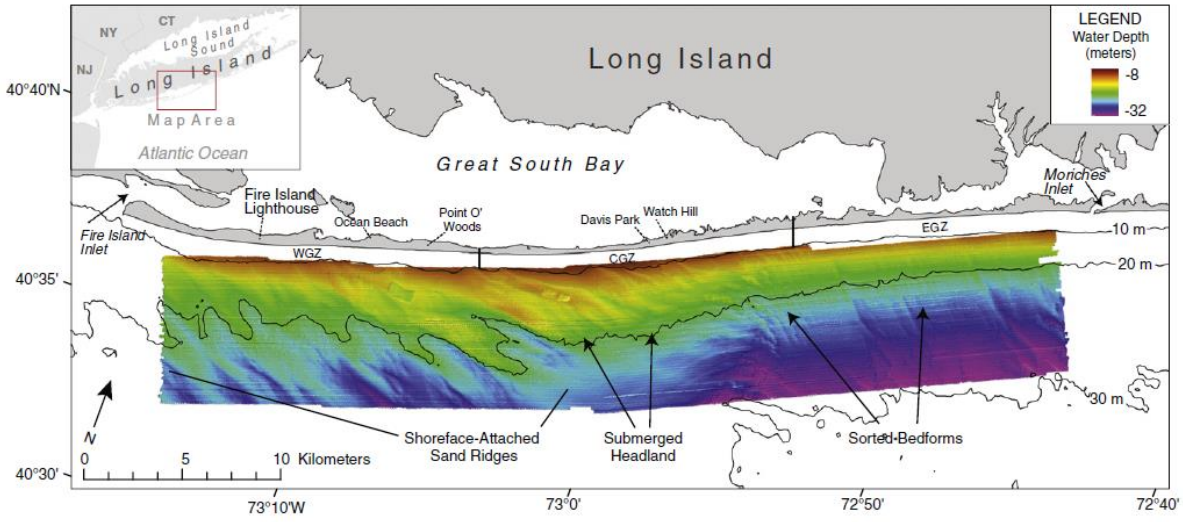
Sand ridges also exist on coasts where they are not the product of tidal hydrodynamics. In such cases they are referred to as sand ridges, shoreface-attached sand ridges, or shoreface-connected ridges (SCRs) (**Figure 4**). SCRs are shoreline-oblique, rhythmic sand ridges. The spacing between each ridge is in the 1-10 km order, their height can exceed 10 m and their length more than 25 km. Found at depths of 5-20 m with their landward end connected to the shoreface (hence the name),they often develop at an angle of around 20^{o} to the shoreline. They occur particularly on storm-influenced shelves and

389 can migrate with the dominant current at a rate of several meters per year (Calvete et al., 2001a,
390 2001b; Falques et al., 1999; van de Meene and van Rijn, 2000a, 2000b; Yoshikawa and Nemoto, 2014).

391 The formation processes of SCRs are rather unclear. Some suggest they are pre-Holocene
392 transgression relics (McBride and Moslow, 1991) while others have focused on the physical processes
393 necessary to create and sustain such bedforms (Calvete et al., 2001a, 2001b; Trowbridge, 1995; Vis-
394 Star et al., 2007). Their dynamics are complex (Browder and McNinch, 2006; Falques et al., 1999; Goff
395 et al., 2015; Guerrero et al., 2018; McNinch, 2004; Schupp et al., 2006) and many are assumed to have
396 formed at shallower depth and been subsequently drowned by rising sea level (Durán et al., 2018;
397 Nnafie et al., 2015, 2014).

398 Other shoreface features have been termed sorted bedforms or rippled scour depressions (RSDs)
399 (**Figure 4**). These are relatively stable features that have been widely documented. RSDs comprise
400 alternating sand and coarse gravel bands with shore-oblique to shore-normal orientation. The coarse
401 gravel is commonly rippled, while the surrounding sand is fine and featureless. Their significance,
402 formation and dynamics are still discussed (Liu et al., 2018; Mielck et al., 2015; Murray et al., 2014a;
403 Rosenberger et al., 2019). Some have been attributed to channelized offshore storm-return flows
404 (Pretorius et al., 2018) and they are thought to exhibit positive sorting feedback (Goff et al., 2005;
405 Green et al., 2004; Gutierrez et al., 2005; Murray and Thieler, 2004). Other smaller sediment structures
406 or sediment heterogeneities have also been described on the shoreface, such as ripples or gravel
407 outcrops (Browder and McNinch, 2006; Inman and Adams, 2005; Schupp et al., 2006).

408



409

410 ***Figure 4.*** *(colour) Example of sedimentary bedforms (shoreface-connected ridges & sorted bedforms) visible on the*
411 *bathymetry from Fire Island (USA) (From: Hapke et al., 2016)*

412 In some instances, formation and evolution of superficial sand bodies is linked to the geological
413 framework. For example, McNinch (2004) and Browder and McNinch (2006) found a correlation
414 between the presence of palaeochannels and shore oblique sandbars. They postulated that
415 inhomogeneity in the geological framework could influence the hydrodynamics and that variation in
416 the nature of the outcropping seafloor could influence the organisation of bedforms. Similarly, Durán
417 et al. (2018) postulate that irregularities in palaeotopography initiate the formation of and influence
418 the orientation and distribution of sand ridges.

419 The sedimentary features present on the shoreface can be an indication of the sediment supply
420 available. Thieler et al. (2014, p. 128) state that 'where modern sediment is relatively abundant, the
421 inner shelf contains shoreface-attached ridges and shoal complexes. Where modern sediment is
422 lacking, the seafloor is characterized by sorted bedforms'. Bedforms and sand bodies, as they develop
423 and evolve under a specific set of conditions, are morphological indicators of the physical environment
424 on the shoreface, such as the hydrodynamic climate or the sediment availability. More importantly,
425 sedimentary features present on the shoreface also seem to influence coastal morphodynamics and
426 shoreline evolution in many ways (see section 8.3).

427

8. Shoreface morphodynamics

8.1 Shoreface zonation: Upper/Lower shoreface

The division between upper and lower shoreface morphodynamics is reflected in the notion of depth of closure (and associated wave base) as described by Hallermeier (1981). A number of studies on depth of closure exist in the literature (Birkemeier, 1985; Li et al., 2005; Nicholls et al., 1998, 1997; Robertson et al., 2008), and a selection of studies in contrasting environments are reviewed here.

Within a micro-meso tidal environment on the Dutch coast where mean tidal range varied from 1.4m to 1.7m and peak tidal velocity reached 1 ms^{-1}, Hinton and Nicholls (1999) analysed a long-term dataset (20 years) of shoreface profiles to identify the depth of closure for various sections. The observed DoC ranged between 5.0 and 9.2m. Hallermeier's DoC (using a 5-year period) was also calculated, and predicted a DoC at 9.2m. Hinton and Nicholls (1999), however, also observed significant morphological variation seaward of this depth of closure. This morphological variability was only observed over longer time scales (> 10^1 years) and at depths greater than 12m. This suggested that the shoreface displays two distinct behaviours occurring at different time scales, highlighting the different behaviour of the upper and lower shoreface zones.

In the Mediterranean Sea, a low-energy micro tidal environment, Frihy et al. (2008) examined the seasonal response of the Abu Qir Bay, Egypt, calculating monthy DoC (using Hallereier's formula) and depths of wave base. While the DoC (*dl*) varied between 6 and 14m the wave base (*di*) varied between 16 and 29m. Maximum *di* varied between 45m in winter and 23m in summer. A previous study (Naffaa et al., 1995) of the DoC that encompassed the same area calculated observed DoC (using historical survey map data) as well as Hallermeier's *dl* and *di*. The observed DoC calculated for over 65 years was 25m. Naffaa et al. (1995) calculated the *di* with the same 65 years return period and obtained a value of 25.7m using Hallermeier's equation. The Frihy et al. (2008) study of Abu Qir Bay showed seasonal patterns attuned to the wave climate and in particular to the fair/storm weather alternation and, for

452 the same area the work of Naffaa et al. (1995) showed, that the lower shoreface was morphologically
453 active on a multi-decadal time scale (>10^1 years).

454 Valiente et al. (2019, 2017) calculated the inner and outer depths of closure, according to various
455 formulations, for the macrotidal, high energy embayments of the coast of Cornwall and Devon (UK)
456 using a 4-year wave time series, giving results ranging between 18.8m and 23.3m for *dl* and between
457 33.6m and 50.1m for *di*. Both *di* and *dl* decreased from south to north according to the decrease in
458 incoming wave heights. Using nearshore wave conditions obtained through hydrodynamic modelling,
459 Valiente et al. (2019) also looked at spatial variability of the DoC within each embayment as well as
460 the average DoC for each embayment. When compared to observed DoC (obtained from repeated
461 shoreface profiles over a 6-year period at one of the embayment), they found that offshore wave
462 conditions over-estimated the DoC and that the use of wave conditions at the 20m contour returned
463 results more comparable to the observed DoC. Overall their results showed that Hallermeier's DoC
464 (calculated with nearshore wave climate) remains a valid approximation for the identification of the
465 limit of significant seasonal morphological change. They also looked at bed shear stress as an indicator
466 of the boundary for sand motion (DoT, depth of transport; see section 4) and when taking tides into
467 account DoT reaches 30 to 50m for macrotidal Cornwall. This showed that intense sediment transport
468 occurs deeper than previously considered (especially in macrotidal environments) meaning that the
469 volume of reworked and transported sediment on the lower shoreface is most likely to be larger than
470 previously thought and is certainly non-negligible.

471 While both upper and lower shoreface zones are active, it appears that the time scales at which
472 morphological change operates in each zone are significantly different. Further, even if the upper
473 shoreface dynamic is closely linked to nearshore wave conditions, both upper and lower shoreface
474 morphodynamics are the result of the combined effect of multiple drivers and interactions that modify
475 the net shape of the shoreface across different temporal scales.

476 ***8.2 Drivers of shoreface morphology***

8.2.1 Lower shoreface

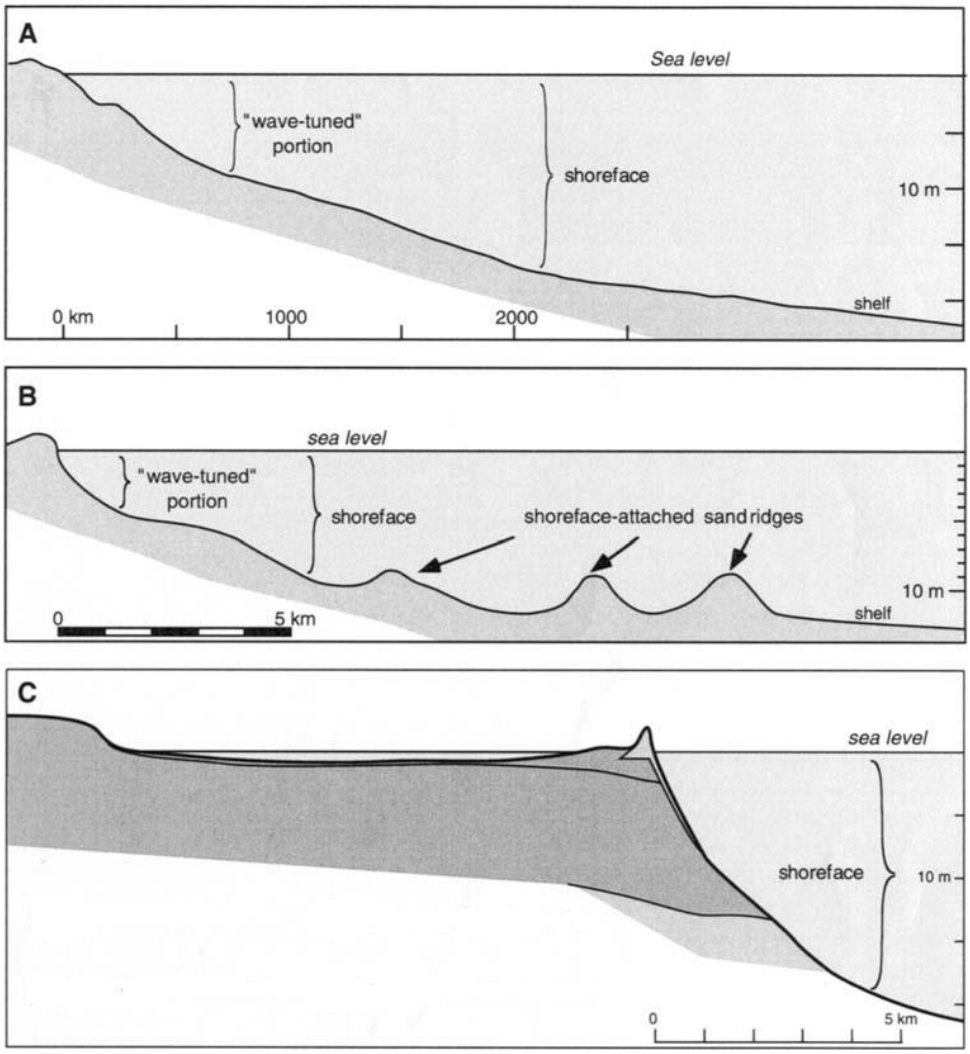
Morphological change on the lower shoreface is driven by the effect of waves and currents combined, with most variability occurring during storm events when currents tend to transport sediment from the foreshore/upper shoreface onto the lower shoreface (Niedoroda and Swift, 1981; Niedoroda et al., 1984; Swift et al., 1985; Vincent et al., 1983). As morphological variability on the lower shoreface is relatively small compared to the upper shoreface it is only observable during energetic events or over longer (> 10^1 years) time periods. The relative importance of each of the drivers impacting lower shoreface morphodynamics is still relatively unknown. For example, Backstrom et al. (2015; 2000) found different morphological responses to a storm on two adjacent shorefaces on the high energy coast of Northern Ireland despite the storm characteristics being similar for both shorefaces. The difference in response of the two shoreface (accretion was observable on one shoreface whilst only minor changes on the other) was attributed to geological shoreface characteristics such as antecedent morphology, slope and the presence of available sediment (sand bodies). Indeed, for the same event the more dissipative shoreface was only slightly affected while the steeper shoreface, with nearby sand bodies, was significantly more accreted.

From the limited set of observations available in the literature, it seems that lower shoreface morphodynamics are influenced both by the hydrodynamic climate and geological factors (i.e. underlying geological framework, sediment supply) reflected in its morphology (**Figure 5**). The relative importance of internal (hydrodynamics) versus external constraints seems to be variable and dependant on the time scale involved, for example external constraints vary with seasonal or inter-annual patterns.

There is also marked spatial variability in shoreface morphology and factors such as slope (Aagaard, 2014; Bowen, 1980) influence their behaviour. Patterson and Nielsen (2016) examined sand transport on the lower shoreface in the Gold Coast region of western Australia. The site encompassed the sediment-rich lobe of the Nerang River, as well as adjacent shorefaces which are less affected by the

local sediment discharge. The long-term evolution (multi-decadal) of the river lobe was interpreted in terms of sediment transport and, as the lobe is in disequilibrium, significant change was observed between 1966 and 2012. Rates of shoreward sand transport from 12.8 m^3/m/yr at 18m to 89.1 m^3/m/yr were calculated and shown to be dependent on wave conditions, as well as water depth and shoreface slope.

Through numerical modelling, Preston et al. (2018) investigated the role of slope and wave energy in nearshore morphodynamics. They showed that for an embayed beach, shoreface slope exerted a control on beach recovery. Specifically, they suggest that beaches with steeper shorefaces are less likely to recover from storms than beaches with more gently sloping shorefaces because (p.2429) "bathymetric slope is a major control on the nearshore sediment budget under calm climatic conditions". However, during storms, hydrodynamics seem to have more control over sediment loss than the slope (Preston et al., 2018).



514

515 ***Figure 5.*** *Diagram of various possible shoreface morphologies (Modified from: Clifton et al., 2005) A: Unconstrained*
516 *morphology B: Morphology influenced by sediment features C: Morphology constrained by geologic framework (thin layer of*
517 *sand over bedrock)*

518 ### *8.2.2 Upper shoreface*

519 As a reminder, we propose to define the 'shoreface' as corresponding with the literature definition of
520 the lower shoreface, while the 'nearshore' is defined as a combination of the upper shoreface and surf
521 zone. Although several definitions of the upper shoreface include the (fairweather) surf zone, it is
522 excluded from consideration here since surf-zone processes are distinctive and different (involving
523 wave breaking, secondary wave motions, etc.) Sediment transport on the upper shoreface can be
524 attributed to the temporary extension of the surf zone to the upper shoreface during storm events
525 (**Figure 6**). Morphological variations of the upper shoreface are thus linked to the dynamics of the
526 outer surfzone bar and the spatio-temporal variability of the position of the surf zone (Lee et al., 1998;

Ruggiero et al., 2016). Antecedent morphology and storm grouping also seem to influence upper shoreface variations. The cumulative effect of successive storms prompts a larger change in upper shoreface morphology because of insufficient recovery between storms.

Anthony (2008, p. 243) similarly argued that "Another important aspect of the sediment dynamics of the inner and outer shoreface link is the role of surf zone sediment dynamics in mediating spatial patterns of shoreface elevation change and net losses offshore during storms". In other words, Anthony (2008) suggests that further potential longshore modulation/variation in morphodynamics due to the influence of the surf zone morphodynamics which due to bars and rip effect etc. has a strong three-dimensional component.

Aleman et al. (2015) documented the morphological variability of the coast along the Gulf of Lyon (Mediterranean, France). The depth of closure for the area was calculated at 7.6m and the DoC observed from successive profile surveys varied between 1.4m and 7.4m (Sabatier et al., 2005). Here the DoC gives an indication of the extent of the shoreface that is most affected by changes in incoming wave energy, i.e. the upper shoreface. The offshore distance of the DoC put the morphologically active zone (from shore to depth 7.6m) at distances up to 2.5km offshore for the more dissipative and rock platform-constrained shoreface profiles. Conversely, for reflective shoreface morphologies, the limit of the active zone was less than 200m from the shoreline. This illustrates that a wide array of shoreface configurations can exist for relatively similar offshore wave conditions and tide regimes. It also shows that for different morphologies the extent of the shoreface affected by waves, (the shoaling zone) can vary considerably and with it the stock of sand available for onshore transport.

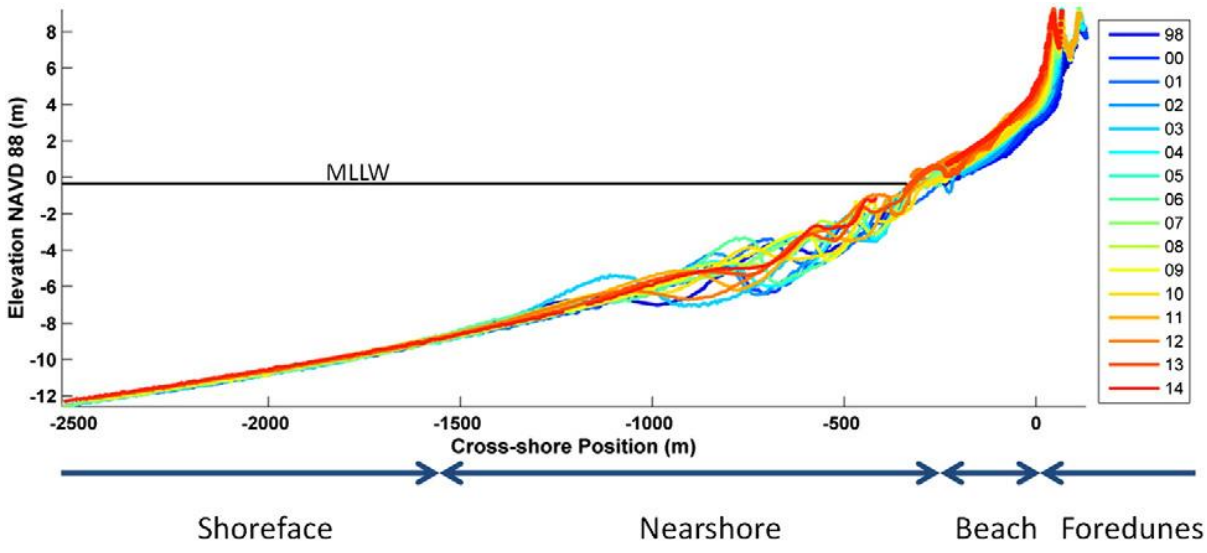


Figure 6. *(colour) Example of profile morphological variability in the Long Beach Peninsula (USA) (From: Ruggiero et al 2016). Each profile corresponds to a survey (one per year from 1980 to 2014).*

8.2.3 . Shoreface sediment supply and exchanges

Although "long-term observations of shoreface bathymetry change are rare" (Cowell and Kinsela, 2018), sediment transfers to and from the shoreface and within the shoreface itself have been noted in several studies. These transfers have been linked to multiple spatiotemporally variable processes which are primarily episodic in nature (Figure 7). Excluding wave and tide-driven sediment motions (see section 5), the following have been reported:"

1. Variation in surf zone extent. Periodically, under high wave conditions, the upper shoreface is affected by surf zone processes, which lead to appreciable changes in bed level. This is illustrated in the well-known seasonal dynamics of nearshore bars (Aubrey, 1979) and by the seasonal changes in beach/ surf zone states (Wright and Short, 1984; Wright et al., 1985). For example, wave breaking typically occurs further offshore during winter than summer. Indeed large wave events are responsible for transfer of sediment between the shoreface and surfzone/beach. Often this is evidenced in onshore bar migration at seasonal timescales (Aagaard et al., 2010, 2004). Offshore bar migration in the surf zone usually ends with bar disintegration and sediment dispersal at the seaward limit of the surf zone (Shand et al., 1999).

2. Generation of rip currents. Rip currents that originate in the surf zone can carry sediment offshore and deposit it on the shoreface where it may be sequestered or return shoreward under constructional wave action (Castelle et al., 2016; Loureiro et al., 2012a).

3. Onshore-directed flows during storms. Extreme waves have frequently been documented to transfer shoreface sediment onshore where it is deposited as washover fans (Aagaard and Kroon, 2019). These sediments originate in the surf zone and sites to seaward, including the shoreface (Schwartz, 1975). Recent studies (Pham et al., 2017) also show that some tsunami-generated overwash sediments originated from shallow marine (shoreface) environments. Fruergaard et al. (2013) documented the deposition of large volumes of sand transported from deep water to the shoreface during and after a 1:1000-year storm in the North Sea. Subsequent emergence led to generation of ephemeral spits and barrier islands.
4. Offshore-directed flows. Following storms (Pretorius et al., 2018; Siringan and Anderson, 1994) and tsunami (Richmond et al., 2011; Slootman et al., 2018) deposition of sediment has been recorded on the shoreface. Liu and Goff (2018) documented lobe structures on the stable (prograding) shoreface of Fire Island (New York, USA) that were attributed to storm return flows. Channelized storm return flows have been linked to the formation and maintenance of SCRs (Pretorius et al., 2018).
5. Longshore sediment movement on the shoreface has recently been documented after storm events, both for sand-rich features and sorted bedforms. For example, on Fire Island's shoreface a retreating area showed erosion and subsequent redistribution of sediment via longshore processes (bedforms migration) while nearby sediment rich areas with SCRs also showed signs of longshore sediment transport (Liu and Goff, 2018). Other studies in the same area also documented bedform migration driven by hurricane events (Goff et al., 2015; Schwab et al., 2017).

6. Shoreface erosion. Evidence of shoreface-derived sediment in the coastal system is widespread, especially on transgressive shorefaces where underlying rock is periodically exposed and subject to wave erosion (ravinement) (Riggs et al., 1995). Contemporary erosion of bedrock outcrop on the shoreface has recently been documented by repeat side-scan sonar surveys in the Baltic Sea (Schwarzer et al., 2014). This sediment contributes locally to the shoreface sediment budget.
7. Shoreface nourishment. Experiments with shoreface nourishment have shown that sediment is exchanged between shoreface and surf zone (Grunnet and Ruessink, 2005; Ojeda et al., 2008), but also that the change in morphology created by nourishments in the form of a bar on the shoreface has a shielding effect during energetic conditions (Barnard et al., 2009; Huisman et al., 2019).

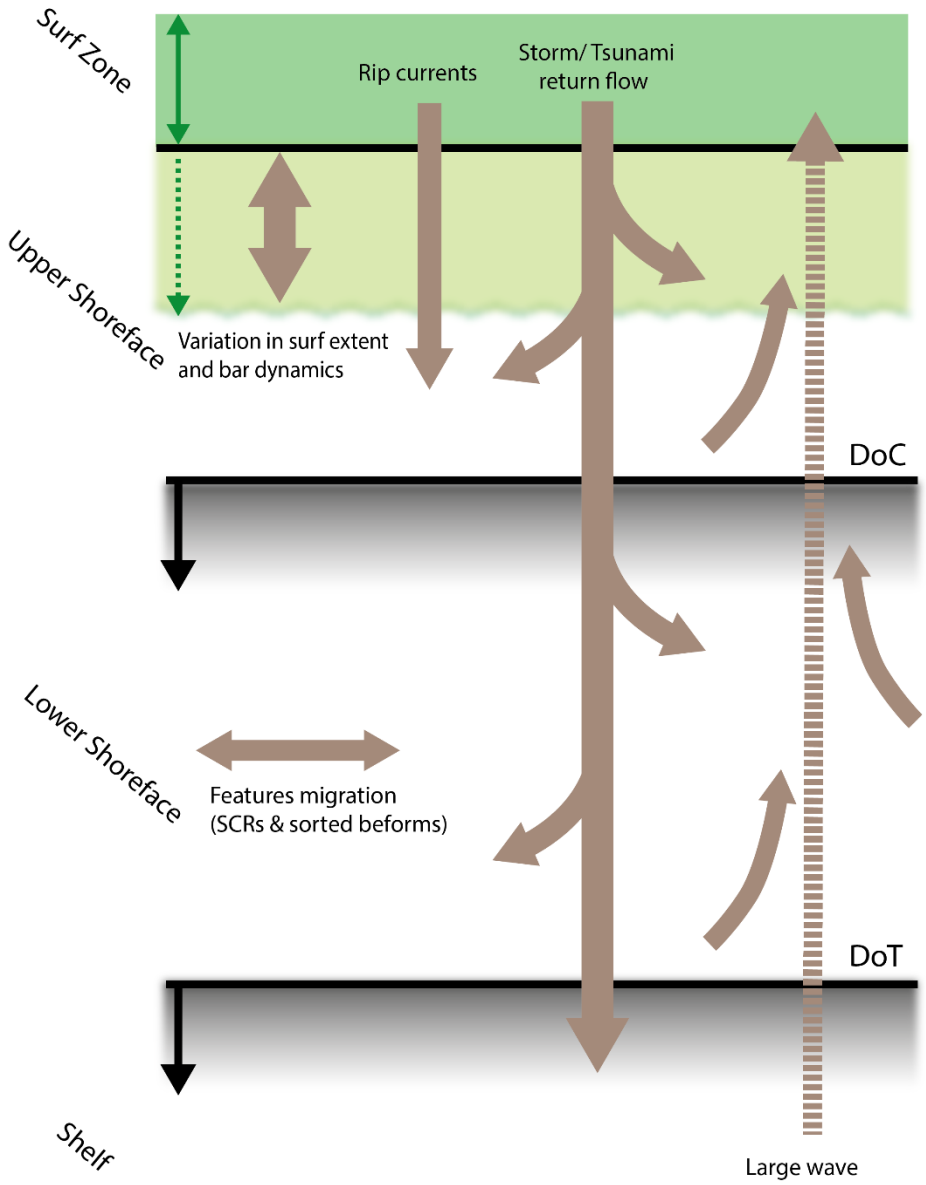


Figure 7: *Schematic representation of sediment exchange over the shoreface. With, The surf zone (in dark green) and upper shoreface (light green) are depicted with the upper and lower shoreface limits (solid black lines; see Figure 9) The schematic represents episodic expansion of the surf zone during energetic events. In brown, the arrows depict the sediment transport mechanisms across the shoreface.*

8.2.4 Sea Level Rise

Understanding the response of shoreface morphology and behaviour to sea-level change over decadal to century timescales (10^1 to 10^2 years) is hampered by the lack of long-term datasets and the difficulty of isolating the sea-level signature from that of shorter-term dynamic processes, a problem that persists even on the much more data-rich beach environments (Cooper and Pilkey, 2004). Information about mesoscale shoreface behaviour is, however, sometimes available from discontinuous historical

613 records (e.g. bathymetric charts) (Cooper and Navas, 2004; Hanna and Cooper, 2002) and, for longer
614 (geological, >10^2 years) time scales, seismic and coring data. The dynamics have also been simulated
615 (and sometimes predicted) via models. A wide array of behaviour models has been developed, from
616 statistical- to process-based models that describe shoreface reaction to sea level rise. Early
617 approaches to long-term shoreface evolution assumed time-invariant shoreface profiles (Cowell et al.,
618 1999), while later models allowed for shoreface profile evolution according to the presence of an
619 erodible or non-erodible substrate (Thieler et al., 1995). Some of the most recent studies point to the
620 importance of the slope as a strong control for long term evolution (Ciarletta et al., 2019; Deng and
621 Wu, 2020; Ribó et al., 2020).

622 The conceptual behaviour model for barrier shorefaces as explained by Cooper et al. (2018) describes
623 three modes (keep-up, give-up and catch-up) of shoreface response to sea level rise (**Figure 8**). In this
624 model, the rate of sea-level rise is a key determinant of shoreline behaviour but other internal factors
625 (sediment supply, sediment volume and basement erodibility), also play a role. At mesoscale the
626 geological inheritance dominates shoreface morphology and conditions the response to forcing
627 through sea-level change and storm events. Knowledge of the contribution of shoreface dynamics to
628 the nearshore stratigraphy, and in particular the processes that contribute to development of wave
629 ravinement (erosion) surfaces (Zecchin et al., 2019), has important implications for determination of
630 former sea levels from seismic stratigraphic records as shown by Plets et al. (2019). Shoreface deposits
631 are frequently found in the geological record in preference to beach units. This shows their propensity
632 for preservation by burial or overtopping.

633 The focus of this paper is the mesoscale (10^1 to 10^2 years), there is, however,(despite their
634 propensity for preservation by burial or overstepping) few studies utilizing such deposits to infer the
635 morphodynamics of the shoreface over longer timescales.

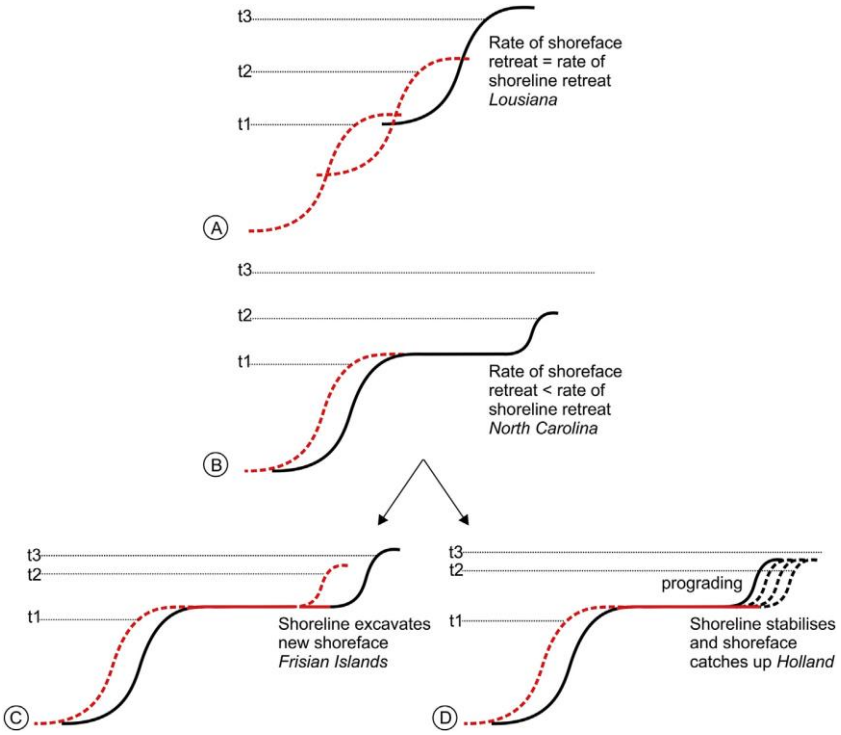


Figure 8. *(colour) Diagram of barrier-shoreface retreat mode. A. Barrier and shoreface retreat at equal rates (keep-up). B. Shoreface retreat slower than barrier C&D. Shoreface "catch-up" with t1 t2 t3 the successive stages of sea level rise. From Cooper et al. (2018)*

8.3 Shoreface-shoreline interactions

During storm events sediment can be transported offshore to the lower shoreface (Niedoroda et al., 1984; Preston et al., 2018), and since it is then beyond the depth of closure, it is then considered lost to the coastal system. However, advances in the study of the role of the shoreface in the coastal system and sediment budget suggest that shoreface sediment (even beyond the depth of closure) might in fact play a key role in nearshore morphodynamics. Indeed, several studies (Billy et al., 2013; Brunel et al., 2014; Lazarus and Murray, 2011; Safak et al., 2017; Schupp et al., 2006; Thomas et al., 2011; Valvo et al., 2006) have linked shoreline shape and mesoscale variation patterns to adjacent shoreface setting.

Frihy et al. (1991) examined the link between shoreline evolution and nearshore bathymetric change in the Nile Delta, Egypt, and found no significant relationship between shoreline change and nearshore (i.e. shoreface) sediment texture, slope or depth. They did, however, attribute erosion and accretion

patterns to movement of eroded sediment or sediment from offshore sources, suggesting that the shoreface might contribute to shoreline dynamics.

Wijnberg (2002), found no correlation between decadal coastal behaviour and sediment characteristics or offshore hydrodynamics on the Dutch coast but instead linked spatial changes in decadal behaviour to variation in shoreface morphology, especially the presence of “terrace like features” on the shoreface linking shoreface morphology and coastal behaviour. Similarly, Cooper et al. (2004) linked nearshore behaviour to sediment accumulation and associated bathymetric change on the shoreface.

Miselis and McNinch (2006), explored the relationship between shoreline variations and nearshore sediment volume in North Carolina, and established that the volume of sediment stored on the shoreface was correlated with decadal shorelines over decadal timescales. Sediment-poor shorefaces were associated with smaller shoreline change rates than sediment-rich shorefaces.

The shoreface of Fire Island, has been the subject to multiple studies regarding the role of the shoreface in the coastal system, for example Goff et al. (2015) examined the impact of hurricane Sandy on the shoreface and found that “sand ridges and sorted bedforms appear to act as a regulator of storm-forced erosion of material beneath the modern sand layer” thereby limiting erosion of older sediment as a form of supply of modern sediment available for transport. Hapke et al. (2016), applied statistical analysis to correlate shoreline response of Fire Island with potential controls, especially storm processes and underlying shoreface geology. The correlation analysis suggested (p.52) that: “[long term shoreline] patterns result from an unresolved combination of, or feedback between, storm processes and framework geology (bathymetric variability and sediment availability)”.

Fire Island has also been the subject of several studies examining underlying geological control of the area and evidence of onshore-directed sediment transport across the shoreface (Hapke et al., 2010; Schwab et al., 2013, 2000). Those studies show that the shoreface acts as a source and/or a conduit for onshore sediment flux and has a role in foreshore long-term (>10^1 years) morphodynamics; there

678 is an apparent link between the volume of shoreface sediment and foreshore behaviour. Safak et al.
679 (2017) subsequently used a three-dimensional, hydrodynamic-based model to examine the effects of
680 shoreface connected sand ridges on shoreline variability at Fire Island. Their results show that
681 hydrodynamic processes and sediment flux are controlled by the effects of offshore geologic
682 framework and that cross-shore flows between the shelf and the coast (i.e. flow on the shoreface)
683 influence coastal evolution and shoreline variability at decadal time scales.

684 In the larger framework of the Mid-Atlantic Bight, McNinch (2004) highlighted a 'suggestive spatial
685 alignment' between shoreline erosional hotspots and shore-oblique bar and associated outcrops. A
686 subsequent study of the same area by Schupp et al. (2006), correlated areas of high shoreline
687 variability with shore-oblique bar and gravel outcrops but with a spatial offset. These observations
688 reinforce the idea that the shoreface morphology (which seems to reflect sediment volume and
689 geological control) constrains or influences foreshore dynamics in some way. Similarly in South
690 Carolina, Denny et al. (2013) linked stable shorelines to adjacent sediment-poor, erosion-resistant
691 shorefaces, whereas the more variable portions of the shoreline occur around portions of the coast
692 rich in sediment and with large sand bodies. Another study (Oakley et al., 2019) on a North-American
693 shoreface (Rhode Island, US) did not find the shoreface sediment volume to be a good predictor of
694 mesoscale shoreline variation, but hypothesize that the alongshore variations in shoreline variation
695 patterns relates to the antecedent morphology of the shoreface and its heterogeneous nature.

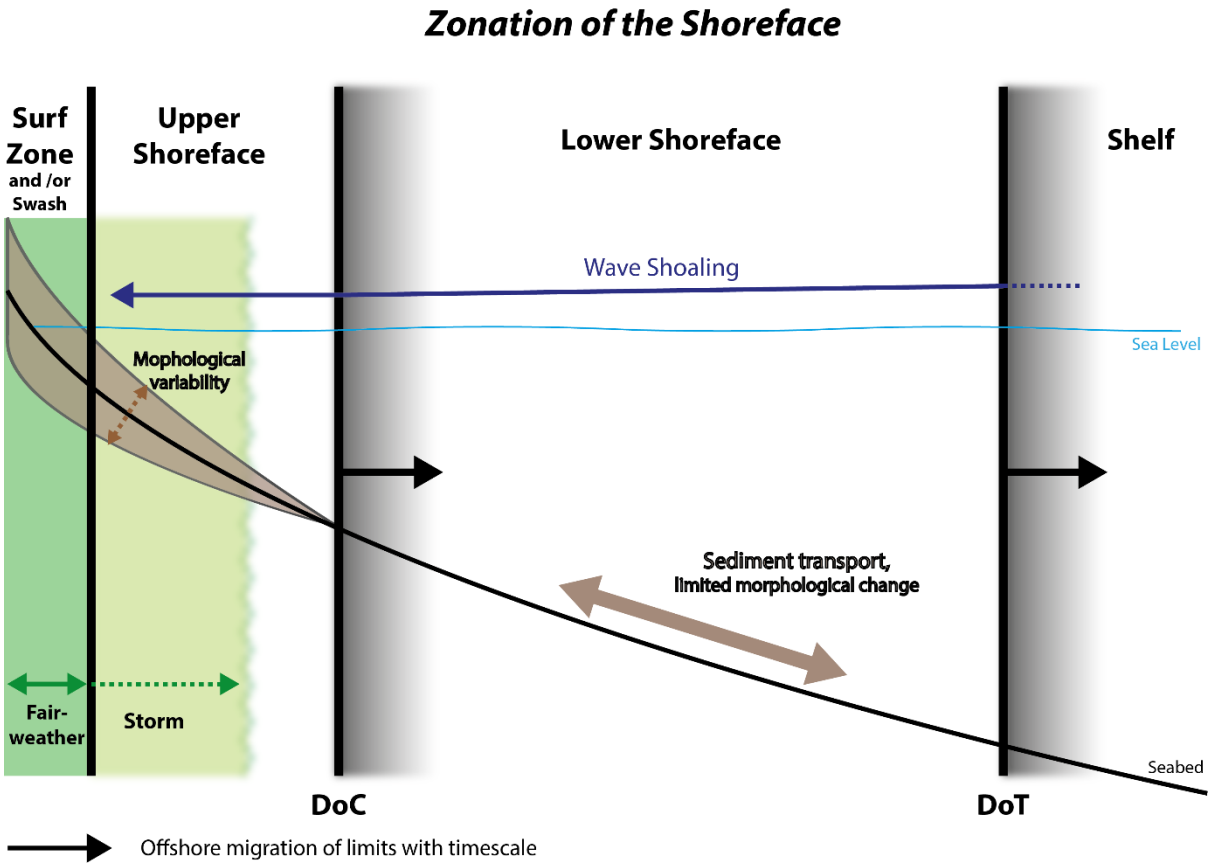
696 Shoreface sand bodies have been shown to exert a strong influence on foreshore change. Sand banks
697 have been shown to decay onshore and provide sediment from the shoreface to the beach-dune
698 system (Aagaard et al., 2004; Anthony, 2013; Anthony et al., 2006; Héquette and Aernouts, 2010;
699 Héquette et al., 2013; Thomas et al., 2011; Verwaest et al., 2020). The opposite phenomenon (offshore
700 bar breakdown followed by transport of sediment from the beach to the lower shoreface) has been
701 documented by Aagaard (2011) on the Danish North Sea coast.

On a longer time scale (>10^3 years), Kinsela et al. (2016) showed through modelling of south-eastern Australian beaches, that the shoreface contributed around 80% of the sand involved in the growth of Holocene strand plains. They suggest that similar processes might still contribute to contemporary shoreline stability. In the same region, Cowell et al. (2001) found that 'the annual net supply residual (of offshore sand) is cumulative and would grow to exceed the beach-cut/fill volume after only several decades', supporting the idea that the shoreface can be an important source of foreshore sand in the longer term.

9. Discussion

9.1 Shoreface definition and morphodynamics

The shoreface is a transition zone between surfzone/beachface and the shelf within which wave shoaling and sediment transport take place. In the face of multiple and conflicting definitions and terminology in the literature, the definition of the shoreface requires standardization. We propose that its landward limit be the outer edge of the surf zone (or the outermost bar) during fairweather/modal conditions, and where this is absent, the seaward edge of the swash zone (or the beach step). This can be justified on the marked difference in nature of wave-driven morphodynamics, pre- and post-breaking. The surf zone is dominated by wave breaking, (and associated processes such as reformation and secondary wave motions or rip currents etc.) (Dally, 2005) and is therefore morphodynamically distinct from the shoreface, not a part of it (**Figure 9**). Although energetic conditions may periodically extend the surf zone seaward, the most persistent process in the upper shoreface is wave shoaling.



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724 ***Figure 9:*** *Definition of the shoreface,* showing *upper and lower shoreface limits. The surf zone is represented in dark green,*
725 *with light green areas in the upper shoreface representing episodic expansion of the surf zone during energetic events. Brown*
726 *arrows depict bidirectional movement of sediment in the shoreface and resulting morphological variability.*

727 The shoreface is made up of two morphodynamically distinct units, namely the upper and lower
728 shoreface. They are divided by the DoC and the outer limit of the lower shoreface is the threshold at
729 which waves start to significantly impact the bottom at the mesoscale. We propose to use the Valiente
730 et al. (2019) depth of transport (DoT) as a calculable boundary for the outer limit of the shoreface at
731 the mesoscale. Upper and lower shorefaces each exhibit distinctive morphodynamics (**Table 1**). The
732 shoreface role in adjacent nearshore (surfzone and swash zone) dynamics is manifest in (a)
733 attenuating waves as they move onshore and (b) as both a sediment source and sink at variable
734 (largely event-driven) timescales.

735 Upper and Lower shorefaces can be differentiated by their dynamics which play out at different spatio-
736 temporal scales. Whilst the upper shoreface is active on a short (seasonal) temporal and small spatial
737 scale ($<10^0$ to 10^1 years and around 10^0 km), the lower shoreface is active over larger scales (10^1 to 10^2

738 years and over 10^1 Km). Because of its mesoscale behaviour, the lower shoreface has often been
739 considered inert and non-influential in terms of coastal dynamics (Hoekstra et al., 1999). The upper
740 shoreface is dominated by wave shoaling and periodic seaward excursions of the surfzone. This is
741 responsible for the morphological variability that occurs on the upper shoreface at seasonal to decadal
742 timescales. The lower shoreface is dominated by oscillatory wave motions that transport sediment
743 but which create morphological change only at long timescales ($>10^1$ years).

744 All shoreface morphodynamic limits exhibit spatio-temporal variability as they can be influenced by
745 water level variations (tides, surges etc.), and variation in storminess with an offshore displacement
746 of the previously cited limits during energetic events, for example. This highlights the interconnectivity
747 between the shoreface zones and adjacent environments. Linkages and overlaps exist between the
748 different zones, particularly at their boundaries. At the junctions between the zones sediment
749 exchanges occur as the boundaries shift. For example, Aagaard (2014) quantified the supply from the
750 lower to the upper shoreface and found it to be coherent with bar migration and aeolian accretion
751 rates. Future targeted research might be able to provide quantifiable limits for the shoreface
752 boundaries based on ratios of morphological change, for example, but thus far, this is not feasible
753 because of the paucity of observations.

754 ***9.2 Sediment supply and shoreface morphology***

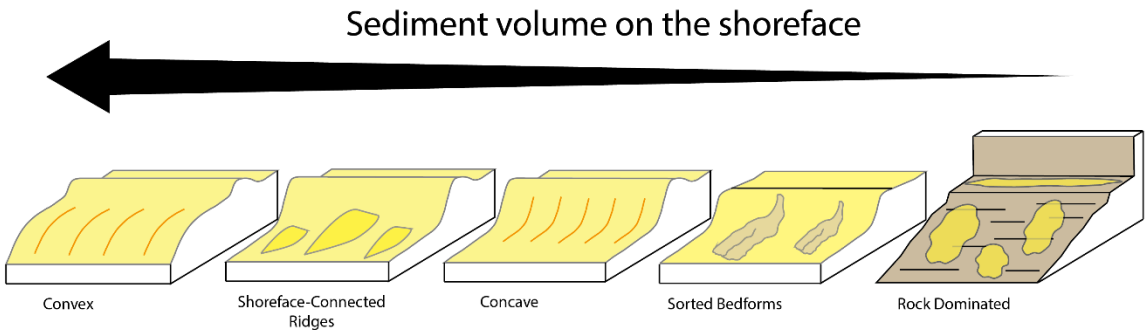

***Figure 10**: Shoreface morphological classification as a function of sediment supply. Typical morphologies associated with, (increasing) sediment volume on the shoreface from a sediment starved erosive (far right) shoreface to a prograding shoreface (far left) with intermediate morphologies indicating the augmentation of available shoreface sediment.*

The literature shows that a wide variety of shoreface morphologies exist in nature (Athanasiou et al., 2019; Kinsela et al., 2020). It is unlikely these variations are linked solely to variations in wave conditions. One of the major factors controlling this variability in morphology seems to be sediment supply (or availability) and more specifically, the sedimentary bedforms associated with sediment-poor and sediment-rich shorefaces. Bedforms are an indicator of the presence (or deficit) of sediment on the shoreface. Offshore of Fire Island (Locker et al., 2017; Schwab et al., 2017) increasing sediment abundance is reflected in the successive development of sorted bedforms and SCRs. Banks, ridges and similar features are indicators of sediment-rich areas while sorted bedforms are more typical of sediment-depleted areas. For example, some shoreface attached sand ridges have been determined to originate from relict tidal deltas (McBride and Moslow, 1991) supporting the hypothesis that these features originate in sediment-rich areas.

We propose a tentative classification of shoreface morphologies based on increasing sediment abundance (**Figure 10**). Shorefaces that lack sediment are likely to be much more erosional. With the addition of a limited amount of sediment, sorted bedforms develop with underlying lag deposits exposed between sandy areas of positive relief. Increased sediment supply can then lead to development of SCRs either from relict sediment sources (e.g. submerged ebb deltas) or

775 contemporary inputs. Evidence of the link is present in the longshore variation in shoreface
776 morphology offshore of Fire Island, for example where onshore sediment abundance and accretion is
777 mirrored by offshore SCRs at the downdrift limit of the barrier. We surmise that increased sediment
778 abundance ultimately creates convex shoreface morphologies. At longer timescales (>10^1 to 10^1 years)
779 shorefaces might evolve along this spectrum according to whether sediment supply is abundant or
780 absent (Deaton et al., 2017; Raff et al., 2018; Shawler et al., 2019). The equilibrium shoreface profile
781 in this conceptual scheme is one of a spectrum of possible morphologies.

782 The influence of shoreface morphology on the overall mesoscale coastal morphodynamics, however,
783 is still unclear, especially in terms of its contribution to the overall sediment budget. While in some
784 locations the shoreface acts as a net sediment source (Denny et al., 2013; Ruggiero et al., 2016;
785 Verwaest et al., 2020), for other regions it seems to be a sink (Aagaard, 2011; Finkl, 2004). It is clear,
786 however, that sediment exchanges within and between the shoreface and adjacent environments is
787 strongly episodic and likely strongly influenced by geological setting (sediment supply and antecedent
788 topography). In addition to the implicit geological control on such exchanges, these sediment transfers
789 are likely to be spatially and temporally variable. Whilst the impact of the variable shoreface
790 morphologies on the foreshore is still not fully understood, several studies have established links to
791 observed behavioural shoreline patterns, demonstrating the potential importance of shoreface
792 morphology to the overall coastal dynamics (Latapy et al., 2020; Verwaest et al., 2020; Wijnberg,
793 2002). Unfortunately, the paucity of datasets of shoreface morphology, preclude detailed insights into
794 the mesoscale dynamics of these shoreface morphologies and the adjacent coasts.

795 The evidence shows that the shoreface is morphodynamically distinct but it does play an important
796 role in nearshore sediment exchanges and is an important conditioner of processes in the surfzone
797 and landward.

798 ***9.3 Future challenges and opportunities***

The field of coastal geoscience has developed around the principle of equilibrium, with the first mention of equilibrium profile dating as far back as the early 19th century (Fenneman, 1902). Since then, research on coastal morphology and processes has mostly been focused on the beach and surf zone through which the dynamic principles that control beach and surf zone morphology have been developed. The Wright and Short (1984) model, which classifies beach state on a spectrum ranging from reflective to dissipative according to their morphology, was subsequently refined and adapted to account for further driving factors such as tides (Jackson et al., 2005; Loureiro et al., 2013; Masselink and Short, 1993; Short and Jackson, 2013)

Whereas the drivers of beach morphology have been well–constrained, the same is not true of the shoreface, possibly because of the gap in both spatial and temporal data that once existed. Spatial data availability is now being considerably enhanced by technical advances in data acquisition but the spatiotemporal scale at which shoreface dynamics operate still poses problems for empirical studies.

Much in the same way as for beaches and surf zones the shoreface can assume a spectrum of forms (Kinsela et al., 2020) that do not result from just grain size and wave climate forcing but rather a multiplicity of factors. Such a conceptual model would help create a much more realistic representation of existing shoreface morphologies worldwide.

Our proposed shoreface classification based on sediment supply provides one model that may prove to be universally applicable and could serve to coordinate studies focusing on very different spatiotemporal scales. Future approaches to understanding shoreface dynamics should consider this and the diverse factors that ultimately influence them at meso- and geological scales. This will improve scientific, engineering and management understanding of shoreface dynamics *per se* as well as add important new insights into their association with beach and shoreline behaviour.

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827

828 Declarations of interest: none

829

	Upper Shoreface		Lower Shoreface
Limits	Fairweather Surf/Swash zone	Morphologically active zone	Long-term limit of intense sediment transport
Calculation		Hallermeier's DoC	Valiente's DoT
Characteristic time scale of morphodynamics	Seasonal to annual $<10^0$ years		Mesoscale > 10^1 years
Main drivers of morphology	Hydrodynamics (wave, wind, tide…)		Sediment supply/availability/size Geological framework & morphology
Hydrodynamics	Wave shoaling, and wave breaking during storm conditions		Wave shoaling
Other drivers of the morphology	Geology & sediments		Hydrodynamics
	Antecedent morphology, slope Storm grouping, ocean patterns Biological processes		

Table 1. *Major characteristics of shoreface components.*

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