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Coastal Processes and Landforms

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Abbreviated title: Coasts

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Abstract

This chapter charts developments in the study of coastal processes and landforms in the period between the 1960s and the end of the millennium, focusing on efforts to better understand sandy beaches, barriers and barrier islands, deltas and estuaries, tidal flats and marshes and coral reefs. The period saw the emergence of a dual focus on, first, the elucidation of landscape history from morphological and, later, stratigraphic evidence; and second, the relationships between shoreline morphology and processes of sediment movement. Particularly noteworthy was the integration of a broad spectrum of space and time scales in a conceptual framework that became known as 'coastal morphodynamics'.

> ‘*Why was this period [the 1970s – 1990s] such a productive one for coastal geomorphology? Our somewhat simplistic answer is that this was an age of discovery characterized by excitement, energy, and enthusiasm among a relatively small community of coastal researchers who wanted to explore the (largely) unknown – the nearshore. There was the prospect of making new discoveries with each new field or lab experiment because there were precious few measurements of nearshore processes with which to validate theoretical speculation and conceptual reasoning. The field was wide open and there was much to be done.’* (Bauer et al. 2020, 125-126)

Introduction

In this chapter, we explore what Bauer et al. (2020) have termed the ‘golden era of coastal geomorphology’, starting with its early roots in WWII and immediate post-WWII coastal research. Acknowledging these historical contexts, we here chart developments in the study of coastal processes and landforms in the period between the 1960s and the end of the millennium. In the space available we cannot hope to describe the full range of coastal research undertaken in this timespan; such detail is expertly provided in texts by, *inter alia*, Bill Carter (1988), Paul Komar (1998) and Colin Woodroffe (2003). We are aware that we do not discuss either coastal sand dunes or rocky coasts in depth but late twentieth century research on these topics has been covered by Bauer and Sherman (1999) and by Trenhaile (1987) and Sunamura (1992) respectively. Rather, we restrict ourselves to an exploration of some emerging themes in relation to the understanding of sandy beaches, barriers and barrier islands, deltas and estuaries, tidal flats and marshes and coral reefs. French (2004) has noted that most accounts of this period draw attention to the emergence of a dual focus on, first, the elucidation of landscape history from morphological and, later, stratigraphic evidence; and second, the relationships between shoreline morphology and processes of sediment movement. We broadly explore these two strands here and, in the case of the second, concur with Church (1996) that this led to a re-focusing on spatial scales of enquiry commensurate with the time scales of observable processes. However, what was particularly noteworthy in coastal geomorphology was the integration of a broad spectrum of space and time scales in a conceptual framework that became known as ‘coastal morphodynamics’.

Beginnings

By the middle of the twentieth century, coastal research was still largely typified by authoritative, regional monographs, such as those on the Mississippi delta (Fisk, 1944), the Danish coast (Schou

1945) and the coastline of England and Wales (Steers 1946). These broad scale descriptions were theoretically framed by the remarkable enduring influence of Douglas Wilson Johnson's (1919) *Shore Processes and Shoreline Development*. For V.P. Zenkovich, author of another major coastal text (Zenkovich 1967), *Shore Processes* 'remains to this day the most complete theoretical study' (Zenkovich 1962, 7), whilst Alfred Steers commented that 'For the first time the study of coasts was co-ordinated and put into a form which at once showed the possibilities of the subject' (Steers 1971, 11). Embarking on the newly established degree course in Geography at Cambridge in October 1919, 'Alfred [Steers] bought his copy [of Johnson (1919)] at the beginning of the Michaelmas Term, and he still had it, much scored and annotated (unusually for him), when he died [March 1987]' (Stoddart 1987, 11).

The 'lone scholar' tradition was transformed in the aftermath of World War II by the development of inter-disciplinary research teams, the emergence of significant funding streams and new fora for research dissemination (including the First Conference on Coastal Engineering, held at Long Beach, California in October 1950). This less descriptive and more process-based coastal geomorphology assimilated advances in oceanography (Kinsman 1965, Komar 1976, Bascom 1980) and marine geology (Shepard 1948, Guilcher, 1958), and the understanding of littoral processes was enhanced by the advent of more quantitative approaches to sedimentology (Krumbein & Graybill 1965, Griffiths 1967) and the first experimental studies of sediment movement by wave and currents (Bagnold 1947, Inman & Bowen 1962). More systematic monitoring of coastal dynamics in this period was facilitated by the datasets provided from the multispectral scanner on the Earth Resources Technology Satellite (ERTS; subsequently re-named LANDSAT-1) (Frihy et al., 1994) although it was the improved resolution and imaging capabilities of LANDSAT-3, (launched in March 1978) that really initiated the use of satellite imagery for documenting coastal change (see also Chapter 2, this volume). The ability to quantitatively analyse rates of shoreline change in a Geographical Information System (GIS) framework came with the development of the USGS Digital Shoreline Analysis System (DSAS) launched in 1992, originally linked to the Apple Macintosh GIS, MapGrafix (Danforth & Thieler 1992), but then re-coded to operate as an add-in to ESRI ARC/INFO and then ARC/GIS in 1999.

Alongside this increased emphasis on contemporary coastal dynamics, the period also saw wider application of an expanded range of radiometric dating methods to the investigation of past coastal environmental change, most notably Uranium-series dating (e.g. Veeh 1966, Broecker et al. 1968, Chappell 1974; and see Chapter 3, this volume). Much of the activity in this field was organized around formal international research collaborations, as typified by Projects 61, 200 and 274 (1974-

1993) of the UNESCO/IUGS International Geological Correlation Programme (IGCP) (synthesized by Carter & Woodroffe 1984).

Coastal fieldwork in the 1960s was largely still characterized by long-established optical survey methods, with beach transects (e.g. Emery 1961, King & Barnes 1964) surveyed from terrestrial benchmarks, current flows established with floats or by hand-held / boat-deployed impellor-based current meters, and basic surface sediment sampling and analysis. During the 1970s, however, research groups in the USA, Canada, Australia, Japan and the UK began to develop electronic sensors to monitor wave, surf zone and swash processes and seabed responses (e.g. Sonu et al. 1973, Kanwisher & Lawson 1975, Brenninkmeyer 1976), hard-wired to ink chart recorders and primitive computers (e.g. Downing et al 1981). Despite this march of technology, the human effort to data return ratio was still immense. During the latter part of the period, theoretical advances in coastal oceanography and the explosion in digital computing stimulated a new focus on numerical modelling of hydrodynamic processes (Matthews & Leavatsu 1975) and the emergence of simulation models for a variety of coastal landforms and coastal types (e.g. King & McCullough 1971, Komar 1973, Hanson & Kraus 1989, De Vriend et al. 1993a). By the end of millennium, a steady increase in the sophistication of coastal process and landform modelling had been facilitated by developments in field instrumentation, which greatly increased the range of phenomena that could be measured, the time and space resolution and duration of measurement and, not least, the safety with which such research could be undertaken (Short & Brander 2020).

Aspects of the shifts in research focus highlighted above had earlier origins. In the United States, the geomorphological underpinnings of coastal engineering were expanded under the auspices of the Beach Erosion Board (BEB). Formed in September 1930 (Quinn 1977), in the immediate post-war period the BEB began to sponsor additional external research on coastal processes at US Universities and oceanographic institutions and develop close linkages with the US Army Corps of Engineers and its Waterways Experiment Station at Vicksburg, Mississippi (Quinn 1977; it always amused R.J. Chorley that a waterways station had once (on 3 October 1960) been burnt to the ground). Particularly influential was the synthesis published as BEB Technical Report No. 4 (1954) entitled *Shore Protection Planning and Design*; this subsequently evolved into the go-to bible of coastal engineering, the *Shore Protection Manual*, which was first published in 1973 by the BEB's successor organization, the US Army Coastal Engineering Research Center (CERC 1973). In 1977, the Corps opened its Field Research Facility at Duck, Outer Banks of North Carolina, USA. This included a 560 m long research pier, a 43 m high research tower and a three-legged mobile survey platform, the CRAB (Birkemeier & Holland 2000). Collaborations with the United States Geological Survey, Oregon State University, and series of international research programmes throughout the 1980s followed.

Importantly, an unexpected consequence of this research was the realization that time-exposure imagery might be used to capture intertidal morphodynamics, through the development of the 'Argus' system of videography (Holman & Stanley, 2007).

Another important stepping stone in the immediate post-war period was the establishment in 1946 of the United States Office of Naval Research (ONR). For coastal geomorphology, ONR research was energised by the formation of a Geography Branch and the appointment of its Director, Evelyn L. Pruitt in 1948. At Louisiana State University, Pruitt and Richard J. Russell established the Coastal Studies Institute (CSI), informally from 1951 and then more formally from 1954. At about the same time in the UK, the Government's Department of Scientific and Industrial Research established the Hydraulic Research Organisation in 1947 and subsequently in 1951 the setting up of an experimental facility, the Hydraulics Research Station, at Wallingford on the River Thames; its links to coastal research are explored further below. In The Netherlands, the catastrophic 1953 storm surge was the catalyst for the formation of the Delta Committee, the adoption of the Delta Plan and major ongoing advances in hydraulic engineering (Gerritsen 2005). However, even with these organisational shifts, the role of key individuals in driving coastal research agendas continued to be important; in the USA many such pioneers were from the so-called 'golden generation', individuals who experienced 'Depression Era deprivation, to daring service in WWII, to post-war boom and building' (Thom 2015).

Coastal morphodynamics

Morphodynamics can be broadly defined as the complex and mutual co-adjustment of processes and forms (Figure 1a). We trace the history as to how coastal geomorphology reached this defining synthesis in more detail below but highlight here what we believe to be *the* key coastal paper of the period 1965-2000, that by Don Wright and Bruce Thom on 'Coastal deposition landforms: a morphodynamic approach' published in the first volume of Progress in Physical Geography in 1977 and subsequently elaborated upon by Thom and Peter Cowell (Cowell & Thom 1994; Figure 1b). The paper was 'a reaction in part to the heavy emphasis by those with a 'fluid dynamics' bent, which did not appreciate the interaction with sediment movements at different time and space scales' (Thom to Spencer, pers. comm., 30 September 2021). In looking back, Thom (2014, 1-2) has recalled:

> *'what we tried to do was bring them [a set of principles] together to capture the complex physical nature of coastal systems. For instance, we expressed how different coastal depositional systems could reflect either positive (self organising or change in state) or negative (self regulating or equilibrium tendency) feedback loops. The stochastic nature of external processes driving change to a given coastal system was placed in the context of inherited or antecedent conditions. We stressed the need to appreciate relaxation times and non-linear behaviour of the interacting processes and forms leading to*

> *uncertainty in outcomes. One distinction we were able to make was between changes driven by "internal" morphodynamic adjustments as sediment is moved within an area compared to changes driven by external or boundary conditions such as eustatic sea level rise or a change in wave climate.' and 'These principles have been applied … over time scales from days to millennia'.*

Post-2000, the Cowell and Thom (1994) space-time hierarchy has been formally extended to coral reef and mangrove forest environments; here we simply note the pioneering studies of Thom (1967) in linking delta dynamics to the distribution of mangroves on the Tabasco and Campeche coast of Mexico (and see also Thom 2020a) and David Stoddart's (1980) exploration of reef development and reef-top mangrove distributions on the northern Great Barrier Reef. (Stoddart had been instrumental in encouraging Thom to study at LSU (as he himself had been encouraged by Steers to visit his good friend Russell) (Thom to Spencer, pers. comm. 23 September 2021). At the end of this chapter, we highlight the increasing use of mathematical models to understand and predict coastal morphodynamic behaviour, and the challenges of developing models capable of providing useful prediction of future coastal landform evolution at the spatial and temporal scales relevant to shoreline management.

Sea-level rise

Alongside Wright, Thom and Cowell's powerful intellectual framework emerged a wider societal and institutional concern: the 'sea-level rise issue' of the 1980s (Carter & Woodroffe 1994). In 1980, the US Senate Committee on Energy and Natural Resources held a hearing on the effects of carbon dioxide build up in the atmosphere, which included maps of how US coastal cities would be affected by a rise in sea level of up to 25 feet (7.6 m). The alarming graphics provided by the Committee's analysis were taken forward by the US Environmental Protection Agency (EPA) who held a conference on their findings in March 1983, shortly followed by several papers at that year's Coastal Zone '83 conference (San Diego, June 1983) and a widely disseminated synthesis on likely societal and economic impacts (Barth & Titus 1984). At this stage, sea level projections were derived from the EPA Report of Hoffman et al. (1983). Although the authors were careful to point out the huge uncertainties present in future sea level predictions and gave a 'most likely' range of 144 – 217 cm of sea-level rise by 2100, most commentators latched onto their 'cannot be ruled out' figure of 345 cm. Revised EPA projections in 1986 (Hoffman et al. 1986) showed an even higher upper bound (367cm), on the assumption of massive increases in iceberg calving in Antarctica (Thomas 1985, 1986; Thomas' own assessment from the same year envisaged an upper bound of 230 cm (published in Thomas 1987)). A highly distinguished panel assembled by USA National Research Council (NRC) arrived at a more conservative estimate of 50 – 150 cm of sea-level rise by 2100 (NRC 1987). In this febrile atmosphere, however, much of the press coverage was still envisaging widespread coastal landscape drowning, with climate change refugees fleeing disappearing coral atolls. On reading the local newspaper headline 'Costa-del-Cambridge by 2050: experts warn' (Cambridge Evening News, 11 April 1989), Cambridge geomorphologist R.J. Chorley marched into the Department of Geography tea room and announced that he was applying for the city's beach deckchair concession. Within the science community, however, a very different narrative was already starting to take shape, set in train at a workshop at the Climate Research Unit, University of East Anglia, UK in September 1987 and then taken forward at a SCOPE workshop in Bangkok in late 1988. Modelling by Tom Wigley, Sarah Raper and Hans Oerlemans was then incorporated into the First Assessment Report of the Intergovernmental Panel on Climate Change (IPCC) (Warrick & Oerlemans 1990). This research provided a radically different picture of future sea-level change, with a new 'best estimate' of 66 cm (range: 31 – 110 cm) by 2100 (Figure 2). When these IPCC figures were then themselves revised in 1992, the resulting sea-level rise estimates were reduced further to 48 cm and with a narrower uncertainty band (15 – 90 cm) (Wigley & Raper 1992 (and see also Wigley & Raper 1993)) and that is where they stayed in the IPCC's Second Assessment Report, published in 1996 (Warrick et al. 1996). Figure 2 shows the broad continuation of this projection, through the IPCC's Third (Church et al., 2001) and Fourth (Bindoff et al., 2007) Assessment Reports, published in 2001 and 2007 respectively. IPCC-generated estimates of global mean sea level rise still typically fall between 0.43 m (0.29–0.59 m, likely range; emissions scenario RCP2.6) and 0.84 m (0.61–1.10 m, likely range; RCP8.5) by 2100, relative to 1986–2005 (Oppenheimer et al. 2019; but see also Nicholls et al. (2011) for higher predictions).

These re-calibrations of likely future sea-level rise found their way through to geomorphology in the mid-1990s. Thus Spencer was able to conclude 'exercises which compare rates of coral growth and reef accretion with rates of predicted sea-level rise to the year 2100 suggest that coral reefs are unlikely to be catastrophically drowned, as was thought likely in the 1980s' (Spencer 1995, 61) but pointed to the complexities of comparing 'biological' and 'geological' measures of reef performance and the complications introduced by possible changes in ocean basin storminess, coral bleaching and non-linear 'phase shifts' (Done 1992) between 'hard' carbonate-accumulating reefs and 'soft' algal reefs. In the case of wetlands, vertical accretionary deficits in US Gulf Coast saltmarshes were linked to rapid and potentially accelerating rates of coastal land loss (Baumann et al. 1984, Penland et al. 1990), although in more strongly tide-dominated settings with adequate sediment availability, both empirical evidence (Stevenson et al. 1986) and modelling (French 1994) indicated greater resilience of marsh elevations (and by implication marsh area) to the kind of sea-level rise scenarios envisaged at this time.

Coastal classifications

Coastal settings, landforms and the plate tectonic revolution

The regional monographs of the early 20th century were often accompanied by a variety of classificatory schemes. Although some of these schemes were descriptive (e.g. McGill 1958), the majority were based upon the application of the Davisian concept of landscape stage (ranging from youthful to mature) or the distinction between submergent and emergent coasts; over time, these latter schemes became more sophisticated, with multiple sub-divisions of coastal type (e.g. Valentin 1952, Cotton 1954). Doug Inman (Plate 1) recalls these approaches with reference to the work of his PhD supervisor Fran Shepard:

> *'what he did, and to some extent it becomes a bit boring, but in his books [e.g. Shepard 1948], for example, he'd take you on tours around the coasts of the world and what the shelf width was here and what it was there, and what the sediment was. And to that extent, he set the necessary empirical relations upon which the science could be built. You couldn't build the science without that information'* (Doug Inman to L. Harkewicz, La Jolla, 8 May 2006 (Regents of the University of California 2006)).

'Building the science' changed utterly with the application of plate tectonic theory (see Chapter 2, this volume), underpinned as it was by 'the golden age' of the exploration of the ocean basins in the 1960s (Menard 1986; and detailed in Spencer et al. 2008, in Volume 4 of the History of the Study of Landforms). For Inman:

> *'I happened to be sitting here [Scripps Institution of Oceanography] and that's why I wrote this paper on the tectonic classification of coasts [Inman and Nordstrom 1971] because it suddenly snapped clearly why the coasts are like they are, and you can't really do something like this until you finally know what was going on. And so when you recognize that you have certain sections where the coasts are colliding, and mountain ranges forming … And you look at places like the east coast of the Americas, where they have quite different coastlines, it suddenly all falls into place. It's obvious why you have these different things.'* (Inman to L. Harkewicz, La Jolla, 8 May 2006).

It took longer for the plate tectonic revolution to be applied to the geology and geomorphology of the islands of the ocean basins and to finally provide a mechanism – lithospheric cooling and plate sinking - for Charles Darwin's great deductive subsidence theory for the transformation of volcanic islands with fringing reefs, into barrier reefs and sea-level (i.e. non-elevated) atolls (Scoffin & Dixon 1983). Regionally, in the Hawaiian Islands – Emperor Seamounts, Ricky Grigg linked island vertical subsidence, lateral reef migration into progressively cooler waters and declining rates of coral calcification to determine the island chain's 'Darwin Point', the location at which sea-level atolls are transformed into submerged seamounts. From this emerged a hypothesis that all island chains must have such points where they move out of (or indeed into) the reef seas and thus where reef development ceases (or begins) (Grigg 1982, Grigg & Epp 1989). As well as these ocean basin-scale dynamics, McNutt & Menard's (1978) benchmark paper showed how the weight of 'young' (ca. 1 Ma) volcanoes at magmatic hotspots results in regional uplift of neighbouring sea-level atolls. Geomorphology can thus be used to calibrate fundamental models of how the lithosphere flexes under loading (for the modern re-interpretation of the remarkable raised reef limestones of the South Pacific – the so-called 'makatea' islands – see Spencer et al. 2008).

Process-based classification

One of the major shortcomings of all these morphogenetic approaches was, and remains, the rudimentary treatment of coastal processes. The pioneering, pre-internet attempt of Jack L. Davies in the 1970s (revised in Davies 1980) to map global and tidal environments was remarkable although Davis and Hayes (1984) subsequently argued that it is the relative roles of waves and tides that is important rather than their absolute values. Central to these exercises have been ternary classifications of sedimentary coasts. In the early 1960s, the Coastal Information Program at CSI in Louisiana acquired data on 200 variables for 50 world deltas, placing individual deltas in a space defined by fluvial, tidal and wave energies (Wright & Coleman, 1973; Coleman & Wright 1975; and see also Galloway 1975), an approach subsequently applied to barrier coasts (Hayes 1979). Envisaging a continuum between deltaic and estuarine systems, Boyd and Dalrymple (Boyd et al. 1992; Dalrymple et al. 1992) extended this approach by adding a temporal perspective; in their scheme, classificatory triangles were stacked to form a three-dimensional prism (Figure 3). Landform characteristics may change, or transitions between landform type occur, along this time trajectory. As with beach morphodynamics (see below), generality is achieved through the description of 'end-member' types, between which transitional cases can be readily identified.

Characteristic forms and static equilibria

With BEB / ONR funding between 1944 and 1948, the Waves Investigation Group (led from the Institute of Engineering Research, Berkeley by Morrough (Mike) O'Brien, J.W. Johnson and Richard G. Folsom) surveyed 40 Pacific coast beaches, between Coronado, California and Grenville Bay, Washington State, a distance of over 2,000 km. Some 500 cross-shore profiles (dune foot to 9 m below sea level) and over 600 sand samples allowed Bascom (1951) to explore, for the first time, the relationships between quantitative measures of beach morphology, sediments and wave energy levels; Wiegel (1964) subsequently extended the analysis by adding data from the lower energy beaches of the eastern seaboard of the USA (New Jersey, North Carolina and Florida) (Figure 4). Complex variations in sediment parameters (grain size and sorting) were evident across individual beach profiles and comparison between beaches was facilitated through the use of a 'reference point' (that part of the beach face subject to wave action at the approximate mid-tide level, defined relative to the preceding high and low water marks). This provided the consistency necessary to establish statistical relationships between beach variables. Shepard (1963) extended these relationships to material as large as 16 mm in diameter (medium gravel) and both Krumbein & Graybill (1965), for coarse sands, and McLean & Kirk (1969), for mixed sand – shingle beaches, later demonstrated the importance of sediment sorting as a control on percolation and beach slope. Nested within Bascom's (1951) broad-scale study was a more local study of Halfmoon Bay, California (Figure 4). The Bay is partially sheltered by a rocky headland and beach sediment size increases away from the headland, to the south. But incident wave energy also increases in this direction so it is clear that the beach slope is a function of not only sediment characteristics but of the spatial differences in wave energy levels.

Subsequent experimental wave tank (e.g. Rector 1954), field (e.g. Harrison 1970) and combined laboratory and field studies (e.g. King 1972) all pointed to a link between wave steepness and beach slope; this was expressed by Dalrymple and Thompson (1976) in a dimensionless variable incorporating deep-water wave height and period and the sediment settling velocity. However, doubts remained as to whether the influence of grain size is exerted directly, via sediment transport processes, or indirectly, through its control on beach permeability and slope (Komar 1998).

Cross-shore equilibria and the 'Bruun Rule'

The idea that the shoreface adjusts towards some kind of equilibrium profile originated in the work of Fenneman (1902), and although heavily critiqued when applied at the shelf scale (Dietz 1963), was more readily applied to beaches. In the late 1940s, from the fields of fluid mechanics and sedimentology respectively, Garbis Keulegan & Bill Krumbein (1949) argued that the theoretical stable cross-shore profile results when incident waves are modified in such a way as to dissipate all their energy without breaking. From this theorizing, the Danish engineer Per Brunn (1954) showed that natural sandy beach profiles, from the Danish North sea coast and Mission Bay, California, could be characterized by exponential curves. Further support was subsequently provided by Dean (1977) from a database of over 500 beach profiles from around the US coastline. The next stage in Bruun's analysis was to define upper and lower limits for the equilibrium beach profile (the latter being the 'depth of closure' for sediment movement under waves) and to devise a conceptual model of shoreline adjustment to a rise in mean sea level; this became commonly referred to as the 'Bruun Rule' (Bruun 1962; Figure 5). At the heart of this model is the idea of the maintenance of an equilibrium water depth, whereby the erosion of material from the shoreface is deposited immediately offshore. The equilibrium profile is thus maintained but translated upwards and shorewards; the shoreline recession is of the order of 100 times the rise in sea level. With the advent of concerns about future accelerated sea-level rise (see above), interest in this model was reinvigorated in the 1980s, offering as it did the prospect of projecting changes in shoreline position at future timesteps from an estimate of the likely rate of sea-level rise. Although widely applied (Dean 1991, Dean & Dalrymple 2002), including in modified form to saltmarsh (Phillips 1986) and soft-rock cliff (Bray & Hooke 1997) responses to sea-level rise, there were powerful critiques from Dubois (1992) and Pilkey et al. (1993) who drew attention to the strongly three-dimensional character of volumetric change on open coasts with strong geological controls to which an equilibrium profile-based model is less well suited. Furthermore, the equilibrium condition refers to a hypothetical long-term average profile achieved under a given wave climate and sedimentology (Schwartz 1982), one that rarely, if ever, exists in nature (Kriebel et al. 1991). The difficulty in validating the Bruun model, its neglect of geological constraints and lack of detail on the processes by which sediment is transferred from the beach berm or dune front to the submerged nearshore profile led to calls for its abandonment (notably by Cooper & Pilkey 2004). However, its core assumptions have underpinned more sophisticated simulations of shoreline evolution (e.g. by Cowell et al. 2004) and, coastal engineering, it has endured as a means of obtaining indicative semi-quantitative insights into shoreline recession due to sea-level rise (Ranasinghe et al. 2012; D'Anna et al. 2021).

Planform equilibria

The planform geometry of equilibrium beaches also began to attract interest from both geomorphologists and engineers from the 1960s. Lewis (1938) and Munk and Traylor (1947) had already highlighted the interaction between bathymetry, wave refraction and sediment movement, and a connection between wave refraction and beach planform had already been alluded to by, amongst others, Johnson (1956). The resulting shoreline planform equilibrium can either be static, in the case of a beach characterised by no net longshore sediment transport, or dynamic, where the morphology is adjusted towards a balance between sediment supply and sediment redistribution by waves (Tanner 1958). Given the complexity of the linkage between wave energy and longshore transport, early studies concentrated on the geometry of static equilibrium forms. In particular, there was much interest in headland-bay beaches, where the presence of a rocky headland interrupts the longshore transport of sediment. In such settings, the beach has a seaward-concave plan shape in which the radius of plan curvature becomes greater with increasing distance from the headland; it was Krumbein (1944), again at Halfmoon Bay, who first noticed this characteristic as being similar to the increasing radius of curvature found in a logarithmic spiral. This idea was subsequently formally tested by Yasso (1965) and extended and given a more mechanistic foundation by LeBlond (1979). LeBlond combined analytical expressions for refraction and diffraction with known empirical relationships between wave energy, grain size and beach slope as originally elucidated at Halfmoon Bay by Bascom (1951). The logarithmic spiral model of equilibrium bay planform found application as an engineering tool (most notably by Silvester 1970 and Silvester & Ho, 1972). Other studies found that a parabolic model gave a better fit to observed bay planforms; this was first noted by Mashima (1961) who applied it to the analysis of Japanese bay beaches. The parabolic model was developed and applied more widely by Hsu et al. (1989) and Tan & Chiew (1994).

A final example of a static equilibrium form is that of tidal inlets. In 1969, Mike O'Brien revisited an earlier analysis (O'Brien, 1931) that showed the mouth cross-sectional area of a geographically extensive sample of inlets to be directly related to the tidal prism. Furthermore, this relationship could be expressed by a simple power function for artificially stabilised inlets but with a linear area-prism relationship for unconstrained entrances (O'Brien 1969; Figure 6). Since this time, similar analyses have been undertaken for a wide range of tidal inlets worldwide, including lagoon entrances and estuary mouths (e.g. Jarrett 1976, Townend 2005). The envelope of variability is large, in part because of the imprecision in much of the underlying data, but also because of the complicating effect of tidal hydrodynamics in very shallow systems (e.g. Gao & Collins 1994) and littoral sediment transport in the case of more exposed inlets (Bruun & Gerritsen 1966). As an engineering 'rule of thumb', the idea of an equilibrium inlet area-prism relation has enjoyed widespread appeal and has provided the basis for geomorphological evaluation of tidal wetland restoration schemes (e.g. Williams et al. 2002). Of course, the derivation of empirical relationships of this kind is heavily dependent on the sample systems being close to equilibrium and free to adjust their morphology. Inlets that show strong geological control typically exhibit much weaker correlation, as has been noted for some Japanese inlets (Shigemura, 1980).

Hydrodynamics and mechanics of sediment transport

> *'Well, I've often preached that scales are very important and when you can scale something then of course you have a chance of beginning to duplicate it. If you can't scale it properly then you don't. And most hydraulic modelers don't give sufficient thought to the scaling problems so that mostly they're measuring something but it may or may not have a reality in what's out there. … But if you can scale, and we did scale, and Guza's thesis was on the generation of beach cusps, for example. Tony Bowen's was on the circulation around rip currents. Then if you can scale them and work up from there it gives you insight. And then you go out in the field and measure them in the field, and you begin to know how much interaction and reality there is between the two. And in seeking that reality and the change in scales you get a much better insight into the phenomena.'* (Inman to L. Harkewicz, La Jolla, 16 May 2006).

Into the 1960s, coastal geomorphology benefited increasingly from fundamental research on hydrodynamics and the mechanics of sediment movement. French (2004) identifies the development of 'rational' approaches in this period, whereby theoretical models of coastal processes were combined with quantitative field measurements to yield insights beyond those afforded by observation alone. One important framework was formalised by Michael Longuet-Higgins, 'one of the towering figures of theoretical fluid dynamics in the twentieth century' (Sajjadi & Hunt 2018, 251), who had been a member of Group W (waves) at the Admiralty Research Laboratory, Teddington, UK under the leadership of the oceanographer George Deacon; the group was formed on the day before the D-Day landings with a view to forecasting ocean swell for future military operations in the Pacific (Longuet-Higgins 2010a). Longuet-Higgins had used the wave flume in the Department of Geography, Cambridge University to study pressure variations under both standing and reflected waves, allowing him to calculate the amount of energy reflected from different kinds of beaches and submerged structures (Longuet-Higgins 2010b). In the summer of 1956, he had accompanied W. Vaughan Lewis to measure sediment transport and beach cusp formation on the great gravel barrier of Chesil Beach, Dorset, UK (Longuet-Higgins & Parkin 1962). Whilst subsequently at the UK National Institute of Oceanography, Longuet Higgins was a visitor at the University of California San Diego in 1961-62 and 1966-67 (having previously spent part of the academic year 1951-52 at Scripps with Walter Munk). In a series of papers in the 1960s, following a visit to Vancouver, Longuet-Higgins and the Canadian Bob Stewart, who had studied turbulence at Cambridge University under the applied mathematician G.K. Batchelor, theorized on the excess flux of momentum due to the presence of waves – or radiation stress – in onshore-offshore, longshore and oblique directions (e.g. Longuet-Higgins & Stewart 1964). This theoretical model predicts a lowering of water level, or 'set-down', outside the surf zone and a rise in level, or 'set-up', shoreward of the breakers. In 1961, Inman was awarded a Guggenheim Fellowship, which he used to spend time at the Hydraulics Research Station, renewing his acquaintance with R.A. Bagnold (who had helped to set up the Station) and using the Wallingford UK facility as the template to establish, with Roger Revelle, the SIO Hydraulics Laboratory (Figure 8), opened in 1964. This experimental facility allowed Longuet-Higgins' work on radiation stress to be taken forward in California by Tony Bowen; Bowen had been at the Hydraulics Research Station in England before moving to La Jolla to undertake research on shoreface circulation and rip currents, supervised by Inman (his first PhD student; thesis completed in 1967).

Bowen et al. (1968) generalised earlier Scripps observations (Shepard & Inman 1950) within this theoretical framework, demonstrating a remarkable fit between theory and experimental results. Bowen (1969) was able to demonstrate that as higher waves break in deeper water than lower waves, the degree of set-down and, crucially, the set-up inside the surf zone is greater shoreward of larger breakers, thereby generating a longshore pressure gradient and current, augmented by the additional longshore component of radiation stress (Figure 7). The result is a cell circulation system landward of the breakers. Bowen & Inman (1969) were also able to show that 'regular systems of rip currents' at El Moreno, Gulf of California, where, unlike at the Scripps beach, there was no bathymetric control or variation in wave fields to explain the interaction between incident waves and edge waves. The latter are shore-normal standing waves, first investigated theoretically in California by the distinguished physicist Carl Ekhart and subsequently in the UK by the wartime refugee Fritz Ursell in the 1950s (Ursell 1952). A monochromatic, unidirectional wave train transfers energy to edge waves of lower frequency through a weak, non-linear interaction. In the initial formulation the edge waves were assumed to be synchronous with the incident waves although subsequent measurements showed them to be sub-harmonic (with a period twice that of the incident waves) in nature (Guza & Davis 1974, Huntley & Bowen 1973).

Sub-harmonic edge wave control was also invoked by Bob Guza and Inman to explain the most common generation mechanism for beach cusps (although wave tank experiments also showed that cusps can be the product of synchronous edge waves). They further argued that edge wave amplitude decreases as the cusps grow such that a negative feedback mechanism limits cusp growth (Guza & Inman 1975). The body of work on cusps by Inman, Bowen and Guza provided a powerful set of arguments for the importance of edge waves in explaining cusp generation and spacing. In later years, however, a rather broader model was proposed, in which cusps simply reflect a form of self-organization (e.g. Werner & Fink 1993, Masselink et al. 1997) rather than them being initiated by incidence wave – edge wave interaction. Finally, the resonance between incident waves and edge waves requires high reflection of energy from the beach face and this is lost when the incoming waves break 'cleanly' on the beach.

Interest in the emergence of rhythmic shoreline morphology extended beyond beach cusps to a variety of larger-scale features. Crescentic bars had been extensively studied by the 1960s (King & Wiliams 1949, Hom-ma & Sonu 1963), and Bowen & Inman (1971) linked the formation of these to edge waves with long periods in the infragravity range. Holman & Bowen (1982) similarly invoked the superimposed edge waves with different modes to account for the formation of oblique 'transverse' and 'welded' bars. However, there was also much debate on whether such features are product of or a generator of edge waves (e.g. Kirby et al. 1981).

The tripartite classification of 'spilling', 'plunging' and 'surging' breakers first gained wide currency through Wiegel (1964). The term 'collapsing' breaker, intermediate between 'plunging' and 'surging' types, was introduced by Cyril Galvin. From a frame-by-frame analysis of films of over 1400 waves in the 29.3 m CERC wave flume between September 1963 and August 1965, Galvin was able to demonstrate that breaker transitions relate to wave and beach characteristics and formalised these relationships in a breaker coefficient: 'breaker type goes from spilling to plunging to collapsing and surging as the breaker steepness decreases or the beach slope increases' (Galvin 1968, 3657). Guza & Inman (1975) began to explore a 'reflectivity parameter' and captured this transition through a 'surf zone scaling factor' that could be related to the way in which shoaling waves break in shallow water. Importantly, they were subsequently able to map wave type transitions onto thresholds in the value of the surf zone scaling factor, which was subsequently followed by a range of similar parameters that can be used to discriminate between different breaker types (Woodroffe 2003, Table 6.1). This research in turn laid the building blocks for extension of these ideas to the morphodynamics of beaches – as we detail below.

Form – process relations: the road to coastal morphodynamics

At around the same time as the science of coastal hydrodynamics and sediment transport was developing on the US west coast, research on the US east coast was beginning to explore the quantitative relationships between nearshore processes and beach morphology. Whilst general 'winter' v. 'summer' profiles had been quantified by W. Vaughan Lewis, Shepard and Cuchlaine King (Plate 2) amongst others in the post-WWII period, more detailed field studies of beach dynamics only began to appear in the 1960s.

Much of this research was associated with the Coastal Studies Institute (CSI) at Louisiana State University. In February - early March 1962 (before the field experiment was destroyed by the Ash Wednesday storm of 5-9 March), using a combination of traditional field survey techniques and more innovative repeat photography on the Outer Banks, North Carolina, CSI's Bob Dolan (Plate 3) investigated the relationships between surf zone processes (wave height, period and direction and still water level) and beach characteristics. He demonstrated that 'large waves with high water levels cause rapid reductions in beach thickness, width, and slope. Small waves with low water levels are associated with thicker, wider and steeper beaches' (Dolan 1966, 699). The experiment was re-instated in December 1963 and ran until May 1964, by which time over 1,300 beach profiles had been recorded, at a semi-diurnal interval. Analysis was taken to the next level by the North Korean-born, South-Korean and Japanese-trained coastal engineer at CSI, Choule Sonu (Short & Jackson 2013). Sonu & van Beek (1971) sub-sampled 291 profiles from the extensive Core Banks dataset for further analysis, generating a beach state model with six characteristic beach cross-sections. The cross-sections were linked by accretionary or erosional pathways; looping through all stages – a beach cycle – took place over a period of four months. Subsequently, Sonu & James demonstrated that the time history of beach geometry could be described as a first-order Markovian process, concluding that 'beach profiles do not alter in a completely random fashion. The time series of beach configurations depends on its own history as well as the time variation in causal processes' (Sonu & James 1973, 1471).

Sonu had already begun speculating on what we would now call nearshore 'hierarchical rhythmic topography' whilst at the University of Tokyo. Analysis of repeat aerial photography of the entire Japanese coastline was used to investigate the dynamics of longshore bars and their relation to shoreline dynamics (Hom-ma & Sonu 1965). Once in the USA, he took these ideas forward both in the continued field campaign on the Outer Banks in 1963-1964 (Sonu & Russell 1966) and in new investigations in 1967 and 1968 at Seagrove, Gulf coast of Florida. In this latter area of complex bottom topography, variations in breaker type over bars and troughs, and pulsatory flows, varied with changes in tidal level, even under a microtidal regime (for an illustration of the data collection technology see Plate 4). In an extension of the US west coast research, Sonu (1972) started to explore the feedback loops between processes and topography, demonstrating that cell circulation can redistribute beach sediments and create troughs and bars, which in turn then influence the cell circulation, constraining its response to changes in incident waves and edge waves. These approaches were further developed by Brian Greenwood and Robin Davidson-Arnott on the New Brunswick coast of Atlantic Canada, adding the ability, through box coring, to link three-dimensional morphologic change in underwater bars to associated fluid and sediment dynamics (Greenwood & Davidson-Arnott 1979).

In June-July 1971, asking 'how do waves, currents and beach topography respond to the excitations produced by a diurnal land-breeze system?', Sonu extended the research at Seagrove through Project SALIS (Sea-Air-Land-Interaction-System). The expanded CSI team included the oceanographers Stephen Murray and Joe Suhayda, and the meteorologist S.A. Hsu. Additional technical support came from Norwood Rector, a former underwater demolitions expert with the US Navy who had first encountered the CSI staff while running a sports diving operation on the Outer Banks. The enlarged team deployed an extraordinary range of air, water level and wave and swash sensors. To establish synoptic current patterns, injected dye trajectories were obtained from repeat aerial photography, a radio-controlled camera being operated from a helium balloon tethered 120 m above the surf zone (Sonu et al. 1973; Plate 5). The insights and expertise gained from these operations was critical to the scientific advances on Australian beaches that followed.

Beach and Barrier Morphodynamics

The catalyst for the so-called 'Australian' school of beach research can be traced back to the University of Sydney where, under the supervision of Trevor Langford-Smith, a number of students, including Bruce Thom, Don Wright and Andy Short, began more systematic nearshore investigations. All three went on to complete PhDs on coastal topics in the Coastal Studies Institute at LSU; both Wright and Short (briefly) participated in Sonu's field campaigns at Seagrove. After Florida, and now joined by the physical oceanographer Bill Wiseman, CSI re-located research operations to the north slope of Alaska (Short 2020a). The approach was as follows:

> *'At the start of the Alaskan project, Coleman [Jim Coleman] covered an entire wall of CSI with a montage of 1:24,000 topographic maps of the entire north coast of Alaska, this was the field site. Next, Short oversaw the digitising of the entire coastal bathymetry and running of a wave refraction program across the bathymetry to calculate breaker wave power along the coast. This was all laboriously undertaken with rather primitive digitising and computer technology. The wave dynamics were then combined with coastal morphology to divide the coast into 'morpho' 'dynamic' segments. This was coastal geography on a grand scale…'* (Short 2012, 142).

Thom returned to Australia in 1971, to ANU, and initial beach monitoring campaigns there centred on beaches embayed between rocky headlands south of Sydney. Roger McLean and Thom established the Bengello (Moruya) Beach surveys in 1972 (Plate 6). The monitoring programme captured a period of severe beach erosion and foredune scarping (Thom 2016) from the impact of a succession of extratropical cyclones between May and June 1974 and this was followed by further storms up to 1978; thereafter the beach started to recover, returning to its previous state by 1983 (Thom & Hall 1991, McLean & Shen 2006). Recent morphostratigraphic research at this location has demonstrated that the 1974 event, whilst unusual on an historical timescale, can be seen as a recurrent phenomenon over millennial timescales (Tamura et al. 2019). Short's surveys at Narrabeen Beach, north Sydney followed in 1976 (Short 1979; Short completed 335 surveys up to August 2006 (Short 2020b) when continuation of the programme was taken over by Ian Turner (Turner et al. 2016)).

Wright and Thom (with Langford Smith) established the Coastal Studies Unit (CSU) in the Department of Geography, University of Sydney in 1976, with Short joining, from Macquarie University, in 1977; this was the point at which beach process studies, initially at Bengello, were grafted onto the monitoring programme, with the Unit's manifesto being the broader morphodynamic framework as outlined by Wright & Thom (1977). Short has outlined the data collection challenges:

> *'the pods [home-made concrete structures to hold the surf zone sensors] moving, summersaulting [sic] and/or being buried, the current meters clogging, and the wires tangling or breaking; while on the beach the chart recorders would run out of ink and the computer [originally a Tektronix 4051 with just 13 kb of RAM] malfunction, run out of memory or overheat'*. (Short 2016, 2).

The great merit of this research was that it fused the theoretical and laboratory approaches emanating from Scripps (with direct collaboration with Guza on sabbatical in Sydney in 1980; see Wright et al. 1982; at the time, Guza was not convinced that 'morphodynamics' was a real word and persuaded Wright and Short to drop 'morpho' from the paper title (Wright to Spencer, pers. comm., 28 September 2021)) with the larger-scale process-form model, and concomitant field instrumentation, developed in Louisiana. In particular, the ideas of 'reflective' and dissipative' energy environments and the transitions between them as exemplified by the surf scaling factor (Guza & Inman 1975) were then applied at the 'whole beach' scale (Figure 8).

The early instrumentation was deployed on reflective, intermediate or low-energy dissipative beaches (Wright et al. 1979) but by 1980 field campaigns had been extended to high energy dissipative beaches (e.g. Goolwa Beach, South Australia; Wright et al. 1982, Wright & Short 1984; Plate 7) and then extended to the entire Australian coast (Short 2020c). At the same time, Pat Hesp extended this approach to the morphodynamics of foredune systems and, in a more quantitative manner than previously, established, with Short, a model of wave – beach – dune interaction (Short & Hesp 1982); these studies were further advanced at a symposium of the IGU Coastal Commission / Association of American Geographers held in Portland, Oregon in April 1987 (Psuty, 1988). And as the early micro- to meso-tidal beach surveys were extended to macro-tidal systems, the wave-dominated model was then followed by a tide-modified model (Masselink & Short 1993). And in an extension for gravel barriers, Orford et al. (1996) presented a morphodynamic conceptual model in which transitions between distinctive 'domains' of barrier process and morphology may be separated by episodes of instability and abrupt transition rather than predictable evolutionary progression.

Delta Morphodynamics

In the exploration of potential hydrocarbon reservoirs, the post-WWII period saw a growing dissatisfaction with a solely stratigraphic approach as the basis for the interpretation of fossil delta sedimentary sequences. Using 'rational lines of reasoning', Charlie Bates put forward an entirely new line of argument for understanding patterns of sedimentation in developing delta systems. After WWII, where he was involved in the prediction of wave heights for the D-Day landings, Bates worked both as an oceanographer in the US Navy Hydrographic Office and for a private company providing marine climate forecasts for the offshore oil industry, then taking a year's sabbatical to complete a PhD on the 'Rational theory of delta formation', the same title as his benchmark 1953 paper (Bates 1953). The work was heavily influenced by advances in the understanding of turbulent flow phenomena, especially theoretical constructs on the dispersion of 'free jets' into still water bodies (e.g. Albertson et al. 1950)). Bates introduced the modifications of a rotating fluid system and the addition of a hydraulic head at the orifice. Various combinations of jet and basin geometry were identified and patterns of idealised sediment distribution predicted. A three-way classification of delta-types was envisaged, based on the distinction between a circular or slot-like orifice and the density contrast between the effluent and receiving basin water. In the 1960s and 1970s, CSI began to revisit these ideas, at first in relation to river mouth processes in the Mississippi Delta (e.g. Wright & Coleman 1971) but then more widely within their database of deltas worldwide (Wright and Coleman 1973). This work built on the classic study of the Ganges-Brahmaputra delta by Jim Morgan and Bill 'Mac' McIntire (1959) from fieldwork in 1955- 1956 (with A.I.H. Rizvi, University of Dacca), Coleman's early field visits in 1964 to the Burdekin Delta, Australia and the Klang Delta, Malaysia (Thom 2020b), supplemented by process-based field research in, amongst many other cases, the joint ANU – LSU project on the Cambridge Gulf - Ord River delta (Wright et al. 1973). In a masterful synthesis, Wright (1977, 857) argued that Bates' 'primary forces and their depositional products are modified to varying degrees by tides and waves'.

Thereafter in the Mississippi Delta Plain, alongside research into the emerging Atchafalaya delta (e.g. Roberts et al. 1980), interest moved to the outer slopes of the birdsfoot delta. In August 1969, three offshore platforms were damaged during Hurricane Camille; it was found that the damage was caused by submarine landslides. From the late 1970s, improved side-scan sonar technology allied to precision positioning and sub-bottom profiling allowed the preparation of delta front mosaics detailing not just landslides but a wide variety of mass movement modes (including the previously described 'mudlumps' of diapiric origin (Morgan et al. 1968). Furthermore, this remarkable imagery required a complete conceptual re-think of the traditional model of slow (of the order of a few mm a^{-1}), continuous delta front progradation by the settling of suspended sediments towards models of delta front growth by bulk sediment transport (Kenyon & Turcotte 1985). In the Mississippi Delta, the slides and mudflows are extremely active with movement rates of several hundred metres per year; more than 30 m of sediment has been deposited by these mechanisms since 1875 (e.g. Prior & Coleman 1982). Traditional slope stability analysis suggests that such movements on very low angle (0.5 – 1.7^0) slopes cannot be explained by gravity flows alone. Rather, movement is facilitated by the reduction of sediment strength properties by large internal pore water pressures, generated by rapid sedimentation at distributary mouths, loading under hurricane and storm-generated waves and the internal generation of biogenic methane gas (e.g. Prior & Suhayda 1979).

Coral Reef Morphodynamics

Few studies before the 1970s looked at reefs and reef-associated communities in the context of the energy fields to which they are exposed and which they themselves modify – a classic set of morphodynamic relations. However, Harry Roberts and co-researchers at CSI were able to quantify the energy attenuation efficiency of Caribbean coral reefs, showing it to be an order of magnitude greater than that expected for a similarly sloping sand shelf of comparable width. It was shown that 75% of this attenuation comes from the passage of flows over, and through, the shallow reef crest but a further 25% comes from the topographic roughness of the shelf itself, often expressed in the form of shore-normal coral ridges separated by sand filled gullies (Roberts et al. 1975). The 'natural breakwater' role of such fore-reef structures had first been described at Bikini Atoll, Marshall Islands, Central Pacific Ocean by Munk and Sargent (1954); in taking these ideas further, Roberts (1974) demonstrated that the spatial variability and geometry of both shallow and deep terraces on the shelf encircling the island of Grand Cayman, Caribbean Sea could be explained by the nearly two orders of magnitude difference in wave power between eastern, windward and western, leeward reefs (Figure 9). Roberts went on to show that the balance between wave forces and current forces varies across the shelf with the highest wave forces experienced at the shallow reef crest and the strongest current flows along the shelf edge; these differences are reflected in coral growth forms, with massively branched, bladed and encrusting corals at the reef crest and delicately-branched, platy and hemispherical corals being found on the deep shelf margin (Roberts et al. 1977).

Morphostratigraphy

Research on coastal evolution over very long timescales was transformed in the 1970s by a revolution in basin stratigraphy whereby a completely new data source, in the form of high-resolution seismic records, allowed the identification of onlap/offlap sedimentary sequences on passive continental margins and thence the generation of a global eustatic record. This was the so-called 'Vail curve', developed by Peter Vail and colleagues at Exxon (e.g. Vail et al. 1977), and subsequently revised for the post-Triassic period by the Pakistani-American oceanographer Bilal Haq (e.g. Haq et al. 1988). Whilst the long, first-order sea level cycles (10^7-10^8 timescales) have been widely accepted, the amplitude and frequency of the third order and higher sea level fluctuations (10^6 – 10^3) timescales have been highly contested, not least because independent checking of commercially sensitive data has not been possible (for a Kuhnian analysis see Miall & Miall 2001). For the Pleistocene epoch, Spencer et al. (2008) documented the realisation from the late 1960s that the oxygen isotope record from deep sea core could be read, in part, as a record of palaeo ice volumes. With core dating utilising the palaeomagnetic timescale, this allowed for the development of much more precise Pleistocene sea level histories. These histories were then corroborated against dated sequences of raised reef limestones from Barbados, New Guinea, Timor and Atauro and the Ryukyu Islands, Japan from which assumed monotonic rates of uplift had been removed (Spencer et al. 2008). And for the Holocene epoch, sea level science went from a position of assuming that there was one 'right' global sea level (and therefore by implication that some curves were 'wrong') to embracing a diverse set of sea level histories resulting from the 'space-and time variant contribution of eustatic and isostatic processes over a deformable Earth' (Spencer et al. 2008, 904; Walcott 1972, Clark et al. 1978). This development was important for coastal geomorphology, not least because it showed that on some coasts present sea level had been reached at around 6 k yr BP (e.g. Australia) but in other locations had only been attained in the last 1 k yr BP with all that this implies for coastal landform adjustments to wave and tidal forcing and sediment supply. Within this context, we now explore a series of case studies concerned with the synthesis of stratigraphic and morphological information, or morphostratigraphy. These studies were aided enormously by the application of motorised auguring and vibrocoring technologies, allied to particle size analysis, mineralogical and faunal studies and extensive radiocarbon dating of core materials recovered from shoreline barriers and reefs, and seismic profiling and improved bathymetric mapping on adjacent shelf areas. It was not until the 1990s that high resolution ground-penetrating radar began to reveal further complexity in barrier substrates and sedimentary body geometries (e.g. Fitzgerald et al. 1992).

Coastal sandy barriers

For the postglacial transgression, depositional sequences in coastal sandy barriers take two basic forms, transgressive and regressive (Kraft & John 1979), each mode representing a balance between three fundamental drivers: the rate of relative sea level change, the rate of sediment influx and the rate of dissipation of wave and tidal energy (Thom 1984). To these controls we might add underlying basement topographies - and human interventions. The relations between the first two drivers, and thus the possibility of switches between modes, was elegantly captured by Joe Curray (1964). Classic transgressive sequences from the eastern seaboard of the USA were described by Chris Kraft (1971); type site regressive sequences from Galveston Island, Gulf Coast (Bernard and Le Blanc 1965) and Nayarit, Mexico (Curray et al. 1970) and complex transgressive – regressive sequences from The Netherlands coast by L.M.J.U. Van Straaten (1965). Drilling of coastal sandy barriers in Australia was initiated by Bruce Thom in the 1970s; this drew upon his PhD research at LSU on the evolution of the barrier shoreline of South Carolina (synthesised in DuBar et al. 1974) and his introduction to drilling techniques and morphostratigraphy by the South Carolina State Geologist Henry Johnson (Thom 2020a; the term 'morphostratigraphy' was borrowed from Frye and Willman (1962) (Thom to Spencer, pers. comm. 30 September 2021)). Thom's research showed that on the wave-dominated coast of eastern Australia, where present sea level was reached at ca. 6.5 k yr BP, barriers in bedrock embayments exhibit a range of prograded, stationary, receded and episodic transgressive dune types (Thom 1984).

Barrier Islands

Many sandy coasts barriers are characterised by barrier Islands. As coring of barrier island sequences started to become commonplace in the 1960s, alongside reconstructions of regional sea level histories, so earlier ideas on the origin of barrier islands started to be once again debated. The theories of WM Davis and DW Johnson, that barrier islands were the product of the buildup of offshore bars, and that of GK Gilbert, who advocated the segmentation of spits into inlet-separated islands, were challenged in an elegant paper by John Hoyt (Hoyt 1967), who argued for the 'engulfed beach ridge' hypothesis. This provided stratigraphic corroboration for ideas developed in WJ McGee's 'submergence theory' of the 1890s (McGee 1890) and provided convincing evidence that back-barrier lagoons are not floored by the open water sediments that would be expected if barriers had originated by either of the other two theories. Furthermore, Hoyt showed how lagoon widths and depths might be related to antecedent topography, the slope of the coastal plain and the degree of land submergence and sea-level rise (Hoyt, 1967). Hoyt's paper generated considerable debate, with re-examinations of earlier ideas and the introduction of several false leads (Cooke 1968, Fisher 1968, Hoyt 1968, Otvos 1970). If barriers are largely seen as being young, transgressive structures formed under the general driver of Holocene sea-level rise, then the rate of onshore migration has clearly been determined by Hoyt's three broad controls. Kraft (1971) showed that sediments beneath a transgressive barrier are re-worked when sea-level rise rates are relatively low. Overwash then becomes important as a mechanism for transferring sediment from the shoreface to back-barrier environments, a process that, at the barrier scale, is termed 'barrier rollover' or 'shoreface retreat' (Swift 1968). However, Field and Duane (1976) identified a series of linear sand shoals on the inner shelf of the US eastern seaboard that might be interpreted as relict barriers. On the inner shelf of Long Island, offshore from the Fire Island barrier, Sanders & Kumar (1975) and Rampino & Sanders (1980) described the presence of back-barrier sands, dated to between 7 and 3 K yr BP, and in some locations preserving the entire sedimentary sequence up to 8 m thick; they concluded that 'it is difficult to reconcile the preservation of the deposits with the idea that the shoreface has retreated by continual erosion' (Rampino & Sanders 1980, 1071). Their alternative hypothesis was one of 'in-place drowning' of the barrier and landward 'overstepping' of the surf zone to the landward edge of the former lagoon. This model generated considerable debate and rejoinders (e.g. Swift & Moslow 1982, Leatherman 1983a, 1983b, Rampino & Sanders 1982, Otvos 1986). Fundamentally, the arguments revolved around the question as to the threshold at which a barrier loses its integrity, of considerable significance on barriers accommodating large resident human populations and important infrastructure.

Deltas

The 'delta cycle' applies the idea of a fluvially dominated regressive phase and a marine-dominated transgressive phase to delta growth and decay (Roberts 1997). Whilst is it widely recognized that the fluvially-dominated Mississippi Delta is atypical, and the modern bird's foot delta atypical even within this regional setting, the remarkable large-scale, detailed sedimentary archive established through the juxtaposition of hydrocarbon exploration and university-based field research provided one of the most compelling stories of coastal geomorphology in the second half of the twentieth century. The delta landform complex also provides a prime example of the nested space-time hierarchies that are so characteristic of coastal morphodynamics more generally.

Richard Russell came to Louisiana State University (LSU) in 1928, at the establishment of the Department of Geography and Anthropology. Prior to that, at Berkeley, he had been under the tutelage of Carl Sauer and had met with WM Davis. It was Davis who issued Russell with the challenge 'try and find out why the Mississippi River follows such a straight course below New Orleans' (Anderson 1975). In 1928, serious geologic studies of the Mississippi Delta were becoming possible through the availability of new topographic quadrangle maps and aerial photographs. Davis told Russell that the subdued topography exhibited by the Lower Mississippi maps was a classic example of an old age landscape but of course the exact opposite turned out to be true (R.J. Chorley to Spencer, pers. comm., Cambridge 2 July 1999). Walker (2008) has detailed how Russell and Hal Fisk established the three-dimensional structure of the Mississippi delta, which was then subsequently elaborated upon by these authors and others in the 1950s and 1960s; Coleman (1988) provides an extensive bibliography. It was Russell (1940) who established the linkage between abandoned Mississippi River courses in the alluvial valley and the deltas they built further downstream. From this emerged the realization that the delta plain was composed of a series of offset and overlapping delta complexes, with the pattern of individual complex growth and decay described by the term 'delta switching'. A rough initial chronology was worked out through archaeological relationships (McIntire 1954), culminating in the iconic map of Kolb and Van Lopik (1958). Subsequently, with the advent of radiocarbon dating, the record was refined initially by McFarlan (1961), then by Kolb and Van Lopik (1966) and, most notably, by David Frazier (1967). In his classic paper, which to this day provides the stratigraphic framework for understanding the history of the Delta, Frazier, using detailed facies analysis from hundreds of shallow borings calibrated by over 100 radiocarbon dates, identified 16 separate delta lobes formed by the Mississippi River in the past 6,000 years. He then grouped these lobes into a series of five delta complexes: 14 lobes were ascribed to the Teche, St. Bernard, and Lafourche delta complexes; the remaining two include the present bird's foot delta, which is an extension of the older initial lobe of the Plaquemines-Modern complex. The complexes overlap in time; as one starts to shut down another starts up. An individual complex has a typical duration of 1000-2000 years, supports wetlands that covered up to 15,000 km^2 and develops sedimentary sequences up to 30 m thick on the inner shelf. Within the delta lobes within each complex, subdeltas are established at breaks in the natural levees of major distributaries. Coleman and Gagliano (1964) showed how stacks of overlapping subdeltas lead to the infill of the inter-distributary bays; individual subdeltas have sediment sequences typically < 10 m think, cover areas of 300 km^2 at time of maximum extension and typical have a lifespan from initiation to decay and return to open water of 150-200 years. Nested within the subdeltas are smaller levee break and overwash deposits; these are typically a few metres thick and survive for a few decades (Roberts 1997).

Whilst Frazier (1967) was able to describe the relations between the regressive and transgressive phases of delta evolution, it was not until the mid-1980s that Shea Penland, Ron Boyd and John Suter provided a detailed description – with an extensive use of historical maps and charts - of transgression in abandoned delta complexes and an evolutionary model to link the different stages of delta decay and loss. Penland et al.'s (1988) model has three stages: stage 1 involves the transformation of an abandoned delta complex into an erosional headland with flanking barriers; in stage 2, the submergence and separation of the stage 1 barrier shoreline from the mainland lead to the formation of a transgressive barrier island arc; and, in stage 3, with further submergence, the island arc sand body is reworked into a sand shoal on the shoreface and inner continental shelf (Figure 10). It is interesting to note that the flanking barriers produced in stage 1 illustrate the spit-segmentation barrier island model proposed by Gilbert (1885) and the separated barriers of stage 2 echo the submergence hypothesis for barrier island genesis advocated by Hoyt (1967). Finally, Penland et al. (1988) note that in an environment characterised by high rates of relative sea-level rise due to deltaic submergence, low gradient continental shelves with limited local sand sources, and a process environment dominated by episodic hurricanes and a micro-tidal regime, the sand shoal sediments of the transgressive phase become volumetrically significant, contributing up to 50 percent of the composite sequence thickness. Finally, Reed (2002) has shown how wetland geomorphic and vegetative processes evolve during both the regressive (active delta building) and transgressive (Stages 1–3) phases of the delta cycle.

Coral Reefs

Over very long timescales, the Vail-type oscillations in sea level were evident in the sedimentary record of slowly subsiding mid-oceanic carbonate platforms. Here, the Haq et al. (1988) interpretation appears to fit best, with the record of > 100m falls in sea level in the Early to Mid-Miocene and the high frequency oscillations of the Pliocene being particularly noteworthy (Wheeler & Aharon 1991). In a coral reef context, these more recent studies complemented the successful testing of Darwin's subsidence theory from deep drilling at Pacific Ocean atolls in the 1950s and the realization in the 1960s that the long-term subsidence record contained evidence for alternating periods of atoll emergence (calcitic limestones with karstified surfaces characterized by pocket soils containing terrestrial pollen and mollusca) and submergence (uncemented aragonitic sands) (Spencer et al. 2008). For the continental margin reefs of the Great Barrier Reef, interpretation of coralline material in bores retrieved from Michaelmas Cay (1926) and Heron island (1937) confused researchers for over 40 years (Chivas et al. 1990); it was not until the 1970s that similar unconformities to those seen in the Pacific atoll cores were recognised in the Great Barrier Reef cores (Davies 1974).

David Stoddart (1973; Plate 8) was one of the first coral reef scientists to grasp what the new sea level records for the Pleistocene might mean for the history of coral reefs, pointing out that the 'saw-toothed' sea level curves showed that sea level had been at its present level for all but 2000-3000 years, only some 0.1 per cent of the duration of the Pleistocene. Furthermore, early applications of Uranium/Thorium radiometric dating from the mid-1960s (see Spencer et al. 2008) revealed a sharp break – the 'Thurber Discontinuity'- between near-surface (typically < 20 m) reef limestone dates of clear Holocene age and dates in excess of 100 k yr BP at greater depth, the latter being broadly correlative with those of many raised reef limestones in all the major coral reef provinces. For Stoddart, there was 'a strong presumption that the topography of modern reefs is essentially an inherited karst landscape formed during the complexity of the Pleistocene and now veneered by recent coral growth' (Stoddart 1973, 321). Stoddart himself pointed out that karstic control might explain the spectacular 'blue holes' of the Caribbean, first described by Northrop in the 1890s from the Bahamas (the 140 m deep blue hole on Lighthouse Reef, Belize featured in Stoddart's own PhD thesis research (Stoddart 1962)), and the enclosed basins and bases of pinnacles and patch reefs within coral reef lagoons (e.g. Truk Atoll, Caroline Islands (Shepard 1970)). The conceptual ground for this interpretation had earlier been staked out by Japanese geomorphologists (synthesised by Tayama 1952) and F. Stearns MacNeil (1954), who argued that emergent limestone platforms would weather to produce a marginal rim and central depression. But these papers were largely ignored; they did not specify the solutional processes involved and the work pre-dated the classifications of tropical karst landscapes by German geomorphologists.

It was in this context, and with the advent of continuous seismic profiling, that Ed Purdy comprehensively argued that the patterning of reefs on the Belize Barrier Reef was karst-determined (Purdy, 1974). Seismic profiling, calibrated against local drill cores, revealed the pre-Holocene topography to be a solution etched, fault- and joint-controlled surface that could be divided into two regions. The northern shelf lagoon is generally less than 5 m deep and dominated by large dish-like depressions; Purdy interpreted this bathymetry as a drowned doline karst, similar to the dry tropical landscape of the northern Yucatan Peninsula. The southern and central lagoon is much deeper (to –27 m), studded with carbonate shoals and floored by muds. Purdy interpreted the outer barrier as a drowned cone karst, analogous to the 'cockpit country' of NW Jamaica, and the lagoon as a drowned karst marginal plain, with isolated limestone towers, the result of aggressive runoff from the non-carbonate Maya Mountains to the west. Crucially, these karst-determined morphologies were then subsequently accentuated by vertical postglacial reef growth on the platform highs, the accentuation preserved because of the thinness (generally less than 20 m) of the Holocene sequences.

Subsequently, by looking at the relations between platform depth and rates of sea-level rise, it was possible to place Purdy's Belizean case study in a wider suite of responses to Holocene sea-level rise (Hopley 1982). Platforms at depth experienced high rates of sea-level rise early in the Holocene transgression and were vulnerable to being 'drowned out' and then subsequently being at too great a depth for reef re-establishment when sea-level rise eventually slowed. This explained the largely coral-free carbonate platforms at > 30 m water depth in the Caribbean basin (Stoddart 1977). Surfaces at more intermediate depths were initially drowned but when sea level rise slowed were at depths where rapid, and accelerating, vertical growth towards modern sea level was still possible. Finally, basements very near sea level have only been flooded very recently at relatively low rates of sea-level rise such that reef accretion has been capable of filling the vertical accommodation space and then promoting lateral reef growth. These three different modes of reef responses to sea-level rise were termed 'give up', 'catch up' and 'keep up' in a paper delivered by Ian Macintyre during a tropical downpour under a palm-thatched roof at the 5th international Coral Reef Symposium in Tahiti in May 1985 and arrestingly captured in Conrad Neumann's cartoon (Figure 11; originally published in modified form in Neumann & Macintyre 1985). In the Caribbean Sea, Walter Adey & Randy Burke (1976) were able to define the transition between 'catch up' and 'give up' reefs by showing that platforms at -15 to – 18 m generally support reefs whereas surfaces at -20 to -25 m do not. The reefs on the Belize Barrier Reef fall within the 'catch up' category and their vertical growth explains the accentuation of the underlying basement topography. Conversely, the platform infilling achieved by 'keep up' reefs means that the signal from the underlying karstic basement is lost, although it may be revealed with seismic profiling (e.g. Nick Harvey (1977) on the Great Barrier Reef). It is perhaps not surprising that this infilling is particularly characteristic of Indo-Pacific reefs where sea level reached its current level ca. 6 k yr BP. The fusion of shallow reef drilling, seismic surveys and radiometric dating of reefal materials came to particular fruition on the Royal Society – Universities of Queensland Expedition to the northern Great Barrier Reef in 1973 (Stoddart 1978, Thom et al. 1978). With subsequent research by John Chappell and others, these studies showed that as a result of hydro-isostatic tilting of the shelf, inner reefs kept up with sea level during the postglacial transgression whereas shelf edge reefs showed catch up behaviour (Chappell et al. 1982). Finally, micro-erosion meter studies in both the Indian Ocean (Aldabra Atoll: Trudgill 1976) and the Caribbean Sea (Grand Cayman: Spencer 1985) have shown that rates of surface downwearing of emergent reef limestones are very low. The implication is that karst basements are likely to have been formed over multiple sea level cycles over the Pleistocene (Hopley 1982). Alternatively, underlying basements may have supported reef-derived sediments which may then have formed the foci for subsequent reef accumulations. The most dramatic example of such depositional (as opposed to erosional) control, albeit inferred from aerial photography, is the deltaic patterning to the ribbon reefs of the northern Great Barrier Reef (Maxwell 1970).

Coastal systems, modelling and management

We have already seen that, in parallel with geomorphology as whole, the 1960s saw a sustained shift from largely descriptive physiographic studies of coastal landforms to a more dynamic science founded not only on geology and stratigraphy but also driven by advances in oceanography and fundamental fluid mechanics. Thinking such as Ludwig von Bertalanffy's General Systems Theory (Chorley 1962; and its incorporation into new ideas on landform evolution (Hack, 1960) and fluvial geomorphology (Leopold et al., 1964)) became increasingly important as a means of reconciling new insights into the mechanics of sediment transport with the broad range of scales relevant to the understanding of landform evolution (Schumm & Lichty 1965, Cambers 1976, Brunsden & Jones 1980), and also provided an organising framework for a new generation of coastal geomorphology textbooks, most notably that by John Pethick (1984). As Barbara Kennedy (1997, 419) later recalled:

> *'it was scarcely surprising that the first reading of Schumm and Lichty produced - if not quite the impact of Chapman's* Homer *on John Keats - a sense of both relief and excitement. … the message was, it seemed, clear: you* could *(as I had suspected) have Wooldridge (or even Davis)* and *Strahler (or Gilbert): it all depended on the scale.'*

This was also the time of the emergence of the concept of 'integrated coastal zone management' and the engagement of the geomorphology community with coastal management and shoreline planning (e.g. King 1974, Hails 1977, Clark 1978). Indeed, by 1994, Karl Nordstrom (1994a, 508) was able to point out:

> *'Recent research on the geomorphology of developed coasts has provided ample evidence of divergences in the form, surface cover and rate of change of human-altered landforms relative to natural landforms but has largely ignored the conclusions that natural landscapes are a myth, that human agency is not an intrusion into the coastal environment so much as it is now a part of the coastal environment and that human-altered landscapes can and should be modelled as a generic system'.*

and to go on to argue for an entirely new field of research into the morphodynamics of 'developed coasts' (Nordstrom 1994b, 2000).

Early work on coastal sediment systems focused on sandy beaches and, in a classic study, Bowen & Inman (1966) quantified sources, fluxes and sinks of sand along 105 km of the Californian coast between Pismo Beach and St. Augustin. They derived sediment budgets for a series of littoral cells, defined with reference to prominent headlands and bays, with additional corroboration of catchment sediment sources and transport pathways from earlier analyses of the heavy mineral fractions of the beach sands by P.D. Trask (Trask 1952, 1955). The concept of the littoral cell was further developed and formalised by Inman & Frautschy (1966; Figure 12) and Davies (1974) and, in conjunction with sediment budgets, proved to be a useful basis for understanding the dominant drivers of spatial variation in shoreline accretion and erosion, notably by James Allen at Sandy Hook, New Jersey (Allen 1981) and by Doug Inman and Bob Dolan on North Carolina's Outer Banks (Inman & Dolan 1989).

From the 1960s onwards, efforts were made to simulate the complexities of coastal landform evolution although these invariably encountered the severe constraint of the limited computing resources available at the time. The ingenuity of some of the earliest coastal modellers is remarkable when viewed from a present-day perspective. The Spitsym model of recurved spit growth was developed at the University of Nottingham, England by Mike McCullagh and Cuchlaine King (McCullagh & King 1970; King & McCullagh 1971). Inspired by Hurst Castle spit on the south coast of England, Spitsym employed a novel probabilistic approach to determine the formation of spit features in the cells of an approximately 0.5 km resolution grid. The Fortran code occupied just 4 K 48-bit words of storage on the English Electric KDF9 computer at Nottingham, which weighed in at nearly 5 tonnes; a complete spit simulation had a pretty respectable execution time of less than a minute.

At a more fundamental level, the basic equation for the sediment mass balance of an arbitrary control volume had earlier been established through the pioneering studies of the Austrian meteorologist and geophysicist Felix Maria von Exner-Ewarten (Exner 1925, 1927), and made more accessible to the English-speaking world by Serge Leliavsky (1955). Sediment mass balance subsequently became central to a new generation of morphodynamic models developed for a variety of coastal landforms. In essence, these combined the Exner equation for sediment mass conservation with an estimation of the alongshore sediment transport rate based on wave and nearshore current conditions. Pelnard-Considere (1954) had already derived a simplified analytical solution for the combined sediment continuity and transport equations and the temporal evolution of an idealised shoreline subject to the blocking action of a groyne. An early application of this kind of model to real beaches was to understand the alongshore diffusion of beach nourishment material (e.g. LeMehaute & Soldate 1977). Similar analytical models were used to simulate the growth of idealised cuspate deltas (Pelnard-Considere 1954, Grijm 1964, Bakker & Edelmann 1964, Refaat et al. 1992).

Such analytical models quickly became quite complex even for highly idealised coastal configurations, and neglected much of the spatial and temporal variation in natural wave conditions and river sediment input. These constraints were eased with the development of numerical models that implemented finite difference solutions to the sediment continuity equation. The general approach here was to assume that the beach was comprised of a sequence of cross-shore profiles that were always in equilibrium and which migrated over time to give the evolution of the shoreline planform. Probably the first computer-based implementation of this so-called 'one-line' model was published by Price et al. (1972), who simulated the effects of a groyne constructed on a model beach fabricated in the experimental wave basin at the Hydraulics Research Station in England. Numerous similar studies followed, but Nick Kraus and colleagues were amongst the first to predict the shoreline planform evolution of a real beach, at Oarai on the Pacific coast of central Japan, using a one-line model (Kraus & Harikai 1983, Kraus et al. 1985). This model was subsequently coded in a more generalised form by Swedish coastal engineer Hans Hanson, and the GENESIS model (Hanson & Kraus 1989) became adopted by the US Army Corps of Engineers as their standard shoreline evolution model. A fundamental assumption of one-line models is that the beach profile does not change over time and is always at equilibrium. The exact form of the profile is not particularly relevant but an idealised function of the kind formulated by Bruun (1954) and Dean (1977) is used to determine the beach slope, which is a term in most alongshore transport equations. The shoreline planform thus evolves only under the influence of varying alongshore sediment transport. GENESIS and other one-line models have been widely used for engineering applications, such as predicting the performance of offshore detached breakwaters and beach nourishments (e.g. Hanson & Kraus 1991). However, the assumption of profile equilibrium and neglect of cross-shore fluxes prompted forceful criticism from coastal geologists, especially Orrin Pilkey at Duke University, USA (Pilkey et al. 1993, Thieler et al. 2000**).** Changes in cross-shore profile were already well documented, not least through the multi-year monitoring of beach profiles at Scripps Pier by Sheperd & Inman (1950), which highlighted the strongly seasonal variation between 'summer' and 'winter' profiles. Modelling profile responses, including the effects of individual storms, and the first models adopted a semi-empirical approach to simulate the movement of break-point bars and the berm; a good example is the widely used SBEACH (Larson & Kraus 1989). Later models, such as those developed by Hedegaard et al. (1991) and Nairn & Southgate (1993) incorporated more physically based representations of wave shoaling, breaking and decay and separate treatments for suspended and bedload sediment transport.

Sediment mass balance is also crucial to the evolution of depositional landforms, and the 1980s saw widespread concern over an apparent imbalance in many coastal wetlands between rates of vertical sediment accretion and sea-level rise. Comparison between measured sedimentation and relative sea level trends obtained from tide gauge records revealed vertical 'accretionary deficits' on the US Gulf coast (DeLaune et al. 1983), and these were highlighted as important drivers of coastal land loss, especially in Louisiana (Baumann et al. 1984, Penland et al. 1990). Geographically extensive studies along the US East Coast by Stevenson et al. (1986) showed a strong positive relationship between marsh accretion and tidal range, while in a northwest European context, French (1994) showed that both tidal range and background sediment supply are important in controlling the ability of tidal marshes to keep pace with sea-level rise. The adjustment of marsh elevations with respect to tidal levels and changing mean sea-level is a classic example of a morphodynamic system and, given the non-linearity of the feedbacks, modelling provides considerably greater insight than short-term observation. Probably the first attempt to model this system was by Ray Krone in the Department of Civil Engineering at Davis, California, who devised a simple but elegant box model to simulate the adjustment of historic marsh elevations in San Francisco Bay (Krone 1987). At around the same time in the UK, John Allen (1990) and Jon French (1991, 1993) independently developed models based on essentially the same approach and performed the first exploratory simulations of vertical marsh growth under various sediment supply, autocompaction and sea-level rise scenarios. These studies demonstrated that tide-dominated marshes have a degree of resilience to sea-level rise, but with potential for loss due to inundation above a critical rate of sea-level rise (French 1994).

In the latter part of the period considered here, developments in both the systematic mapping of shoreline change and coastal landform modelling started to feed into shoreline management programmes, partly in response to increasing concern over the prevalence of erosion on sandy beach, dune and soft rock cliff coasts around the world (e.g. Bird 1985, Williams et al. 1990). In the USA, following the recommendations of the National Research Council (NRC, 1990), and building on the pioneering work of Dolan et al. (1985) and Griggs and Savoy (1985), the USGS National Assessment of Shoreline Change has embarked on a region by region assessment of shoreline change, using standardized methodologies for the geo-referencing, digitization and DSAS analysis of historic maps ('T-sheets') and aerial photography and brought up to the present with LiDAR-derived coastal topography; the first of these reports was for the Gulf of Mexico, published in 2004 (Morton et al. 2004). At the sub-regional scale, the littoral cell concept, first proposed by Bowen and Inman (1966), has provided an overarching geomorphological framework. A prime example of its application at a national scale was the mapping of major cells and sub-cells around the entire coast of England and Wales by Motyka and Brampton (1993), followed by the development of a set of 42 Shoreline Management Plans for the sub-cells (MAFF 1995). Prediction of future coastal morphological change at this scale proved very difficult, however. A variety of scale hierarchies was proposed, which invariably emphasised a correlation between the space and time scales of coastal processes and landform morphodynamics (e.g. Kraus et al. 1991, Stive et al. 1991, Fenster et al. 1993). These conceptualisations all acknowledged the relatively complete grasp of fundamental processes at one end of the scale spectrum as well as the increasingly detailed appreciation of the geological context for Quaternary coastal landform evolution at the broadest scale; the difficulty lay in the inadequacy of both understanding and modelling capability at what Cowell & Thom (1994) termed the 'engineering scale' (and see Figure 1, this chapter). Terwindt & Battjes (1990) used the term 'Large Scale Coastal Behaviour' to refer to timescales measured in decades and spatial scales of the order of tens of kilometres, which they considered relevant to the management of beach and dune erosion on the Dutch coast. Reporting on a colloquium on this topic held in Amsterdam in July 1989 under the auspices of the Royal Dutch Academy of Arts and Sciences, they noted considerable pessimism from the coastal specialists present on the potential for large scale coastal behaviour models to be able to achieve any meaningful predictive skill. Much of the difficulty was seen to lie in the nonlinearity of interactions between hydrodynamics, sediment transport and morphology and the fact that morphological change typically results from small residual effects that emerge from much larger oscillations due to, *inter alia*, tides and seasonal variation in wave climate (De Vriend 1991).

Despite this initial pessimism, pragmatic ways forward soon emerged and De Vriend et al. (1993b) were able to present alternative strategies for simulation of LSCB based upon input reduction, model reduction and behaviour-oriented modelling. The first of these attempts aimed to reduce the complex variability of driving processes, such as tides and waves, to a single 'dominant' tide or of tide/wave condition. This drew upon work by Bernard Latteux at the Electricité de France, Laboratoire National d'Hydraulique (summarised in Latteux 1995) but had conceptual similarities with the idea of a 'morphologically significant event' of the kind envisaged much earlier by Wolman & Miller (1960). The second approach, model reduction, makes use of the difference in time scale between the hydrodynamics (which accounts for most of the computational effort) and the sediment bed response, typically by extending the number of morphological timesteps at which the hydrodynamic forcing is re-computed. Lastly, the behaviour-oriented approach pragmatically avoids the intractable representation of every known fine-scale process in favour of building a model that captures the essential morphodynamic behaviours that emerge at the scale of interest. Peter Cowell and co-workers at the University of Sydney led the way in this field with the Shoreface Translation Model (Cowell et al. 1992, 1995). This was founded on the assumption of sediment mass conservation and a set of geometric rules for shoreface and active sand-body morphology. Fine-scale dynamic processes were parameterised to define the geometry of the active cross-shore profile, which can be used to represent the kinematic behaviour of an entire coastal cell. The emergent properties of the Shoreface Translation Model were shown to provide new insights into the evolution of the southeastern Australian coast and shelf during the postglacial marine transgression. As French et al. (2016) have argued, matching the level of mechanistic understanding with the scale of the problem is fundamental to any modelling endeavour. Not surprisingly, therefore, the development of behavioural models that adopt a less reductionist, and more synthesist, approach has become a very active area of early 21st century coastal geomorphology.

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In memoriam: we dedicate this Chapter to David Ross Stoddart (1937-2014) one of the truly gifted and inspirational coastal geomorphologists of the second half of the twentieth century.

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Figure Captions

Figure 1: (a) Primary components involved in coastal morphodynamics. 'The feedback loop between form and process is responsible for fundamental complexity in coastal evolution' (Cowell & Thom 1994, 34); (b) Definition of spatial and temporal scales involved in coastal evolution, with typical classes of sedimentary features (from Cowell &Thom, 1994) (reproduced with permission of Cambridge University Press through PLSclear).

Figure 2: History of global eustatic sea-level rise projections, 1982 – 2007. Key to projections: (1) Hoffman et al (1983); (2) Hoffman et al. (1986); (3) Thomas (1987); (4) Warrick & Oerlemans (1990); (5) Wigley & Raper (1992); (6) Wigley & Raper (1993); (7) Warrick et al. (1996); (8) Church et al. (2001); (9) Bindoff et al. (2007).

Figure 3: Evolutionary classification of coastal environments. Following Galloway (1975), the three corners of the triangular prism correspond to depositional conditions dominated by fluvial, wave and tidal processes. The long axis of the prism represents relative time with reference to changes in sea level and sediment supply: transgressions and regressions are shown by movement toward the front and the rear of the prism respectively. During a sea-level cycle, a coastal area will track forward and backward through the prism by an amount determined by the interplay between the rate of relative sea-level change, the sedimentation rate and basin size. (1992). Reprinted with permission from Dalrymple, R.W., Zaitlin, B.A. & Boyd, R. (1992) Estuarine facies models: conceptual basis and stratigraphic implications. *Journal of Sedimentary Petrology* 62, 1130-1146*,* SEPM Society for Sedimentary Geology, © 1992.

Figure 4: Beach face slope as a function of the median grain size of beach sediment, including the classic dataset from Half Moon Bay, California (Bascom 1951). Note difference between high energy (US West Coast) and low energy (US East Coast) beaches, indicating the importance of wave energy on beach slope. From Komar (1998; modified from Wiegel, R.M. (1964) *Oceanographical Engineering* by kind permission of John Wiegel and Bascom, W.N. (1951) The relationship between sand size and beach face slope, Eos Trans. AGU, 32(6), 866–874, https://doi.org/10.1029/TR032i006p00866. Reprinted with permission of Eos and the American Geophysical Union.

Figure 5: Influence of sea-level rise on the development of the beach and offshore profile. Reprinted with permission from Bruun, P. (1962) Sea level rise as a cause of shore erosion. *American Society of Civil Engineers, Proceedings and Journal, Waterways and Harbors Division* 88, 117-30, ASCE Press, © 1962.

Figure 6: Equilibrium form of 28 tidal inlets, Atlantic, Pacific and Gulf coasts of the USA. Minimum flow cross section of the entrance channel (throat) measured below mean sea level (square feet) and the tidal prism corresponding to the diurnal or spring range of tide (cubic feet). Key: o = inlets with no jetty; Δ = Inlets with 1 jetty; x = inlets with 2 jetties. Reprinted with permission from O'Brien, M.P. (1969) Equilibrium flow areas of inlets on sandy coasts. *Journal of the Waterways and Harbours Division, American Society of Civil Engineers* 95, 43-52, ASCE Press, © 1969.

Figure 7: Wave set-down and set-up for two different wave breaker heights, laboratory flume study, Scripps Institution of Oceanography. The larger wave height produces greater set down; the water slope of the set-up is similar for both wave heights but water level is higher shoreward of the larger

breakers since set-up is initiated further offshore. From Bowen, A.J. (1969) Rip currents 1. Theoretical investigations. *Journal of Geophysical Research* 74, 5457-5478. © 1969, Reprinted by permission of John Wiley and Sons.

Figure 8: Six major morphologic types (or states) of high-energy dissipative beach / inshore systems (Wright et al. 1979). Reprinted from Marine Geology, Volume 32, L.D. Wright, J. Chappell, B.G. Thom, M.P. Bradshaw and P.Cowell, Morphodynamics of reflective and dissipative beach and inshore systems: Southeastern Australia, 105-140, © (1979), with permission from Elsevier.

Figure 9: (a): Yearly mean shore wave power distribution (in ergs per second per metre of shoreline) for 68 sectors, Grand Cayman Island, West Indies. Note two orders of magnitude difference between windward east coasts and leeward west coast. Reprinted with permission of the first author and the Journal of Marine Research, Sears Foundation (Yale Peabody Museum) from Roberts, H.H., Murray, S.P. & Suhayda, J.N. (1975) Physical processes in a fringing reef system. *Journal of Marine Research* 33, 233-260.

(b) Schematic profiles across high wave energy (upper panel) and low wave energy (lower panel) shelf areas, Grand Cayman. Reprinted with permission of the author and the Australian Coral Reef Society, successor to the Great Barrier Reef Committee, from Roberts, H.H. (1974) Variability of reefs with regard to changes in wave power around an island. *Proceedings, Second International Symposium on Corals and Coral Reefs,* Great Barrier Reef Committee, Brisbane, Australia, 497-512.

Figure 10: Genesis and evolution of transgressive depositional systems in the Mississippi River delta plain summarized by this three-stage geomorphic model. Reprinted with permission from Penland, S., Boyd, R. & Suter, J.R. (1988) Transgressive depositional systems of the Mississippi delta plain. *Journal of Sedimentary Petrology* 58, 932-949, SEPM Society for Sedimentary Geology, © 1992.

Figure 11: Comparing reefs to a group of joggers in a race against sea level. Original cartoon by Conrad Neumann. Reprinted by permission from Springer Nature Customer Service Centre GmbH from Macintyre, I.G. & Neumann, A.C. (2011) Reef classification, response to sea level rise, in Hopley, D. (ed), *Encyclopedia of Modern Coral Reefs*. Dordrecht; Springer Nature, 855- 856, © 2011.

Figure 12: Schematic representation of littoral sediment cells along the southern California coast. Each cell contains a complete sedimentation cycle: sand is brought to the coast by streams, carried along the coast by waves and currents, and lost into deep submarine basins through submarine canyons. Reprinted from Inman, D.L. & Frautschy, J.D. (1966) Littoral processes and the development of shorelines. *Proceedings of the Coastal Engineering Speciality Conference, Santa Barbara, California*. New York: American Society of Civil Engineers, 511-536, with permission from ASCE.

Plate Captions

Plate 1: Doug Inman (1920-2016) photographed in 1971 (SIO Photographic Laboratory records. SAC 0044. With kind permission of the Director, Special Collections & Archives, UC San Diego Library).

Plate 2: Cuchlaine King (1922-2019) surveying on Morsarjokull, S.E. Iceland (photograph: J. Ives, reproduced with kind permission of the author).

Plate 3: Bob Dolan (1929-2016) with recording station on Seaport pier, Outer Banks, North Carolina, USA shortly before the catastrophic Ash Wednesday storm of 7 March 1962 (photographer unknown, with kind permission of Patrick Dolan).

Plate 4: Field caravan containing pre-computer age chart recorders hardwired to waves and current meters at Little Talbot Island, Florida, USA 1973; the same set-up was used in the Seagrove experiment of 1970 (photograph: A.D. Short, reproduced with kind permission of the author)

Plate 5: Choule Sonu (1929-2020) demonstrates the radio-controlled balloon camera and its watertight plexi-glass container, as used in the Florida experiments, 1968-1971 (State Times Advocate, Baton Rouge, LA, 5 February 1969). Reprinted with authorization from Capital City Press / Georges Media Group, and Baton Rouge, LA.

Plate 6: Bruce Thom surveying the beach profile at Moruya, New South Wales, Australia, 2006 (photograph: Roger McLean, reproduced with kind permission of the author)

Plate 7: Don Wright and Felicity Coffey at the beach camp, Goolwa, The Coorong, South Australia, 28 January – 3 February 1980 (photograph: A.D. Short, reproduced with kind permission of the author). The risk of sunburn led to the adoption of innovative headgear (L.D. Wright to Spencer, pers. comm., 28 September 2021).

Plate 8: David Stoddart (1937-2014) in Fiji, early 1970s (photographer unknown, reproduced by kind permission of June Stoddart and family). On his visits to the Coastal Studies Institute, Louisiana State University, the Director, R.J. (Doc) Russell always referred to Stoddart as 'red beard' (B.G. Thom to Spencer, pers. comm., 23 September 2021). Note the camera straps: Stoddart always carried a brace of 'Pentax' cameras in the field, one with 35 mm colour slide and one with black and white film.

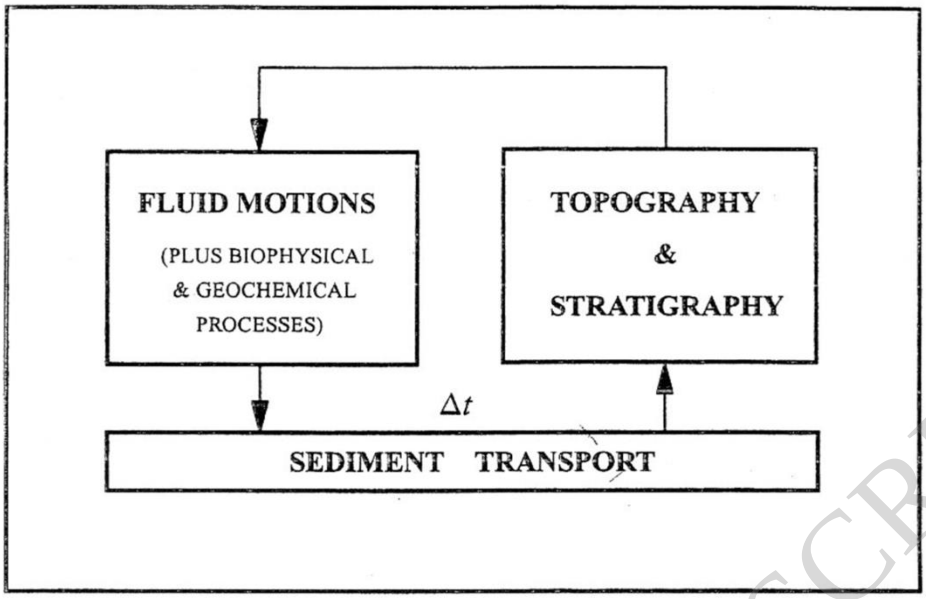

FLUID MOTIONS
(PLUS BIOPHYSICAL
& GEOCHEMICAL
PROCESSES)
TOPOGRAPHY
&
STRATIGRAPHY
Δt
SEDIMENT TRANSPORT

Figure 1

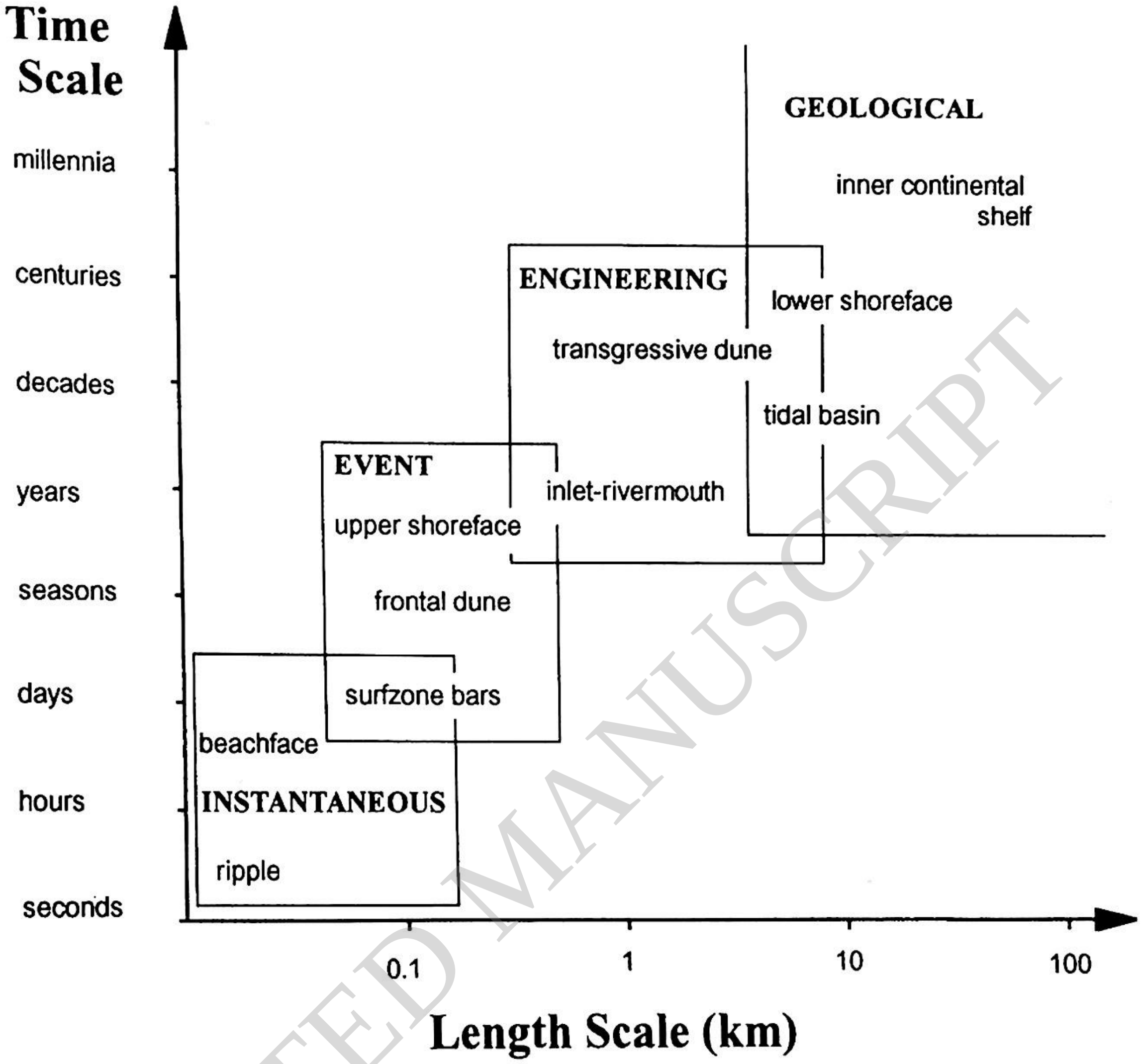


Figure 1(continued)

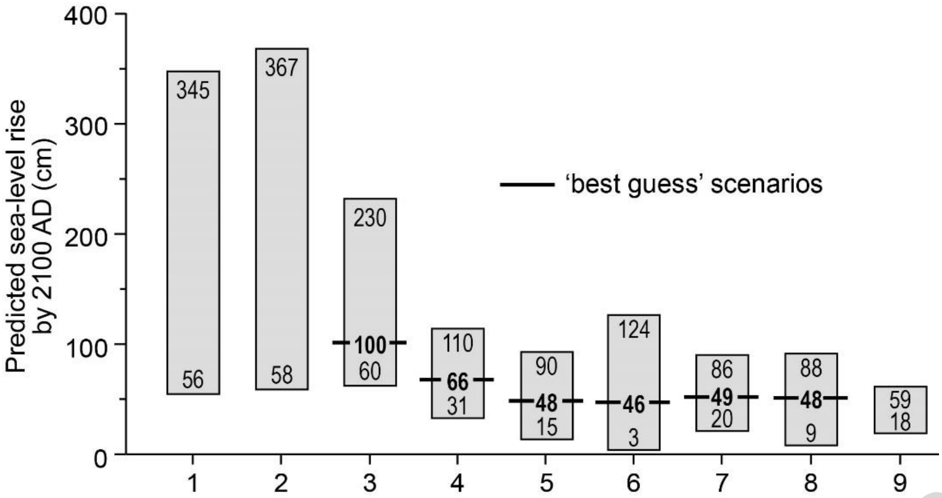

Predicted sea-level rise by 2100 AD (cm)
400
300
200
100
0
345
56
367
58
230
100
60
110
66
31
90
48
15
124
46
3
86
49
20
88
48
9
59
18
'best guess' scenarios
1
2
3
4
5
6
7
8
9

Figure 2

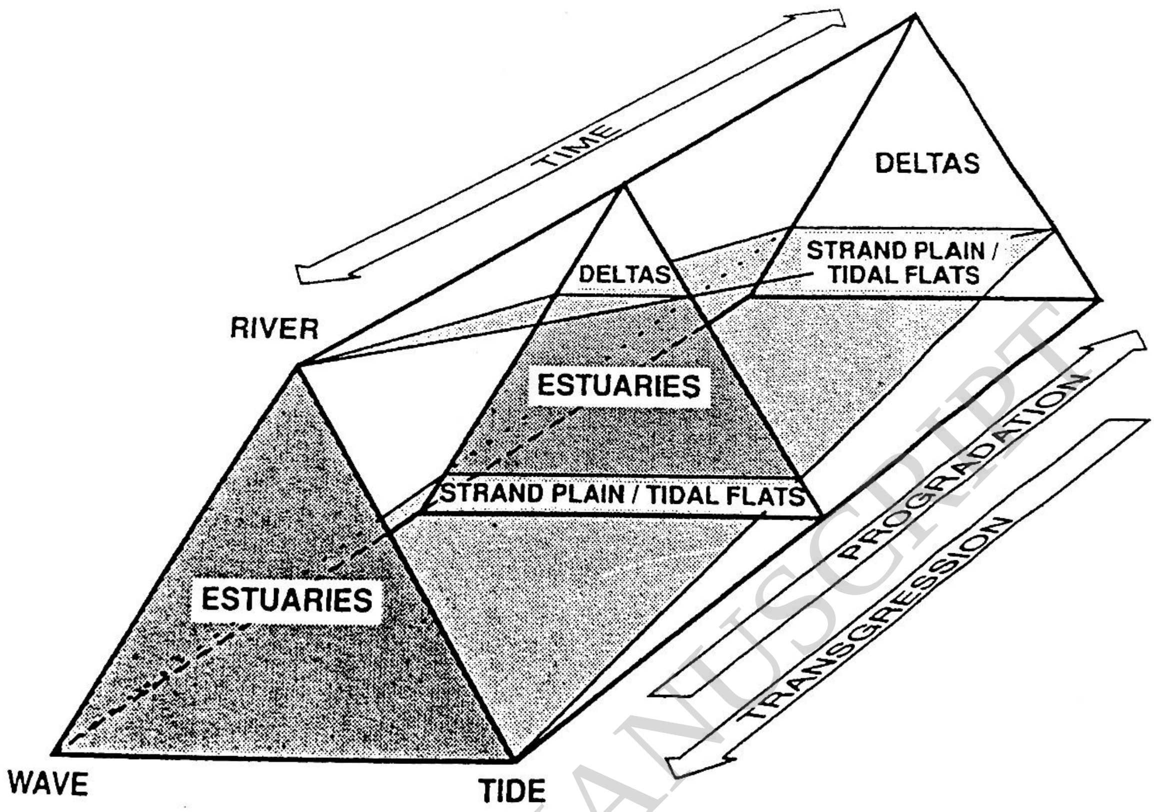
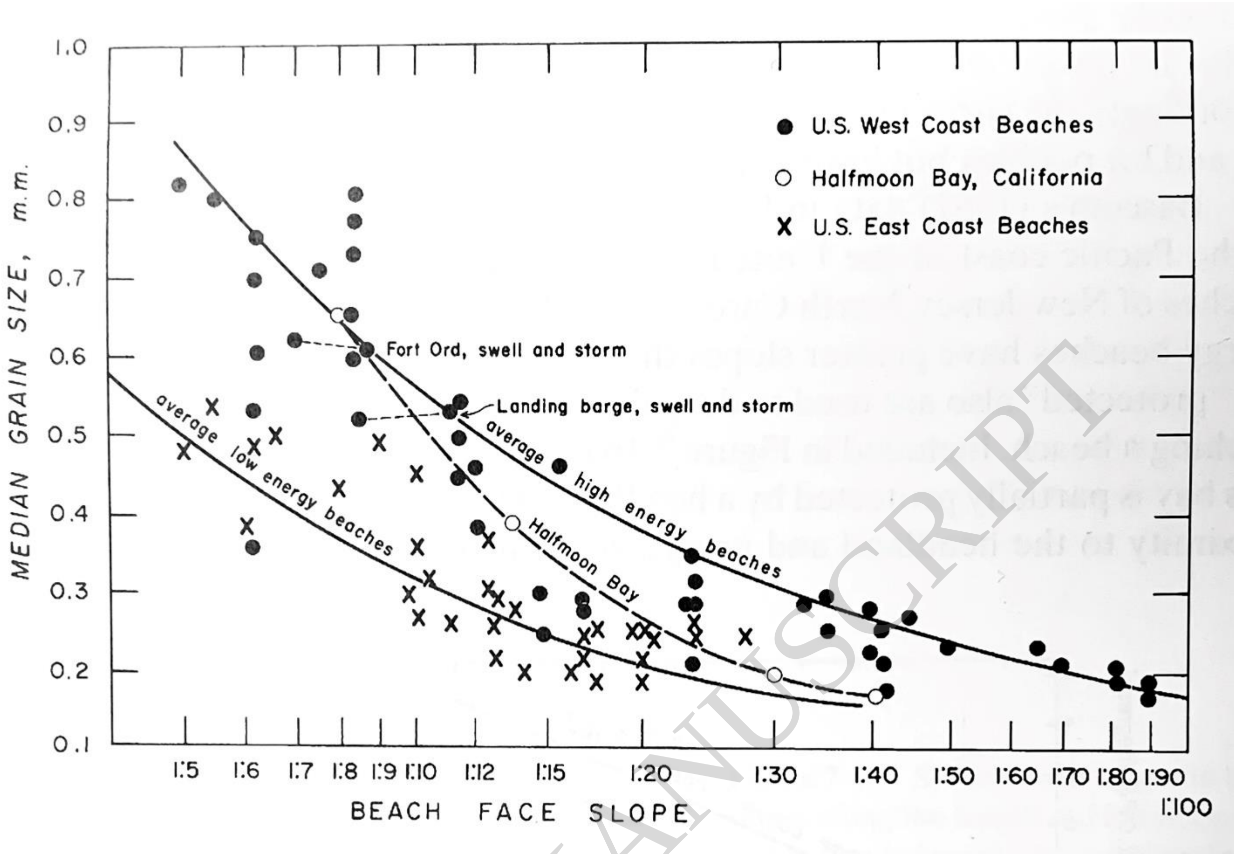

TIME
DELTAS
STRAND PLAIN / TIDAL FLATS
DELTAS
RIVER
ESTUARIES
STRAND PLAIN / TIDAL FLATS
ESTUARIES
PROGRADATION
TRANSGRESSION
WAVE
TIDE

Figure 3



U.S. West Coast Beaches
Halfmoon Bay, California
U.S. East Coast Beaches
Fort Ord, swell and storm
Landing barge, swell and storm
average high energy beaches
average low energy beaches
Halfmoon Bay
MEDIAN GRAIN SIZE, m.m.
BEACH FACE SLOPE
1.0
0.9
0.8
0.7
0.6
0.5
0.4
0.3
0.2
0.1
1:5
1:6
1:7
1:8
1:9
1:10
1:12
1:15
1:20
1:30
1:40
1:50
1:60
1:70
1:80
1:90
1:100

Figure 4

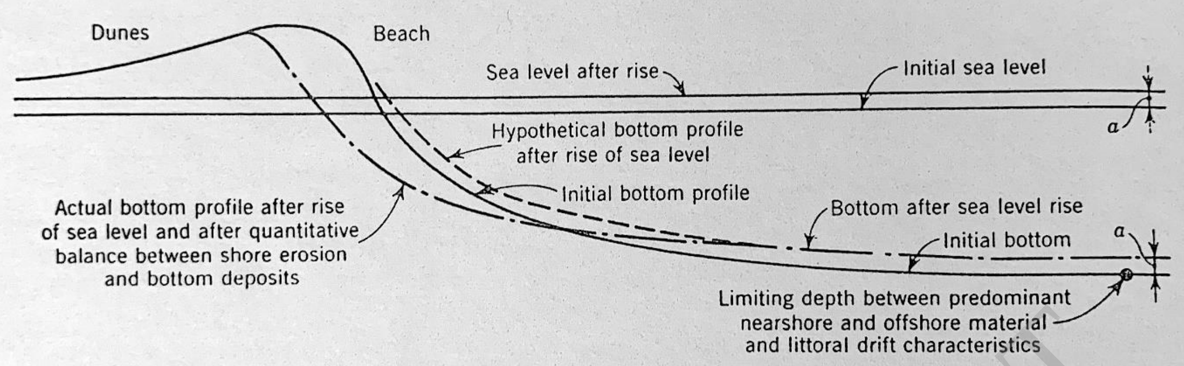


FIG. 4.—INFLUENCE OF SEA-LEVEL RISE ON THE DEVELOPMENT OF BEACH AND OFFSHORE PROFILE

Figure 5

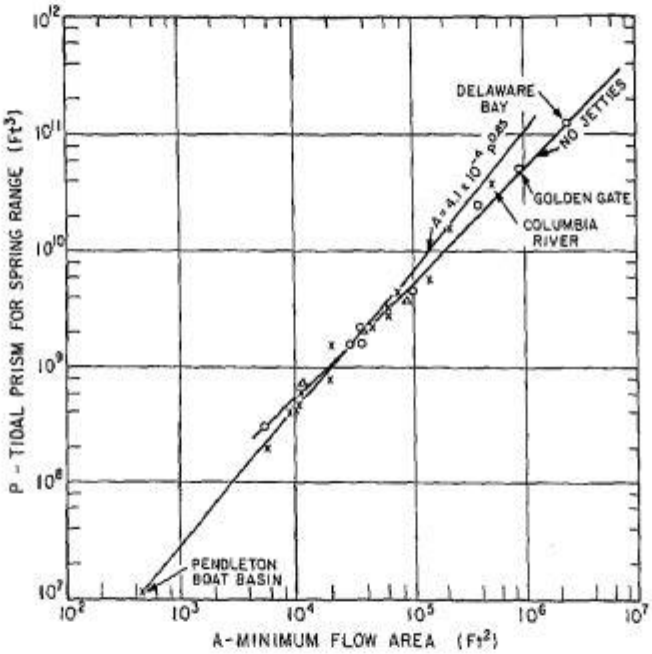

P - TIDAL PRISM FOR SPRING RANGE (Ft³)
A-MINIMUM FLOW AREA (Ft²)
DELAWARE BAY
NO JETTIES
GOLDEN GATE
COLUMBIA RIVER
PENDLETON BOAT BASIN

Figure 6

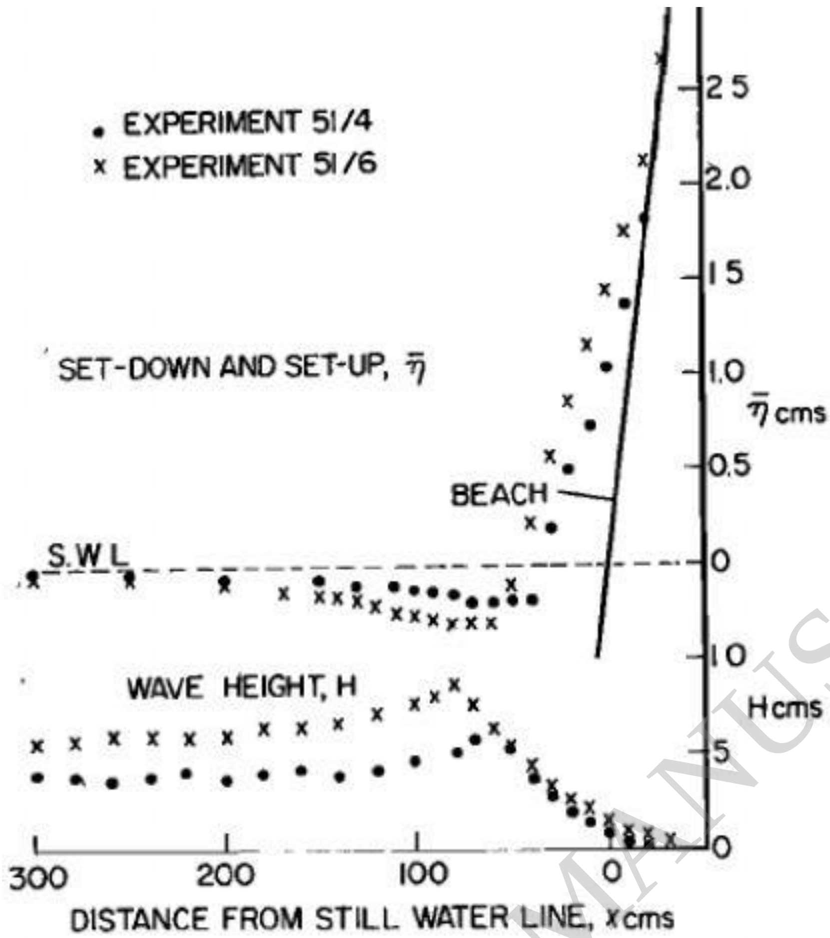

• EXPERIMENT 51/4
× EXPERIMENT 51/6
SET-DOWN AND SET-UP, $\bar{\eta}$
BEACH
S.W.L
WAVE HEIGHT, H
2.5
2.0
1.5
1.0
$\bar{\eta}$ cms
0.5
0
1.0
H cms
.5
0
300
200
100
0
DISTANCE FROM STILL WATER LINE, x cms

Figure 7

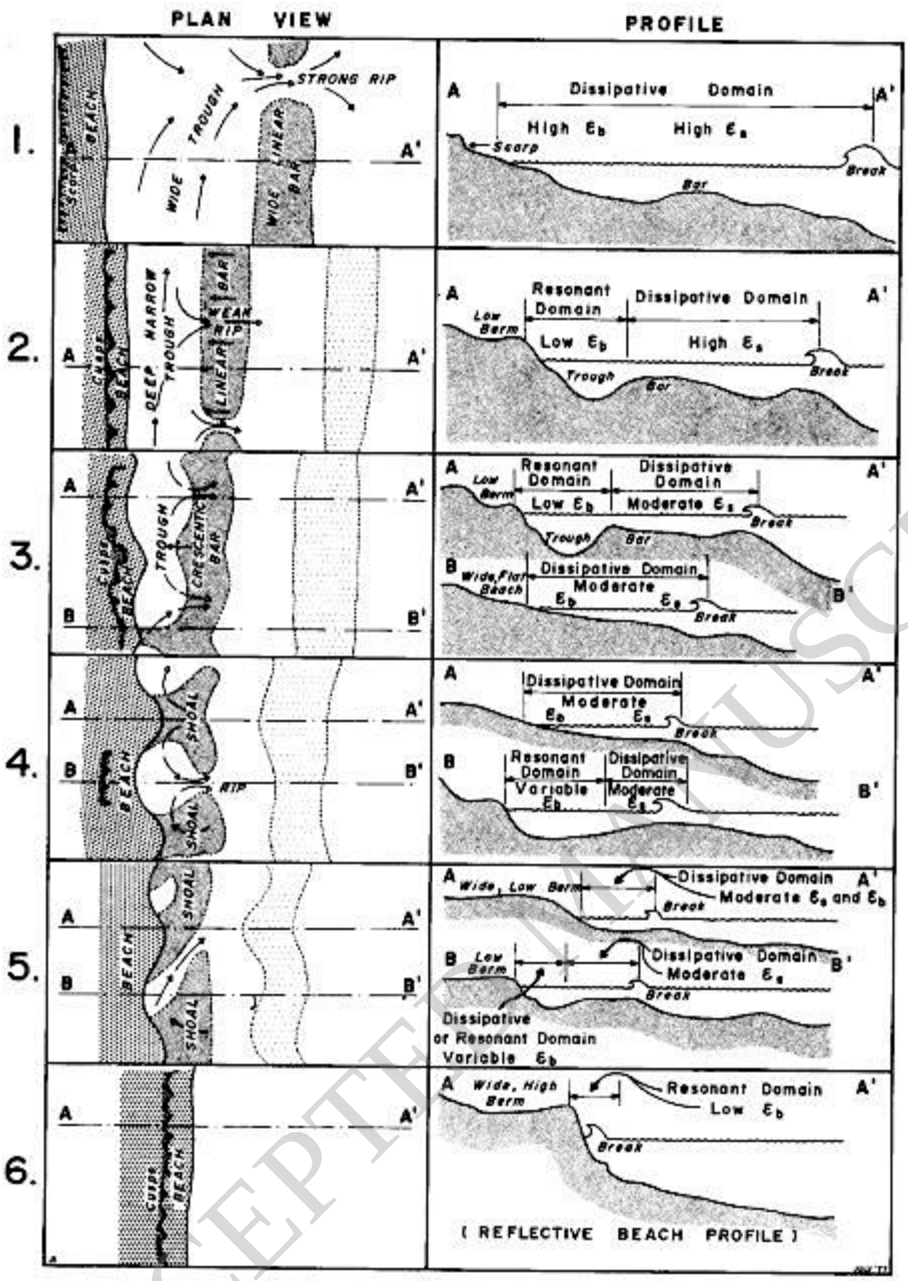

PLAN VIEW
PROFILE
1.
2.
3.
4.
5.
6.
STRONG RIP
TROUGH
WIDE
WIDE LINEAR BAR
BEACH
Dissipative Domain
High ε_b
High ε_s
Scarp
Break
Bar
DEEP, NARROW TROUGH
WEAK RIP
LINEAR BAR
Resonant Domain
Dissipative Domain
Low Berm
Low ε_b
High ε_s
Trough
CRESCENTIC BAR
Moderate ε_s
Wide, Flat Beach
Moderate
SHOAL
RIP
Variable
Wide, Low Berm
Moderate ε_s and ε_b
Dissipative or Resonant Domain
Variable ε_b
Wide, High Berm
Resonant Domain
(REFLECTIVE BEACH PROFILE)

Figure 8

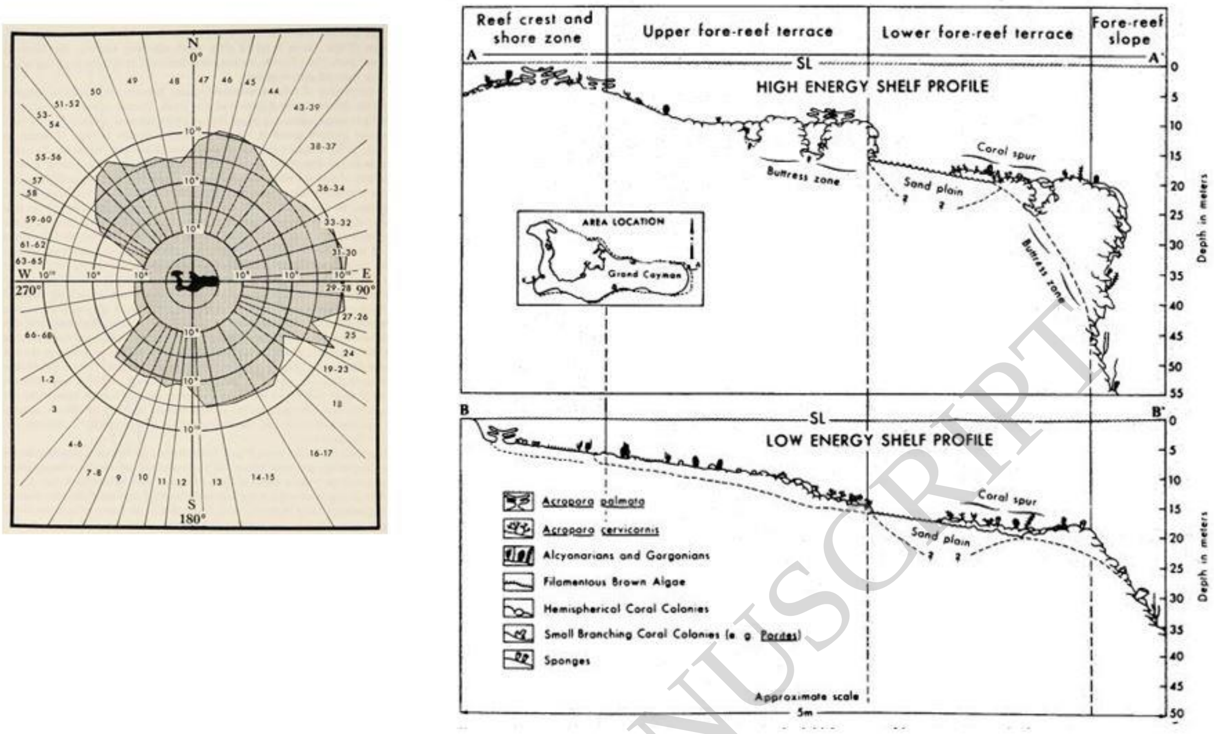

Reef crest and shore zone
Upper fore-reef terrace
Lower fore-reef terrace
Fore-reef slope
HIGH ENERGY SHELF PROFILE
LOW ENERGY SHELF PROFILE
Coral spur
Sand plain
Buttress zone
AREA LOCATION
Grand Cayman
Acropora palmata
Acropora cervicornis
Alcyonarians and Gorgonians
Filamentous Brown Algae
Hemispherical Coral Colonies
Sponges
Approximate scale
Depth in meters

Figure 9

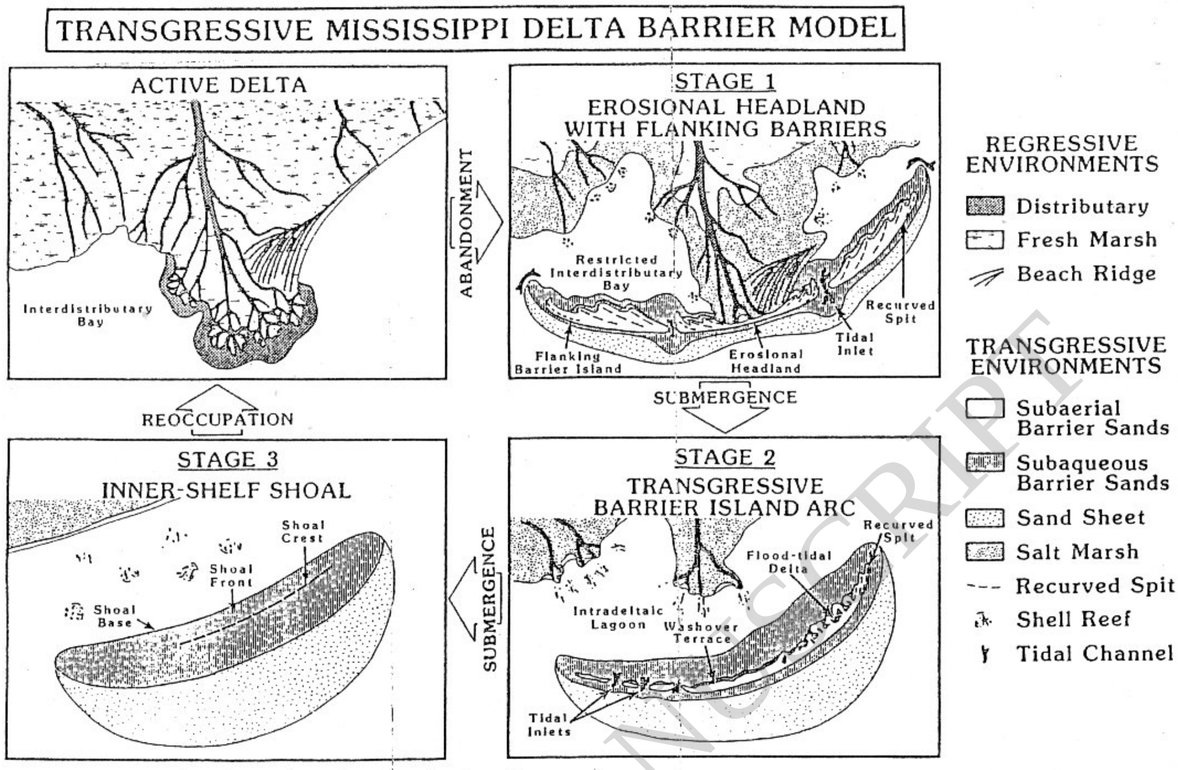

TRANSGRESSIVE MISSISSIPPI DELTA BARRIER MODEL
ACTIVE DELTA
Interdistributary Bay
ABANDONMENT
STAGE 1
EROSIONAL HEADLAND WITH FLANKING BARRIERS
Restricted Interdistributary Bay
Recurved Spit
Tidal Inlet
Flanking Barrier Island
Erosional Headland
SUBMERGENCE
REOCCUPATION
STAGE 3
INNER-SHELF SHOAL
Shoal Crest
Shoal Front
Shoal Base
SUBMERGENCE
STAGE 2
TRANSGRESSIVE BARRIER ISLAND ARC
Recurved Spit
Flood-tidal Delta
Intradeltaic Lagoon
Washover Terrace
Tidal Inlets
REGRESSIVE ENVIRONMENTS
Distributary
Fresh Marsh
Beach Ridge
TRANSGRESSIVE ENVIRONMENTS
Subaerial Barrier Sands
Subaqueous Barrier Sands
Sand Sheet
Salt Marsh
Recurved Spit
Shell Reef
Tidal Channel

Figure 10

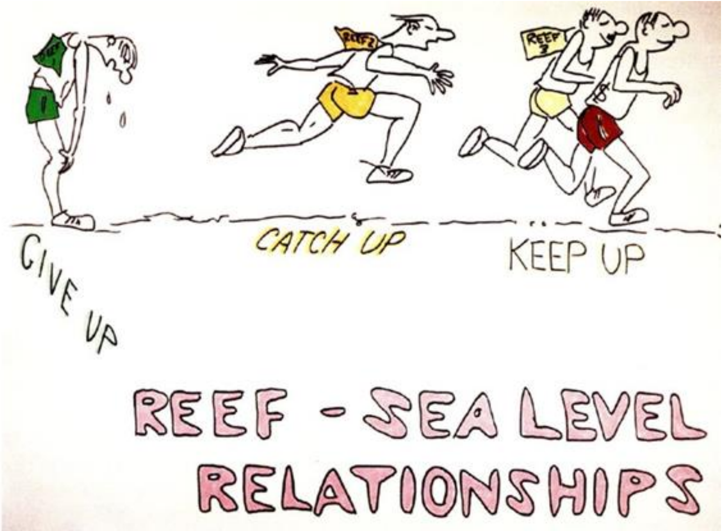
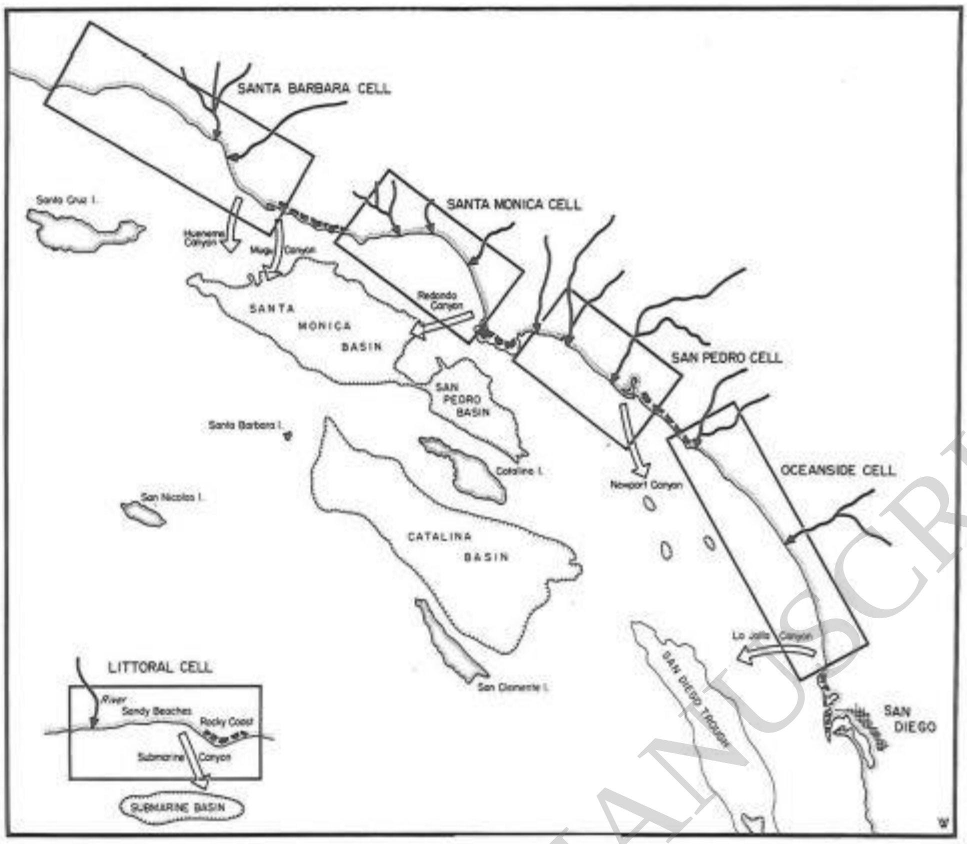


Figure 11



SANTA BARBARA CELL
SANTA MONICA CELL
SAN PEDRO CELL
OCEANSIDE CELL
Santa Cruz I.
Hueneme Canyon
Mugu Canyon
Redondo Canyon
SANTA MONICA BASIN
SAN PEDRO BASIN
Santa Barbara I.
San Nicolas I.
Catalina I.
CATALINA BASIN
San Clemente I.
Newport Canyon
La Jolla Canyon
SAN DIEGO TROUGH
SAN DIEGO
LITTORAL CELL
River
Sandy Beaches
Rocky Coast
Submarine Canyon
SUBMARINE BASIN

Figure 12

Plate 1

Plate 2


COASTAL STUDIES INSTITUTE
LOUISIANA STATE UNIVERSITY
TIME-LAPSE BEACH STUDY
Project of
GEOGRAPHY BRANCH
OFFICE OF NAVAL RESEARCH
US NAVY

Plate 3

Plate 4

Plate 5

Plate 6

Plate 7

Plate 8