The lower shoreface: Morphodynamics and sediment connectivity with the upper shoreface and beach

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Highlights

• The lower shoreface mediates wave energy and sediment delivery to the upper shoreface and beach.

• Lower shorefaces can be sediment-rich or starved, siliciclastic, carbonate, low to high energy.

• Knowledge of lower shoreface is limited by instrumentation, monitoring difficulties, upscaling.

• Seabed mapping and bedform mobility studies document onshore sand supply and coastal accretion.

• There is need for better integration of studies on the lower and upper shoreface.

Abstract

The lower shoreface provides the connection between the continental shelf and the shoreline via its onshore transition called the upper shoreface. Lower shorefaces are diverse, and range from bedrock-controlled, through sediment-starved to sediment-rich, siliciclastic, carbonate, low to high wave-energy, microtidal to macrotidal, and are variably affected by storm and wind-driven flows. The lower shoreface can be a repository for deposits of

terrestrial origin, and a zone of active carbonate production. It can therefore be an important source of sediment for beaches, dunes, estuaries, and tidal basins. There has been progress in the ability to predict suspended sediment transport under non-breaking and shoaling waves across the lower shoreface. However, high-resolution measurement of sediment transport over unknown seabed configurations with unpredictable bed-level changes under hydrodynamic conditions that are unknown at the outset, and especially involving bedload transport, is still faced with significant challenges. Non-linear interactions between processes contributing to sediment transport render calculations and modelling of transport directions and magnitudes uncertain, and the spatial and temporal scales of transport are much larger than those of the upper shoreface. On the other hand, transport rates and morphological change may be much smaller on the lower shoreface compared to the upper shoreface. Another challenge is the upscaling of short-term measurements to explain the long-term morphological evolution of the lower shoreface. This limited understanding implies that current paradigms of lower shoreface dynamics based on morphological equilibrium and disequilibrium relative to the ocean-forcing conditions may be too simplistic, though possibly appropriate over long timescales (decades to millennia), and modelling work and prediction of change no more than exploratory. Over such long timescales, boundary conditions (sea level, wave climate) are likely to change. Making way forward on these issues is important for understanding the connectivity between the lower shoreface and beach recovery after major storm erosion, and for estimating coastal sediment budgets, short- to long-term coastal change and response to natural and anthropogenic perturbations. At geological timescales, the lower shoreface is a central element in tracking shoreline changes.

Progress is needed in measuring sediment transport and upscaling to timescales compatible with lower shoreface change. It is also important to take advantage of on-going rapid progress in seabed and shallow stratigraphic mapping, bed-level changes, including remote-sensing approaches, for a better understanding of lower shoreface morphodynamics and sediment connectivity with the coast. This includes the now routine identification of large subaqueous bedforms, possibly ubiquitous features on the world's continental shelves, their mobility over time, and their potential link with the shoreline. The common relationship between fine sand, dissipative beaches and large aeolian dunes also poses the question of how fine sand is abundantly supplied from the lower shoreface, given the

common perception that it is readily swept offshore on beaches. These multi-theme challenges need to be addressed in order to advance our understanding of the lower shoreface and its connectivity with the upper shoreface and beach.

Keywords: lower shoreface; upper shoreface; sediment transport; seabed mapping; beach; dunes; morphodynamics; continental shelf bedforms; sediment connectivity.

1. Introduction

With very few exceptions associated with some narrow tectonically convergent plate margins (Schellart and Rawlinson, 2010), the world's continental shelves, notably on passive plate margins (Bradley, 2008), play a very important role as long-term sedimentary repositories and transport routes between continents and ocean basins. They act as tracts over which much of Quaternary sea level has fluctuated, and transform and dissipate the energy of wind-generated waves, tides, and tsunami. The transition between the continental shelf and the coast has generally been defined in terms of the *shoreface* (Fig. 1). The term 'shore face' was first used by Barrell (1912) to identify, in a stratigraphic framework, the zone between the subaerial and subaqueous plains of a deltaic topset system, the author considering this zone as one with a relatively steep slope developed by breaking waves. Johnson (1919) later redefined the term 'shoreface' as the zone between the low-tide shoreline and the more nearly horizontal surface of the offshore. While the use of the term has prevailed in stratigraphy, a clear hydrodynamic/morphodynamic conceptualization has emerged in more recent considerations of the shoreface, which has been defined (e.g., Swift et al., 1972, 1985; Niedoroda and Swift 1981; Niederoda et al., 1984; Cowell et al., 2003a) as that part of the continental shelf dominated by sea/swell wave motion, and extending to an undefined offshore limit. Galloway and Hobday (1996) and Hinton and Nicholls (2007) defined the shoreface in terms of three zones: the upper shoreface equated with the inner surf zone, the middle shoreface defined as the portion occupied by breaker bar systems, and the lower shoreface identified as the area seaward where the shoreface merges with the inner shelf. More commonly, however, the zone from the inner shelf to the coast is envisaged as comprising two units: an upper shoreface (US) and a lower shoreface (LS) (Fig. 1). Wright (1995) considered the LS and the inner shelf as being the same, and defined the inner shelf as the 'region immediately seaward of the surf zone where waves normally (or

frequently) agitate the bed'. Clifton (2019) deemed the shoreface as an integral feature of nearly all clastic coasts and forming a surface that slopes away from the low-tide shoreline, comprising an US and a LS, and imperceptibly merging with the flatter inner shelf or basin plain ramp. There is, therefore, no clear separation of the LS and the inner shelf. For purposes of clarity and simplification, we define the LS as that part of the coastal margin where wave agitation of the seabed may lead to morphological change that is comparatively much smaller on an annual timescale than on the US (Aagaard, 2014). The so-called 'depth of closure' (Hallermeier, 1981) separates the morphologically active US and relatively inactive LS on an annual timescale (*3.1*). Generally, the LS is perceived as bounded seaward by wave base (Fig. 1), which is the limit within which waves affect, and are affected by, the seabed and beyond which currents become dominant in moving sediment.

By mediating wave energy delivery to the shoreline (Ardhuin et al., 2003), the LS is a fundamental component of coastal morphodynamics and plays a central role in sediment exchanges between the continental shelf and the beach (Anthony, 2009). The ability to predict sand transport under non-breaking and shoaling waves across the LS is important for understanding the rate of recovery of beaches after major storm erosion, and the behaviour of nourishment sand placed offshore. The LS is, potentially, an important source of sediment for beaches, dunes, estuaries, and tidal basins. In a coastal tract, for instance, a 1-mm lowering of a 10-km-wide LS in one year is equal in volume to an extra 1 m of sediment added to the beach and frontal dune over a width of 10 m. The LS can also provide aggregate for the construction industry and for beach nourishment, a perspective that needs, however, to be gauged carefully in terms of its environmental impacts (ICES, 2016). Studies on LS morphodynamics and sedimentology are fundamental for estimates of coastal sediment budgets and short- to long-term coastal change (e.g., Kelley et al., 2005; Hapke et al., 2010; Ruggiero et al., 2010; Anthony et al., 2019). Such estimates also inform on the fate of diminishing beach and US sand volume following long-term nourishment (e.g., Thieler et al., 2001). The links between short- to long-term geomorphic change, notably in small segments of coast, and the way these are embedded in, and controlled by, larger-scale aspects of coastal change that commonly involve considering the area well offshore of the beach, are not always well apprehended (Gelfenbaum and Kaminsky, 2010; Sedrati and Anthony, 2014). Insight on the processes driving sediment transport and exchanges both alongshore and across the LS/US and on how these influence sediment budgets is important in order to

predict with confidence coastal response to natural and anthropogenic perturbations. A potentially changing wave climate, sea-level change, and significant anthropogenic perturbation of coastal sediment budgets (Pilkey and Cooper, 2014; Ranasinghe, 2016; Anthony, 2019), are, indeed, further incentives for a better understanding of links between the LS and the beach. Understanding the morphodynamics of the LS is also important for various marine activities such as placement of offshore infrastructure (e.g., cables and pipelines), characterization of offshore placers and their mining potential, offshore archaeology, offshore habitats and marine protected areas, offshore contaminant transport and deposition, as well as aiding decision-making regarding marine spatial planning and protection. At geological timescales, studies on the LS shed light on the evolution and changes in continental shelf stratigraphy and sediment budgets with changing sea level (e.g., Coe et al., 2003).

The LS is commonly a repository for terrestrial sediment deposited by rivers and terrestrial runoff, and sometimes by offshore winds, during earlier sea-level low-stands (e.g., Hampson et al., 2008), for sediment derived *in-situ* from carbonate sedimentation (e.g., Michel et al., 2019; Laugié et al., 2019), and/or mobilized from the beach and US and moved offshore during exceptional high-energy events (e.g., Thom and Hall, 1991). The LS is part of the tract across which shoreline translation occurs over geological time, often leaving subsurface and surface evidence of such translation that has proven to be important in the analysis and interpretation of ancient deposits (Anthony, 2009). The LS may exhibit relief due to preserved or reworked relict landforms, including submerged palaeo-shorelines, in-situ carbonate-cemented forms, and contemporary bedforms related to prevailing hydraulic process regimes. The preservation of relict forms would be expected to prevail where hydrodynamic forcing is weak, or where in-situ cementation processes create conditions for preservation even in energetic conditions. The LS is, thus, commonly characterized by large sedimentological variability and facies arrangements (Cowell et al., 1999; Gao and Collins, 2014) reflecting inherited environments, palaeo-shorelines, and bedforms. These are subjected to wave, wind, storm and tidal current activity that results in sediment-sorting, including the commonly observed across-shelf transition between sand and mud (George and Hill, 2008). On sediment-poor LS settings, bedrock may outcrop, or subcrop at depths shallow enough to be of relevance in contributing to topographic variability that affects morphodynamic behaviour and sediment connectivity with the US and beach (e.g., McNinch,

2004; Jackson et al., 2005; Menier et al., 2019). The influence of inner shelf geology on the sedimentary and morphological characteristics of the LS has been highlighted in a large number of studies (e.g., Riggs et al., 1995; Thieler et al., 1995, 2001; Billy et al., 2013; Cooper et al., 2018a; Kirkpatrick and Green, 2018; Menier et al., 2019; to cite but a few). However, even a high wave-energy sediment-rich LS can be subject to the strong influence of inherited topography and geological framework (e.g., Kirkpatrick and Green, 2018). High sustained fine-grained discharge can also lead to the contemporary accumulation of abundant terrestrial mud on the LS, as on the Amazon-influenced coast of the Guianas in South America (Gratiot et al., 2007). Cooper et al. (2018b) have reviewed the large range of constraints, including sand sequestration, imposed by inner shelf and LS geological control on mesoscale coastal barrier behaviour. Where sand supplies from the LS contribute to the formation of coastal barriers, the diversity of resulting barrier types should consistently reflect the influence of the LS (e.g., Zenkovich, 1967; Short and Hesp, 1999; Short and Woodroffe, 2009; Short, 2010, 2013, 2019).

Notwithstanding this recognized importance, research on the contemporary (as distinct from the palaeo-) morphodynamics of the shoreface as a whole (LS and US) shows a strong imbalance in favour of the US, and its connection to the beach has long been an integral aspect in numerous studies. Earlier comprehensive reviews of LS morphology, sediment transport, and dynamics were proposed by Wright (1995), Cowell et al. (1999, 2003a, 2003b) and Kleinhans (2002). Balson and Collins (2007) edited a collection of papers that covered in part shelf sediment transport. More recently, Gao and Collins (2014) reviewed present knowledge on Holocene continental shelf deposits in relation to the processes driving their formation, from a marine sediment-dynamics perspective. Aagaard (2014) provided important insight on LS sediment transport, while Ruggiero et al. (2016) highlighted beach progradation across a range of spatial and temporal scales encompassing the LS. These studies clearly show that a finer understanding of how the US and beaches and barriers evolve, erode or accrete, necessitates integration of the LS, within a 'big-picture' morphodynamic framework that considers sediment redistribution between these domains. In this long-term perspective, individual events and processes may become less important than the "broader context", shifting the focus on morphological patterns produced from sequences of events and the sum of background processes (Ciarletta et al., 2019).

Although studies linking the development of coastal sand barriers to sand supply from the LS date back to at least the 1950s (*5.1*), the idea was conceptualised by models proposed by Roy and Thom (1981) to explain the long-term progradation of sand barriers in southeastern Australia. These models are based on progressive erosion of a sediment reservoir forming an inner shelf bulge or a plane surface to feed barrier progradation (Fig. 2a). They paved the way for exploratory aggregated modelling of dated barrier progradation (Fig. 2b) driven by a LS in disequilibrium (e.g., Kinsela et al., 2016), as well as for barrier morphostratigraphic studies characterized by progradation driven essentially by onshore sediment supply from the LS (e.g., Oliver et al., 2017a, 2017b, 2019; Carvalho et al. 2019). Support for this large-scale coastal approach of linkages or 'connectivity', postulated by Cowell et al. (2003a) in their concept of the coastal 'tract' that encompasses shoreline-to shelf 'sediment sharing', is permeating into sedimentology (Heckmann et al., 2018) and geomorphology (e.g., Wohl et al., 2019).

Waves are the main process driver of coastal change, but tides, wind-generated flows, and episodic high-energy events such as large storms, cyclones, and tsunami can also be important. The available sediment on the LS and inner shelf, the inherited morphology of these domains, and sea-level rise and fall can all play a significant role in coastal erosion or progradation by modulating these drivers. The authoritative reviews of shoreface morphodynamics provided by Wright (1995), Cowell et al. (1999) and Kleinhans (2002) brought to the fore many subsidiary questions regarding the articulations between waves and shoreface morphodynamics, mediated by complex feedback processes and by sediment grain-size. Many of these issues are still inadequately understood or pose questions relating to the long-term big-picture relationship between the beach-US and the LS. Among these issues, some are worth reiterating here (Kleinhans, 2002): (i) high variability in, and uncertainties concerning, the bed shear stress components involved in sediment transport, and poor testing and calibration of datasets; (ii) a poor grasp of interactions between waves and currents; (iii) lack of consensus on definitions of bedforms and their characteristics and genesis; (iv) poor knowledge on the exchange of sediment between the surf zone, the shoreface, and the shelf; and (v) paucity of datasets from measurements of both bedload and suspended load transport and near-bed hydrodynamics at high resolution, and none that allow the probabilistic integration to annual transport on the shoreface. Given the limited understanding of LS processes and sediment exchanges with the US and beach, it

may be necessary to acknowledge that at this point modelling work is necessarily still exploratory. Prediction, with confidence, of coastal behaviour and response to natural and anthropogenic perturbations is still a long way ahead.

This paper is aimed at providing an up-to-date review of research on the morphodynamics (as defined by Wright and Thom, 1977) of the LS and the potential sediment exchanges between the LS and the US/beach. Paradigms and processes of sediment transport and sediment exchange between these two domains in sandy, essentially (but not exclusively) wave-dominated settings, and the ways in which these processes have been mediated in the past, and still continue to be so, involving coastal sediment budgets and shoreline mobility across a range of embedded spatial and temporal scales, are examined. Although the focus of this review is on LS morphodynamics and sediment connectivity with the US and beach at timescales of up to 10^4 years, it is important stating at the outset that these themes are set within the framework of shoreface geology, tectonics, sea level, and stratigraphy. Steep shoreface substrates offer less accommodation, for instance, than low-gradient substrates (Roy et al., 1994; Cattaneo and Steel, 2003; Stolper et al., 2005; Cooper et al., 2018b). Sea-level changes, whether hinged on eustasy, tectonics or climate, control the location and translation of the shoreface and its trajectories in the depositional record (Coe et al., 2003; Hampson et al., 2008; Cattaneo and Steel, 2003; Løseth et al., 2006; Zecchin et al., 2019).

Following this introduction (1), we discuss the following themes in succession: (2) the current state-of-the-art in measurement and modelling of LS sediment transport and morpho-sedimentary change; (3) LS boundaries, sediment transport processes, and the concept of an equilibrium shoreface profile; (4) an overview of long-term (>10^2 years) sediment connectivity between the LS and the US, including aspects of shoreface morphodynamic/morpho-sedimentary equilibrium and disequilibrium; (5) sediment supply from the LS to the beach and vice versa, and implications, in both morphodynamic and sediment-budget terms, for both domains, with examples from the world's coasts; and, finally, (6) future perspectives in LS studies.

2. Sediment transport and morpho-sedimentary change on the LS: measurement and modelling

Sediment transport on the LS, and exchanges with the US and beach have been documented at a range of spatial and temporal scales. Monitoring of these processes is generally carried out through field measurements of waves, currents, and volumes and patterns of sediment mobilization. Measurements of seabed morphological changes provide a longer-term template for indirectly assessing the aggregated outcome of these processes. Increasing use of numerical models has been facilitated by improvements in model skill and computer power, aided by calibration using available, albeit often limited, datasets.

2.1. Instrumented moorings, seabed landers and near-bed field measurements

The beach and US are easily accessible through the subaerial section of the coast for deployment of equipment and shallow enough for the measurement of significant sediment transport during short-term (days-weeks) experiments. Research on beaches and on the adjacent US has, thus, had the greater share of efforts in trying to understand the processes responsible for coastal sediment mobilization and transport, and the associated morphological changes. Much less progress has been made on sediment transport mechanisms on the LS, where field data are vital to hydrodynamic and sediment transport modeling by informing on the often poorly-constrained bottom boundary conditions (Holman and Haller, 2013; Elko et al., 2015). Compared to the US, the spatial and temporal scales of sediment transport on the LS are much larger, while the transport rates (and associated morphological change) may be much smaller. Changes to the LS, though slow on average, can occur in response to events such as cyclones or other large storms that can generate significant bed mobility and morphological change. There are, thus, opportunities for short-term monitoring if we can "capture" such events. But this is still a major challenge, reliant on serendipity. Studies using state-of-the-art technology for monitoring of instantaneous to short-term hydrodynamics and sediment mobilization on the LS, especially during high-energy turbulent events when transport is expected to be larger, are still relatively scarce. A further challenge consists in being able to discern long-term trends from such short-term observations.

Significant sand transport under waves is restricted to the lower few dm of the water column, and measurement of this transport, at high resolution over unknown seabed configurations and with unpredictable bed-level changes occurring under hydrodynamic conditions that are unknown at the outset, is challenging. Such measurements require

deployment on or very close to the seabed, and both deployment and maintenance of instruments on the LS can be difficult and risky. There is also a strong possibility that in many LS environments, sediment movement could involve grain-by-grain mobility of bedforms of various sizes that is not easy to monitor using devices geared towards direct instantaneous or short-term measurements or estimations of suspended sediment concentrations and sediment transport from wave-current data. Determining the relative importance of different processes such as surface waves, winds, tides, and geostrophic currents at sufficiently high resolutions and spatial and temporal scales large enough to address their relative importance as drivers of sediment transport, is, therefore, still a major challenge fraught with uncertainties.

Notwithstanding these hurdles, the necessity of improving methods of deployment under mobile bed or high-energy conditions (e.g., Williams *et al*., 2003) has led to significant efforts being made on the optimisation of deployed instrument packages that are also beneficial to studies of the LS. A number of high-resolution studies of this type have been conducted and show the way forward using either side-by-side deployments of various instruments (e.g., Ferré et al., 2005) or a bottom lander (e.g., Zhang et al., 2016) on which various instruments are deployed, such as ECMs (Electromagnetic Current Meters), CTD probes (Conductivity, Temperature, Depth), OBS (Optical Backscatter Sensors), and ADCPs (Acoustic Doppler Current Profilers), or similar acoustic current meters, often equipped with pressure sensors, and LISST (Laser In Situ Scattering and Transmissometry) grain-size analysers (Fig. 3). Lagrangian field measurements using drifters can also yield useful information on surface currents (e.g., Brown et al., 2015), but currents measured by drifters do not necessarily correspond to actual transport of sediment.

2.2. Particle monitoring, seabed mapping, and shallow stratigraphy

Whereas the monitoring of sediment mobilization at short timescales (instantaneous to a few days) is hampered by logistic constraints, understanding of the morphodynamics of the LS is being enhanced by seabed-mapping techniques. These include monitoring of sediment facies, and measurements of bed-level changes using geophysical methods in particular. The former are generally based on bed sampling using grabs or direct diver collection of samples, and particle-tracking methods. Geophysical methods include sidescan multibeam echosounding, vessel-mounted sub-bottom seismic profilers, and interferometric

sonar. There is, indeed, a need for consideration of geological boundary conditions derived from preserved LS sedimentary records. Such records not only yield valuable information on processes and morpho-sedimentary change over time (e.g., Tamura et al., 2007; Cooper et al., 2018a), but are also easier to measure than modern mm- to cm-scale processes.

Bathymetric changes from geophysical data have been used as a proxy for patterns of short to long-term mobilization of bedforms on the LS (see *4.5*). The problem with this type of inquiry is that the limited temporal resolution hinders identification of the physical conditions that caused transport of sediment. Improvements in mapping are increasingly highlighting the complexity and variability of the seabed on the LS and continental shelf (e.g., Diesing et al., 2014, 2016; Linklater et al., 2019). High-resolution mapping of seabed sediment facies is important in distinguishing hard grounds from mobile sediment, in revealing local-scale sedimentological variability of the US and LS, and in contributing to more accurate determination of sediment connectivity between these two domains (e.g., Kinsela et al., 2020).

The use of underwater vehicles, including increasingly compact and autonomous vessels, in the acquisition of geophysical data and imaging of the bed topography of submarine environments, is also making headway (Wynn et al., 2014; Gafurov and Klochkov, 2015; Sahoo et al., 2019). Other areas of development are the inference of bathymetry from the dynamics of wave propagation using data derived from UAVs (Bergsma et al., 2019a; Simarro et al., 2019), shore-based coastal video monitoring systems (Bergsma et al., 2019b; Thuan et al., 2019), or X-band marine radar observations (Honegger et al., 2019). Airborne bathymetric LiDAR may be applicable in LS areas of clear water (e.g., Aleman et al., 2015), and can be combined with vessel-based multi-beam echosounder surveys to yield high-resolution data (e.g., Kinsela et al., 2020), but optically opaque waters limit its utilization. Another area of development is that of the use of multi- or hyper-spectral satellite remote sensing data to generate inexpensive bathymetry but these data sources are also subject to atmospheric and water clarity constraints (Gao, 2009). They generally provide more reliable bathymetric estimates in depths < 10 m (e.g., Pacheco et al., 2015). High-resolution optical satellite video imagery holds promise for deriving accurate LS bathymetry over large regional scales of up to 100 km^2 based on wave propagation (Almar et al., 2019), while machine-learning techniques are deemed as useful in sorting out, ordering and classifying bathymetric data from large series of satellite images (Sagawa et al., 2019). Overall, there

are significant on-going developments in optimizing sea-bed mapping (Stephens and Diesing, 2016; Diesing et al., 2016; Linklater et al., 2019; Sagawa et al., 2019).

A number of coastal experimental facilities aimed at long-term (multi-decadal) measurement of shoreface bathymetric or shoreface profile change have been operating now for decades. These facilities also support short-term (days to weeks) hydrodynamic and sediment transport monitoring. Their operation has contributed to acquisition of datasets on the morphodynamic and sedimentary coupling of the LS with the beach at short (days to weeks) to long (multi-decadal) timescales. They include the Duck Field Research Facility of the U.S. Army Corps of Engineers near the town of Duck, North Carolina, United States (http://www.frf.usace.army.mil), the Jarkus profile datasets on the Dutch coast (https://publicwiki.deltares.nl/display/OET/Dataset+documentation+JarKus), and the experimental station at Hasaki beach on the Pacific coast of Japan (https://www.pari.go.jp/unit/edosy/en/main-facility/). Various research groups, mainly in Europe, the United States and Australia, have also been active in coastal field experimental work extending down to the LS.

River delta lobe switches often generate discernible bathymetric changes affecting the LS, and, where data are available, such bathymetric changes can be used to establish regional multi-decadal sediment budgets (e.g., Sabatier et al., 2006; Brunel et al., 2014; Patterson and Nielsen, 2016) of the LS and their relationship with shoreline changes. Patterson and Nielsen (2016) analyzed, for instance, 46 years of changes in LS bathymetry (in 10–20 m depths) over an abandoned ebb-delta lobe at the northern Gold Coast, Australia (*5.1*). Opportunistic events, such as large exceptional sediment discharge from river mouths, can serve as a template for obtaining data on short-term LS change (e.g., Maillet et al., 2006). Nearshore nourishments could provide a means of detailed short-term monitoring of shoreface-to-beach sand transfers, generally described as the ‘feeder’ effect (e.g., van Duin et al., 2004; Ojeda et al., 2008), but such projects, including the Sand Engine or Sand Motor in the Netherlands (Stive et al., 2013; Huisman et al., 2019), have, thus far, been limited to the US.

2.3. Numerical and behaviour-oriented modelling

The difficulties of in-situ measurement of both LS hydro- and sediment dynamics and the resulting morpho-sedimentary change, as well as the limitations associated with

integrating measured short-term hydrodynamic and sediment flux data over longer time frames of change, have prompted increasing use of modelling or a combination of modelling with in-situ measurements. Modelling can overcome spatial and temporal limitations provided that good model skill and computer power are combined with sufficient observational data for calibration and validation. Modelling of US and beach sediment transport based on empirical data has long been a popular approach that has also been applied to LS environments (e.g., Héquette et al., 2008; Aagaard, 2014; Simarro et al., 2015; Aouiche et al., 2016; Ruggiero et al., 2016; King et al., 2019; Valiente et al., 2019a, 2019b), aided by hydrodynamic and sediment-transport modelling packages such as Delft 3D (https://oss.deltares.nl/web/delft3d/) and Mike 21 (https://www.mikepoweredbydhi.com/products/mike-21), and supported by existing wave and bathymetric data. Important datasets of this type have been acquired for some regions, such as on the Dutch coast. The problem with this type of inquiry is that the limited temporal and spatial coverage of datasets hampers identification of the physical conditions that cause transport of sediment and, hence, our understanding of the morphodynamics. Without this understanding, there is no chance of getting the modeling effort correct, as we may have no more than a limited grasp of the relevant processes that should be modeled.

At longer timescales ($>10^3$ years), behaviour-oriented modelling of sediment-transport regime and profile kinematics has also been used to derive long-term changes in LS morphology and potential sediment-sharing with the US and the beach (e.g., the Shoreface Translation Model: Cowell et al., 1995, 2003b, 2006; de Vriend, 2003; Dillenburg et al., 2012; BARSIM: Storms et al., 2002; Storms, 2003; GEOMBEST: Stolper et al., 2005; Moore et al., 2010). Such models assume a pre-determined LS equilibrium configuration and model its translation based on mass conservation principles, but ignore explicit sediment transport processes in order to produce simulations of large-scale/long-term coastal behaviour and shoreface profile response to sea level. Numerical experiments with such models have highlighted the importance of substrate slope, a fundamental concern in the concept of the equilibrium shoreface profile (see *3.3*). This type of modelling approach involves simulation of outcomes that are consistent with general hydrodynamic principles affecting the shoreface as a whole, and with beach and barrier morphology and stratigraphy. This is an option that is appropriate for estimating the LS response at depths and timescales that preclude direct measurement of the morphological response. The chronology is constrained

through radiometric dating and sea-level history, but, as noted by Cooper et al. (2018b), the geological controls are difficult to quantify, and are often overlooked, ignored or grossly simplified.

3. LS boundaries, sediment transport processes and the equilibrium profile

3.1. LS boundaries: depth of closure and wave base

The US comprises the breaker and surf zones (Aagaard, 2014; Ruggiero et al., 2016) and, in contrast to the LS, is a morphologically active zone commonly containing one or more nearshore bars, while the latter is typically relatively planar (Fig. 1), and under normal, average or non-storm conditions waves are shoaling rather than breaking (*3.2*). On barless beaches, however, the landward limit of the LS would lie a short distance seaward of the step at the base of the beach face. Interactions between the LS and US and beach involve a range of spatial and temporal scales, from instantaneous sediment transport rates through event-scale to longer time- and space-integrated morphological change. The shoreface is, thus, affected by a hierarchy of process cascades in which coastal behaviour at any intermediate level results from the residual effects of processes operating at timescales of seconds to weeks, or years as we move onto the LS (Cowell et al., 2003a, 2003b). The US and its onshore extension, the beach, commonly exhibit rapid accretion and erosion in response to waves and storms in particular. Changes in bed elevation are generally largest over the beachface and surf zone and progressively decrease offshore, forming an envelope of profile change that pinches out seaward. Over short timescales (< 1 year), the US and the beach may, therefore, behave, as a closed zone, with no significant sediment exchange with the LS. The offshore decrease in the range of vertical bed-level change on the US is caused by: (1) the offshore diminution in the intensity of near-bottom flows, especially seaward of the surf zone (Fig. 4, see *3.2*), and (2) the fact that sand accumulating in the surf zone and in the vicinity of the beach under accretionary conditions is spread out, during erosional periods, over a much wider area offshore and, therefore, forms a much thinner sediment cover. There is, therefore, no distinct morphological manifestation of such sand blanketing offshore within a typical year (Cowell et al., 1999). The envelope of profile change eventually diminishes to the 'depth of closure' (DoC). The DoC was initially defined by engineers to designate a short-term seaward limit of inter-seasonal to annual bed change over the shoreface (Hallermeier, 1981; Birkemeier, 1985). The DoC is usually determined using

Hallermeier's formulations or derivatives thereof, for model applications around the world, sometimes supported by profile surveys that provide approximate long-term bathymetric calibration. Udo et al. (2020) have drawn attention, however, to two major unresolved issues associated with computing the DoC in this way: the generic applicability of the coefficients used to compute the DoC, and the accuracy of the wave data required for reliable DoC computations. More rarely, grain-size data have also been used to identify the DoC (e.g., Aragonés et al., 2018). The DoC has been documented from a geological perspective with regards to wave ravinement (Wallace et al., 2010). Recent studies on the hydrodynamic and morpho-sedimentary implications of the DoC include those of Ortiz and Ashton (2016) and Valiente et al. (2019a, 2019b).

Prediction of the horizontal mobility of the US and beach at timescales of less than a decade generally does not take into account sand exchanges with the LS, and the DoC has been assumed to represent the limit of 'significant' cross-shore sediment transport (Hanson et al., 2003). Indeed, rates of sediment supply from the LS to the US and beach in wave-dominated settings indicated by various lines of evidence reviewed by Cowell *et al*. (2001) are typically on the order of 1 m^3 a^{-1} per metre of shoreline, a volume corresponding to a lowering of the LS by only a few millimetres a year. The difficulties of measuring such rates are clearly evident. Prediction of processes acting over timescales of decades and longer necessitates the hard task of resolving small net changes in a system characterised by large fluctuations (de Vriend, 2003). Since the US and beach system have cross-shore length scales that are typically one to, maybe, two orders of magnitude smaller than those of the LS, changes affecting the latter are associated with disproportionately larger changes on the former due to mass continuity for sediment exchanges between the two domains (Roy et al., 1994; Cowell et al., 1999; 2003b). van Heteren et al. (2011) suggested, for instance, that for a coastal tract extending 10 km offshore, a vertical uncertainty of 0.05 m (which is at the limit of the most accurate seabed surveys) corresponds to a volume uncertainty of 500 m^3/m seaward of the (barrier) coastline. For a 1 km-wide barrier, this volume translates on average into an extra 0.5 m of sand lost or gained across its entire surface.

A number of studies have shown, however, relatively significant rates of LS reworking, thus translating into large rates of onshore migration of the limit corresponding to the DoC. In Skallingen, Denmark, for instance, this limit clearly moves onshore at a rate of 10-15 m/yr (Aagaard et al., 2004), representing an onshore sand supply of 7-8.5 m^3/yr per m

of shoreline (Aagaard et al., 2004). Even larger net annual onshore transport rates have been determined by Patterson and Nielsen (2016). Such changes in shoreface morphology question, thus, the validity of the DoC (as defined by wave parameters) for longer timescales. The idea of a DoC may not even hold on shallow macrotidal sand-rich shorefaces where recurrent storm activity can drive sand banks onshore, across the US-LS (Fig. 5), as on the southern North Sea coast of France (Héquette and Aernouts, 2010; Anthony, 2013), as well as in relatively sediment-limited shorefaces but where deeper large bedforms can be mobilized during storm events (e.g., Backstrom et al., 2015). Examples of such large bedforms are further discussed in *4.4*. Analyses of multi-annual beach profile data at the Duck experimental site extending well offshore on the east coast of the United States have shown that the DoC increases with increasing timescale (Nicholls et al., 1998; Schwartz and Birkemeier, 2004). It is not, therefore, a true cross-shore sediment transport limit over time (Stive and de Vriend, 1995; Hinton and Nicholls, 1998, 2007; Aagaard, 2014), but rather a seaward limit of 'significant' sediment transport gradients on the US.

The DoC may also vary alongshore at length scales of kilometres (e.g., Hicks et al., 2002; Sabatier et al., 2005). Stretching of the active profile beyond the typical DoC may be induced by severe storms, especially when these are closely spaced, occurring in clusters (Inman et al., 1993; Lee et al., 1998). Cheng and Wang (2019) defined a 'short-term' DoC off the low-energy west-central Florida coast, and its offshore extension following the highly obliquely shore-incident Hurricane Irma in September 2017, which generated a negative surge of 1.1 m that was deemed to have significantly lowered wave base. As Cowell et al. (1999, 2003a, 2003b) have recalled, however, essentially nothing is known of fluctuations in DoC over timescales of decades or more. Hinton and Nicholls (2007) documented DoC fluctuations from profile datasets covering over 25 years. Similarly, little is known of how profile shapes, and consequently, profile depth fluctuations, are morphologically related to extreme storms. Thom and Hall (1991) suggested that gradual beach recovery involves re-smoothing of the perturbed profile, as sand eroded from the beach and dunes and deposited on the LS during cyclones is returned shoreward by swell waves over periods of several years.

The seaward limit of the LS is even less amenable to seabed morphological and sediment-transport surveys than the DoC, and harder to define. This limit may correspond to 'wave base' (Fig. 1), but this is, by no means, an agreed definition, and hydrodynamic drivers

other than waves may affect the LS beyond wave base. Wave base generally fluctuates between storm waves and fair-weather waves (Peters and Loss, 2012), although continental shelves with long modal swell waves do not clearly conform to the two categories of storm and fair-weather. In fact, although there is consensus on the term, there is some variation in the literature on its definition. Baker et al. (1966) defined wave base as the depth beyond which 'wave action ceases to stir the sediment', and Carter (1988) as a point where the motion of surface waves just reaches the bed, this being presumably storm wave base. Kinsela et al. (2020) identify, for instance, a break in slope at the outer limit of the LS and a depth of wave base somewhere seaward and deeper than this break in slope.

Considering the definition proposed by Wright (1995), which regarded the LS as the region where ‘waves normally (or frequently) agitate the bed’, the seaward boundary of the LS may be identified as the depth where gravity waves can stir the sand on the seabed under ‘normal’ conditions, which we define here as annually averaged wave conditions. The threshold for sediment stirring by gravity waves is often predicted using the Shields parameter:

$$\theta = \frac{½ f_w u_{max}^2}{(s-1) g D_{50}} \quad (1)$$

where f_w is the wave friction factor, typically of O(0.01) for fine-medium sand, u_{max} the maximum orbital velocity in the wave cycle ($2\sigma_u$, where σ denotes the standard deviation; Thornton and Guza, 1983), (s-1) relative sediment density, g acceleration of gravity and D_{50} the median (sand) grain diameter on the seabed. For typical shoreface sand grain sizes, the threshold Shields number for sediment stirring is $\theta_c \approx 0.05$ (e.g., Nielsen, 1992). The equation yields a critical value of the orbital velocity and linear wave theory can be used to estimate the maximum depth of the LS, given annually averaged wave height/period. For example, assuming D_{50} = 0.2 mm, and annually averaged wave conditions H = 1.5 m, T = 10 s, the critical wave orbital velocity is approximately 18 cm/s. Using linear wave theory, the seaward boundary of the LS is then located at $h \approx 40$ m.

However, although it may be conceptually convenient to consider the shoreface in terms of depth and wave-limit variations (Fig. 1), and occasionally grain size and type, identifying dynamic, morphological, and sedimentological limits across any active shoreface profile over time is, altogether, another matter. Challenges include the potential variability of the inherited shoreface morphology and substrate (potential susceptibility of the

substrate to liberate sediment, variable carbonate production, for instance, and spatio-temporal variability in grain characteristics), and the variability of the ocean climate (choice of representative wave characteristics).

3.2. Wave-driven sediment transport across the LS

The ubiquitous nature of surface gravity waves over the oceans and their transformation across the world's continental shelves (Ardhuin et al., 2003), between wave base (Peters and Loss, 2012) and the beach, underscores their importance in potential bed mobilization. Identification of the sediment transport processes involved in the co-adjustment between a sedimentary shoreface profile and waves is, however, extremely elusive, given the complex nature of sediment transport (Amoudry and Souza, 2011; Wainwright et al., 2015), both in space and time. Hence the use of proxies such as the geometric shoreface profile by Bruun (1954) and Dean (1977, 1991), further reviewed in *3.3*, in which sediment transport is implicitly assumed to occur through gradient diffusion (Wright, 1995), and/or Bagnold-type energetics transport models (e.g., Bowen, 1980; Bailard and Inman, 1981; Aagaard and Sørensen, 2012; Ortiz and Ashton (2016).

It is well known from measurements on the US that net (cross-shore) suspended transport of sediment can be expressed as:

$$q_x = \int_{z=0}^{z} \left(< u_z >< c_z > \right) + < u_z' c_z' > dz \qquad (2)$$

where z is elevation above the seabed, u_z cross-shore fluid velocity and c_z suspended sediment concentration, both determined at z (e.g. Jaffe et al., 1985; Huntley and Hanes, 1987; Aagaard et al., 2013). Angular brackets indicate the time-averages while primes represent oscillatory wave components. The first term represents transport by mean flows, such as onshore-directed (Lagrangian) mass transport due to Stokes drift and wave breaking, and offshore-directed (Eulerian) undertow and rip currents that conserve mass. The second term represents transport due to short and long-wave orbital motions resulting from wave skewness, asymmetry and/or the phase coupling between u' and c' (Jaffe et al., 1985; Aagaard et al., 2013; Aagaard and Hughes, 2017). According to these concepts, advective transport rather than sediment diffusion dominates cross-shore sediment exchange.

The process assemblage is, however, somewhat different on the LS. Transport processes exclusively associated with wave breaking (such as wave asymmetry, rip currents

and mass transport due to breaking waves) are clearly absent on the LS (Fig. 4). Moreover, Stokes drift and undertow tend to be balanced everywhere (Fig. 6) in the vertical (Brown et al., 2015) and the first term in Eq. (2) may be negligible. On the other hand, wind-driven mean flows such as up/downwelling (Fewings et al., 2008), as well as gravity, are likely to assume increased importance in the cross-shore sediment exchange, partly because near-bottom wave orbital velocities and wave skewness decrease offshore as depth increases towards wave base. The increasing depth also implies an increasing proportion of bedload transport since shear stress on the seabed decreases. Theoretically, suspended load and bedload are equivalent when the ratio of friction velocity (u_f) to sediment fall velocity (w_s), $u_f/w_s \approx$ 0.8-1 (Bagnold, 1966; Bowen, 1980) and suspended load dominates for larger ratios. For a wave height H = 2 m, with a period of T = 8 s and a sediment grain size of 0.2 mm, suspended load would dominate in water depths shallower than approximately 20 m (Aagaard and Hughes, 2017).

Generally, the range of processes contributing to shoreface sediment transport and the non-linear interactions between them, as well as vertical inhomogeneity make calculations and modelling of both the direction and magnitude of sediment transport on the LS uncertain, as demonstrated by Eulerian field measurements of suspended sediment transport seaward of the breaker zone using either acoustic Doppler techniques (Lacy et al., 2005; Ferré et al., 2010; Aagaard et al., 2010; Aagaard, 2014), or electromagnetic flow meters and optical sensors (Wright et al., 1991; Ruessink et al., 1998). Further complication arises because field data explicitly describing bedload transport in this domain are rare (e.g., Wright et al., 1991; Ardhuin et al., 2002; Aagaard, 2014; Simarro et al., 2015; Guererro et al., 2018; McCarroll et al., 2018).

Nevertheless, the available field evidence has shown that once suspended into the water column, sediment grains can be transported by a suite of mean and oscillatory fluid motions. Referring back to Eq. (2), onshore sediment transport is primarily controlled by onshore-skewed oscillatory wave motions associated with wave shoaling (Hsu and Hanes, 2004; Ruessink et al., 2009; Grasso et al., 2011) and this is consistent with results from week-to-month-long instrument deployments for measurements of suspended sediment transport on the LS (Ruessink et al., 1998; Aagaard et al., 2012). Offshore transport, on the other hand, is mainly accomplished by potential excess of Eulerian compared to Lagrangian mean flows, downwelling, and gravity (Aagaard and Hughes, 2017). Results of these aspects of transport

from short-term transport measurements can probably be aggregated to represent longer-term tendencies (Fig. 7) as suggested by Aagaard (2014). On the basis of these empirical measurements, Aagaard and Hughes (2017) proposed a model to predict profile and shoreline change for different wave conditions and sea-level trends. It was found that a gently sloping shoreface favours net onshore transport of sediment and shoreline progradation, while the opposite is the case for a steeply sloping shoreface, which is consistent with exploratory simulations from morphological-behaviour models (e.g., Cowell et al., 1995).

While these various studies have been able to resolve suspended sediment transport over significant parts of the water column, instrumentation necessary to determine bedload transport under LS field conditions is still lacking, compounded by the risks and difficulties with deployment of instrument systems in relatively deep water. Such risks and difficulties may even be considered as potential hurdles when funds are sought and need to be justified for this sort of work. And, although the afore-mentioned studies have increased our understanding of the processes driving sediment transport on the LS on short/medium-term time scales, it is unclear how these measurements can be upscaled to explain the long-term LS morphological evolution, which is often inferred from bathymetric, morphological and stratigraphic data (see section **5**).

In addition to the difficulties of conducting sediment transport experiments on the LS, especially during storms when transport rates are expected to be more significant, other aspects that hinder the comprehension of transport processes are: (1) directional changes, from the inner shelf to the beach, of wave incidence angles resulting from large regional spatial-scale refraction processes or generated by 'anomalous' storm directions (see Harley et al., 2017, for instance), (2) sediment mobilization by other mechanisms in addition to waves (see *3.4*), (3) the preponderance in many LS environments of large-scale bedforms (see *4.4*) which are probably migrating due to bedload movement, and (4) the potential influence of multiple sediment fractions, as Huisman et al. (2018) have shown in an assessment of sediment transport on the LS off the Sand Motor in the Netherlands.

3.3. The equilibrium shoreface profile

The principle of morphodynamic feedback between hydrodynamic processes, movable sediment, and shoreface morphology is commonly deemed to generate a shoreface

profile that tends towards an equilibrium shape for any given stable wave climate. Along this profile, local, time-averaged net cross-shore sediment transport is assumed everywhere nil (Bowen, 1980; Inman et al., 1993). Equilibrium over periods of years or decades implies that the sum of all onshore sediment fluxes is balanced by that of all offshore fluxes (Wright et al., 1991). In this context of morphodynamic adjustment, a profile of specific grain size exposed to constant forcing conditions (e.g. wave climate), will evolve into a shape that displays no net change in time, although sediment may be mobilized (Larson et al., 1999).

These principles conform to the concept of an equilibrium concave shoreface profile reflecting the sedimentary shoreface substrate response to the wave climate (Dean, 1991). This concept dates back to Cornaglia (1889) who speculated that equilibrium emerges when onshore transport of sediment due to velocity-skewed incoming waves is everywhere balanced by offshore transport due to gravity. The increasing velocity skewness in progressively shallower water led to the prediction of a concave shoreface shape with progressively coarsening sediment grain sizes towards the shoreline. This aspect of grain size will be discussed in later subsections. The concept of an equilibrium profile shape has underpinned coastal engineering applications for decades since it was refined by Bruun (1954), following the earlier work of Cornaglia (1889), Fenneman (1902), Johnson (1919), Keulegan and Krumbein (1949), among others, to obtain an equilibrium shoreface shape. The equilibrium profile has been considered as 'perhaps the single most important concept in the field of nearshore processes' (Inman et al., 1993, p. 18181). It has been used in many studies to model shoreline change, especially under sea-level rise, using the so-called 'Bruun Rule' (Bruun, 1988), the shortcomings and overly simplistic generalizations of which have been highlighted (Thieler et al., 2000; Cooper and Pilkey, 2004; Cooper et al., 2020).

Several authors have raised objections to the concept of a smooth concave 'equilibrium' shoreface profile shape, as predicted by the models of Bruun (1954) and Dean (1991), notably because underlying geology and substrate characteristics can have an overarching role in determining the shape of this profile (Pilkey et al., 1993; Thieler et al., 1995; Cooper and Pilkey, 2004). There are many shelves where the presence of abundant LS bedforms (some of which may be in a state of long-term morphodynamic equilibrium, see *4.4*), or features inherited from previous sub-aerially exposed terrestrial geomorphic regimes, contribute to profile shapes that are all but concave. Local depositional controls and complex time-varying shoreface sediment transport (see *3.2, 3.4*) also imply that a

theoretical equilibrium profile shaped by wave-driven sediment transport across the shelf and shoreface is rarely attained in nature (Wright, 1995). In this idealised equilibrium configuration, the long-term LS sediment-supply perspective is purely cross-shore, and sediment transport is assured by waves (see *3.2*). But, as shown in *3.4*, in many situations, longshore transport on the LS can play an additional, sometimes determining, role. Cowell et al. (1999) also highlighted the severe shortcomings of the concept, but conceded that insights from equilibrium theory can be used to develop a more general and unified understanding of shoreface variability among the contrasting environmental settings existing in nature. The comparison of a shoreface against a theoretical equilibrium (see *4.1*) provides a means of evaluating shoreface behaviour and time-dependent shoreface change relative to its equilibrium or disequilibrium context (Daley and Cowell, 2013). Large-scale coastal-behaviour models, for instance, use the shoreface equilibrium profile as a fundamental morphological unit that is translated in space to simulate coastal depositional system response to, for example, sea-level oscillations and variability in sediment supply (e.g., Cowell et al., 2003a, Stolper et al., 2005; Lorenzo-Trueba and Ashton, 2014; Patterson and Nielsen, 2016).

3.4. Other drivers of sediment transport on the LS

While the wave-dominated LS is the archetype reflecting the pervasive influence of waves at the global scale, many LS and inner shelves are affected to varying degrees by wind-driven currents and associated upwelling and downwelling, tidal currents, temperature-driven currents and saline density currents from fluvial freshwater flow (e.g., Kleinhans, 2002; Lentz and Fewings, 2012; Grifoll et al., 2015). These flows are influenced by regional characteristics, including LS morphology (e.g., Héquette et al., 2001; Backstrom et al., 2008, 2015; Lavoie et al., 2014). Directions of transport (including offshore) may also vary; the number of studies is still too limited to account for the large directional range of transport possibilities. From the perspective of the sediment exchanges between the LS and the beach, cross-shore processes are the overarching mechanism, while longshore processes can influence sediment mobility and morphological change on the LS, but with no direct cross-shore exchange in principle. However, by moving sediment alongshore on the LS, longshore transport can significantly influence the local to regional cross-shore sediment budget, as shown below.

The sediment transport processes generated in the transition from the LS to the beach are, thus, essentially cross-shore, even where wave approach is oblique. However, tides can be important in both direct shoreface sediment transport through strong tidal currents (Anthony and Orford, 2002) and the indirect cross-shoreface modulation of wave influence by large tidal ranges (Dashtgard et al., 2012; Yang and Chang, 2018). Tidal processes, combined with wind forcing, especially during storms, can have a significant impact on longshore sediment transport on the LS as shown by both field/modelling studies (e.g., Kleinhans and Grasmeijer, 2006; Héquette et al., 2008; Aagaard, 2011) and modelling approaches (e.g., Davidson et al., 2008; Giardino et al., 2010; King et al., 2019; McCarroll et al., 2018; Valiente et al., 2019a, 2019b; Latapy et al., 2020).

Variations in LS gradient and width are a large-scale control on spatial patterns and energy distribution of waves (including refraction) and (tidal) currents (e.g., Anthony and Orford, 2002). Porter-Smith et al. (2004) proposed, for instance, a classification of the Australian continental shelf on the basis of predicted sediment threshold exceedance from tidal currents and swell waves (Fig. 8). In similar manner, King et al. (2019) proposed a classification scheme of the energetic, macrotidal LS off southwest England based on sediment transport magnitude due to wave-forcing, tide-forcing, and nonlinear wave-tide interactions, and suggested that different wave/tide conditions have implications for sediment transport direction and distribution. On macrotidal LS associated with shallow epicontinental tidal seas such as the eastern English Channel, the southern North Sea and the Gulf of Bohai, where tidal currents exceed about 0.5 m s^{-1} and sand is abundant, the sand transport regime exhibits a dominant longshore tidal signal modulated by storms with directions that may vary (Harff and Zhang, 2016). In this context, while strong tidal currents lead to long-term alongshore mobility of sand banks, strong storms associated with onshore winds and waves can drive these banks onshore (Fig. 5) especially during neap tides, leading to oblique shore-attached bedforms, as documented in several studies (Héquette and Aernouts, 2010; Anthony et al., 2010; Anthony, 2013).

In addition to a limit of ‘significant’ morphological change associated with the DoC, and consistent with the foregoing observations, Valiente et al. (2019a) identified a maximum depth of extreme bed activity and sediment transport they termed the Depth of Transport (DoT). Under extreme conditions, the DoT exceeds 30 m depth (Fig. 9) in their study area off the macrotidal, high-energy coastline of SW England. An important aspect of their findings is

that tides can deepen the DoT estimate by ~10 m on macrotidal shorefaces, representing a 30% increase compared to tideless settings. The tidal effect can, thus, have a significant bearing on the local LS sediment transport and budgets (long-term enrichment or depletion in places), and consequently, on the beach-dune sand budgets and accretion or erosion. Sedrati and Anthony (2014) drew attention to the difficulty of using the coastal cell concept (Bowen and Inman, 1966; Carter, 1988; Bray et al., 1995; van Rijn, 2011) as a management tool on the macrotidal coasts of the Dover Strait and the southern North Sea where cell boundaries cannot feasibly be delimited because of a wide and active LS. One reason for the failure of many operations aimed at countering shoreline erosion is that their conception and implementation are not commonly based on a sufficient knowledge of processes shaping the coast at various morphodynamic scales (Sedrati and Anthony, 2014). A better understanding of these larger-scale transport pathways and a perspective based on the consideration of sediment supply from the LS to the beach are, thus, important in the establishment of regional coastal sediment budgets (Sedrati and Anthony, 2014; Valiente et al., 2019b; Kinsela et al., 2020).

On the LS, upwelling and downwelling Ekman currents associated with wind-driven geostrophic flows can contribute significantly to cross-shore sediment transport (Kleinhans, 2002), as well as alongshore transport of suspended sediment (Fewings et al., 2008; Aagaard, 2011; Lentz and Fewings, 2012). Aagaard (2011) found that wind-generated longshore currents were instrumental in removing sand supplied to the LS from the beach and US off the accreting Wadden Sea coast of Vejers, Denmark (see also *5.2*). Indeed, various studies have shown that surf-zone sediment dynamics can mediate spatial patterns of shoreface elevational change and net losses offshore, for example through the formation of transient rip currents (Johnson and Pattiaratchi, 2004; Hally-Rosendahl et al., 2014). Apart from undertow processes just seaward of the surf zone, downwelling currents generated by storm surges are deemed to be an important mechanism of offshore sediment transport (e.g., Wright et al., 1991, 1994; Cudaback and Largier, 2001; Héquette et al., 2001; Goff et al., 2010, 2019). Keen et al. (2003) suggested that storm-induced losses in sediment deduced from bed elevational changes at the Duck site down to a shoreface depth of 13 m were most likely replaced by sediment transported alongshore on the US. Black et al. (2008) identified, from oceanographic measurements made in the surf zone (5.5 m depth) and on the inner shelf (8 m depth, and likely corresponding to the LS) in the monsoonal setting of southwest

India, and from numerical modelling, a closed sedimentary circulation system they described as 'step-ladder'. In this setting, the regional-scale dynamic sediment equilibrium is dominated by an annual net northerly flux of wave-suspended sediment on the US driven by wave-induced currents that is balanced by a net southerly flux of wave-suspended sediment driven by wave-induced flows, but on the LS/inner shelf. The two counter-directional sedimentary pathways are linked by cross-shore bridging transport. Green et al. (2012) showed a strong influence of the Agulhas Current off the high wave-energy shelf of South Africa in modulating sediment transport and deposition, resulting in what may be considered as a mixed storm- and boundary-current-dominated LS. Waves are considered as an important stirring mechanism on these LS, with the mean currents from tidal, wind- and density-driven flows generating bedload transport and suspension transport of the finer sand fractions.

Kleinhans and Grasmeijer (2006) have shown that tide-driven bedload transport is an important portion of the net annual sediment transport rate on many LS, but have also drawn attention to the quasi-impossibility of isolating the fraction of bedload transport associated with waves and currents, respectively. Finally, while the thrust in this review is on cross-shore sediment transport from the LS to the beach and vice versa, offshore transport mechanisms on the shelf, notably turbidity currents, and other types of underwater sediment density flows, not treated here, have been considered as, arguably, the most important flow processes for moving sediment on Earth (Talling, 2014; see also recent collection of papers edited by Asch et al., 2020).

As a result of the complexity of many LS, sand transport mechanisms cannot simply be considered in terms of direct cross-shore exchanges, although this is a convenient frame of study, especially in terms of net sand supply to the benefit of the beach, or vice versa. Denny et al. (2013) documented, for instance, long-term LS bedload transport controlled by the trajectory and intensity of storms, resulting in a dominance of net alongshore flow compared to cross-shore transport. The potential for bedload transport along the LS to bypass bold headlands, and contribute to sediment supply to adjacent embayments, has been underlined (Goodwin et al., 2013; McCarroll et al., 2018; Valiente et al., 2019a, 2019b, 2020). The potentially large range of process combinations prevailing on the LS generates both stochastic and dynamic uncertainty (Cowell et al., 1999). Stochastic uncertainty is attributed by these authors to the joint occurrence probabilities of various types of flows with respect

to their magnitudes, frequencies and directions, while dynamic uncertainty is due to the consequences generated by the non-linear interactions between these flows. Among these consequences is the large range of bedforms on the LS and shelf (*4.4*).

4. LS morphodynamic/morpho-sedimentary equilibrium/disequilibrium over time, and bedforms

In terms of long-term sedimentary coastal development, the overarching consideration in the link between the LS and the beach resides in the potential sediment exchanges between these two domains. Over the long timescale (10^3-10^4 years), this sediment connectivity may be gauged in terms of equilibrium or disequilibrium relative to LS sediment mobilization and morphology (Fig. 10). Long-term equilibrium over a sedimentary shoreface may be achieved: (1) when onshore-offshore sediment transport balances out, as conceptualised in the equilibrium shoreface profile; (2) in a sediment-limited context wherein the LS sediment reservoir is exhausted (probably very rare) or movable sediments lie deeper than a shoreward-migrating wave base; or (3) in a sediment-rich context where dynamic forcing is limited by considerable wave-energy dissipation, sometimes resulting in a tide-dominated situation.

Given the focus of this review, we will leave aside the simple concept of a concave wave-driven equilibrium shoreface profile (*3.3*) and move towards a consideration of the LS as a short- to long-term source or sink for sediment, and how this may translate into morphodynamic (i.e., sediment-transport and morphological) equilibrium over time. We do this without losing sight of the dominant transport mechanisms at play on the LS: onshore transport of sediment mainly due to skewness of shoaling waves, and offshore transport due to downwelling currents and gravity (Fig. 4), which, when balanced, are deemed to yield equilibrium.

Daley and Cowell (2013) describe three 'shelf regime' modes to conceptualise the state of the LS and its potential sediment connectivity with the US and beach: underfit, overfit and graded, using terminology borrowed from fluvial geomorphology. We adopt these modes here (Fig. 10). An underfit regime is morphologically defined as pertaining to a LS that is too deep or steep for equilibrium (Fig. 10a) under given conditions of sea-level, coastal hydrodynamics, and sediment characteristics. It is characterised by positive sediment accommodation wherein the LS is underfilled with sediment, providing opportunity for

offshore transport and deposition of mobilized sediment (Daley and Cowell, 2013). The converse applies on an overfit LS which is too shallow or flat for equilibrium. It is characterized by negative accommodation capacity, and overfilled with sediment, with a tendency for divergent, across-shelf sediment transport towards the shore (Fig. 10b). A graded regime applies by definition when the LS is in equilibrium (Fig. 10c) with the forcing and transport regime for endemic sediments, leading, thus, to neutral accommodation (Daley and Cowell, 2013). To avoid overlap with Swift and Thorne's (1991) complementary accommodation- and sediment-dominated shelf regime models and to focus on the implications for shoreface morphodynamics, Kinsela et al. (2020), and Kinsela (pers. comm., June 2020) adopted these three regimes in terms of shoreface 'morphodynamic states'.

4.1. Underfit disequilibrium

Where the LS is in an underfit morphodynamic state (positive accommodation), being too steep or too deep for equilibrium to prevail (Fig. 10a) offshore-directed supply of sediment from the beach via the US may occur, due to downslope transport by gravity (Aagaard and Hughes, 2017) or downwelling currents. More commonly, terrestrial sediment supply to the LS occurs where prograding river deltas develop clinoforms, which are inclined and normally offshore-dipping horizons occurring over a range of spatial and temporal scales, and which represent sediment pathways to the shelf (Patruno and Helland-Hansen, 2018). Positive LS accommodation associated with sourcing of sediment by shoreline retreat typically occurs with sea-level rise through wave erosion of the US (Zecchin et al., 2019). This 'ravinement' leads to the liberation of sediment that may be deposited over the LS as a transgressive sand sheet (e.g., Roy et al., 1994; Moore et al., 2010), while the beach may be preserved as a retreating barrier (Cooper et al., 2020) or drowned by rapid sea-level rise (Fig. 11a). Sediment depletion of the US, resulting in shoreline erosion and offshore transfer to the LS, can also occur during stable sea level.

For wave-dominated contexts, the extent to which the US and beach are reworked and the beach-barrier shoreline preserved as a retreating entity (the case of many currently retreating barriers where sea-level rise has been prevalent during the Holocene) would depend on the ambient energy relative to the following parameters (Fig. 11b): rate of sea-level rise, sediment supply from cross-shore and alongshore sources, the substrate gradient and composition and its resistance to erosion (Cooper et al., 2018b; Zecchin et al., 2019),

which are the main factors controlling wave ravinement. On the other hand, several studies have documented shoreline progradation in the face of rising sea level due to large alongshore and/or cross-shore sediment supply (Stive, 2004; Aagaard, 2011; Ruggiero et al., 2016; Fruergaard et al., 2018).

4.2. Overfit disequilibrium

Where the LS is too shallow (negative accommodation) for equilibrium to prevail (Fig. 10b) onshore-directed supply of sediment to the beach via the US may occur due to positive gravity wave skewness (Aagaard and Hughes, 2017) or upwelling currents, resulting in shoreline accretion. This implies at the outset that the LS is rich enough in mobile sediment that can be progressively reworked to attain, eventually, an equilibrium morphodynamic state (see also Fig. 2). This regime is typical of shallow sediment-rich LS characterized by large-scale bedforms. It generally excludes sediment-starved or bedrock-dominated shelves, which are more likely to prevail on some steep, convergent plate margins (Schellart and Rawlinson, 2010; Harris et al., 2014). There are, however, notable examples of sediment-limited shelves on divergent margins (e.g., Thieler et al., 2001; Dominguez, 2009; Halla et al., 2019; Kirkpatrick and Green, 2018; Menier et al., 2019). Since wave skewness increases with decreasing water depth, negative sediment accommodation potentially constitutes a favourable condition for LS sediment-sourcing, infill of estuaries and tidal basins, as well as the generation of large aeolian dunes, and thick wave-formed beach-ridge plains that may ultimately become preserved in the depositional record. Supply of sediment from the LS is expected to diminish with time (possibly over millennia, see *4.3*) as the LS sediment reservoir is progressively exhausted through incorporation of sand into coastal deposits via the US and beach (Fig. 2). The LS becomes, thus, deeper. A steepening LS also results in decreased wave skewness and increasingly more limited onshore transport, eventually approaching an equilibrium condition where onshore and offshore sediment transport are balanced.

Onshore sediment supply may be reactivated by sea-level fall that results in decreasing LS depths (Fig. 12a) through the process of 'forced regression' (Posamentier et al., 1992). This can potentially activate morphodynamic disequilibrium, such as along the active plate margin of Japan (Tamura et al., 2007). Sea-level fall on shores undergoing tectonic uplift, such as those subject to isostatic rebound, may have LS associated with

'staircase' beach/dune-ridge complexes that show a decreasing trend and elevation (and age) towards present sea level (Fig. 12b). Enclosed areas of rapid historical sea-level change, such as the Caspian Sea, provide an opportunity for monitoring of wave grading of the LS and eventual onshore supply of sand as sea-level falls. Storms and Kroonenberg (2007) documented rapid beach-ridge growth they attributed to fluvial sediment supply on the Azerbaijan shores of the Caspian Sea in a context of rapid sea-level fall (0.8 m between 1995 and 1999), although they recognized, but did not investigate, that there may have been a potential link also with sand supply from the LS.

4.3. Timescales of LS change and graded equilibrium

Although the entire shoreface profile may be 'active', its response to hydrodynamic forcing takes place over different timescales and the adaptation timescale for the LS may, consequently, be quite long (Stive and De Vriend, 1995). The US can respond to changing water levels in days or weeks (even hours for large tsunami and storms) while the LS may take thousands of years (Fig. 1) to approach equilibrium (Cowell and Thom, 1994; Stive and de Vriend, 1995; Wright, 1995; Cowell et al., 1999, 2001). The rate at which the shoreface morphology tends towards equilibrium decreases with increasing water depth because wave energy driving change is depth-dependent due to attenuation of orbital velocities with depth. In addition to depth-dependent wave energy, grain-size and type considerations are also important in terms of sediment mobilization.

Over time, an equilibrium LS (Fig. 10c) may, theoretically, be approached (Thom, 1984), maintained by a stable sea level, and tantamount to cessation of coupled barrier-shoreline progradation. This may be associated with an equilibrium US and beach where the LS serves as the only sediment source for coastal infill and progradation (see section **5**). Eventual LS morphodynamic equilibrium, while being theoretically possible in some settings, may only be achieved where environmental drivers remain stable over variably long timescales (probably > 10^3 years). Key drivers of profile change are, obviously, hardly likely to remain constant over such long timescales. The fact that it takes a long time for the LS to reach equilibrium may imply, thus, that this state may never be achieved on many of the world's LS, notably because this time typically exceeds global-scale sea-level hiatus. The degree of movement towards equilibrium will, of course, also depend on the two other variables: the ocean hydrodynamic climate and the amount of reworkable sediment, which,

on many shelves, may be augmented by active river supply (Patruno and Helland-Hansen, 2018) or carbonate production (Michel et al., 2019). Regional variations in shelf characteristics and accommodation controlled by geology and tectonics (e.g., Kirkpatrick and Green, 2018; Rangel and Dominguez, 2019; Menier et al., 2019) can also influence sediment connectivity between the LS and the beach, as shown by a fine example from Sodwana Bay, South Africa (Fig. 13). Moreover, many shelf environments are shallow and draped with abundant bedforms that may be in long-term equilibrium with their environment and the hydrodynamic forcing (*4.4*). Even where the LS is deemed to have acted as a significant source of sand for barrier progradation over the last 6k years, the morphology and sediment cover can still be quite variable (e.g., Kinsela et al. (2020). In reality, whether equilibrium is really ever attained in nature in settings where this is theoretically possible remains an open question. The issue needs to be considered in the timescale perspective, as stated above. Graded equilibrium in no way systematically signifies that the LS sediment reservoir has been exhausted, but rather a balance between onshore- and offshore-directed sediment transport vectors. Section **5** provides some rare examples of LS now close to, if not at equilibrium.

4.4. LS bedforms

Many LS exhibit a shallow morphology associated with loose, abundant sediment forming a variety of bedforms (Fig. 14). These bedforms, which sometimes directly impinge on the US, carry a range of denominations, with no standardized nomenclature, as noted by Kleinhans (2002). They have elicited considerable attention since the seminal studies of Houbolt (1968) in the North Sea, and of Duane et al. (1972) and Swift et al. (1972) on the northeastern Atlantic shelf of the USA. Among these bedforms are sand banks and ridges (Berné et al., 1994, 1998; Dyer and Huntley, 1999; Reynaud et al., 2003; Burningham and French, 2011; Liu et al., 2007; Latapy et al., 2019, 2020), sand waves (Bao et al., 2020), giant sand banks (Zhou et al., 2020), sorted bedforms (Trembanis and Hume, 2010; Mazières et al., 2015; Simarro et al., 2015; Guererro et al., 2018; Liu et al., 2018), shoals (Denny et al., 2013; Thieler et al., 2014), shoreface-connected ridges (van de Meen and van Rijn, 2000; Houser and Matthew, 2011; Cawthra et al., 2012; Goff et al., 2015; Nnafie et al., 2015), radial ridges (Xu et al., 2015), rippled-scour depressions (Thieler et al., 2001; Bellec et al., 2010; Davis et al., 2013), and hummocky bedforms (Arora et al., 2018). As mapping of the

world's continental shelves advances in the coming years (Wölfl et al., 2019), such large bedforms may well turn out to be ubiquitous LS features. The abundance and diversity of these bedforms reflect wave-bed interactions, especially during storms, and the influence of wind forcing and tidal currents (*3.3*). Some, however, are not, or not exclusively, formed through accumulation/lateral migration under contemporary hydrodynamic processes. Some contain an old core consisting of channel fill, shoreface, estuarine and/or ebb-tidal delta deposits (e.g., Houbolt, 1968; Laban and Schüttenhelm, 1981; Berné et al., 1994; Thieler et al., 2001).

Many of these features have been considered in terms of 'sorted' or self-organised forms implying significant feedback dynamics (e.g., Murray and Thieler, 2004; Diesing et al., 2006; Coco et al., 2007). Thieler et al. (2014) and Liu et al. (2018) have also highlighted a possible relationship between sediment abundance and bedform type on parts of the northeastern U.S. inner continental shelf. Portions of the inner shelf with relatively high sediment abundance are characterized by shoals and shoreface-attached ridges, and where sediment is less abundant, the seafloor is dominated by sorted bedforms, notably rippled-scour depressions.

These bedforms are pertinent to the question of LS equilibrium/disequilibrium and timescales of change since they constitute sand reservoirs that can potentially be mobilized towards the beach in many settings. However, there are several examples where no connection with the shoreline has been found (e.g., Simarro et al., 2015). Still others are probably fed episodically by sand transported offshore from the beach during storms (e.g., Backstrom et al., 2015). Research carried out in recent years has tended to focus on: (1) timescales of bedform mobility, (2) high-energy events and bedform mobility, (3) and the potential for these bedforms to supply sediment to the coast (*section* **5**). Some bedforms have been shown to exhibit mesoscale (decadal) stability (Diesing et al., 2006), and Simarro et al. (2015) have even argued that despite the impact of thousands of Mediterranean storms, the detached sand ridges they monitored off the Spanish coast showed no evidence of migration or morphological degradation, thus suggesting long-term equilibrium with the storm climate. A similar pattern of stability of rippled scour depressions relative to the ambient forcing was observed by Thieler et al. (2001) on the North Carolina (United States) LS. The possibility of some sort of substrate erodibility control cannot be excluded. This could be the case of the Brown Bank in the North Sea between the Netherlands and the UK

(S. van Heteren, pers. com., 2020). Several other studies (e.g., Héquette and Aernouts, 2010; Denny et al., 2013; Schwab et al., 2013; Latapy et al., 2019, 2020) have, however documented bedform mobility notably driven by storms. The impact of Superstorm Sandy (late October 2012) on the mobility of these bedforms on the LS and inner shelf off New York (Fig. 15) has been thoroughly investigated in a series of papers (Trembanis et al., 2013; Goff et al., 2015; Schwab et al., 2017; Arora et al., 2018; Warner et al., 2018). This issue of mobility poses the question of the link between LS bedforms and onshore fluxes of sand to feed the coast, or offshore fluxes, during storms for instance. There is disagreement regarding whether or not large-scale bedforms on the inner shelf off the east coast of the United States are connected to, and serve (or not) as sediment sources for coastal deposits (e.g., Hapke et al., 2010; Kana et al., 2011; Denny et al., 2013). Examples of this potential link are further reviewed in section **5**.

5. Sediment-supply connectivity between the LS and the beach, and implications for coastal accretion

Direct and reliable evidence of sediment supply from the LS (and potentially the inner shelf) to the coast is restricted by scale constraints and a range of uncertainties in measuring and integrating sediment fluxes over time and linking them with the coastal sediment budget. Nevertheless, there have been numerous inferences in the literature on this sediment supply contributing over long timescales (> 10 years) to coastal progradation (Davies, 1980). More than 60 years ago, fascinated by the breadth of sandy beach-ridge development and by the large coastal lagoons that acted as traps for fluvial sediment along the Côte d'Ivoire coastal margin in the Gulf of Guinea (West Africa), Le Bourdiec (1958) proposed that the sand source for these large beach-ridge plains was the inner continental shelf. A similar conclusion was reached by Tricart (1959) for beach-ridge deposits in Brazil. Bird (1963) made the same deduction for similar deposits in southeastern Australia, paving the way for the conceptual models (Fig. 2) by Roy and Thom (1981) on shoreface sediment reworking to supply sand to barriers that have inspired various other studies over the last few years.

With the exception of cases where: (1) sediment supply for coastal progradation can only be accounted for by the LS, or (2) where productive offshore carbonate systems occur (Fig. 16), other lines of evidence for sediment supply from the LS are more tenuous. They

include coastal sediment budgeting, grain-size mineralogy of typically marine sediments and grain-size shape, with roundness often invoked as a criterion of 'marine' sourcing. The linking of progradational histories of the coast with pulses of, or continuous, sediment supply from the LS can be inferred from geological evidence on the inner shelf (Figs. 12, 13), but such studies are not evenly distributed or are not often framed to investigate this link. Examples of coasts with unequivocal and circumstantial evidence of progradation sourced by sand from the LS and inner shelf are described and the relationship to a potential diverse range of disequilibrium situations examined.

5.1. Sediment supply from the LS to the beach: a common situation?

Coastal barrier histories have been linked to changes in LS bathymetry, including large-scale bedform/nearshore bar mobility, and to onshore sediment supply in areas where the LS constitutes the only plausible source of sediment. There are grounds for arguing that sediment supply from the LS and inner shelf have played, and continue to do so, a relevant role in the progradation of many of the world's sandy shorelines. This is the case especially where these shores face sediment-rich inner shelves exposed to long regular ocean swells that are subject to considerable cross-shelf increase in wave shape/orbital velocity skewness that enhances onshore sediment transport. However, mapping of the world's continental shelves has still some way to go (Wölfl et al., 2019), and there is especially a need for identifying sediment-rich and sediment-poor shelves as a first step in characterising eventual sediment connectivity between the world's clastic shorelines and the LS. Carbonate factories are well identified at the global scale (e.g., Laugié et al., 2019; Michel et al., 2019), and among the four types, the Photozoan T (typically coral reefs) and the Heterozoan C (dominated by bivalves), beaten by swell waves or storm waves, appear to be the most productive (Fig. 16a) for beach and dune sand (J. Borgomano and J. Michel, pers. com, June 2020). While carbonate sand may be largely dominant on many LS devoid of terrestrial sediment supply through rivers during previous lowstands, many other coasts are characterized by a mixture of carbonate and quartz sand (Fig. 16b).

One of the most unequivocal areas where the LS has supplied sand for coastal barrier progradation is Australia. Various studies there have referred to onshore sand supply from the LS/inner shelf to the coast under conditions of moderate to high ambient wave energy and rising post-glacial sea level (Thom, 1984; Roy et al., 1994; Cowell et al., 1999; Short and

Woodroffe, 2009; Short, 2007a, 2007b, 2010, 2013, 2019; Harris and Heap, 2014; Kinsela et al., 2016; Oliver et al., 2017a, 2017b, 2019, 2020; Carvalho et al., 2019; Miot da Silva and Hesp, 2019; Sharples et al., 2020). These studies have shown that most of the coastal sediment around southern Australia is shelf-derived, with abundant carbonate content (Fig. 16b) off the southern and southwestern coasts (Boreen et al., 1993; James et al., 1994; James and Bone, 2017; Short, 2013). Quartz sand is abundant on the southeastern coast (Kinsela et al., 2016), deposited on the shelf during low sea-level stands and reworked onshore by the energetic waves, primarily during the Post-Glacial Marine Transgression (PMT), to form widespread onshore deposits. In an alongshore-connected set of sand barriers, Oliver et al. (2020) identified a supply of both quartz-rich and skeletal carbonate sand transported onshore, the latter becoming an important component of the sediment budget for some of the barriers since ~3k years ago. By and large, this rich set of studies on coastal barrier histories has gone apace with limited coring and chronological work on the Tuncurry embayment and LS/inner shelf in southeast Australia (see Roy et al. (1994).

The carbonate-sourced beach-ridge plains of the north coast of the Yucatan Peninsula have been described by Lowery and Rankey (2017). The wide Yucatan shelf constitutes an important Photozoan-C factory from which large bedforms, well identifiable on satellite images, are seen to weld onto the coast (Fig. 17). These bedforms are likely driven onshore by waves and currents generated by trade winds and sea breezes and by stronger winds generated by cold fronts (called Norte) in winter (Appendini et al. 2012; Lowery and Rankey, 2017; Torres-Freyermuth et al., 2017). Currents are stronger near the coast which accounts for the orientation of the ridges. It is likely that once they reach the US and beach, the bedforms feed an important longshore transport system driven by both the modal, lower-energy trade-wind waves from the east and upwelling currents (Appendini et al., 2012), providing sand for the numerous beach-ridge and spit complexes on this coast.

Dunefields are commonly associated with high-energy beach systems in temperate to arid climates around the south, east, and west coasts of the world (Hesp, 2012), and where such dunes are composed of marine carbonate sands, the shelf is clearly the source as has been demonstrated by Short (2013, 2019) and Miot da Silva and Hesp (2019). LS/inner shelf-carbonate production can, in fact, be an unlimited source of sand for coastal dune accretion in some high-energy settings such as the coast of southern Australia (Fig. 16b).

The southern North Sea is one area of shelf-derived quartz-rich sand. This storm-wave shallow shelf environment with micro- to macro-tidal ranges is rich in sand inherited from the last sea-level lowstand (deposited mostly from glacial outwash and by the various tributaries of the main-stem lowstand Rhine River draining towards the Atlantic) and partly reworked into sand banks and ridges. The LS has been deemed as a particularly important source for large-scale mid-to-late-Holocene aeolian-dune development and tidal-basin and estuarine infill on the west-facing English Channel coasts of France (Fig. 18), and north-facing southern North Sea coasts of France and Belgium (Anthony, 2000, 2002, 2013; Anthony et al., 2006, 2010; Héquette and Aernouts, 2010; Mrani Alaoui et al., 2011; Latapy et al., 2019, 2020), the Netherlands (Beets and van der Spek, 2000; Rieu et al., 2005; van Heteren et al., 2011) and along the Danish Wadden Sea (Aagaard et al., 2004, 2007). In several areas of the southern North Sea and the Wadden Sea, there is a well-established link between shoreward migration of sand bodies and/or nearshore bars, their onshore welding, and coastal accretion. Other European coasts where shoreward migration of sand bodies has been recorded are the high-energy Britanny (Menier et al., 2019) and Aquitaine coasts of France (Klingebiel and Gayet, 1995; Tastet and Pontee, 1998) and north and northwest coasts of Ireland (Cooper et al., 2002; Cooper and Navas, 2004). Many of these coasts are characterized by well-developed aeolian dune systems. Mazières et al. (2015) identified slow shoreward movement of dune-like sorted bedforms on the inner shelf off the coast of Aquitaine, where the large aeolian dunes that have accumulated over the last 5K years have been associated with sand sourcing from the shelf (Klingebiel and Gayet, 1995; Tastet and Pontee, 1998). In South America, a number of sand barriers fringing the wave-dominated coasts of eastern and southeastern Brazil have been considered as sourced by shelf-derived sand (Angulo et al., 2009; Dominguez, 2009; Dillenburg et al., 2009, 2012; Hein et al., 2013). Large areas of the Brazilian shelf east of the Amazon and along the central coast are, however, sediment-deficient (Dominguez, 2009; Halla et al., 2019). In earlier work on the coast of Brazil, Martin et al. (1987) and Dominguez (1996) suggested accumulation of what they considered as inner shelf-derived sand on the updrift flanks of four river deltas (Jequitinona, Doce, Sao Francisco, Paranaibo do Sul), whereas downdrift shoreline growth was deemed to have been generated by sediment supplied directly by these rivers. Supply through delta mouths, and alongshore distribution of sediment, do not preclude, however, sediment supply from the LS benefiting the same coast (Anthony, 2015). On the west coast

of the United States, Cooper (1958) suggested that the extensive dune sheets of Oregon and Washington were sourced from the shelf, and more recent work by Kaminsky et al. (2010), Ruggiero et al. (2010) and Peterson et al. (2020) also indicates a LS source for some of the beaches, sand ramps and cliff-top dunes of the Pacific coast.

Elsewhere, while linkages may seem plausible, the evidence is more tenuous, masked by the proximity of active fluvial sources of sand, as in West Africa (Anthony, 1995). This tenuous link is also the case in LS environments exhibiting variably mobile large-scale bedforms, such as off the coast of the eastern and northeastern United States. Unlike the rich coastal barrier histories alongside the relatively poorly known LS environments of southern Australia, the LS environments of the eastern and northeastern United States have been investigated extensively with regards to their geology, notably through numerous seismic studies (see references in *4.4*). Large portions of the LS off this coast are, however, probably sediment-deficient, characterized by low fluvial sediment input during the Late Pleistocene sea-level lowstand, are devoid of carbonate factories, and still subject to sea-level rise due to isostatic adjustment. The retreating barrier island environments thus formed are not associated with the same large sediment storage as many of the multiple beach-ridge and dune barrier coasts evoked above. Denny et al. (2013) postulated that although the processes mobilizing sediment across the LS of the sediment-limited environment of the Grand Strand coast of South Carolina, United States, are not well known, the thick relict Holocene deposits forming shoal complexes (Fig. 14) composed of moderately sorted fine sand are likely contributing significantly to the beach sediment budget. At Fire Island, New York, Kana et al. (2011) found no evidence for onshore sand transport from the LS to the coast, in contrast to Hapke et al. (2010) and Schwab et al. (2013) who hypothesized that modern marine sands are being advected from the LS and inner shelf sand ridges and from eroded Pleistocene sediments to feed the coast. In the same area, Goff et al. (2015) found that the primary impact of Superstorm Sandy on the LS was to force migration of major bedforms (sand ridges and sorted bedforms) tens of metres WSW alongshore, with migration distance decreasing with increasing water depth. Although greater in rate, this migratory behaviour was no different from that observed by the authors over the 15-year span prior to Sandy. Interestingly, Goff et al. (2015) found, in contrast to Schwab et al. (2013), that migration of the largest bedforms (sand ridges and sorted bedforms) did not involve landward advection, since the observed migration direction was

alongshore to offshore, consistent with the direction taken by Sandy and by northeasterly storms (locally termed nor'easters). On the other hand, sou'westers tended to reverse the dominant current direction, thus potentially enabling sediment transport from the sand ridges to the coast. A similar pattern was noted by Hill et al. (2004) in the course of storms on the coast of Maine. These examples illustrate the variability of potential long-term sediment transport directions on the LS and the stochastic nature of shoreface sediment sourcing of the coast. Establishing firm, unequivocal links between long-term onshore sediment flux from the LS and inner shelf to the coast is hampered by time and space constraints that impose the need for datasets with scope and quality that are currently out of reach for most LS settings. The problem is even more complex due to the open nature of many LS to various directions of potential forcing, including alongshore (Fig. 5).

Abandoned river delta lobes, and to a lesser extent ebb delta lobes, generate, at timescales of 10-10^2 yrs, variability in LS morphodynamics and can, thus, be local to regional sources of disequilibrium. Abandoned delta lobes may be considered as being somewhat akin to fixed mega-bedforms, or to Pleistocene sediment stores being reworked as they are submerged by sea-level rise. River deltas are receiving increasing attention worldwide because these accretionary forms, which commonly tend to prograde onto the LS, are increasingly undergoing erosion under dwindling fluvial sediment supplies (e.g., Besset et al., 2019; Dunn et al., 2019). Many protruding, shallow river-delta fronts and ebb deltas may, thus, represent extensive erodible LS sediment reservoirs. The Old Huanghe (Yellow River) delta in China has, for instance, undergone severe erosion following a change in the location of its river mouth from the Yellow Sea to the Bohai Sea in 1855, and this erosion has since become the main source of sediment to the Yellow Sea and East China Sea (Zhou et al., 2014). These authors have shown that > 790 Mt of sediment a year have been reworked for the last 100 years from depths of < 20 m, with about 25% of this sediment volume deposited offshore in the western Yellow Sea. Delta lobes are commonly more exposed to waves compared to adjacent areas (e.g., Zhou et al., 2014; Sabatier and Anthony, 2015; Patterson and Nielsen, 2016; Ruggiero et al., 2016). Patterson and Nielsen (2016) analyzed, for instance, 46 years of changes in LS bathymetry (in 10–20 m depths) over an abandoned delta lobe at the northern Gold Coast, Australia (Fig. 19). In this example, the sand eroded from the ebb-delta lobe and transported northward has contributed to the massive longshore transport volume (550,000 m^3/year) by an extra 80,000 m^3/year along a 5 km stretch of the

coast. On some coasts, this longshore transport builds up spits (e.g., Penland et al., 1988), notably due to lobe perturbation of the ambient wave field (Anthony, 2015; Anthony et al., 2016).

5.2. Sediment supply from the beach to the LS

As shown in *3.3,* sediment connectivity between the LS and the beach commonly involves more complex trajectories than that of episodic or steady wave-driven onshore sediment supply. Return of sediment from the beach to the LS may be a common occurrence (see for example Fig. 20). The role of storm-induced (mega)rip currents has been highlighted, and significant sediment transport from the beach to the LS has been reported from embayed, high-energy beaches (e.g., Loureiro et al., 2012; McCarroll et al., 2015). On many shorefaces, however, offshore sediment transport is part of the background morphodynamic regime. Thieler et al. (2001) reported how sediment accumulation from over 30 years of extensive beach nourishment at Wrightsville Beach, North Carlina, appeared to have exceeded the local US-beach accommodation, resulting in the 'leaking' of nourishment sand to the lower shoreface.

Longshore sediment transport along the US can integrate long-term cross-shore sediment supply from the LS but also contribute to the return of sediment to the latter. Although the US may be striving towards equilibrium, slopes may become locally too steep due to large inputs from longshore sources, undertow increases sand transport offshore, feeding the LS. Fruergaard et al. (2015) have shown, in their description of the evolution of the Danish Wadden Sea barriers, that progradation was episodic, probably punctuated by supply to the LS from updrift sources discussed in Aagaard (2011). Aagaard and Kroon (2007) and Aagaard et al. (2010) documented both seaward flushing of sand from the US to the LS at Vejers on the Danish North Sea coast through net offshore bar migration (NOM; e.g., Wijnberg and Terwindt, 1995; Walstra et al., 2012), which involves nearshore bars moving offshore on interannual timescales and decaying at the boundary between the LS and the US. On the other hand, transfer of sediment from the LS to the US occurred through net onshore bar migration at Skallingen, located 25 km further south on the other side of the cuspate foreland of Blaavands Huk (Aagaard et al., 2004). This difference is associated with steeper and bigger bars at Vejers and strong seaward undertows (since undertow speed scales with seabed gradient; Aagaard et al., 2002), whereas bars at Skallingen are low and

gently sloping with weak undertows. The difference in bar size and steepness is probably due to alongshore gradients in the longshore drift; Vejers has a large supply of sand from updrift sources (negative longshore transport gradients) while the opposite prevails at Skallingen which has a large positive alongshore transport gradient resulting in downdrift sediment loss (Aagaard and Sorensen, 2013). At Vejers, the residual longshore current driven by prevailing winds transports sand north on the LS, as opposed to the wave-driven alongshore southerly transport on the US. The northward transport links up with the large-scale sand ridges on the LS further up the west coast of Jutland that have been described by several authors (Kuijpers et al., 1993; Anthony and Leth, 2002), producing a spatial longshore transport pattern (Fig. 20a) in analogy with the 'step-ladder' transport concept described by Black et al (2008) (see *3.3*).

Spits associated with river deltas, commonly formed following lobe changes (*5.1*), can also serve as pathways for offshore sediment transport to the LS (Fig. 20b,c). In both the Rhône (Sabatier et al., 2006) and the Volta deltas (Anthony and Blivi, 1999; Anthony et al., 2016), sand transported along the seaward flanks and distal tips of large sand spits formed over the last century, and partially derived from abandoned delta lobes, is being actively redeposited on the LS. In cases where progradation has caused shoreline irregularities, such as salients, there is also a potential for alongshore funneling of sand caused by large-scale gradients in sediment transport.

5.3. LS sediment supply, patterns and rates of coastal progradation

Sediment supplied onshore from the LS has been deposited over the last 6-7k years in tidal basins and estuaries, and as beach ridges and dunes. This accretion has been modulated by rates of sediment supply, available accommodation, and storage over the shoreface tract, including: (1) changing temporal patterns of demand for sediment or, (2) alternatively, sediment-supply connectivity between estuarine/tidal basin and barrier systems, and (3) eventual headland bypassing and alongshore sediment connectivity between barriers. Additionally, increasing area and depth of the LS have entailed changes in hydrodynamic forcing, notably wave power, but also in some areas, change from tide- to wave-dominated sediment transport (e.g., van Heteren et al., 2011), including increased wave-driven alongshore diversion or closure of the inlets of tidal basins and estuary mouths by barriers and spits. At a finer scale, a range of local to regional beach-US and dune

conditions, such as swash dynamics, wind direction and aeolian fetch, and vegetation, have driven significant morphological variability.

It has been shown that the initial stages of coastal development in many areas involved demand for sediment between, on the one hand, estuaries and tidal basins, and, on the other, coastal barriers, and this influenced barrier progradation rates, as shown by studies in the Netherlands (e.g., Beets et al., 2000; Rieu et al., 2005; van Hetern et al., 2011), Australia (e.g., Roy et al., 1980; Sloss et al., 2006; Oliver et al., 2019), and Brazil (Angulo et al., 2009). Barriers did not start to prograde, or exhibited low progradation rates, until neighbouring tidal basins and estuaries, which are generally efficient sediment traps, had been largely infilled. Exceptions to this occurred where swell-driven high longshore sand transport sealed off such infilling tidal basins at an early stage, as on the strongly drift-aligned Bight of Benin coast, West Africa, where this occurred probably as early as 3-4k years ago (Anthony et al., 2002). According to Oliver et al. (2019), the progradation rate of Pedro Beach, Australia, sourced by sand supply from the LS was low (0.30 m/yr) between 7k years and 5.8k years ago, probably because a large volume of sediment, which might have resulted in rapid shoreline progradation (as occurred later in the barrier's history), was trapped in a large flood-tide delta during this time (Fig. 21a). Barrier progradation rate was rapid between 5.8k and 5.2k years ago (1.2 m/yr), presumably following significant infill of the Congo Creek tidal basin, and then slowed (0.38 m/yr) between 5.2k and 3.9k years ago (Fig. 21b), probably as sand supply from the LS diminished (Oliver et al., 2019).

In strongly embayed beach-ridge settings where there is neither alongshore input nor leakage of sand (but where longshore transport and beach rotation may operate) and aeolian transport is weak, the progradation rate should decrease as the barrier US impinges on deeper water. In siliciclastic settings where LS sand supply cannot be renewed (as distinct from settings with LS carbonate factories), this implies LS deepening and steepening and concurrent transmission of higher wave energy onto the US. If we put the limit between the US and the LS at the seaward toe of the outer bar (Fig. 1), at some stage, the surf zone can no longer widen as the LS deepens. This occurs because shoreline and US stationarity is forced by the long-term landward translation of the DoC contour. Landward translation of the DoC contour is associated with a steeper US characterised by stronger undertow and balanced on/offshore sediment transport rates (*3.2*).

Barrier progradation may eventually cease altogether at some stage, even where there is a longshore sediment supply. Sand could be actively accommodated through onshore aeolian transport with little or no progradation, as Bristow and Pucillo (2006) and Oliver et al. (2019) have shown for embayed barriers in southern Australia. Bristow and Pucillo (2006) identified changes in beach-ridge morpho-stratigraphy that expressed the increasing influence of beach encroachment over a deeper more energetic shoreface. Following relative cessation of LS sand supply nearly 4k years ago, barrier progradation in parts of southeast Australia has been shown to have resumed in places, but at much reduced rates, as a result of more recent onset of fluvial-sourcing of longshore transport (Kinsela et al., 2016; Oliver et al., 2019). Kinsela et al. (2016) simulated a LS sand supply in the order of ~10^4 m^3/m (81%) for the Tuncurry barrier (Fig. 2b), and diminishing from 3k years ago, supplemented since then by alongshore sand supply (19%). Carvalho et al. (2019) identified a similar scenario for the Shoalhaven sand barrier, also in southeast Australia. Delivery of sand from the LS following the PMT to drive sustained progradation was subsequently complemented by fluvial sands discharged to the coast through the progressively infilled estuary of the Shoalhaven River. This is also a typical situation with delta-sourced sand barriers in the siliciclastic-dominated setting of West Africa, which have been interpreted as benefiting from both relict LS sands and fluvial sands transported alongshore (Anthony, 1995). In several of these examples, alongshore connectivity of sediment transport involving headland bypassing has been achieved over time. Time frames of alongshore sediment connectivity in the progradation of five sand barriers have been identified by Oliver et al. (submitted) along a 26 km-long sector of coast SE Australia. The adjustment to changing sediment supply from the shoreface is expressed by variations in progradational rates, by the morphology and spacing of individual beach ridges, and by variations in the amount of quartz relative to carbonate sands.

These long-term changes in sediment transfer from the LS to the beach also have implications for both dune and beach morphodynamics. Psuty (1994) demonstrated that dune size increased with shoreline stability, i.e., small dunes for rapidly prograding coasts, and large dunes for stable shorelines, and even large transgressive dunes that build up at the expense of coastal erosion, as in the Netherlands (van Heteren et al., 2011). Progressively decreasing rates of shoreline advance would, therefore, allow for progressively increased duration of dune-building episodes and hence progressively wider and higher dunes. This

may explain, for instance, the high outer foredunes that bound some sandy barriers, such as Pedro Beach, in southeast Australia (Fig. 21b). Oliver et al. (2019) attributed the high outer foredune of Pedro Beach to increased sand input resulting from human interventions on catchments. This foredune morphology may, however, also well reflect in-situ US-beach morphological adjustment to longshore sediment inputs once the degree of plan-form embayment curvature following progradation had allowed for sand bypassing across the headlands bounding Pedro Beach.

Over the long-term, modal beach morphodynamic states (Wright and Short, 1984) on coasts sourced in sediment by a deepening and widening LS adapt to increased incident wave energy. They do so by having an US and beach sand volume adequate to dissipate wave energy and maintain equilibrium, i.e., involving minimal net transport offshore (to the LS) or onshore (to the US). Beaches exposed to high waves tend to be of the dissipative type with mild slopes and subdued bars, and are typically composed of fine sand (Short 2007b; Splinter et al., 2014). On such gentler-sloping beaches, waves break farther offshore, resulting in wide barred surf zones that effectively dissipate wave energy, undertow is weak, and these conditions result in less wave energy available to move sand onshore/offshore, thus maintaining stability (Aagaard et al., 2014).

5.4. Fine sand from the LS, dissipative beaches, and large aeolian dunes

In considering sediment connectivity between the LS and the US, the issue of grain size and how it affects and feeds back on the morphology, merits attention. In the context of onshore-offshore sand exchange across an equilibrium shoreface, net offshore transport of the fine sand fraction appears to be the norm, in conformity with the traditionally assumed relationship between grain size and (wave) energy/flow speed, because sand size often decreases offshore into deeper water. There is a link, however, between dissipative beaches rich in fine sand and large active dunes (Short and Hesp, 1982; Hesp and Smyth, 2016). This highlights a real conceptual problem. Given the afore-mentioned common perception that fine sand is swept offshore, this relationship poses the question regarding the source of all the fine sand contained in dunes.

Many moderate-high energy dissipative US-beach and associated aeolian dune systems can be permanently sourced from offshore carbonate factories, as mentioned in *5.1*. Regarding fine quartz sand, its abundance on the world's LS reflects: (1) source

catchment weathering processes and the characteristic downstream fining in fluvial systems, and down to the shelf during sea-level lowstands (Woodroffe, 2003; Anthony, 2014), and (2) long-term hydrodynamic sorting in bedforms (*4.4*), with preferential onshore transport and accumulation of fine sand. The LS sediment reservoir thus reflects both geological heritage and subsequent hydrodynamic sorting, and these conditions influence the grain-size composition of the US, beach and dunes. Anthony and Héquette (2007) showed from an analysis of 665 samples collected from the LS to the dunes in the shallow storm wave-dominated macrotidal setting of the southern North Sea (Fig. 10), how progressive hydrodynamic sorting led to a mixed, coarse sand- and gravel-dominated LS cover and an increasingly medium-to-fine sand-dominated sediment type on the beaches, with aeolian sorting further leading to fine sand accumulation in the dunes. We may assume that onshore sand transport from many of the world's LS potentially involves a range of fractions from coarse to fine because of common exceedance of threshold velocities. The equilibrium shoreface model of Aagaard and Hughes (2017) actually predicts onshore movement of suspended fine sand (wave-driven onshore sediment transport rates on the LS being inversely proportional to sand grain size), although the model does not consider bedload. In cases where sand supply is assured, a beach and dune system probably strives towards equilibrium not only by maintaining a shallow barred dissipative system (commonly rich in fine sand) that reduces slope and offshore transport through undertow, but also by wind recycling of fine sand to the dunes. Under cessation of shoreward sediment supply, and in the absence of longshore sand supply, the maintenance of a low sloping dissipative US rich in fine sand and incorporating a foredune may also be an inherent feedback condition under increased incident wave energy density in the surf zone resulting from a deepening LS. Once initiated, transgressive dunes can be effective in sequestering sand, notably through landward transfer, even in a shoreline erosional regime (e.g., Tamura et al., 2019).

To summarize, bringing into consideration the LS suggests that onshore or offshore transport of fine sand from beaches and the mechanisms of grain-size segregation may not be as simple as embodied in current thought. This consideration brings out ideas that merit further study. The LS may be deemed as an eroding substrate and/or as one of carbonate production, both processes yielding sand grain sizes that ultimately determine those characterizing the US and beach. In addition to a geological shoreface control, LS

hydrodynamic sorting can lead to bedforms in which fine sand is concentrated and potentially mobilized onshore.

6. Future perspectives

We have sought to synthesize various aspects of the morphodynamics of the LS and its sediment connectivity with the US and beach using as much as possible a multi-problem approach that has attempted to integrate contemporary morphology and recent evolution, hydrodynamics, sediment transport, and grain size. LS morphology and dynamics can be complex and extremely variable, depending on environmental context, antecedent conditions, thresholds, and morphodynamic feedback. LS are diverse, ranging from bedrock-controlled, through sediment-starved to sediment-rich, siliciclastic, carbonate, low to high wave-energy, microtidal to macrotidal, and exhibit variable sediment connectivity with the US and beach, acting as sources but also sinks for coastal sediments. All of these conditions provide a basis for eventual LS multi-criteria classification, but render attempts at setting up broadly applicable concepts, such as depth of closure based on wave parameter inputs, and models, such as the equilibrium profile, controversial. Notwithstanding this diversity and complexity, progress is being achieved, on the one hand, albeit unevenly, on morpho-sedimentary and bedform characterization aided by rapidly evolving seabed-mapping techniques, including through remote sensing. On the other hand, our understanding of near-bed hydrodynamics and sediment transport on the LS is hampered by the lack of adequate instrumentation, and the difficulties and risks of deploying increasingly sophisticated (and often expensive) but sensitive instruments in relatively deep waters. Limited substantive progress has been achieved in understanding near-bed suspended sediment transport and its translation into LS morphological change since the earlier reviews of Wright (1995), Cowell et al. (1999) and Kleinhans (2006). It is known that significant transport is restricted to the lower part of the water column near the bed (Wright et al., 1991; Ruessink et al., 1998; Lacy et al., 2005; Ferré et al., 2010; Aagaard et al., 2010; Aagaard, 2014). However, high-resolution measurements are faced with the challenge of integrating a knowledge of the overall seabed configuration and bed-level changes under a range of hydrodynamic conditions, including water-density conditions, that also need to be known at the outset. These problems are compounded by the inherent difficulties in actually monitoring processes, sediment transport rates and morphological change in high

resolution, including pertinent deployment strategy in space and over time (at sites or along transects/gradients, 24 hr or in tidal cycles, before, during and after extreme events). In fact, the inherent risks and difficulties may even be considered as a limitation to funding, since the successful outcome of experiments especially aimed at high-resolution monitoring of short-term events over a reasonable time frame is commonly not guaranteed.

The unsurmountable problem of bedload measurement in the field, still largely insoluble even in more accessible fluvial and coastal environments, comes out as a fundamental constraint. While datasets on bedload are scarce, there is a strong possibility that the LS is characterized by significant bedload movement, either through individual grain movements, or more likely through bedform mobility. Under the difficult, and sometimes extreme, conditions on the LS, non-quantitative observations are no doubt needed as they may provide guidelines for subsequent quantitative work.

While there may have been too much, but nevertheless justified, emphasis on wave-graded LS, given the ubiquity of waves over the world's shelves, it has become increasingly clear that processes hitherto investigated mainly from hydro-ecological or pollution perspectives, such as upwelling and downwelling, as well as temperature and density currents, can also play a complementary role in sediment transport. In this regard, the non-linear interactions among processes driving sediment transport can render calculations of transport directions and magnitudes uncertain. One important obstacle in understanding how the LS evolves, especially under non-storm conditions, is that of the large spatial and temporal scales of transport, which significantly exceed those of the US, but for which transport rates or morphological change may be much smaller, posing a problem of adequate resolution, but also one of eventual upscaling. Insight into this small scale of change is important as far as minimizing errors in upscaling are concerned. The upscaling issue needs to be addressed extensively, emphasizing the importance of the preserved sedimentary record of not just the lower but also the US. Measurement representativeness is also an issue that should be borne in mind. How representative are measurements at a given point in a LS environment that may comprise uneven bedform development (a form of patchiness and heterogeneity)? Hence the importance of developments in optimizing sea-bed mapping (Diesing et al., 2016; Sagawa et al., 2019). There has been significant progress in the mapping and monitoring of LS bathymetry and of large bedforms in numerous continental shelves of the world over the last decade, with progressive gain in the overall

mapped area (Wölfl et al., 2019) as well as increasing higher-resolution mapping. However, understanding of the processes and conditions under which bedform mobility occurs, and the rates of such mobility, are still far from certain. Now routinely identified in many studies, these bedforms draw attention to aspects of LS equilibrium and disequilibrium, as some seem to be mobilized shoreward (e.g., Schwab et al., 2013), while others are not (e.g., Simarro et al., 2015). Seabed mapping needs to be conducted and analysed in terms of the variable external conditions that may have acted as drivers. Areas of improvement include the determination of uncertainty, better field validation of acoustics, and the use of external drift in interpolation of data points. There is a need, also, for classification of the wide range of LS bedforms and for determining to which extent, and the conditions under which, bedforms are mobile, and their eventual connectivity with the shoreline. Long-term field measurements on several different sites are needed as these enhance a finer understanding and reduce measurement errors, together with the subsequent integration of data for a comprehensive analysis. But can we also use legacy data?

Modelling offers an increasingly explored approach for attempting to understand LS morphodynamics as it is deemed capable of overcoming spatial and temporal limitations. But appropriate modeling results are determined not only by good model skill but also need sufficient observational and in-situ data for calibration and validation, which brings us back to the point of departure, characterized by the limited availability of datasets, especially on bed morphology and near-bed transport. Devoid of proper data for calibration, model output remains purely speculative. The stochastic nature and range of processes could imply that LS modelling is not just limited by data. Despite numerous laboratory experiments and observations of surf-zone and beach processes, we are still far from achieving reliable long-term modelling of beach morphodynamics. Characterizing the changing LS is difficult enough, so how will we ever be able to explain and predict its behaviour? To make progress, exploratory simulations from morphological-behaviour models need to integrate sensitivity analyses using assumed scenarios of transport and/or bed-level change (even at a rate of 0.1 mm per year) to determine to which extent model runs do not get out of control. Geological observations (thickness of the active layer, depth to the top of the non-reworked (Pleistocene) surface) provide important boundary conditions that need to be incorporated into these models; importantly, models that include the LS should be run at least for millennial, not decadal, change.

Integrating the morphodynamics of the LS with that of the US also highlights a number of conceptual problems, one of which is the common link between fine sand, dissipative beaches and large aeolian dunes. This link poses the question of fine sand supply from the LS and its retention on the US, given the common perception that fine sand is readily swept offshore on beaches. Dissipative beaches display mild slopes, and waves break further offshore, resulting in wide barred surf zones that effectively dissipate wave energy, render undertow weak, and create conditions where less wave energy is available to move sand onshore/offshore. In such systems, adaptation to increasing incident wave energy density with LS deepening may involve not only this barred morphology rich in fine sand but also preferential aeolian recycling of fine sand onto dunes, rather than offshore. Many large dune systems can also be sourced infinitely in fine sand from offshore carbonate factories. Another question concerns the effect of increased storminess and sea-level rise. Will this result in a shift of the DoC and the DoT seaward, thus adding a sediment source to the system that had gotten out of reach?

To summarize these perspectives, progress is needed in at least four areas in order to improve our insight into the morphodynamics of the LS and its sediment connectivity with the beach: (1) better and adapted instrumentation and strategies for high-resolution monitoring of near-bed hydrodynamics and sediment transport, (2) validated datasets with quantified uncertainty and linked to driver-monitoring on sediment transport and bathymetric change, (3) the feasibility of upscaling of data in the short- and long-term with minimization of measurement errors, and (4) reliable modelling. Tackling these challenges may require that our science community comes together and develop initiatives that can be carried out at key study sites, aided by firm national government commitments in terms of funding and logistical support.

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FIGURES AND FIGURE CAPTIONS

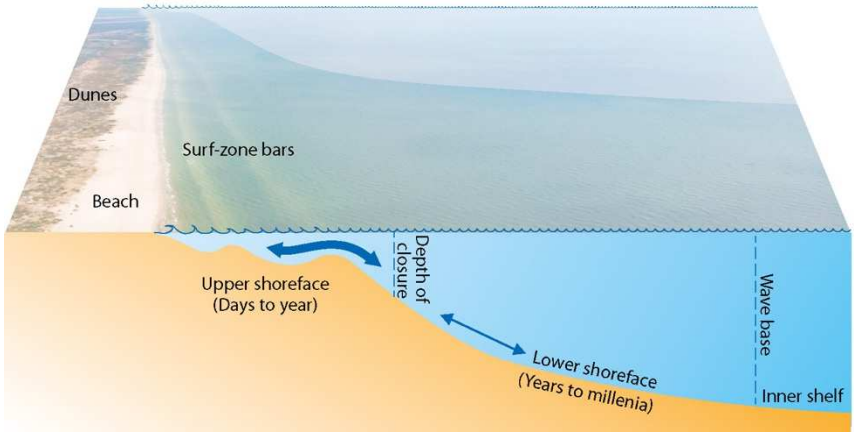


Figure 1. The interlinked beach and shoreface system, conceptual shoreface boundaries, and relative timescales of morphological change. Arrows represent relative magnitudes of sediment transport rates.

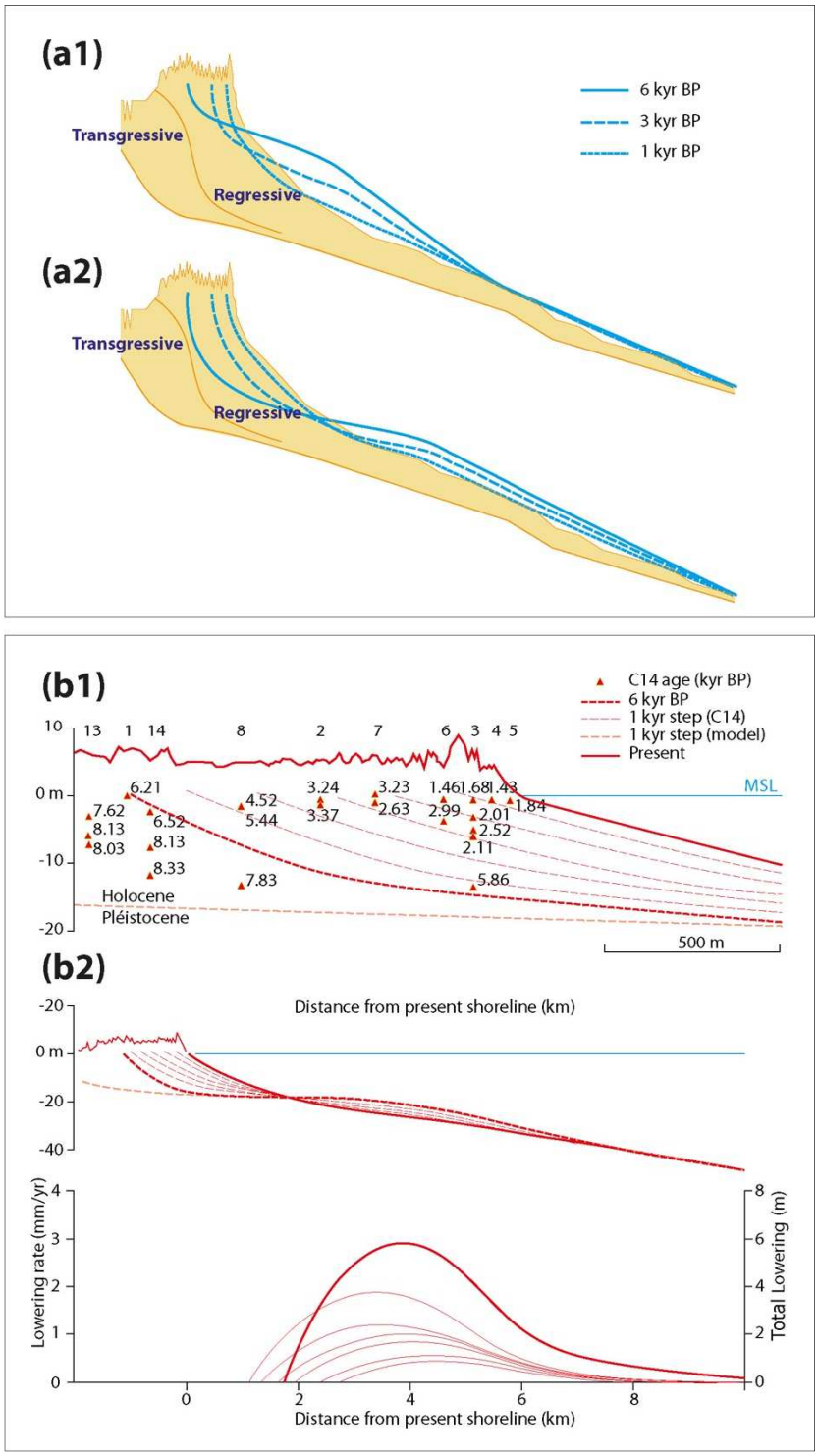


Figure 2. (a) Conceptual models proposed by Roy and Thom (1981) to explain long-term barrier progradation (regressive barriers) in response to lowering of the shoreface following the mid-Holocene sea-level stillstand: (a1) the 'sediment bulge' hypothesis, in which a barrier is supplied with sand from a convex sand body situated on the upper to middle shoreface; (a2) the 'uniform shelf lowering' hypothesis, in which sand is supplied to the barrier by near-uniform planing of the inner-shelf surface; (b) simulated strandplain growth at Tuncurry tuned to the radiocarbon chronology for stable sea-level simulations through time: (b1) 1-kyr time step shoreface surfaces interpreted from C14 chronology; (b2) simulated underlying shoreface sand supply driven by disequilibrium morphology (81%) with

a secondary external sand supply from the alongshore transport system (19%), the former, of the order of ~10^4 m^3/m, diminishing from 3 kyr B.P. onwards, to be supplemented by the latter. Adapted from Kinsela et al. (2016), with permission from Elsevier.

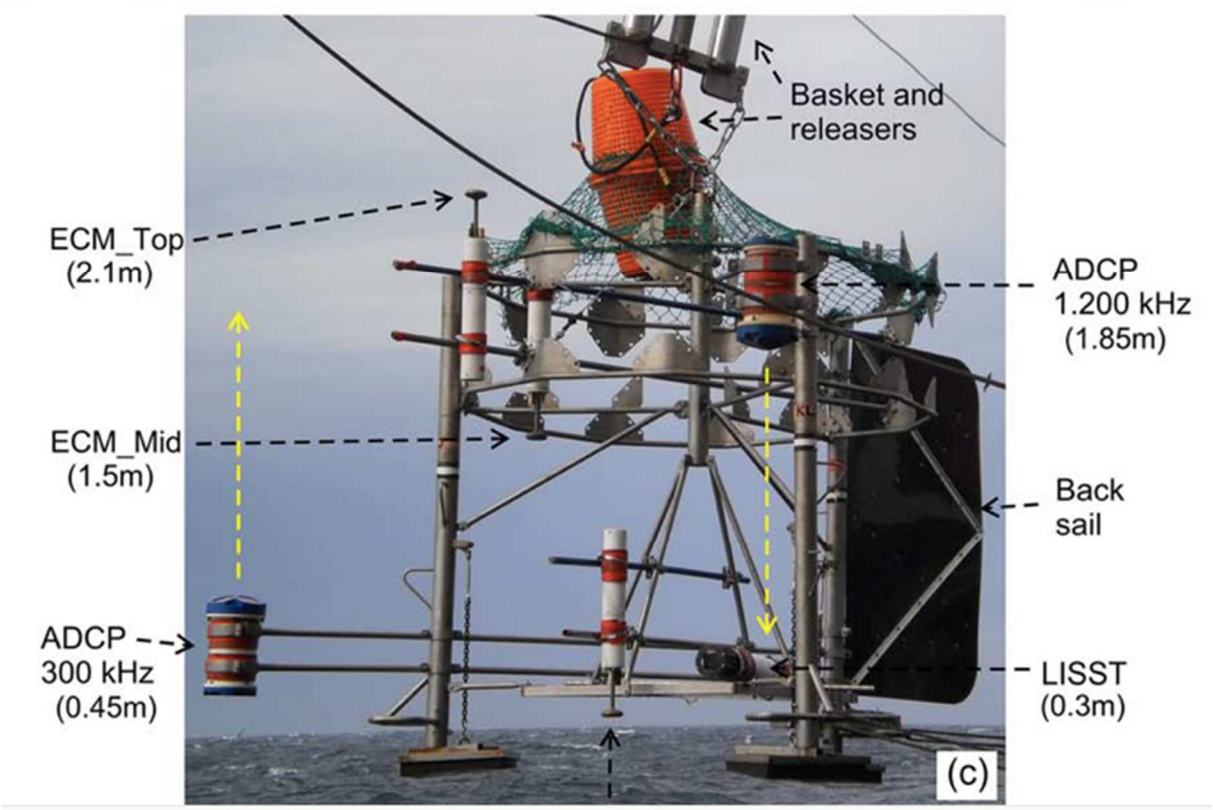


Figure 3. Photograph of a seabed lander and its instrumentation. Distance of each instrument from the bottom of the lander is also indicated. From Zhang et al. (2015), with permission from John Wiley and Sons.

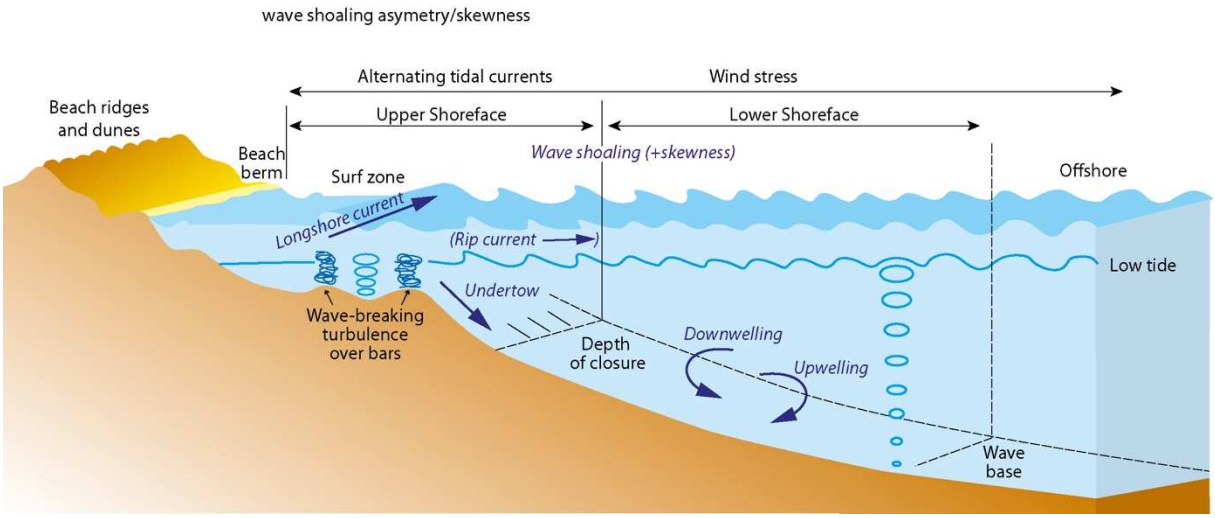


Figure 4. Schematic of waves and currents on the US and LS marked by offshore diminution in the intensity of near-bottom flows, especially seaward of the surf zone, and the

dominance of wave skewness, gravity flows, and upwelling/downwelling currents on the LS. Block diagram inspired from an original illustration by Cowell et al. (1999).

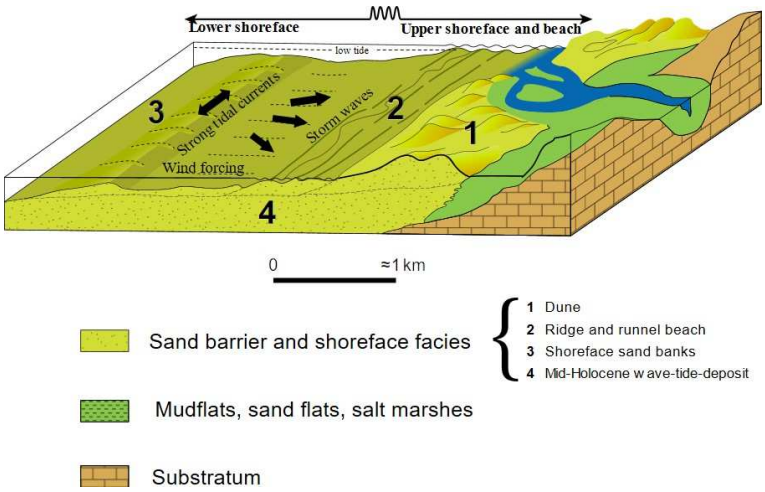


Figure 5. Schematic of process domain zonation and modulation on a shallow, sand-rich mixed storm- (wind and wave activity) and tide-dominated macrotidal shoreface in the southern North Sea, where identification of a DoC based on the classic transition from the US to the LS is not feasible. The LS is characterized by shore-parallel sand banks subject to strong alongshore tidal currents (flood-tide currents can be reinforced by the dominant southwesterly winds). Storm waves, especially coinciding with neap tides, can drive these sand banks onshore, connecting them as oblique features (not shown here) onto the US and wide adjoining bar-trough beaches, creating an intertidal reservoir of sand that can be mobilised by winds to source aeolian dune accretion.

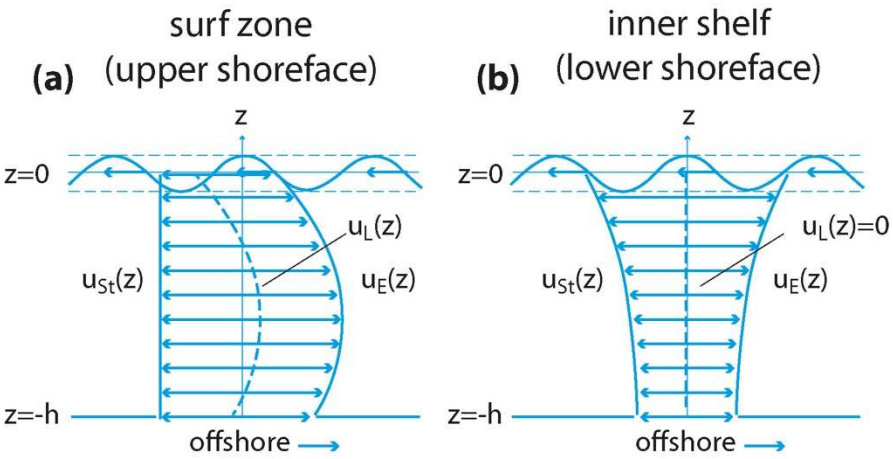


Figure 6. Two-dimensional (x, z) wave-driven circulation typically observed (a) inside the surf zone (US) where the onshore Lagrangian Stokes drift uSt(z) is uniform in depth and the offshore wave-driven Eulerian undertow uE(z) is parabolic, resulting in a vertical imbalance in the net mean flow, uL(z) 6.0 (dashed line); (b) over the inner shelf (LS), where the onshore Lagrangian Stokes drift uSt(z) and the offshore wave-driven Eulerian undertow uE(z) cancel,

resulting in a vertical balance and zero net mean flow (uL(z)) 5 0 (dashed line) at all depths. The black wavy line represents the instantaneous sea surface. From Brown et al. (2015), © American Meteorological Society. Used with permission.

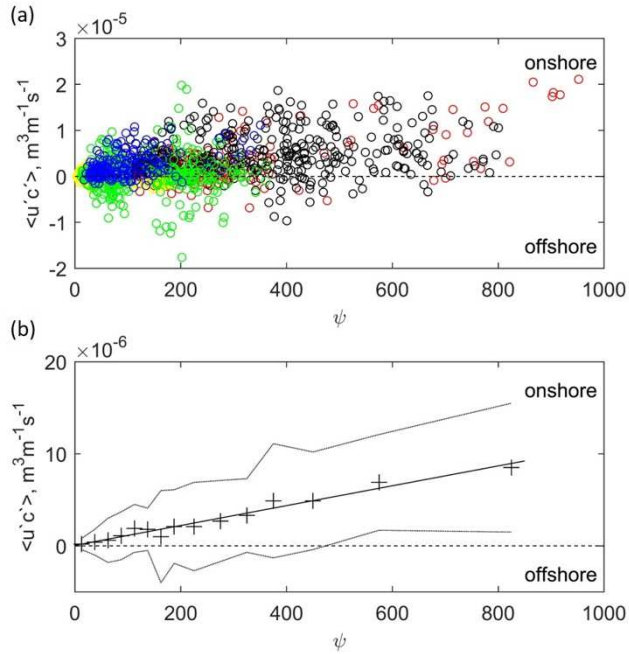


Figure 7. Wave-driven (oscillatory) transport (*u'c'*) as a function of the mobility number, ψ. Panel (a) shows measured transport rates on the LS at Vejers Beach, Denmark (red symbols), Fanoe Beach, Denmark (black symbols), Pearl Beach, NSW, Australia (green symbols), Staengehus Beach, Denmark (blue symbols) and Skallingen, Denmark (yellow symbols). Panel (b) shows the average transport rates within separate ψ-bins. The best-fit relationship (solid line) has an R^2 = 0.968 and a root-mean-square error of 1.651 10^{-6}. The dotted lines indicate ± 1 standard deviation about the means. From Aagaard (2014), with permission from John Wiley and Sons.

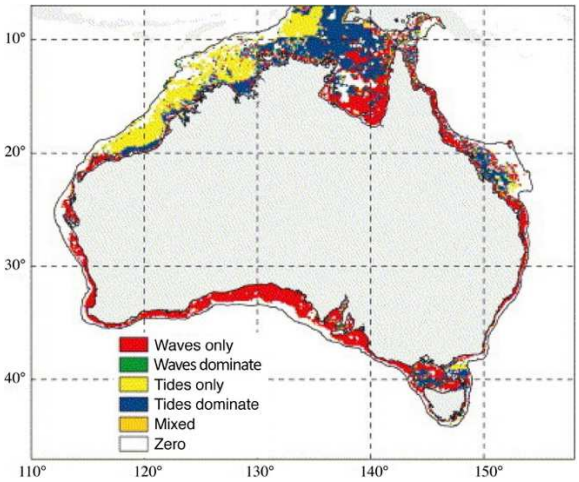


Figure 8. Regionalization of the Australian continental shelf based on estimated exceedance of the velocity thresholds for sediment mobilization caused by waves and tides. Adapted from Porter-Smith et al. (2004), with permission from Elsevier.

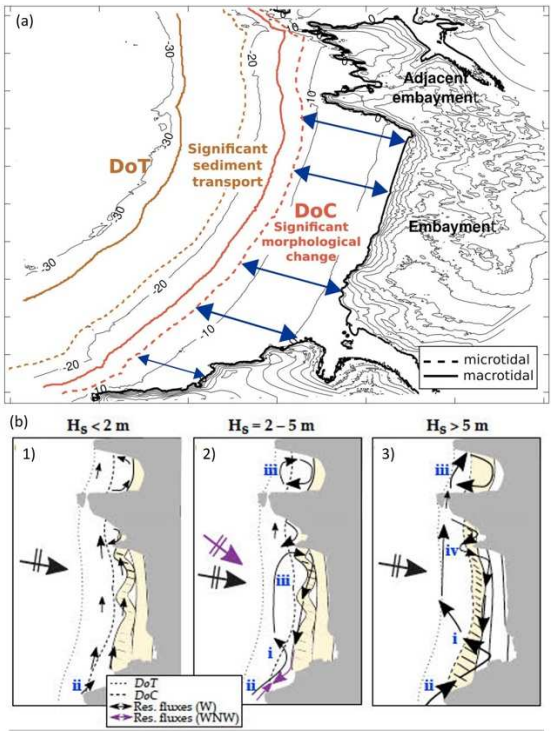


Figure 9. Plan-view showing the DoC, the DoT, and zone of significant morphological change on a 15-km stretch of the embayed high-energy macrotidal coast of north Cornwall, SW England (a); conceptual diagram showing the variability of major alongshore sediment flux pathways (b) under conditions of: 1) low waves, 2) moderate-high waves and, 3) higher than average waves. Arrows (size increases with increasing magnitude) indicate predicted residual fluxes based on Delft 3D model output. Accretion due to cross-shore fluxes is shown in beige. Dotted and dashed lines show, respectively, idealized limits of the active shoreface Depth of Transport (DoT) and Depth of Closure (DoC). (a) adapted from Valiente et al. (2019), and (b) adapted from Valiente et al. (2020), with permission from Elsevier.

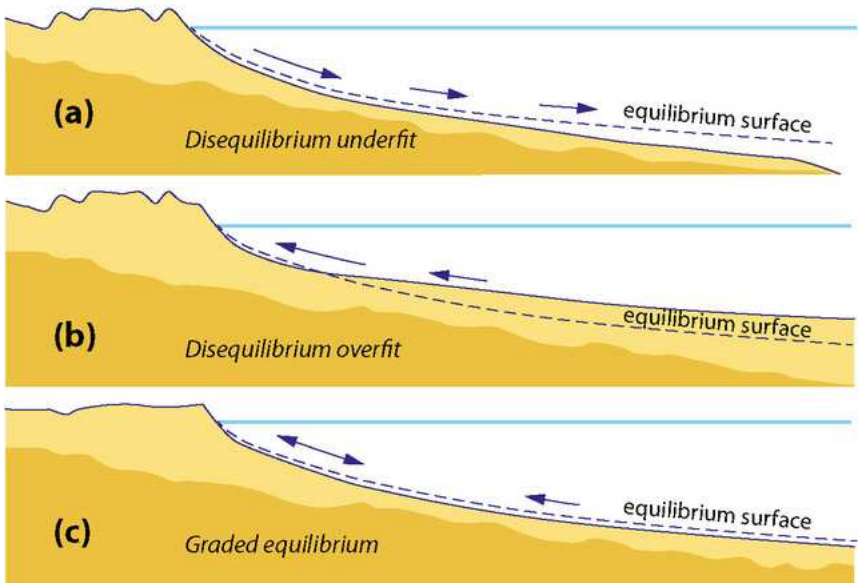


Figure 10. State of shoreface morpho-sedimentary equilibrium relative to an equilibrium surface after a period of sea-level stillstand and profile evolution: (a) shoreface shallower than the equilibrium surface (disequilibrium overfit), causing negative accommodation; (b) shoreface steeper than the equilibrium shoreface (disequilibrium underfit), thus providing sediment accommodation offshore, with sediment eventually supplied by river deltas; (c) fully adjusted shoreface (graded equilibrium). Adapted from ideas developed by Cowell et al. (2003b), Daley and Cowell (2013) and Kinsela et al. (2020).

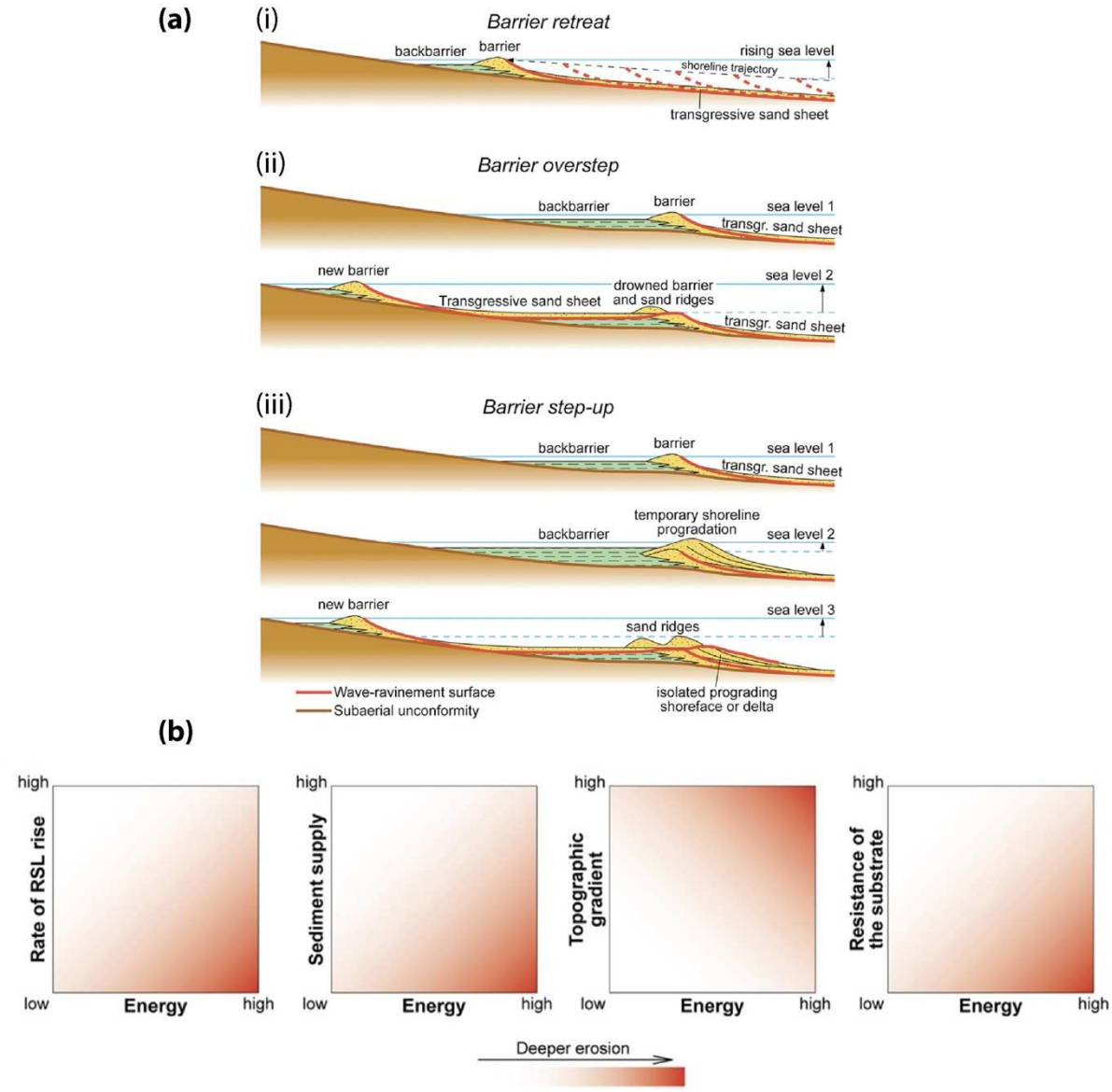


Figure 11. (a) Transgressive barrier retreat models involving a continuous landward retreat of the US resulting in the deposition of a transgressive sand sheet over the LS with its profile maintained (a1); it is worth noting that in some areas, this this sand sheet is actually a residue/lag formed by removal of certain transportable fractions; 'barrier overstep' transgressive model, where a barrier that initially retreats or grows vertically fails to keep pace with rising relative sea level and/ or does not receive enough sediment supply, resulting in barrier overstepping and a new barrier forming landward (a2); barrier mode

associated with high sedimentation rate, temporary phases of slow relative sea-level rise or relative sea-level stillstand, interrupted by renewed rapid relative sea-level rise and shoreline transgression; (b) dependence of wave-dominated US erosion during sea-level rise on rate of sea-level rise, sediment supply, substrate gradient and substrate resistance to erosion. From Zecchin et al. (2019), with permission from Elsevier.

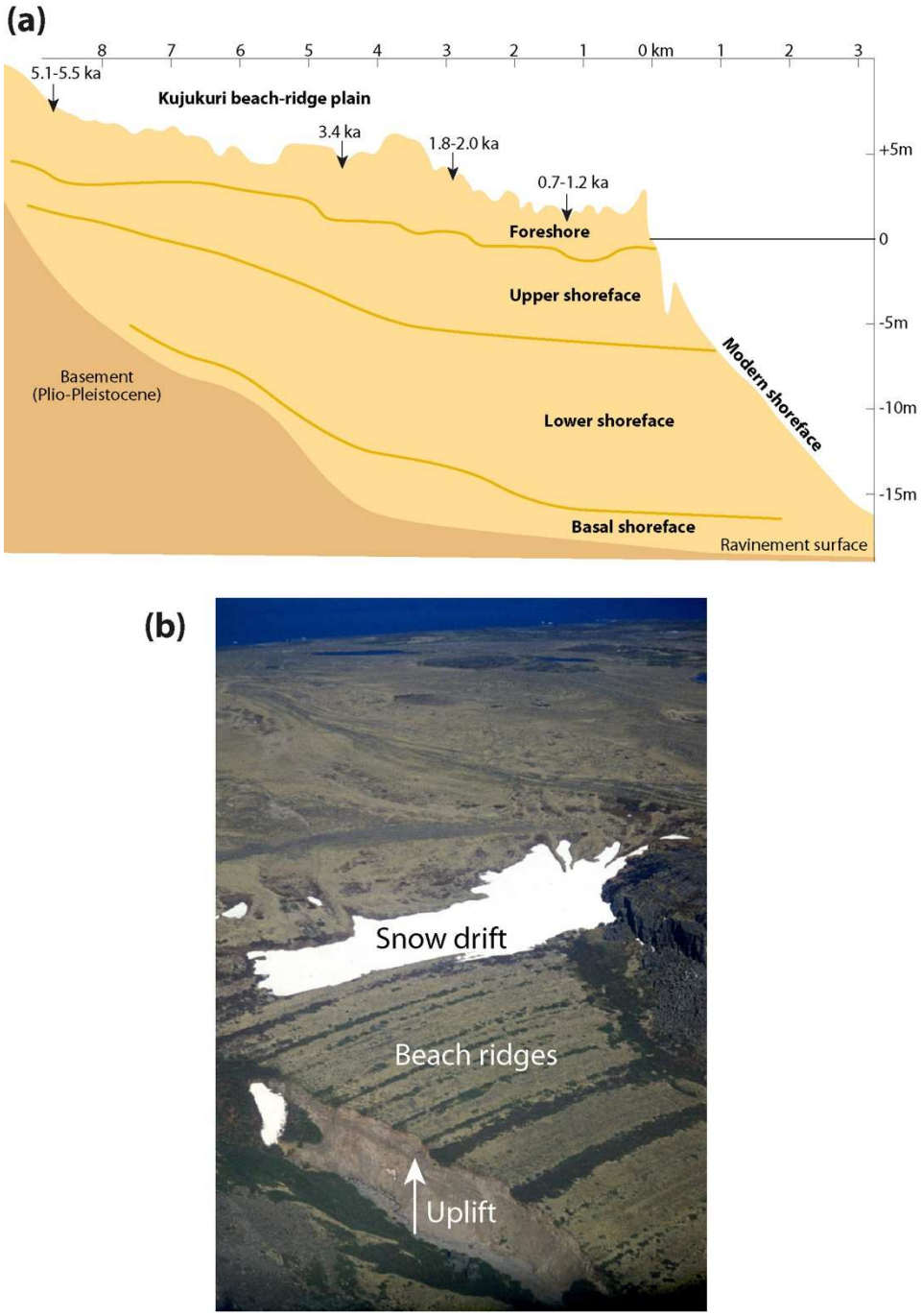


Figure 12. Simplified example (the numerous borehole and radiocarbon dates reported by the authors have been removed) of complex LS expression of sea-level falls, Kujukuri beach-ridge plain, Pacific coast of Japan, in response to tectonic uplift. Adapted from Tamura et al. (2007) (a); successive Holocene 'stair-case' raised beach ridges due to uplift resulting from

glacio-isostatic rebound following ice removal at Kativik, Hudson Bay, Canada (b). Photo credit: Jean-Marie Dubois, Sherbrooke, Canada.

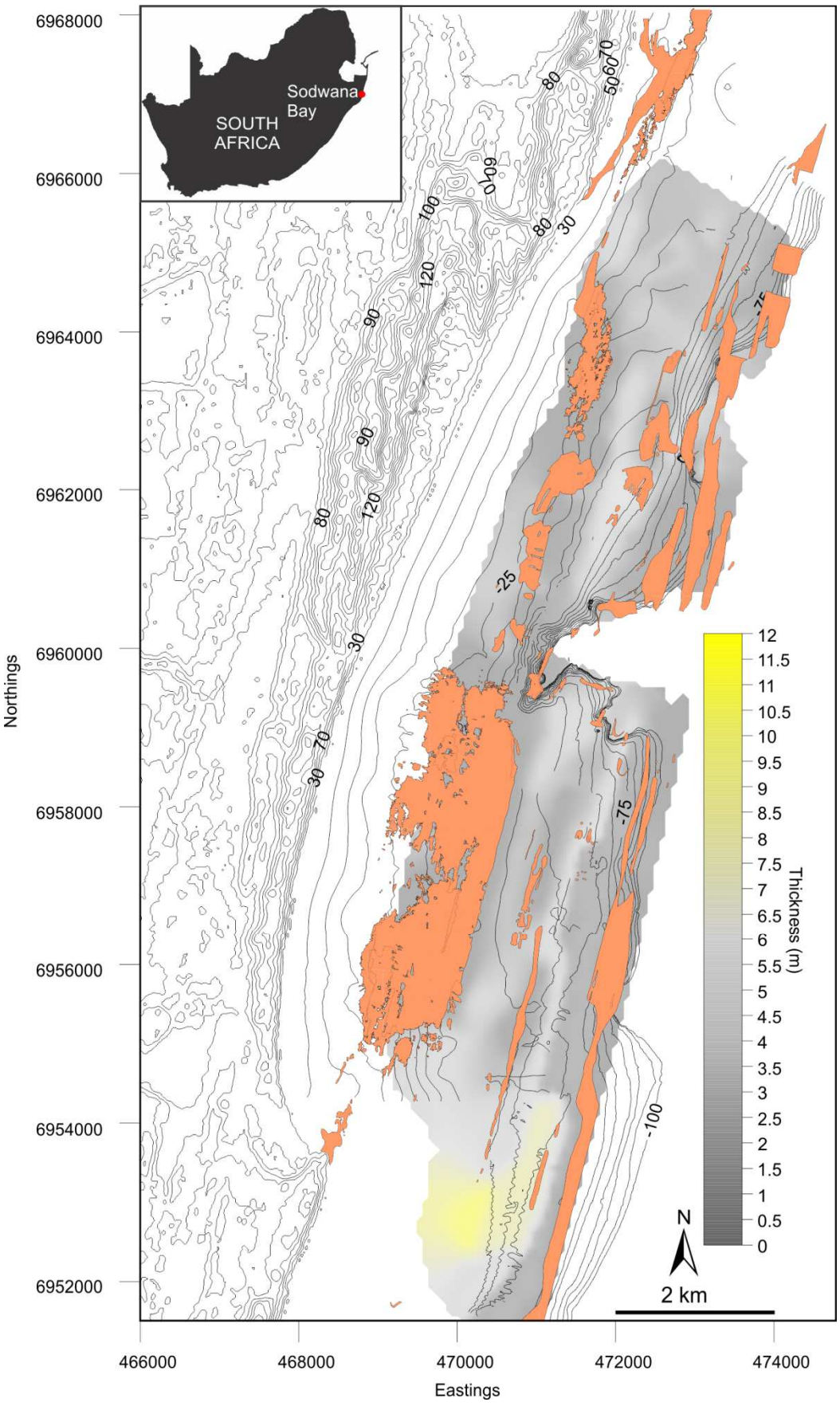


Figure 13. Geological influence on sediment thickness on the Sodwana continental shelf and LS, South Africa, and on sediment supply to the coast. Much of the volume of inner shelf sediment is impounded behind a coast-parallel outcrop of beachrock situated at 25 to 40 m depth (orange outcrops). Where alongshore gaps in the beachrock exist, material has been transferred from the LS to the US, forming 2-3 m-thick localised depocentres in ~10-15 m of water. In these areas, the adjoining coastal dunes supplied with sand from the LS reach heights of up to 120 m above sea level. Conversely, the dunes that front the beachrock-impounded shoreface are smaller (80 m high) and narrower due to the influence of the beachrock on onshore sediment supply. The very steep outer to middle shelf was subject to active ravinement during the Post-Glacial Marine Transgression (PMT), rendering the outer shelf relatively deficient in sediment and concentrating the material in the post-ravinement shoreface (Salzmann et al., 2013). Since the PMT, sediment supply on the LS has been assured by some carbonate production (see *5.1*), and this, together with the voluminous sediment entrained from the thick shoreface wedge, probably explains the large dunes on this coast. There is no production to seaward, the bioclastic debris that stretches to the shelf edge is mostly palimpsest. The dunes are among the tallest vegetated coastal dunes in the world and reach up to 200 m in elevation above mean sea level (Credit: Andrew Green).

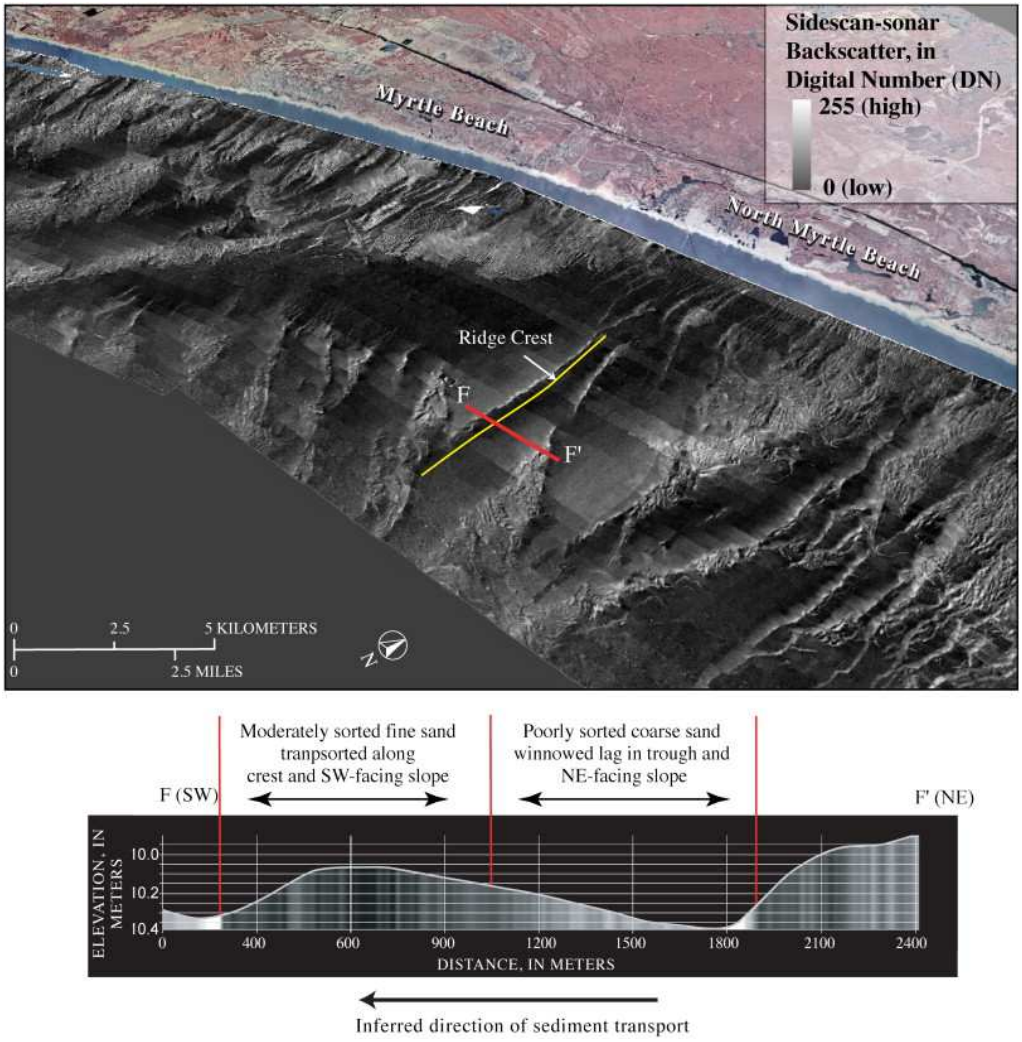
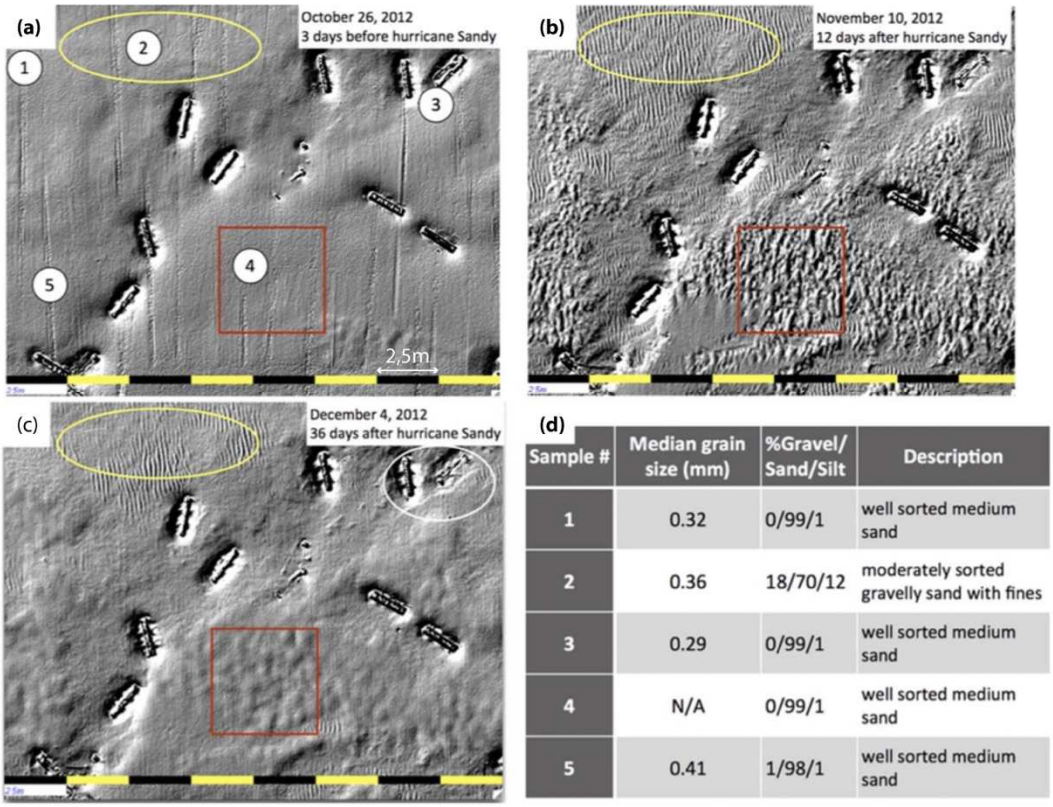


Figure 14. Top: Perspective view of the inner shelf offshore of Myrtle Beach, South Carolina, USA, showing sidescan-sonar imagery draped over bathymetry. Vertical exaggeration is 200. Bottom: Shore-parallel bathymetric profile across a characteristic low relief LS ridge (see top image for location). Backscatter and grain-size variations are shown in relation to ridge morphology. The geomorphic and textural variations suggest a long-term southwest transport of sediment that feeds the beach. Adapted from Denny et al. (2013), with permission from Elsevier.



Sample #	Median grain size (mm)	%Gravel/ Sand/Silt	Description
1	0.32	0/99/1	well sorted medium sand
2	0.36	18/70/12	moderately sorted gravelly sand with fines
3	0.29	0/99/1	well sorted medium sand
4	N/A	0/99/1	well sorted medium sand
5	0.41	1/98/1	well sorted medium sand

Figure 15. A 25 cm/pixel gridded shaded relief bathymetric map close-up of a portion of the New Jersey continental shelf, USA, showing the impact of Hurricane Sandy on sediment mobility: (a) 26 October 2012, three days before Sandy, showing a generally undisturbed seabed of fine sand supporting smooth hummocky beds (red square); (b) 10 November 2012, 12 days after the passage of Sandy, showing dramatic changes in the seabed morphology. Large symmetrical wave orbital ripples are evident in an area composed of moderately sorted gravelly sand with fines (yellow oval and sample 2 in d), while scour pits have become more pronounced and expanded around the scattered subway cars (these are old scrapped New York underground trains that have been dumped on the seabed to create artificial reefs for fishes and crustaceans). Areas of smooth well-sorted medium sand before the storm (red square in a) underwent important scour and erosion (red square in b). (c) 4 December 2012, 36 days after the passage of Sandy, illustrating partial recovery and return of the seabed to the pre-Sandy configuration. Note the return to a partially smooth hummocky morphology in the red square in c, and smoothing of some of the large wave orbital ripples (yellow oval in c), while some of the ripples mapped after Sandy are still

present. (d) Five sediment samples taken within the area illustrated in a–c. Adapted from Trembanis et al. (2013), with permission from John Wiley and Sons.

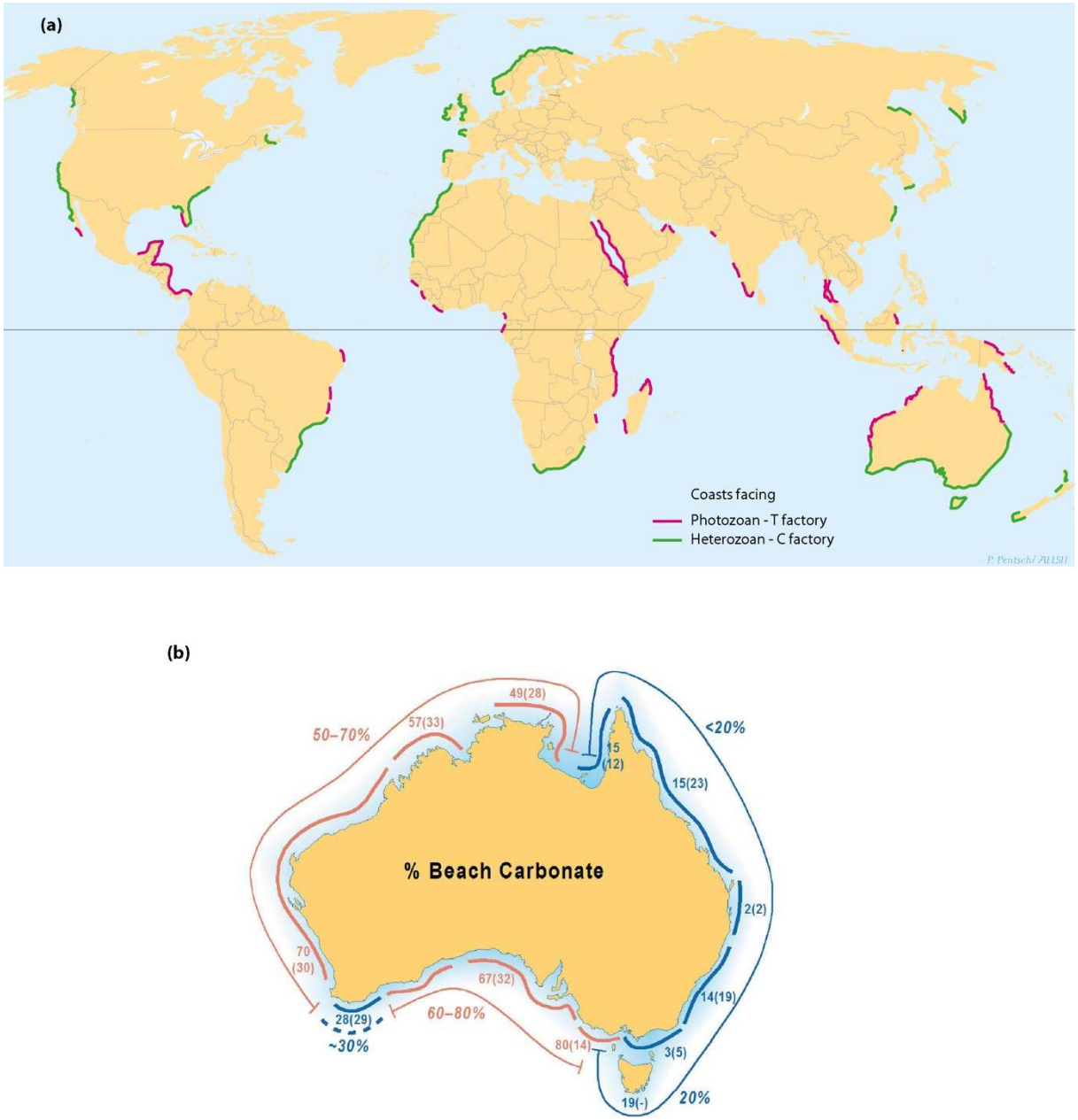


Figure 16. Coasts facing sand-producing LS carbonate factories (Photozoan-T and Heterozoan-C) exposed to mid-to high-latitude storm and swell waves (a); example of distribution of carbonate beach sands around Australia, essentially derived from the shelf (b). Numbers indicate mean regional percentage of carbonate beach sand, with standard deviation in brackets. Carbonate sands dominate across the southern and western coasts, derived from continental shelf biota, and in low energy areas, from seagrass meadows, while in the northwest, fringing reefs are a major source; with permission from Short and Woodroffe (2009).

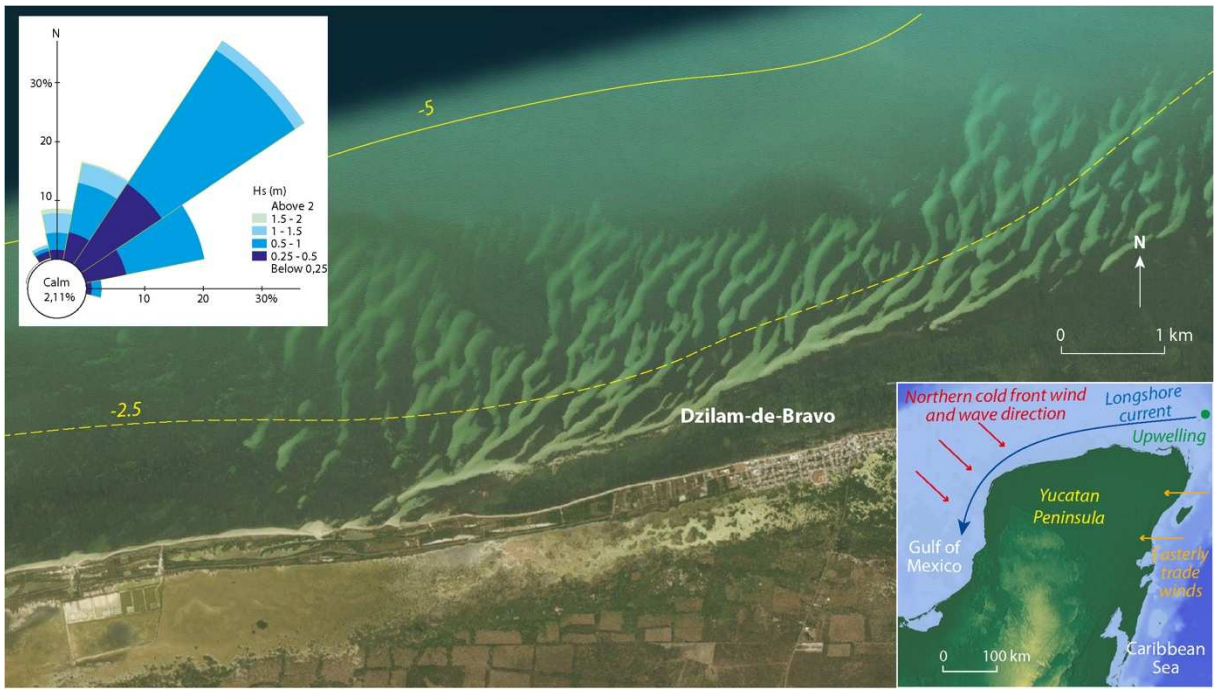


Figure 17. Google Earth image of the north coast of the Yucatan Peninsula, Mexico, showing the welding, on the US and beach, of bedforms (2-3 m high?) of carbonate sediment derived from a LS carbonate factory. Top left inset (from Appendini et al., 2012) shows wave heights measured at Dzilam-de-Bravo using an acoustic Doppler current profiler, while bottom right inset (from Lowery and Rankey, 2017) shows wind-generated hydrodynamic forcing circulation that includes significant upwelling caused by easterly winds. LS bedform mobility and onshore welding are probably driven mainly by high waves generated by cold northerly fronts. These welded forms feed the wave-driven US longshore transport system that contributes to the construction of the numerous beach ridges and spits characterizing the northern Yucatan coast.

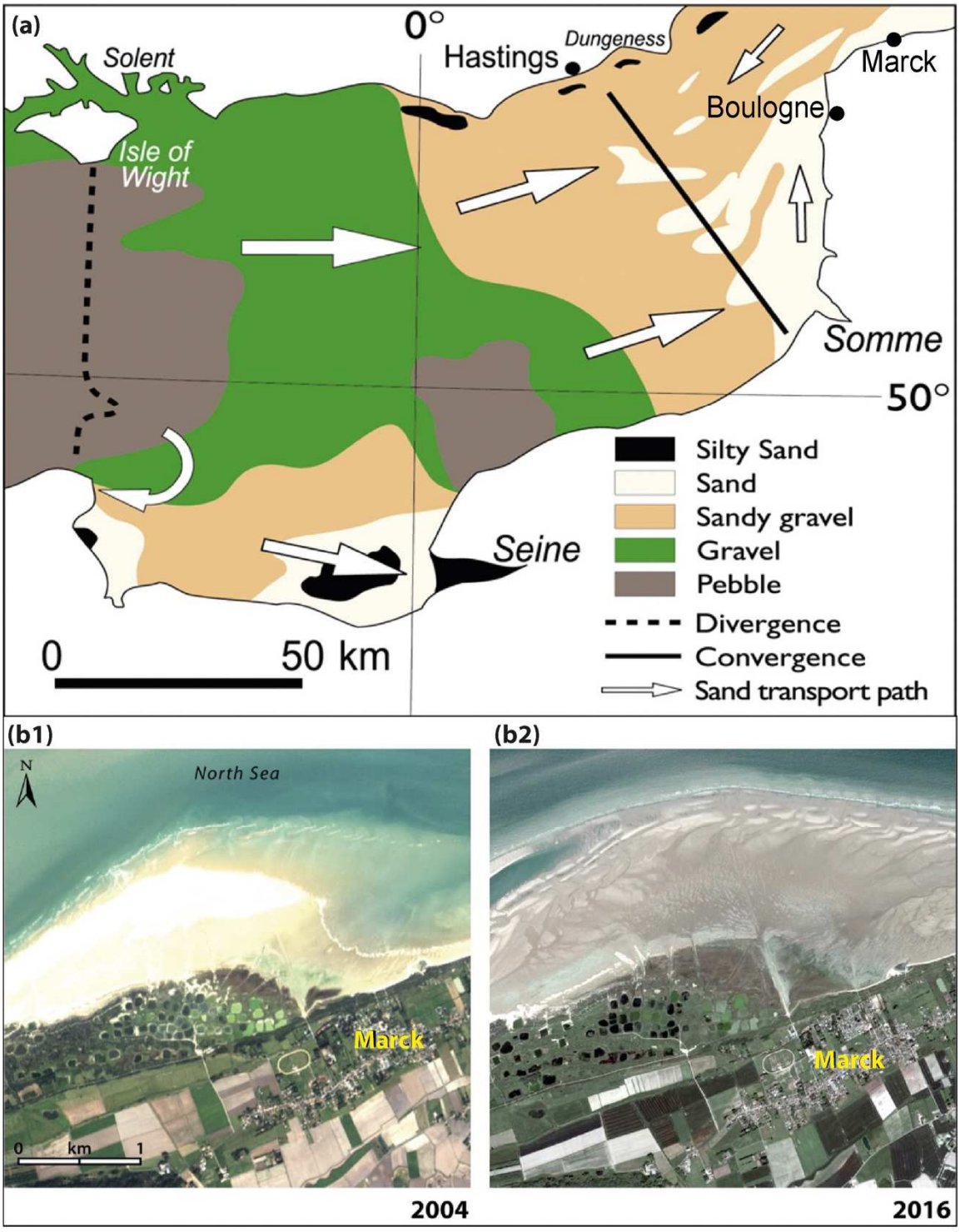


Figure 18. Sand transport convergences and pathways on the continental shelf of the eastern English Channel and southern North Sea resulting from storm, wind and tidal sorting of loose sediment deposited by rivers and glacial outwash during the Late Pleistocene lowstand, and reworked following the Post-Glacial Marine Transgression (a); sand depleted in the western part of the Channel seabed to expose an increasing gravelly substrate is piled up along a coastal sediment-transport pathway corresponding to the LS north of the Somme River estuary in France. This pathway has also constituted an important reservoir of sand

banks and ridges (see also Fig. 5) that are locally driven onto the US and beach, sourcing the large dunes on the west-facing French coast over the last 5K years; Google Earth images in 2004 and 2016 showing the eastern limit of a prograding sandflat shore that developed from the welding of a shoreface sand bank at Marck in the course of the 20th century (b1, b2).

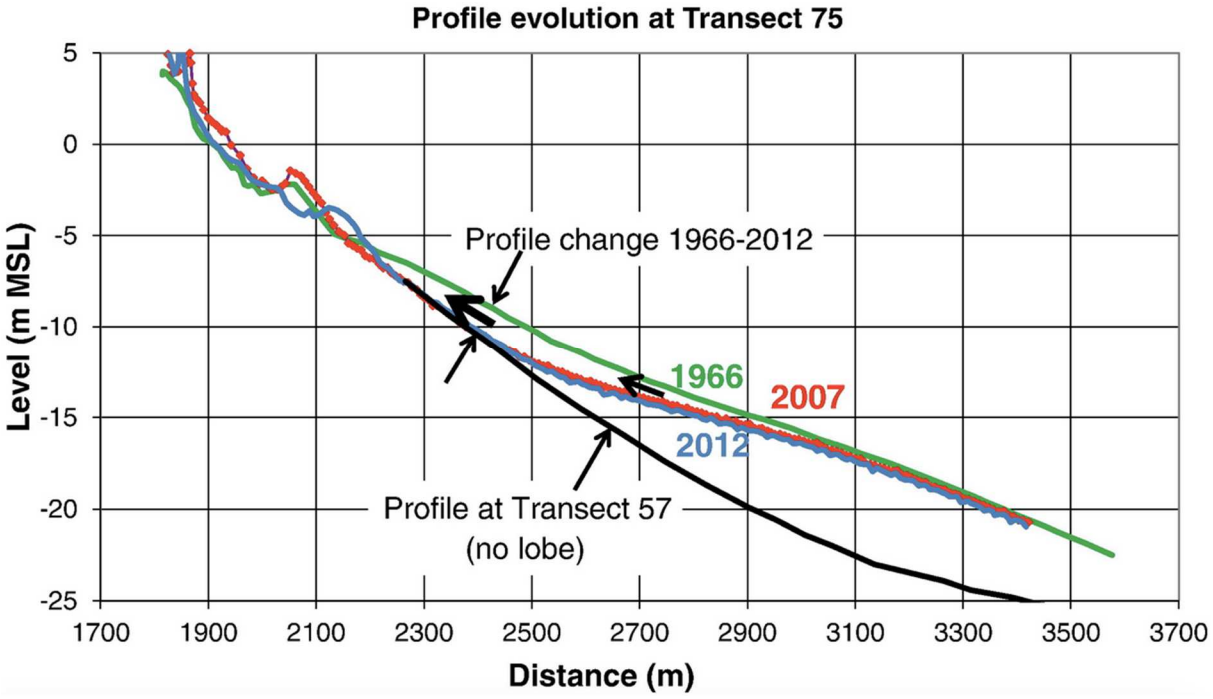
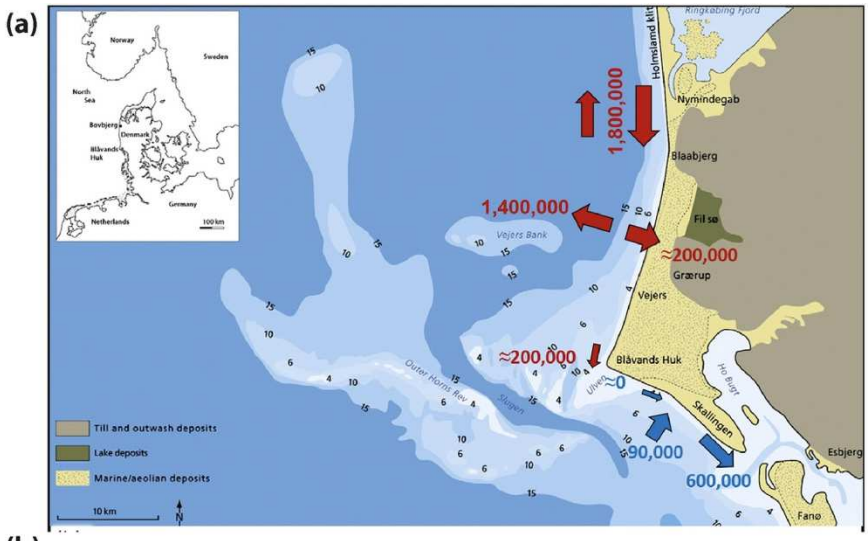
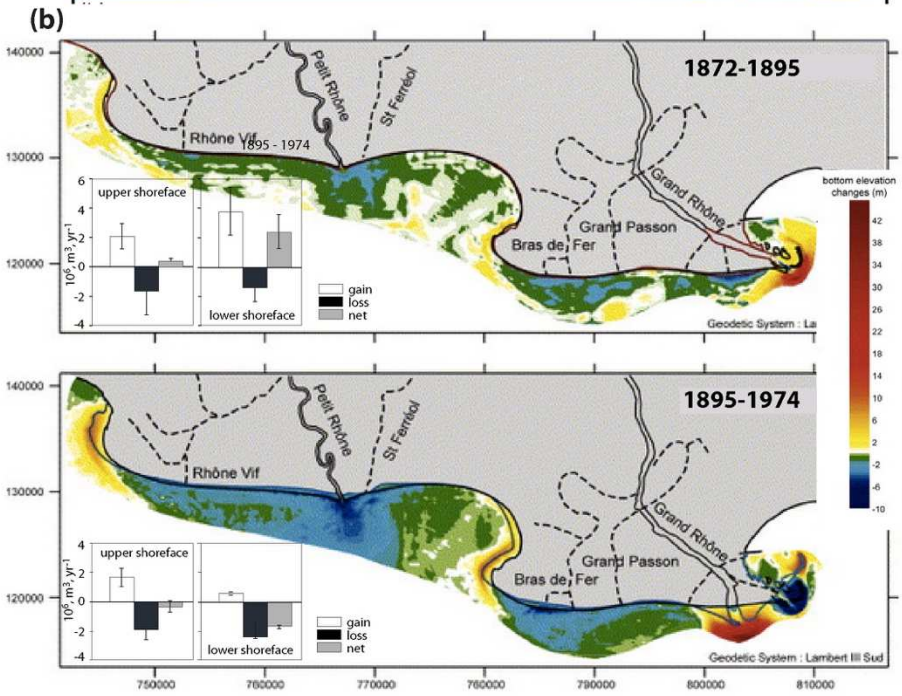


Figure 19. Progressive reworking over 46 years (1966-2012) of the abandoned delta lobe of the Nerang River, Gold Coast, Australia, on the LS, marked by onshore sand supply that significantly supplemented the longshore transport system. From Patterson and Nielsen (2016), with permission from Elsevier.


(a)
Norway
Sweden
North Sea
Bovbjerg
Denmark
Blåvands Huk
Germany
Netherlands
100 km
Holmsland Klit
Ringkøbing fjord
Nymindegab
Blaabjerg
Fil sø
Vejers Bank
Grærup
Vejers
Blåvands Huk
Ho Bugt
Outer Horns Rev
Skagen
Skallingen
Esbjerg
Fanø
1,800,000
1,400,000
≈200,000
≈200,000
≈0
90,000
600,000
Till and outwash deposits
Lake deposits
Marine/aeolian deposits
10 km
N


(b)
1872-1895
1895-1974
Petit Rhône
St Ferréol
Rhône Vif
Grand Rhône
Grand Passon
Bras de Fer
upper shoreface
lower shoreface
gain
loss
net
bottom elevation changes (m)
Geodetic System : Lambert III Sud
750000
760000
770000
780000
790000
800000
810000
140000
130000
120000

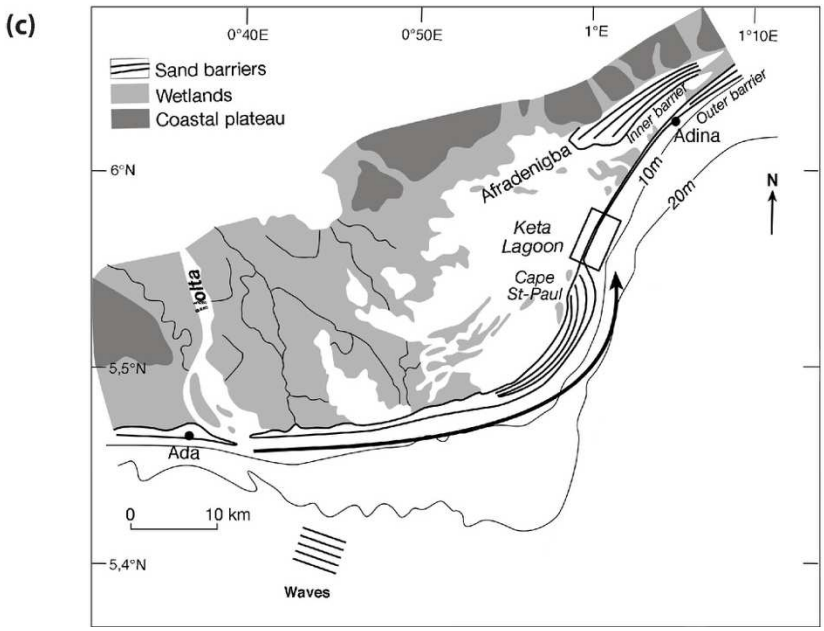

(c)
0°40E
0°50E
1°E
1°10E
Sand barriers
Wetlands
Coastal plateau
6°N
5,5°N
5,4°N
Afradenigba
Inner barrier
Outer barrier
Adina
10m
20m
N
Keta Lagoon
Cape St-Paul
Volta
Ada
0
10 km
Waves

Figure 20. Examples of connected LS-US sand transport pathways and sediment budgets: (a) the shoreface sediment budget near the Blaavands Huk salient at the northern edge of the Wadden Sea region in Denmark. Red arrows refer to the Vejers coastal cell and blue arrows the Skallingen cell. Numbers are stated in m^3/yr; (b) the Rhône delta where abandoned lobes are being progressively eroded, resulting in landward translation of the LS (blue areas) whereas deposition on the LS occurs in the presently active Grand Rhône mouth and in the distal tips of spits (yellow to red areas); (c) the Volta delta spit which has been actively sourced by longshore transport from the Volta River but also probably by the reworking of an abandoned LS delta lobe. Sand transported along the spit spills over at its distal tip (arrow head) onto the LS. (a) adapted from Larsen (2003) and Aagaard (2011); (b) adapted from Sabatier et al. (2006); (c) modified, after Anthony et al. (2016).

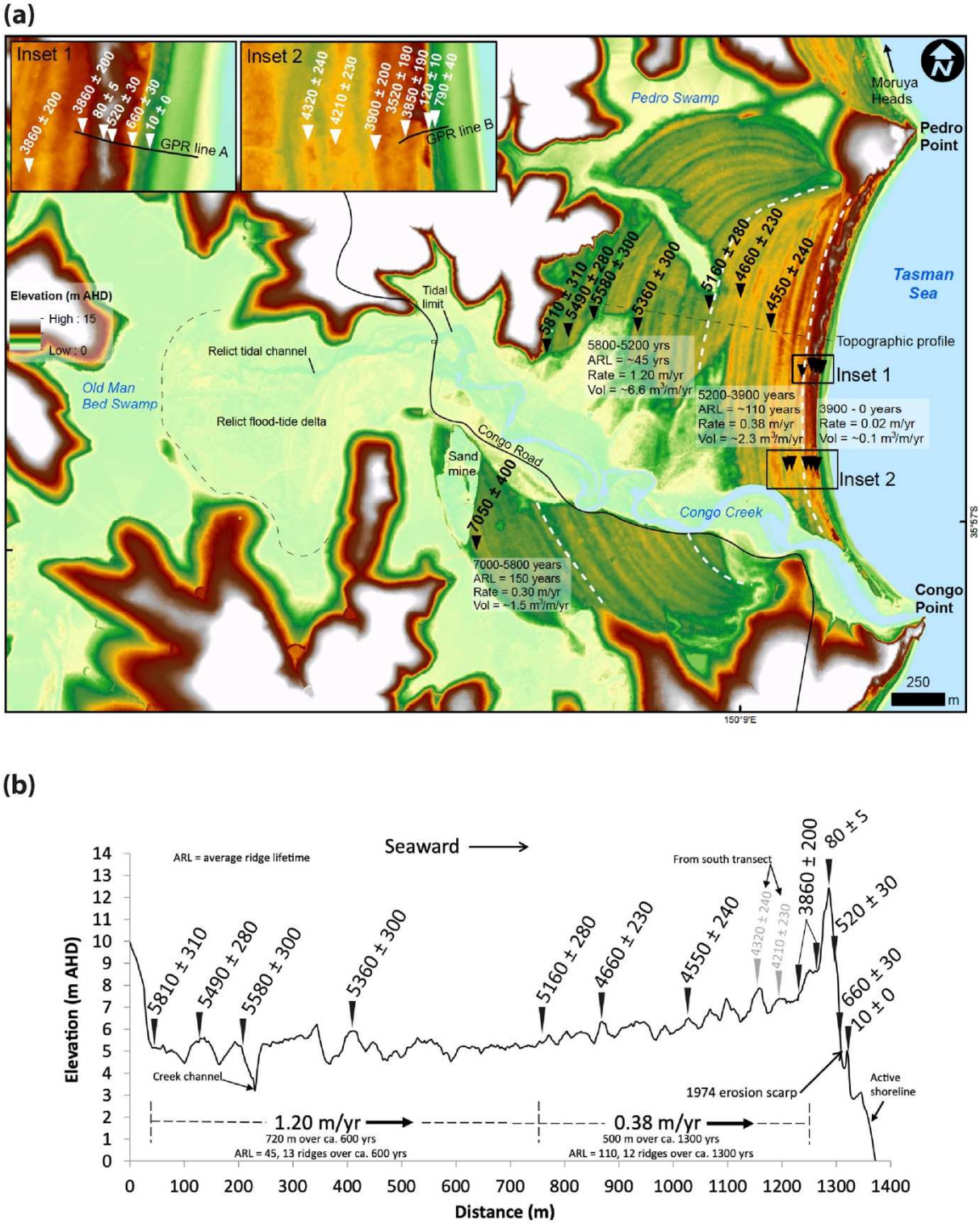


Figure 21. Shore-normal topographic profile across the centre of the Pedro Beach barrier, southeastern Australia, showing OSL ages in relation to beach ridges (lighter shaded ages are from another transect). Progradation virtually ceased nearly 3.9K years ago, to be replaced by important vertical dune ridge accretion. Adapted from Oliver et al. (2019), with permission from John Wiley and Sons.