1 Introduction to Coastal Environments and Global Change

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1.1 Setting the scene

1.1.1 What is the coastal zone?

At the outset of this book, it is important to articulate clearly what we mean by 'coast', because the term means different things to different people. For most holidaymakers, the coast is synonymous with the beach. For birdwatchers, the coast generally refers to the intertidal zone; while for cartographers, the coast is simply a line on the map separating the land from the sea. Coastal scientists and managers tend to take a broader view.

According to our perspective, the coast represents that region of the Earth's surface that has been affected by coastal processes, i.e. waves and tides, during the Quaternary geological period (the last 2.6 M years). The coastal zone thus defined includes the coastal plain, the contemporary estuarine, dune and beach area, the shoreface (the underwater part of the beach), and part of the continental shelf and, in areas of isostatic or tectonic uplift, fossil raised shorelines (Fig. 1.1). At a first glance, it seems rather arbitrary and perhaps odd to take such a long-term view of the timescale involved with coastal processes and geomorphology. However, as we will see later (Chapter 2), the Quaternary was a period characterized by significant changes in sea level. In the past, eustatic, or global, sea level has been considerably lower than at present (>100 m) during cold glacial periods, but also somewhat higher (up to 10 m) during some of the warm interglacial periods. This implies that coastal sediments and landforms have the potential to extend considerably beyond the zone of contemporary coastal processes. In areas of former glaciations, where isostatic processes have caused crustal uplift, fossil coastal landforms can be found far above the present shoreline (Fig. 1.2a). Similarly, in tectonically active coastal areas, fossil shorelines can also be significantly displaced (Fig. 1.2b). In a lateral sense our definition means that the coastal zone can span hundreds of kilometres, especially

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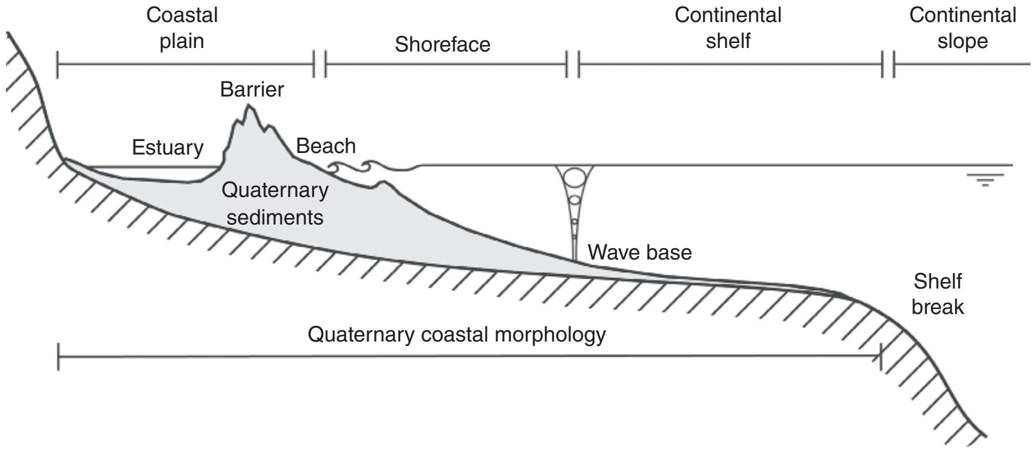

Fig. 1.1 Spatial extent of the coastal zone, including the coastal plain, shoreface and continental shelf. Note that the widths of these zones are globally highly variable. (Source: Masselink et al. 2011. Reproduced with permission of Hodder & Stoughton Ltd.)

Fig. 1.2 (a) Postglacial raised beaches at Porsangerfjord, Finnmark, Norway; (b) fossil coastal notch in Barbados formed in the last interglacial (c. 125,000 years ago) and raised above sea level by tectonic processes; and (c) view from Prawle Point (south Devon, UK) looking east, showing an apron of periglacial solifluction deposits emplaced on a raised shore platform presumed to date to the last interglacial. The fossil interglacial sea cliff is also visible. (Source: Photographs by Roland Gehrels.)

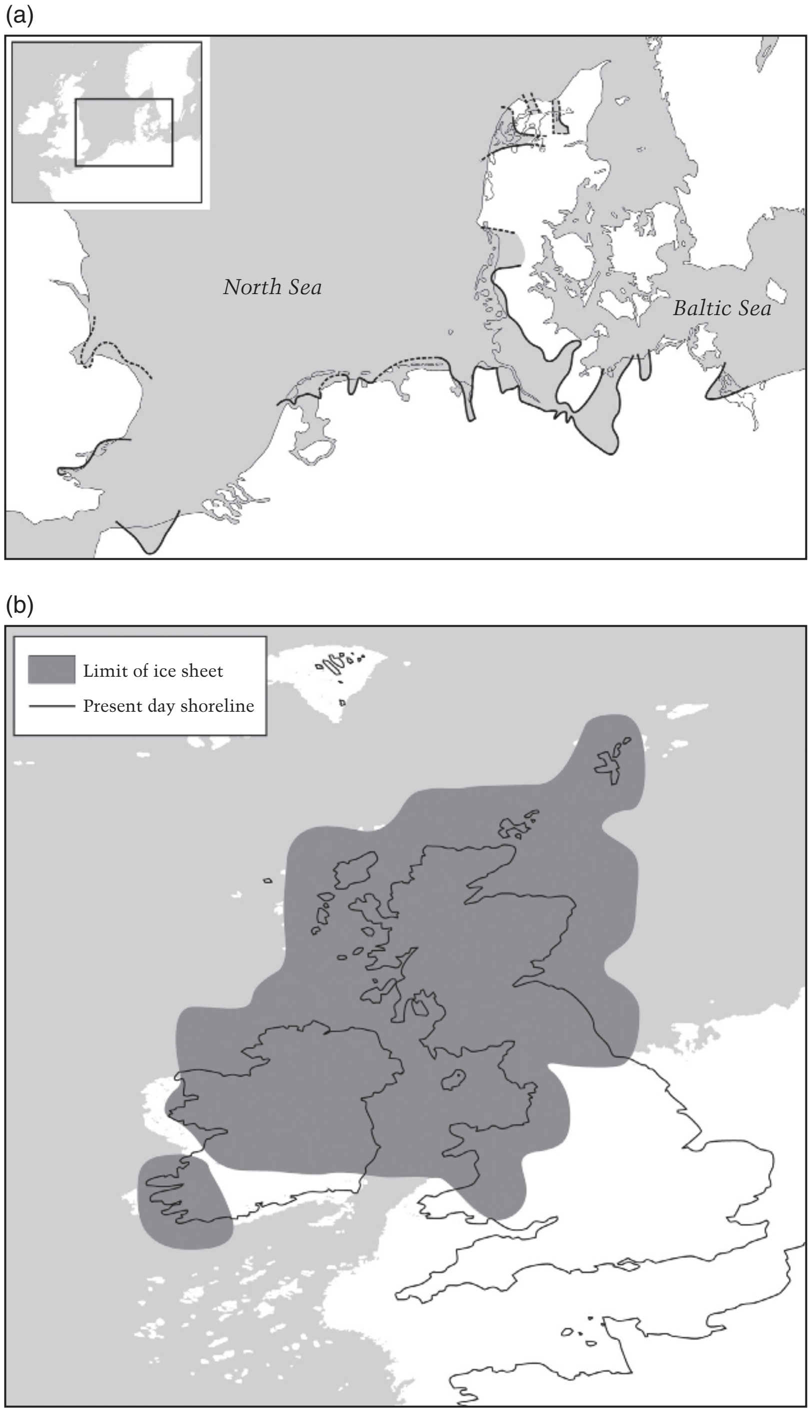

Fig. 1.3 (a) Coastline around the North Sea during the last interglacial, around 125,000 years ago (Source: Adapted from Streif 2004. Reproduced with permission of Elsevier); and (b) land area (in white) around the British Isles during the Late Glacial Maximum, around 20,000 years ago (Source: Adapted from Brooks et al. 2011).

in areas with broad continental shelves and shallow seas. For example, Fig. 1.3a shows the position of the coastline in northwest Europe during the last interglacial when sea level was several metres higher than today. During the Last Glacial Maximum the shoreline was close to the present-day continental shelf edge (Fig. 1.3b). Because coastal evolution is cumulative, i.e. the contemporary coastal landscape is partly a product of coastal processes and landforms in the past (Cowell and Thom, 1994), we need to take this long-term perspective.

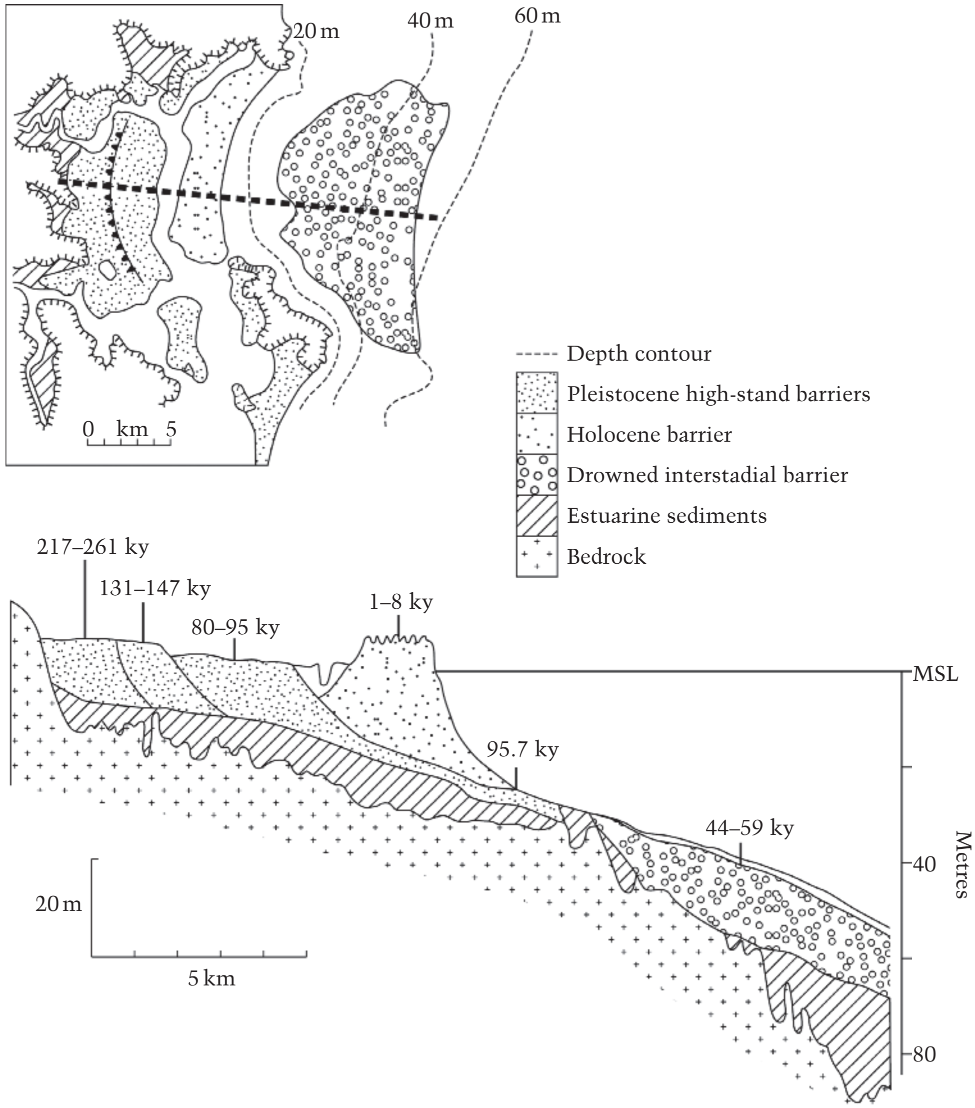

Fig. 1.4 Coastal morphology of the Tuncurry embayment, New South Wales, Australia, showing the presence of five barrier systems: the contemporary barrier, a drowned barrier on the inner shelf, and three high-stand barriers. Each of these barriers is of a different age and formed at a different relative sea level. MSL, mean sea level. (Source: Adapted from Roy et al. 1994. Reproduced with permission from Cambridge University Press and Masselink et al. 2011.)

Figure 1.4 shows an interpretive map and cross-section of the Tuncurry embayment in New South Wales, Australia. Here, research has demonstrated the presence of at least five coastal barrier systems of various ages (see Chapter 8), each of which is associated with a different sea level (Roy et al., 1994). In addition to the contemporary barrier system, there are three so-called highstand barriers to the landward (ages c. 240ky, 140ky and 90ky BP) and one drowned barrier system to the seaward on the continental shelf (age c. 50ky BP). To understand fully the dynamics of the present barrier system, in addition to contemporary coastal processes and sea level, the evolution and configuration of these older barriers also have to be taken into account. For example, the drowned barrier system can supply (and probably has supplied) sediment to the contemporary barrier, whereas the highstand barriers have provided the substrate on which the present-day barrier has developed.

Figure 1.2c shows a scenic view from Prawle Point in Devon, UK. At this location, periglacial solifluction deposits (locally known as 'head') were emplaced during the last glacial period on a raised shore platform that formed during the preceding interglacial when sea level was several metres higher than present. The 'head' is an important sediment source for contemporary beaches, while rocky shore platforms are re-occupied during consecutive interglacial highstands. So here also, present-day coastal geomorphology is significantly affected by past coastal processes and landforms. In fact, erosional coastal features, especially when carved into resistant rocks, are often polygenetic (i.e. the product of more than one sea level) and rocky coast morphology can rarely be explained solely in terms of contemporary processes and sea level (Trenhaile, 2010).

1.1.2 Coastal zone and society

The coastal zone, representing the interface between the land and the sea, is of interest to a range of coastal scientists, including geographers, geologists, oceanographers and engineers. Societal concern and interest are, however, concentrated on that area in which human activities are interlinked with both the land and the sea. This area of overlap is referred to as the 'coastal resource system' and is of great societal importance, often serving as the source or backbone of the economy of coastal nations. The most obvious use of the coastal zone is providing living space, and the coast is clearly a preferred site for urbanization. For example, 23% of the global population currently live within 100 km of the coast and less than 100 m above sea level. Population density in coastal areas is three times larger than average, and projected population growth rates in the coastal zone are the highest in the world (Small and Nicholls, 2003). In addition, 21 of the 33 megacities (cities with more than eight million people; the projected top five for 2015 are Tokyo, Mumbai, Lagos, Dhaka and Karachi) can be considered coastal cities (Martinez et al., 2007). It is worth pointing out, however, that the dynamic definition of the coastal zone at the start of this section (based on sediments, sea-level history and coastal processes) is different from the static definition generally used by planners and demographers, based on some arbitrary distance from the coastline and/or elevation above sea level.

Human occupation is, however, but one of many uses of the coastal resource system and an extraordinarily wide range of resources and activities essential to our society take place in the coastal zone, including navigation and communication, living marine resources, mineral and energy resources, tourism and recreation, coastal infrastructure development, waste disposal and pollution, coastal environmental quality protection, beach and shoreline management, military activities and research (Cicin-Sain and Knecht, 1998). Unfortunately, there can be fierce competition for coastal resources by various users (or stakeholders) and these may result in conflicts, and possible severe disruption, or even destruction, of the functional integrity of the coastal resource system. Such conflicts are especially prevalent in the case of incompatible uses of the coastal zone (e.g. land reclamation versus nature conservation; coastal protection versus tourism; waste disposal versus fisheries).

The dramatic growth in coastal population and uses has placed increased pressure on the coastal resource system and has led, in many cases, to severely damaged coastal ecosystems and depleted resources. In addition, overdevelopment of the coast in terms of urbanization and infrastructure has significantly increased our vulnerability to coastal erosion and flooding, whilst at the same time the increased reliance on hard coastal engineering structures for coastal protection has reduced our resilience. To make matters worse, global climate change resulting in a rise in sea level and potentially an increase in storminess (or at least a change in wave climate) will provide additional pressure on the coastal zone. An integrated approach is required for the management of activities and conflicts in the coastal zone (Integrated Coastal Zone Management, ICZM; see section 7.4 and Chapter 17), but what is also essential, is a thorough understanding of the key processes driving and controlling coastal environments.

1.1.3 Scope of this book and chapter outline

The focus of this book, therefore, is to provide a description of the various coastal environments, including their functioning and governing processes, and also to evaluate how they might be affected by global change and how coastal management may assist in dealing with coastal problems arising from climate change. To provide the theoretical framework and the scope of this book, this chapter will first discuss the dominant paradigm for coastal research ('morphodynamics'). This is followed by a summary of the dominant elements of climate change relevant to the coastal zone and finally a description of the various approaches used for modelling coastal change.

1.2 Coastal morphodynamics

1.2.1 Research paradigm

In science, the term 'paradigm' refers to the 'set of practices that defines a scientific discipline at any particular period of time' (Kuhn, 1996). It relates to the overall research approach adhered to by the majority of the researchers in a certain scientific discipline and encompasses a large number of elements, including methods of observation and analysis, the types of questions asked and the topics studied, the theoretical framework of the discipline, and even mundane issues such as the key scientific journal(s) of the discipline. In the vernacular, it can simply be translated as the most common way to study a subject or, even, the way a subject should be studied ('exemplar'). As a

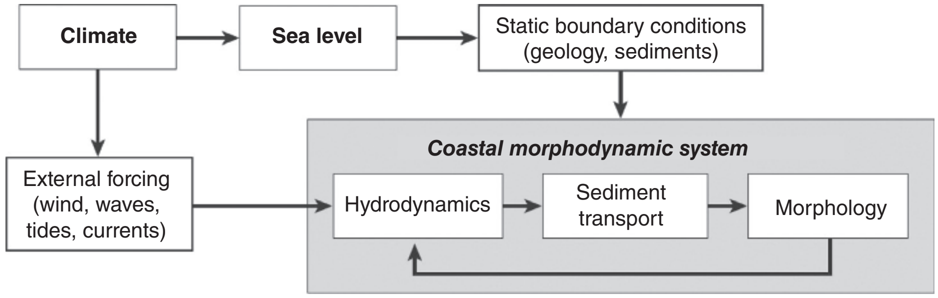

Fig. 1.5 Conceptual diagram illustrating the morphodynamic approach, showing the coastal morphodynamic systems and the environmental boundary conditions (sea level, climate, external forcing and static boundary conditions). (Source: Masselink 2012. Reproduced with permission from Pearson Education Ltd.)

discipline evolves over time, it is imperative that our knowledge and understanding thereof increases, concurrent with an increased sophistication of the research tools and analysis methods. As this happens, the relevant questions and methods of addressing these are likely to change as well; in other words, the paradigm changes. Thomas Kuhn (1922–1996), a leading philosopher of science, argued that science progresses by means of abrupt paradigm shifts, generally initiated by key scientific discoveries and/or novel research tools shedding new light on hitherto unobservable phenomena.

The dominant paradigm in coastal research up to World War II was observation and classification of coastal landforms, mainly in the context of geology and sea-level change, with coastal scientists primarily being concerned with describing and mapping the coast. During the 1950s and 1960s, the emphasis changed from observation to explanation, and this required a better understanding of the actual processes involved in driving and controlling coastal landforms and evolution. This development occurred right across the disciplines of geomorphology and physical geography, and is referred to as the process revolution (Gregory, 2000). A key tool of this paradigm was conducting actual measurements of (coastal) processes, either in the laboratory or in the field, and formulating empirical models and theories to explain these observations. Coastal landforms were very much considered the mere product of the processes, but it quickly became apparent that not only is the morphology shaped by processes, but it also provides feedback to these processes. In other words, the geomorphology is an active player, rather than a passive responder to the forcing, and has some degree of control over its own development. This notion initiated a new paradigm, referred to as the 'morphodynamic approach', and this approach was eloquently and comprehensively introduced to coastal geomorphologists by Wright and Thom (1977) in a benchmark paper in *Progress in Physical Geography* (ironically, a journal now rarely used as an outlet for coastal research).

There have been subsequent developments in geomorphology and physical geography that have contributed to a refining of the morphodynamic paradigm, involving concepts such as chaos theory and non-linear dynamics (Richards, 2003). However, these are all directly reliant on the key notion of mutual feedback between process and form, and are therefore not fundamentally different from the morphodynamic approach. It has been argued that the most current paradigm involves interactions between physical and socio-economic systems, and has materialized in a new scientific field: Earth System Science. Others maintain that this is merely a rebranding of the old discipline of Geography (Pitman, 2005). We leave such musings behind and focus on what the morphodynamic paradigm represents.

1.2.2 Coastal morphodynamic systems

According to the coastal morphodynamic paradigm, conceptualized in Fig. 1.5, coastal systems (e.g. salt marsh, beach, tidal basin) comprise three linked elements (morphology, processes and sediment transport) that exhibit a certain degree of autonomy in their behaviour, but are ultimately driven and controlled by environmental factors (Wright and Thom, 1977). These environmental factors are referred to as 'boundary conditions', and include the solid boundary (geology and sediments; Chapter 3), climate (section 1.3) and external forcing (wind, waves, storms, tides and tsunami; Chapters 4 and 5), with sea level (Chapter 2) serving as a meta-control by determining where coastal processes operate. When contemporary coastal systems and processes are considered, human activity should also be taken into account. In fact, along many of our coastlines human activities, such as beach nourishment, construction of coastal defences, dredging and land reclamation, are more important in driving and controlling coastal dynamics than the natural boundary conditions and can therefore not be ignored

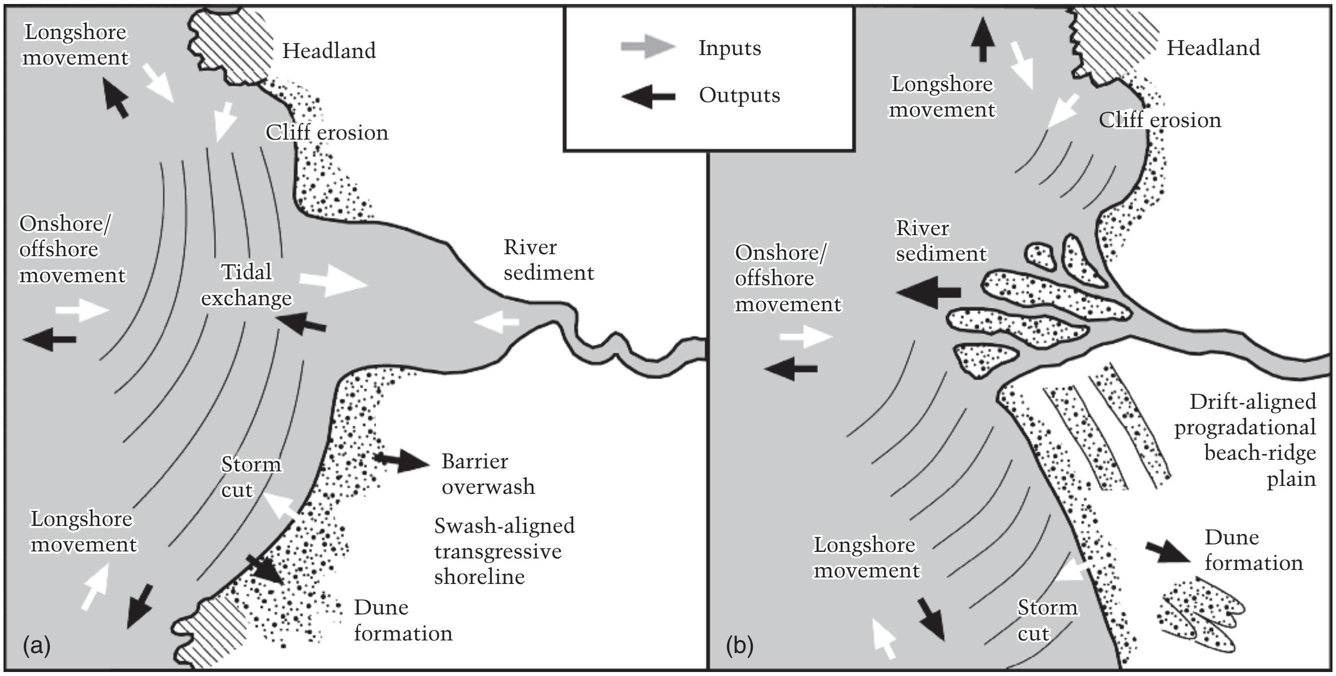

Fig. 1.6 Sediment budgets on: (a) estuarine; and (b) deltaic coasts. (Source: Masselink et al. 2011. Reproduced with permission of Hodder & Stoughton Ltd and adapted from Carter and Woodroffe 1994 with permission from Cambridge University Press.)

(Chapter 17). Moreover, through climate change, humans are altering the boundary conditions themselves (sea-level rise and changes to the wave climate).

Unless long-term coastal change (centuries to millennia) is considered, the boundary conditions can be viewed as given and constant, although it should be borne in mind that external forcing is stochastic (random), and the dynamics of coastal systems arise from the interactions between the three linked elements:

(1) Processes: This component includes all processes occurring in coastal environments that generate and affect the movement of sediment, resulting ultimately in morphological change. The most important of these are hydrodynamic (waves, tides and currents) and aerodynamic (wind) processes. Along rocky coasts, weathering is an additional process that contributes significantly to sediment transport, either directly through solution of minerals, or indirectly by weakening the rock surface to facilitate mobilization by hydrodynamic processes (Chapter 15). In addition, biological, biophysical and biochemical processes are important in salt marsh (Chapter 10), mangrove (Chapter 11) and coral reef (Chapter 16) environments. River outflow processes are important in deltas (Chapter 13).

(2) Sediment transport: A moving fluid imparts a stress on the bed, referred to as 'bed shear stress', and if the bed is mobile this may result in the entrainment and subsequent transport of sediment. The ensuing pattern of erosion and deposition can be assessed using the sediment budget (Fig. 1.6). If the sediment balance is positive (i.e. more sediment is entering a coastal region than exiting), deposition will occur and the coastline may advance, while a negative sediment balance (i.e. more sediment is exiting a coastal region than entering) results in erosion and possibly coastline retreat. This makes quantifying the sediment budget a fundamental means for understanding coastal dynamics, as well as providing a tool for assessing and predicting future coastal change.

(3) Morphology: The three-dimensional surface of a landform or assemblage of landforms (e.g. coastal dunes, deltas, estuaries, beaches, coral reefs, shore platforms) is referred to as the morphology. Changes in the morphology are brought about by erosion and deposition, and are, in part, recorded in the stratigraphy (section 1.2.4).

It is worth emphasizing that the morphodynamic approach is scale-invariant, i.e. the approach can be applied regardless of the spatial scale of the coastal feature under investigation. For example, at the smallest scale, the approach can be applied to wave and tidal bed forms; at the largest scale, to tidal basins or entire delta systems. Importantly, the spatial and temporal scales of coastal morphodynamic systems are related (Fig. 1.7): the larger the spatial scale of the coastal system, the longer the timescale associated with the dominant process(es) and the associated coastal morphodynamics. The spatio-temporal relationship is, however, not linear: some coastal systems respond faster than one would expect on the basis of their

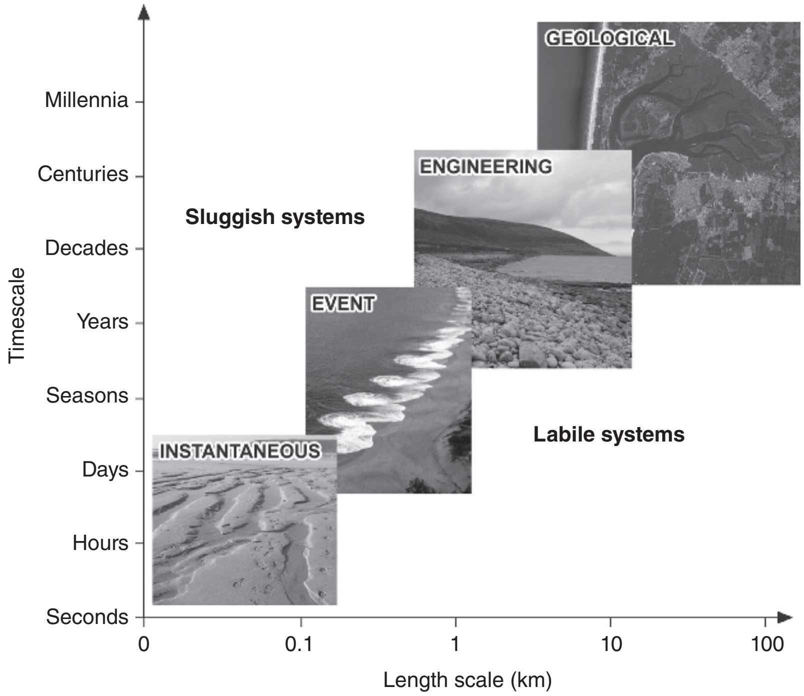

Fig. 1.7 Relationship between spatial and temporal scales of coastal systems. Sluggish and labile systems are those that respond relatively slow and fast, respectively. (Source: Adapted from Cowell and Thom 1994. Imagery © 2013 Terrametrics. Map data © 2013 Google.) For colour details, please see Plate 1.

size (labile systems; e.g. sandy barriers without dunes), whereas other coastal systems exhibit a relatively slow response (sluggish systems; e.g. rocky coasts). The timescale of the response of a coastal system also depends, of course, on the magnitude of the forcing, and the classic magnitude-frequency concept (Wolfman and Miller, 1960) is as relevant now as it was when it was introduced in geomorphology.

1.2.3 Morphodynamic feedback

A characteristic of coastal morphodynamic systems is the presence of strong links between form and process (Cowell and Thom, 1994). The coupling mechanism between processes and morphology is provided by sediment transport and is relatively easy to comprehend. There is, however, also a link between morphology and processes to complete the morphodynamic feedback loop.

As an example, under calm wave conditions sand is transported on a beach in the onshore direction resulting in beach accretion and the construction of a feature known as the 'berm' (Fig. 1.8). During berm construction, the seaward slope of the beach progressively steepens and the top of the berm increases in elevation relative to sea level through accretion; both morphological developments

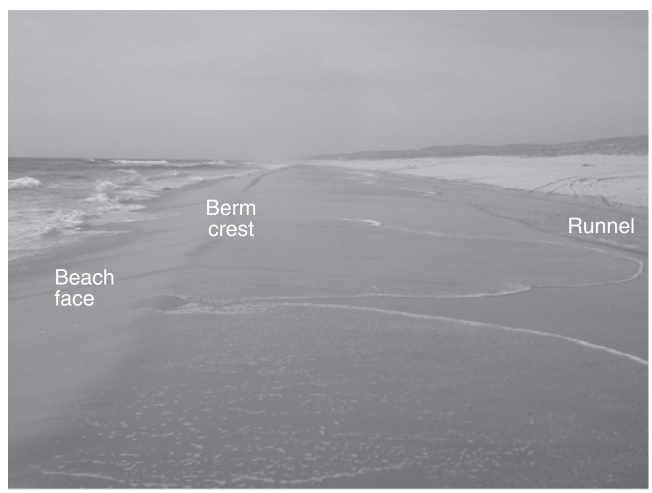

Fig. 1.8 Photograph of a developing berm on a sandy beach. Berms are swash-formed features that usually develop as part of beach recovery following storm erosion. On tidal beaches they are found just above the high-tide level. This particular berm formed after a period of energetic waves and is well defined with a small runnel located to the landward. The photo was taken at high tide and the berm is still being overtopped by swash action and is therefore still being constructed. (Source: Photograph by Gerd Masselink.)

have profound effects on the wave-breaking processes and sediment transport (Masselink and Puleo, 2006). The steepening of the beach makes it increasingly difficult for the onshore-directed uprush flow to transport sediment up-slope, whilst at the same time the down-slope transport by the offshore-directed backwash flow is enhanced. Additionally, the wave breaker type may change from energetic plunging, which entrains large amounts of sediment that become advected into the uprush promoting onshore transport, to surging, which is less favourable to the uprush. The increased elevation of the berm reduces the frequency of waves reaching the top of the berm, leading to a progressive reduction in the vertical accretion rate. At some stage during beach steepening, the hydrodynamic conditions may be sufficiently altered to stop further onshore sediment transport, and berm construction will cease.

The berm development discussed above is a relatively simple example of morphodynamic feedback and other examples of feedback between morphology and processes include estuarine infilling and tidal currents, foredune development and aerodynamics, delta lobe growth and hydraulic gradient, salt marsh accretion and tidal inundation frequency, mangrove establishment and sedimentation processes, and coral reef development and wave attenuation. In all cases, due to the close coupling between process and form, cause and effect are not readily apparent. This gives rise to the 'chicken-and-egg' nature of coastal morphodynamics whereby it is often not clear whether the morphology is the result of the hydrodynamic processes, or vice versa. In a developing morphodynamic system, process and form co-evolve, and this is one of the key factors that make it so difficult to predict reliably long-term coastal development: small errors in predicting either the morphological change or the hydrodynamic processes end up magnifying dramatically over time. In more technical parlance, coastal morphodynamic systems are therefore also described as 'complex' or 'non-linear'.

The feedback between morphology and processes is fundamental to coastal morphodynamics, and can be negative or positive.

- **Negative feedback** is a damping mechanism that acts to oppose changes in morphology and is a stabilizing process, eventually resulting in equilibrium. An example of negative feedback is the berm development discussed previously. However, morphological adjustment involves a redistribution of sediment and this requires a finite amount of time. The time it takes to attain equilibrium defines the relaxation time and is a measure of the morphological inertia within the system (de Boer, 1992). The relaxation time depends on the volume of sediment involved in the morphologic adjustment (i.e. the spatial scale of the landform) and the energy level of the forcing that controls the sediment transport rate. For large coastal landforms, the relaxation time generally exceeds the time between changes in environmental conditions, and in these cases it is unlikely that equilibrium is ever reached.
- **Positive feedback** pushes a system away from equilibrium by modifying the morphology such that it is even less compatible with the processes to which it is exposed. A morphodynamic system driven by positive feedback seems to have a 'mind of its own' and exhibits self-forcing behaviour. An example of positive feedback is the infilling of deep estuaries by marine sediments due to asymmetry in the tidal flow. In a deep estuary, flood currents are stronger than ebb currents and this tidal asymmetry results in a net influx of sediment and infilling of the estuary. As the estuary is being infilled, the tidal asymmetry increases even more as friction and shoaling effects are enhanced by the reduced water depths (Friedrichs and Aubrey, 1988). In turn, the increase in tidal asymmetry speeds up the rate of estuarine infilling. This constitutes positive feedback between the estuarine morphology and the tidal processes, resulting in rapid infilling of the estuary. Eventually, intertidal salt marshes and tidal flats start developing in the estuary and this marks a reversal in feedback. As the intertidal areas become more extensive, the flood asymmetry of the tide progressively decreases so that the estuarine morphology approaches steady state as sediment imports during flood and exports during ebb equilibrate.

One of the most powerful and exciting explanations for coastal features to have emerged from the last two decades of coastal morphodynamic research is the notion of self-organization, or emergence, which refers to the development of morphological features with a specific shape and/or spacing that has arisen from the mutual interactions between form and process. In other words, the template for the morphology is not directly related to that of a specific hydrodynamic phenomenon, but has emerged from the morphodynamic interactions (i.e. feedback). The notion of self-organization has now become well established in a wide range of disciplines (Gallagher and Appenzeller, 1999), including geomorphology (Murray et al., 2009), and a range of coastal features are now interpreted as being self-organizing features, including rhythmic features such as wave ripples, beach cusps, bar morphology and cuspate shoreline features (Coco and Murray, 2007; Fig. 1.9). One of the main challenges of research into self-organization has been to identify the dominant length scales of the rhythmic shoreline features and, in addition to empirical techniques, the dominant tool has been the application of numerical modelling (section 1.4). An example of the application of a numerical model to explain cuspate features in coastal lagoons is discussed in Box 1.1.

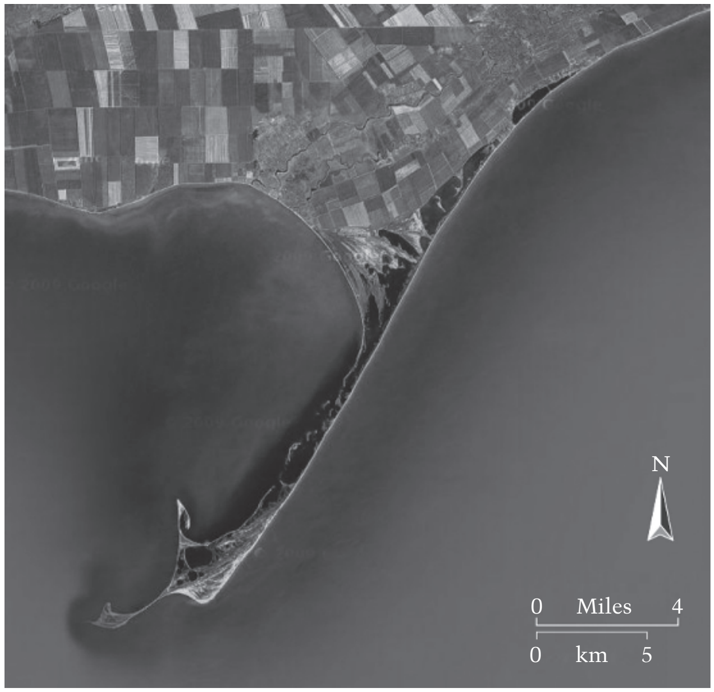

Fig. 1.9 Flying spit in the Sea of Azov, Ukraine. The formation of these features has intrigued coastal scientists for decades, but numerical modelling by Ashton and Murray (2006a, b), based on the relation between the longshore sediment transport rate and the deep-water wave angle (see Box 1.1), seems to have provided a satisfactory explanation for their formation. (Source: Image © 2013 Terrametrics. Map data © 2013 Google.)

CONCEPTS BOX 1.1 Self-organization of elongate water bodies

The long axis of some elongate water bodies (e.g. coastal lagoons) exhibit wave-formed features, such as sandy spits and capes, and in some instances a series of almost-circular lakes outline a larger basin, suggesting that opposing cuspate shoreline features have joined, segmenting the lake along its long axis (Fig. 1.10). Zenkovich (1967) suggested a qualitative model whereby the formation of cuspate forms and the eventual segmentation of elongate water bodies could be attributable to waves generated by winds blowing across the long fetch parallel to the main axis, arriving with crests at angles greater than 45° relative to the long coastlines. Recent numerical studies (Ashton and Murray, 2006a, b) have investigated how such high-angle waves lead to the initial formation and subsequent self-organization of cuspate features, and further work by Ashton et al. (2009) has suggested how the growth of cuspate shoreline features in elongate water bodies may eventually lead to a segmentation of the water body into smaller, round water bodies (Fig. 1.10).

The physical basis of the models of Ashton and co-workers is shown in Fig. 1.11. It is based on the notion that the rate of longshore sediment transport is maximized when the angle between the crests of deep-water waves and the shoreline is approximately 45°; thus, for both smaller and larger wave angles, longshore transport rates decrease away from this 'flux-maximizing' angle. When the angle between the deep-water wave crests and the shoreline is greater than 45°, the sediment flux along the convex-seaward crest of a perturbation decreases in the flux direction, because the angle between the waves and the local shoreline is increasing, moving progressively farther away from the flux-maximizing angle (Fig. 1.11a). The resulting sediment accumulation at this location will lead to a growth of the perturbation (positive feedback). When deep-water waves approach from smaller angles, the sediment flux along the convex-seaward crest of the perturbation increases in the flux direction, because the angle between the waves and the local shoreline is moving progressively closer to the flux-maximizing angle (Fig. 1.11b). This results in erosion at this location, leading to a smoothing out of the perturbation and a straightening of the coastline (negative feedback).

According to the model of Ashton et al. (2009), a large number of small cuspate features initially develop in elongate water bodies, but, as the morphology evolves and feedback between the different cuspate forms start to become significant, the number of cuspate features decreases, while their size increases (left panels of the

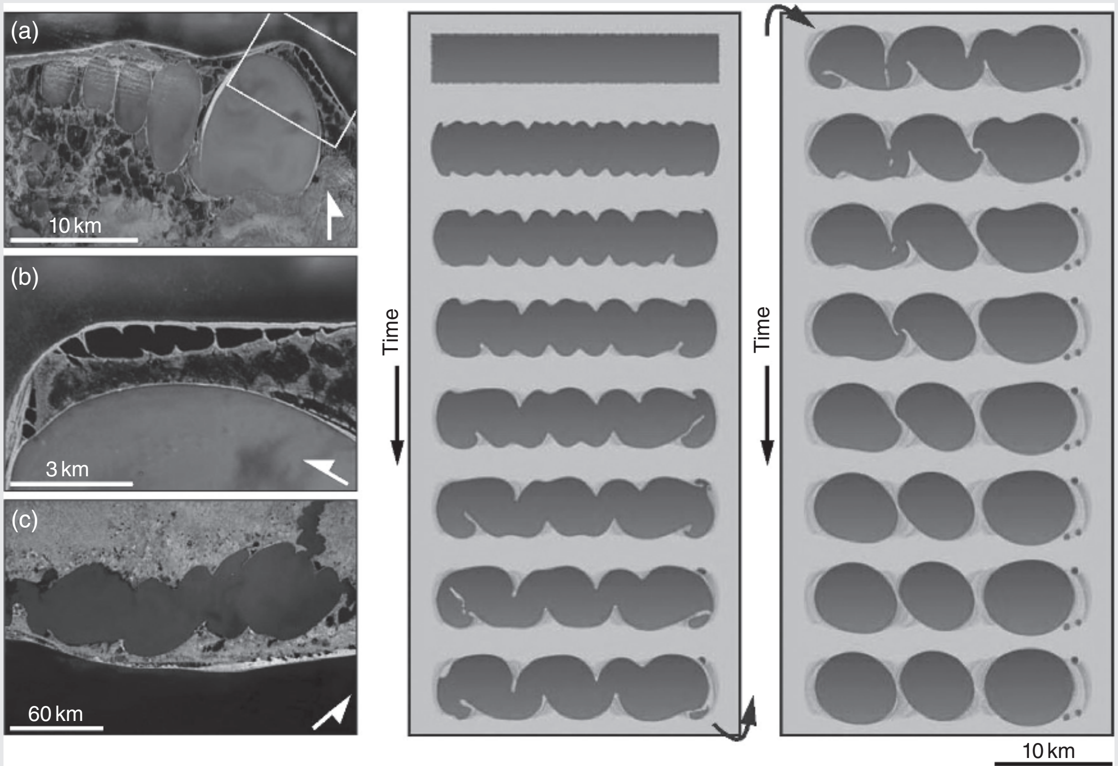

Fig. 1.10 Natural examples of enclosed water bodies with cuspate features and segmented water bodies. (a) Laguna Val'karkynmangkak, Russia; (b) inset of (a); and (c) Lagoa Dos Patos, Brazil. The results of a numerical simulation of the formation of cuspate features and segmented water bodies are shown in the right panels. (Source: Ashton et al. 2009. Reproduced with permission of the Geological Society of America.) For colour details, please see Plate 2.

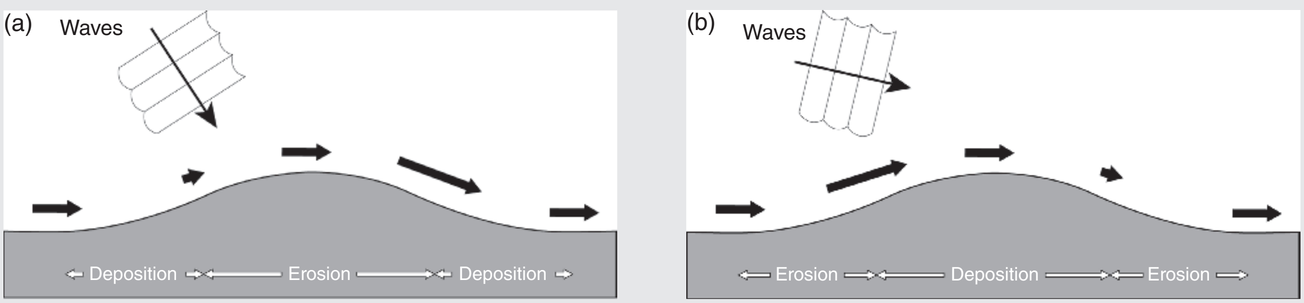

Fig. 1.11 Schematic illustrations of shoreline change caused by high-angle and low-angle waves. When the angle between the deep-water wave crests and the shoreline is less than the flux-maximizing angle (<45°) shoreline perturbations will be smoothed out, whereas for angles greater than the flux-maximizing angle (>45°) the perturbations will grow. (Source: Adapted from Coco and Murray 2007. Reproduced with permission of Elsevier.)

model simulation in Fig. 1.10). Such increase in spacing and amplitude is a well-known phenomenon of numerical self-organization models. Once the cuspate features extend significantly offshore (approximately half-way across the water body), opposing cuspate features start affecting each other by providing shelter from waves that propagate along the long-axis of the water bodies. A new dynamic emerges and opposing cuspate features that are initially offset, grow together, eventually merging and segmenting the water body into smaller, round water bodies (right panels of the model simulation in Fig. 1.10). Conditions conducive to the development of segmented elongate water bodies are relatively shallow water bodies with an energetic wind regime and non-cohesive (e.g. sand or gravel) shores. Inhibiting factors are shoreline vegetation (stabilizing the shoreline), significant tidal flows (which would become faster as the flow is constricted) and low sedimentation rates (cuspate spit growth may be too long to occur compared to other long-term environmental changes).

1.2.4 Coastal evolution and stratigraphy

As the coastal system evolves over time, its evolution is recorded in the sediments (clay, silt, sand and gravel) in the form of the stratigraphy. It is important to realize that stratigraphic sequences are a record of the depositional history and that erosional events are only represented by gaps or discontinuities in the stratigraphic record. Stratigraphy is the realm of geologists and sedimentologists, but, because it provides insights into the geomorphological evolution of coastal landforms as well as the history of the governing coastal processes, it is also of considerable interest to coastal scientists in general. The stratigraphy can be particularly useful when dating has provided the age of certain coastal deposits; this information can then be used to quantify rates of accretion, and also to help with the reconstruction of the sea-level history.

As an example, Fig. 1.12 shows the stratigraphy of a salt marsh in southern New Zealand (Gehrels et al., 2008). Accumulations of salt-marsh sediment in this part of the southern hemisphere are generally very thin, because sea level during much of the middle and late Holocene was only slightly higher than at present and little accommodation space was available for salt-marsh deposits to fill. The intertidal sands that form the substrate of the salt marsh were

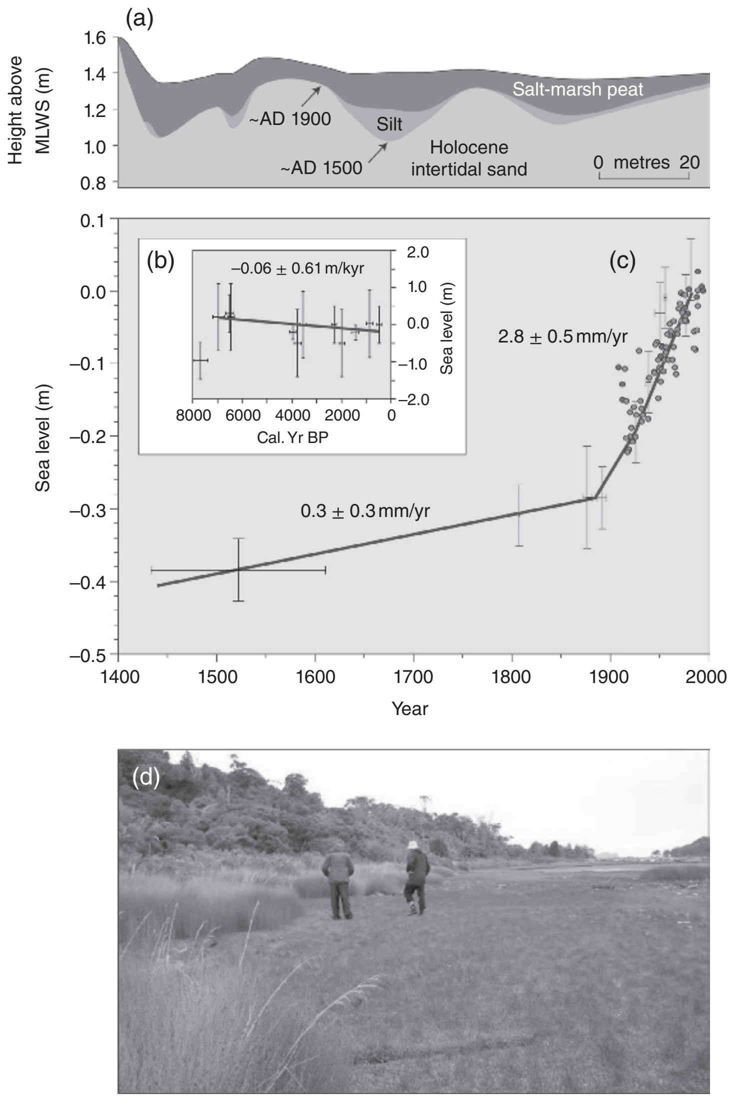

Fig. 1.12 Deriving sea-level history from salt-marsh stratigraphy. (a) Stratigraphy of a salt marsh in southern New Zealand. The marsh developed in the past half millennium on a substrate of late Holocene intertidal sands. The sands were deposited in the middle and late Holocene when sea level was slightly higher than present. MLWS, mean low water spring. (b) Since about AD 1900, accumulation has been very rapid as a consequence of the accommodation space provided by the sharp sea-level acceleration (c). The crosses in (b) and (c) represent dated samples of shells and plant material, respectively, which can be related to former sea levels (with vertical and age uncertainties). Different coloured dots in (c) represent annual measurements of sea level from two nearby tide gauges. Cal. Yr BP, calibrated years before present. (d) The photo shows an overview of the marsh, which can be found on the Catlins coast in southeastern New Zealand, near the village of Pounawea. (Source: Adapted from Gehrels et al. 2008. Reproduced with permission of John Wiley & Sons.) For colour details, please see Plate 3.

deposited during this time. By around 500 years ago, a salt-marsh environment had developed on the sands. The microfossils in the salt-marsh sediments show that the silts were deposited in an upper salt-marsh environment, close to the limit of the high spring tides. These microfossils are single-celled organisms (protists) called foraminifera. They are particularly useful in this context because they live in narrow vertical niches in the intertidal zone and can be precisely related to sea level. The transition therefore signifies that, following the deposition of the intertidal sands, the sea level must first have dropped to a low stand, before rising to a level that allowed upper salt-marsh grasses to colonize the sandy substrate. Thus, there is a significant time hiatus between the deposition of the intertidal sands and the formation of the salt marsh. Salt-marsh accretion was initially very slow; in 400 years the surface of the salt marsh only rose by about 10 cm. Around AD 1900, however, a remarkable change occurred. This change is reflected in the sediments as a transition from silty to highly organic salt-marsh deposits. The microfossils show that the surface of the marsh remained close to the highest spring tide level, indicating that about 40 cm of sea-level rise took place after c. AD 1500, but 30 cm of this occurred in the last 100 years. The rising sea level has preserved the organic salt-marsh sediments very well, whereas the sediments that were deposited during the preceding centuries have lost their organic content due to frequent subaerial exposure.

This example clearly shows how: (1) sea-level change can control the stratigraphy and sediment types of the coastal zone; (2) sea-level rise provides the accommodation space in which sediments can accumulate; and (3) the sediments provide an archive from which sea-level changes can be reconstructed. A slowly rising sea level allowed salt marshes to colonize emerged tidal-flat deposits, first slowly, but in the last 100 years very rapidly. This rise is being recorded by various tide gauges (Hannah, 2004), but the sediments in the coastal system also bear witness to the sea-level acceleration. The rapid sea-level rise, which commenced around the beginning of the 20th century, appears to be a worldwide feature (Gehrels and Woodworth, 2013) and is due to climate change (Woodworth et al., 2009; Mitchum et al., 2010).

1.3 Climate change

1.3.1 Quaternary climate change

Throughout Earth's history, climate has always been changing, but at the onset of the Quaternary, about 2.6 million years ago (Gibbard et al., 2009), the closure of the Isthmus of Panama appears to have triggered a major change in the world's ocean circulation (Sarnthein et al., 2009). Since that event, the Earth has known over 50 glacial-interglacial cycles. The most complete record of these cycles is preserved in the marine sedimentary record. Analyses of oxygen isotopes in marine sediment cores have shown that the Quaternary contains 103 marine isotope stages (Raymo et al., 1989; Gibbard et al., 2009); the evenly numbered stages are cold (glacials and stadials), the odd-numbered stages are warm (interglacials and interstadials). This subdivision is a far cry from the 'classic' four glacial and interglacial periods that had been recognized in Europe and North America by the end of the 19th century. Climate change, glaciations and sea-level change are clearly the defining features of the Quaternary.

The most conspicuous consequence of Quaternary climate change that is relevant to the coastal zone is the growth and demise of ice sheets and the resulting changes in the level of the world's oceans, with amplitudes of up to 150 m. Glaciations have also produced significant vertical changes in the level of the solid earth surface through the loading and unloading by ice and water. These isostatic changes affect both the land (glacio-isostasy) and the sea floor (hydro-isostasy) and they can produce vertical shifts of the coastal zone of up to 500 m.

Sea-level changes during the past million years have been reproduced by model simulations (Fig. 1.13). Model results compare well with the longest Quaternary sea-level record hitherto obtained, that from the Red Sea, which spans 470,000 years (Fig. 1.13). The Red Sea is an evaporative basin, separated from the Arabian Sea by a shallow sill, which turns highly saline when sea level drops. Oxygen isotope ratios of seawater are sensitive to salinity changes. Because deep-sea foraminifera take up their oxygen from seawater, the oxygen isotopes in shells of foraminifera preserved in cores is a good measure of the level of the Red Sea and it allows the reconstruction of sea-level changes over several glacial-interglacial cycles. Both modelled and proxy records show that over millennial timescales, sea level behaves remarkably predictably, with lowstands during the coldest periods of 120 ± 10 m below present sea level, and highstands during the peak of interglacials, to within 10 m of present sea level. What is less certain, however, is the behaviour of sea level on centennial timescales, particularly during sea-level highstands. It has been suggested that during the last interglacial, when sea level was up to 9 m higher than present (Kopp et al., 2009), sea-level fluctuations were very rapid, with rates of rise of, on average, 1.6 m per century (Rohling et al., 2008). If correct, sea level during the last interglacial (marine isotope stage 5e) may be a reasonable analogue for future sea-level changes, when sea-level rise is predicted by some authors to be of similar magnitude (e.g. Vermeer and Rahmstorf, 2009). Marine isotope stage 11 may be the best analogue for future climate, because orbital (Milankovitch) forcing was broadly similar to present and near-future conditions. Perhaps reassuringly, sea-level behaviour during stage 11 was less erratic than during stage 5e (Rohling et al., 2010), but further research into sea-level changes during previous interglacials is needed, especially from a wider range of archives, to determine which sea-level behaviour is 'typical' for global conditions that are a few degrees warmer than the present.

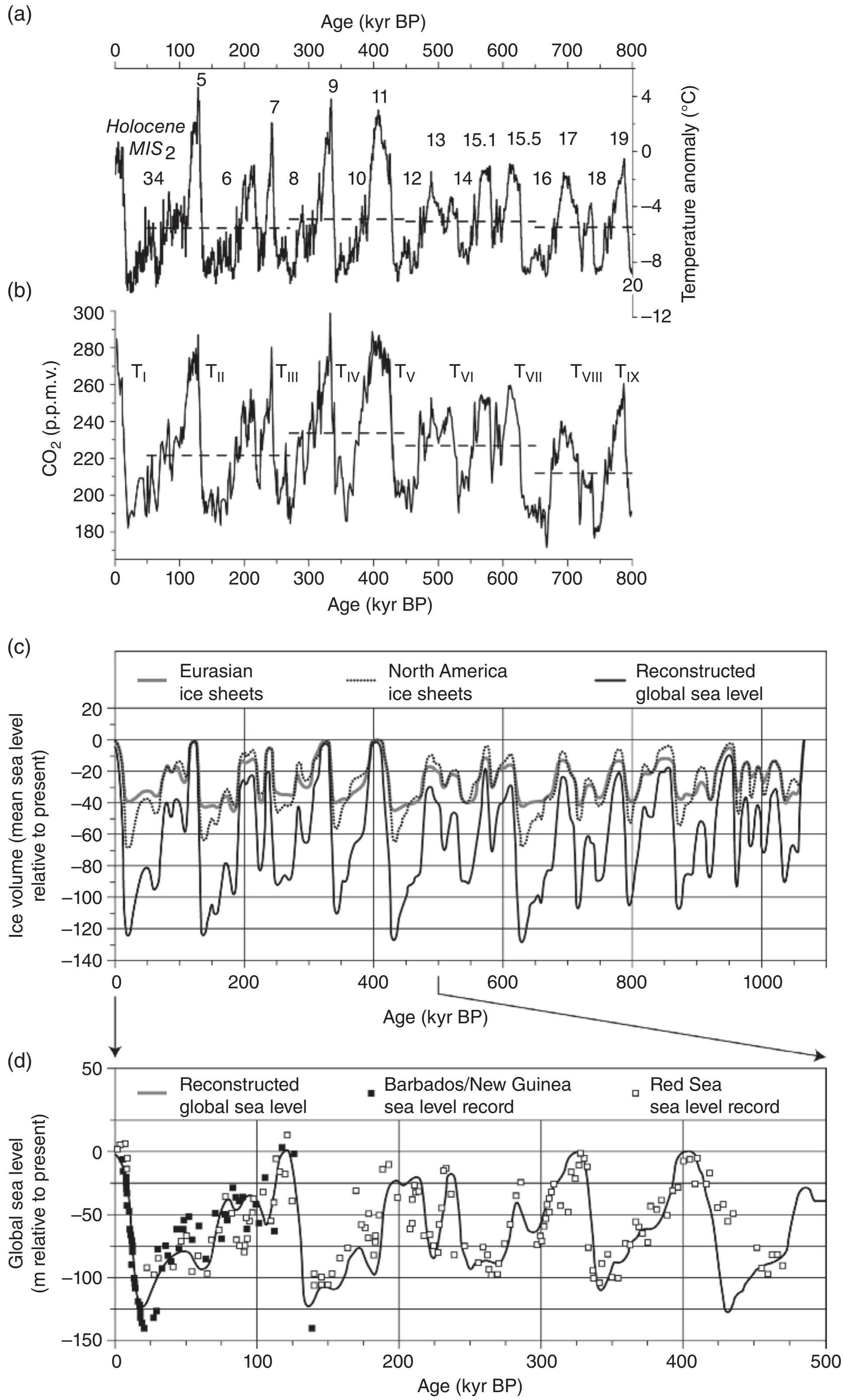

Fig. 1.13 Temperature (a) and carbon dioxide record (b) from Antarctica. (Source: Lüthi et al. 2008. Reproduced with permission of Nature Publishing Group.) (c) Modelled global sea-level record for the past 1M years, with contributions from Eurasian and North American ice sheets (Source: Bitanja et al. 2005). In (d), the model output is compared with the longest Quaternary sea-level record in the world, the record from the Red Sea (Source: Siddall et al. 2003), and with coral reef data from Barbados and New Guinea (Source: Lambneck and Chappell 2011).

1.3.2 Present and future climate change

Although sea-level change is an ultimate driver of coastal change (in the sense that it controls the position of the coast on the Earth's surface), it is not the only factor related to climatic change that affects the coast (Nicholls et al. 2007). Global warming raises the temperature of coastal waters, produces ocean acidification, changes storm patterns, and increases precipitation with major effects on coastal systems. Many climate-driven changes to our coasts are already underway (Table 1.1).

Temperature change and atmospheric greenhouse gas concentrations are intrinsically linked (Box 1.2). Future impacts of climate change depend to a large extent on the

Table 1.1 Climate drivers and their effects on coasts. (Source: Adapted from Nicholls et al. 2007.)

Climate driver	Trend	Effects
CO_2 concentration	Rising, 0.1 pH unit since 1750	Ocean acidification
Sea-surface temperature	Rising, 0.6°C since 1950	Circulation changes, sea-ice reduction, coral bleaching and mortality, species migration, algal blooms
Sea level	Rising, 1.7 ± 0.5 mm/yr since 1900	Flooding, erosion, saltwater intrusion, rising groundwater table and impeded drainage
Storm intensity	Rising	Erosion, saltwater intrusion, coastal flooding
Storm frequency, storm tracks, wave climate	Uncertain	Altered storm surges and storm waves
Run-off	Variable	Alterations in flood risk, water quality, fluvial sediment supply, circulation and nutrient supply

CONCEPTS BOX 1.2 Climate change and radiative forcing

The energy derived from the Sun controls the Earth's climate, but solar energy is reflected, absorbed and re-emitted by the Earth's surface and its atmosphere. The properties of the Earth's surface, through albedo effects, and the composition of the atmosphere, primarily through concentrations of greenhouse gases, play a critical role in regulating the Earth's temperature.

Climate change occurs because all three controlling mechanisms (the Sun's energy, the properties of the Earth surface, and the composition of the atmosphere) are subject to change on various timescales. The amount of solar energy that reaches the Earth varies with changes in the orbit of the Earth (e.g. Milankovitch cycles). The Earth's albedo (or its reflectivity) changes with the waxing and waning of ice sheets, which is an example of positive feedback in the climate system. Solar energy is reflected and absorbed in the atmosphere by dust particles and aerosols, which also have an effect on cloudiness, producing cooling through feedbacks. The most important greenhouse gases are water vapour, carbon dioxide (CO_2) and methane (CH_4), and their concentrations are subject to change through natural causes (e.g. volcanic gas emissions and exchange with the ocean) and through human emissions. Water vapour is the strongest greenhouse gas. Its concentration in the atmosphere depends on surface temperature and is therefore also prone to feedback.

The term 'radiative forcing' is used to describe how certain factors can alter the balance between incoming and outgoing energy. It is expressed in Watts per square metre (W/m^2) and is positive if a factor causes warming and negative if it causes cooling. The Intergovernmental Panel on Climate Change (IPCC) reports radiative forcing relative to a pre-industrial background at 1750. The total contribution of greenhouse gases to radiative forcing during a certain time period is determined by its change in concentration and by its strength, or effectiveness, in affecting the balance between incoming and outgoing energy. For example, in 2005 CO_2 had a radiative forcing of 1.49 to 1.83 W/m^2, whereas cloud albedo effects generated a cooling of –1.8 to –0.3 W/m^2. The net total contribution of anthropogenic factors was 0.6–2.4 W/m^2 (90% confidence range), mostly due to the emissions of greenhouse houses since the Industrial Revolution.

Some greenhouse gases (e.g. CO_2, CH_4 and nitrous oxide, NO_2) are stable and persist in the atmosphere for decades or longer. Changes in their concentrations over time have been accurately measured in gas bubbles preserved in ice cores (e.g. Fig. 1.13b) and, since the 1950s, by instruments. The concentration of atmospheric CO_2 has increased from 280 ppm (parts per million) in pre-industrial times to 400 ppm in 2013. As a consequence, the average global temperature during the 20th century increased by 0.74 ± 0.18 °C (Solomon et al., 2007), while sea level rose by 0.17 ± 0.03 m (Church and White, 2006). The IPCC states that sea level will continue to rise for centuries or millennia, even if radiative forcing were to be stabilized.

Table 1.2 Selected features of the Intergovernmental Panel on Climate Change (IPCC) emission scenarios from the Special Report on Emission Scenarios (SRES). Data are for the year 2100 and are from Nakićenović et al. (2000). Temperature forecasts are from Solomon et al. (2007). (Source: Data from Nakićenović et al. 2000.)

Family			A1		A2	B1	B2
Scenario group	1990	A1F1	A1B	A1T	A2	B1	B2
Population (billion)	5.3	7.1	7.1	7	15.1	7	10.4
World gross domestic product (GDP) (trillion 1990$US/yr)	21	525	529	550	243	328	235
CO_2 emissions from fossil fuels (GtC/yr)	6.0	30.3	13.1	4.3	28.9	5.2	13.8
Percentage of carbon-free energy usage	18	31	65	85	28	52	49
Range of projected temperature increase (°C)	0	2.4–6.4	1.7–4.4	1.4–3.8	2.0–5.4	1.1–2.9	1.4–3.8

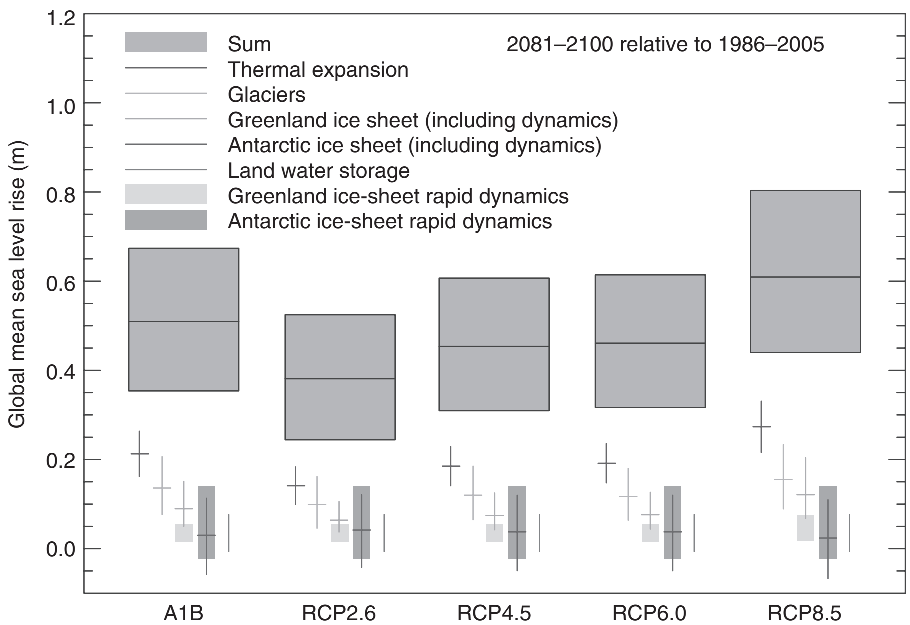

Fig. 1.14 Projections from process-based models with likely ranges and median values for global mean sea-level rise and its contributions in 2081–2100 relative to 1986–2005 for the four RCP scenarios and scenario SRES A1B used in the AR4. From Church et al. (2014). (Source: Meehl et al. 2007. Reproduced with permission of the IPCC.)

effects of human activities and the amount of carbon dioxide (CO_2) and other greenhouse gases that are likely to be emitted. In their assessment reports, the Intergovernmental Panel on Climate Change (IPCC) developed a range of emission scenarios that reflect a range of potential development projections of our planet. The set of scenarios that was used in the Fourth Assessment Report (AR4) was published in the Special Report on Emissions Scenarios (SRES) and are termed the SRES scenarios (Nakićenović et al., 2000). For these projections, the future state of the world was described in demographic, economic and political terms as 'storylines', resulting in four families of projections (Table 1.2). The A1 family has three 'marker scenarios'; the A2, B1 and B2 family have one each. The six marker scenarios are used by climate modellers to drive their climate models. For sea-level change in the year 2100, this resulted in projections that range between 0.18 m for the lower end of the B1 scenario to 0.59 m for the upper end of the A1F1 scenario (excluding rapid ice-sheet dynamics). These changes were driven by projected temperature rises of 1.1 °C for the low end of the temperature range forecast under the B1 scenario, to 6.4 °C for the maximum rise considered possible under the A1F1 scenario. By the time of the publication of the Fifth Assessment Report in early 2014 modelling of ice-sheet dynamics had improved. The latest IPCC projection for the high-emission scenario is 0.52–0.98 m (relative to 1986–2005), producing rates of sea-level rise of up to 16 mm/yr (Church et al., 2014). The emission scenarios, except for A1B, have been replaced by so-called representative concentration pathways RCPs; (van Vuuren, 2011; see Chapter 2; Fig. 1.14).

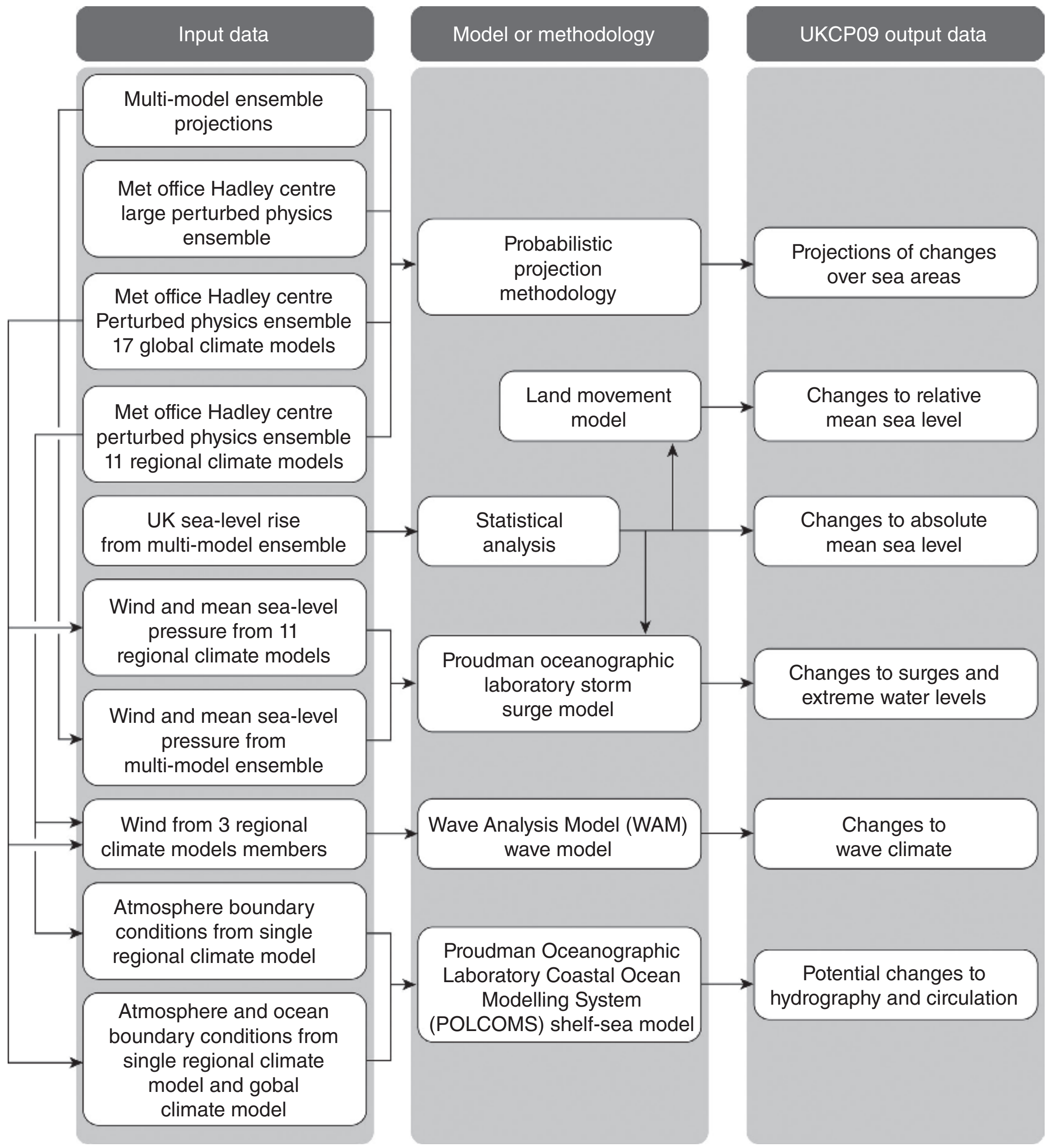

Fig. 1.15 Example of a regional approach to predicting sea-level changes. UKCP09, United Kingdom Climate Projections 2009. (Source: Adapted from Lowe et al. 2009. © UK Climate Projections, 2009.)

IPCC projections, including those for sea level, are global in scope. For coastal impact assessment they should ideally be downsized to a regional scale that is of practical use to coastal planners and managers (Gehrels and Long, 2008). This has been done in the UK, for example, where the UK Climate Impact Programme (UKCIP) has translated the SRES storylines into national and regional scenarios relevant to the UK economy. Moreover, the projections for sea-level change also include processes that act on a regional and local scale, including local land movement, tides, wind and wave climate (Fig. 1.15; Lowe et al., 2009).

Since the publication of the IPCC AR4 in 2007 there has been much debate about the accuracy of the sea-level predictions, mainly because they failed to include adequately dynamical glaciological processes such as ice-stream acceleration, basal lubrication and shelf breakup. The IPCC-estimated range of 0.09–0.17 m in the AR4 was almost certainly too low to account for these processes, but realistic modelling is notoriously difficult. Alternative semi-empirical projections that bypass the modelling difficulties include those based on the relationship between historical temperatures and sea level (Vermeer and Rahmstorf, 2009) and those based on palaeodata from the last interglacial (Rohling et al., 2008). The maximum sea-level predictions for the year 2100 based on these approaches are 1.6–1.9 m. In the UK, the H++ sea-level scenario (Lowe et al., 2009), which estimates that regional sea-level rise could be as high as 1.9 m by the year 2100, is partly based on the last interglacial analogue of Rohling et al. (2008). How accurate these projections are remains to be seen, but it is interesting to note that since IPCC predictions began, in 1990, global sea level has followed a path than overlaps with the upper range of their predictions (Rahmstorf et al., 2007).

1.4 Modelling coastal change

1.4.1 Need for adequate models

Climate represents a key environmental boundary condition for coastal systems. Climate change is therefore expected to have a major effect on coastal processes and coastal morphology. The two most important consequences of climate change are sea-level change (Chapter 2) and increased storminess (Chapters 4 and 5), both resulting in coastal erosion and flooding. There are, however, many other consequences, including the melting of permafrost cliffs and reduction in ice cover, resulting in increased erosion rates along cold coasts (Chapter 14), changes in precipitation affecting cliff instability and cliff erosion (Chapter 15), and the increase in sea-surface temperature causing coral bleaching (Chapter 16).

The ability to forecast confidently the consequences of climate change to the coastal zone is of paramount importance, not only for mitigating any adverse changes, but also to help reduce our vulnerability to environmental changes in the coastal zone through planning (Kay and Alder, 1999). This is not an easy task because predicting climate change effects to the coast comprises a number of linked steps: (1) consider appropriate greenhouse gas emission scenarios arising from our behaviour; (2) application of coupled ocean–atmosphere models to predict climate change; (3) evaluating the effect of climate change on sea level and wave climate; and (4) predicting the effect of the change in coastal drivers on nearshore sediment transport and morphological change. During each step, the feedback between drivers and responders needs to be considered, and with each step the amount of uncertainty in the predictions increases. We will focus here on the final step: models that link the coastal processes to geomorphology and evolution.

Any model is a representation, and therefore a simplification, of the real world, but the degree of abstraction (or its reverse: the level of complexity) varies hugely amongst models. In this section we consider, on a scale from simple to complex, different types of models: conceptual, empirical, behaviour-oriented, and process-based morphodynamic models. A special class of models to be discussed are physical models, which are scaled-down versions of the real world. The terminology used here is somewhat loose; for example, all models can be considered conceptual and the two most sophisticated models both have a strong empirical basis, but through the examples shown here the main characteristics of the different types of models will be made clear. For ease of comparison, the examples used to illustrate the different modelling approaches all pertain to the same coastal process: the response of barrier systems to storms and sea-level rise.

1.4.2 Conceptual models

Conceptual models provide a qualitative description, often in graphic form, of coastal systems and their main governing processes and functioning. They are generally developed by synthesizing generalities from a large number of field observations and are the result of inductive reasoning. They are often linked to classifications where they help identify different states (e.g. Australian beach state model; Wright and Short, 1984), and can also be used to predict qualitative changes in the environment by recognizing sequential stages of development (e.g. coral-reef island formation model; Kench et al., 2005). They are useful pedagogic tools when they can help bring across complex issues and enable case studies to be placed in a more general scientific framework (e.g. ternary delta model; Galloway, 1975). Conceptual models also help to identify key processes that can then be formalized in more sophisticated models to be used for predictions. They are, however, significantly oversimplifications of the real world and their practical use is generally limited to describing the current, but not the future, state of coastal systems. If used for prediction, conceptual models can at best predict the direction of change, but not the rate of change.

Figure 1.16 represents a conceptual model of the response of (gravel) barriers to increased wave conditions and raised water levels (Orford et al., 2003). According to this model, the critical factor in determining the response of barriers to increased hydrodynamic forcing is the difference in height between the elevation of the crest of the barrier and the wave run-up level, known as ‘freeboard’. Positive freeboard occurs when the maximum run-up does not reach the barrier crest, and this will result in a relatively minor morphological change to the seaward face of the barrier. When the freeboard is zero or has a small negative value, the maximum run-up just reaches the crest of the barrier; this is referred to as ‘overtopping’ and causes

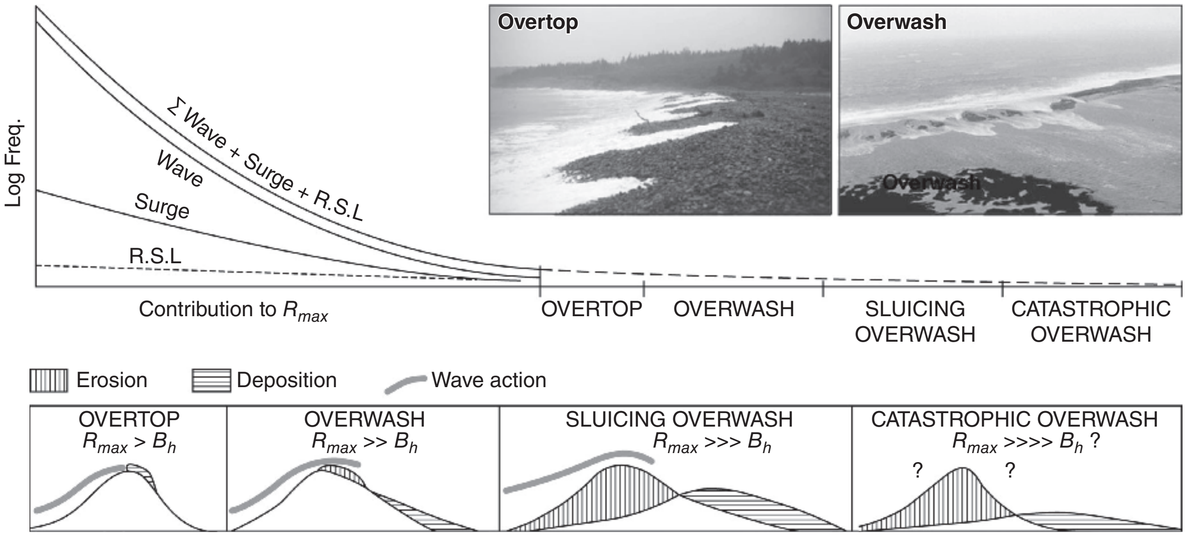

Fig. 1.16 Conceptual model showing the different stages of barrier response to increasing wave and water-level conditions. R_{max} and B_h refer to maximum run-up height and barrier height, respectively. For positive freeboard $R_{max} < B_h$, and for negative freeboard $R_{max} > B_h$. The maximum run-up height is the summation of storm surge and wave run-up, but for long timescales the relative sea-level rise also needs to be taken into consideration. The photographs represent overtopping of a Grand Desert gravel barrier in Nova Scotia, Canada and overwashing at Hurst Castle spit in Hampshire, UK (Source: Andrew Bradbury). RSL, relative sea level. (Source: Adapted from Orford et al. 2003. Reproduced with permission of American Society of Civil Engineers.)

sediment accretion on the top of the barrier, leading to an increase in the barrier crest elevation. A relatively large negative freeboard is accompanied by 'overwashing' of the barrier, with run-up events frequently extending across the crest of the barrier and running down the landward face of the barrier. This causes flattening of the barrier crest and deposition of sediment behind the barrier in the form of overwash. Increasing the negative freeboard even more may result in 'sluicing overwash' or even the wholesale destruction of the barrier feature. This conceptual model usefully illustrates the different stages of barrier response as a function of the intensity of the forcing, but does not indicate the exact thresholds for the morphological responses, nor the rate of morphological change.

1.4.3 Empirical models

In contrast to conceptual models, empirical models express relations between different elements of coastal systems in quantitative terms through the use of equations or parameters. Their formulation generally relies on a statistical analysis of numerical data collected in the field or the laboratory. Empirical models are particularly useful when they are combined with conceptual models, to provide the latter with a more quantitative foundation (e.g. the use of the dimensionless fall velocity and the relative tide range for classifying beaches; Masselink and Short, 1993). When the empirical correlations are of a more generic form, for example the relationship between tidal prism and cross-sectional area of tidal inlets in estuarine environments (Townend, 2005), these equations can be used in more comprehensive models. The most widespread use of empirical models for predicting future coastal change is to quantify current change using statistical techniques (cliff recession, coastal retreat, salt-marsh accretion), and extrapolating this change into the future. By necessity, such models are site-specific.

When the interest is in understanding the functioning of coastal systems, empirical models are generally based on comparing or combining hydrodynamic and geomorphological parameters. In the case of barrier response to storms, the barrier breach model proposed by Bradbury (1998) is a good example. The model shown in Fig. 1.17 is based on the notion that the likelihood of a barrier breaching depends on the balance between the disturbing forces (parameterized by the wave steepness) and the resisting forces (parameterized by an inertia parameter based on barrier geometry and wave height). This notion is similar to that encapsulated by the conceptual model discussed earlier. However, the conceptual model has been taken one step further by parameterizing the disturbing and stabilizing forces, and using field observations to identify the thresholds between overwashing, overtopping and no change to the barrier crest.

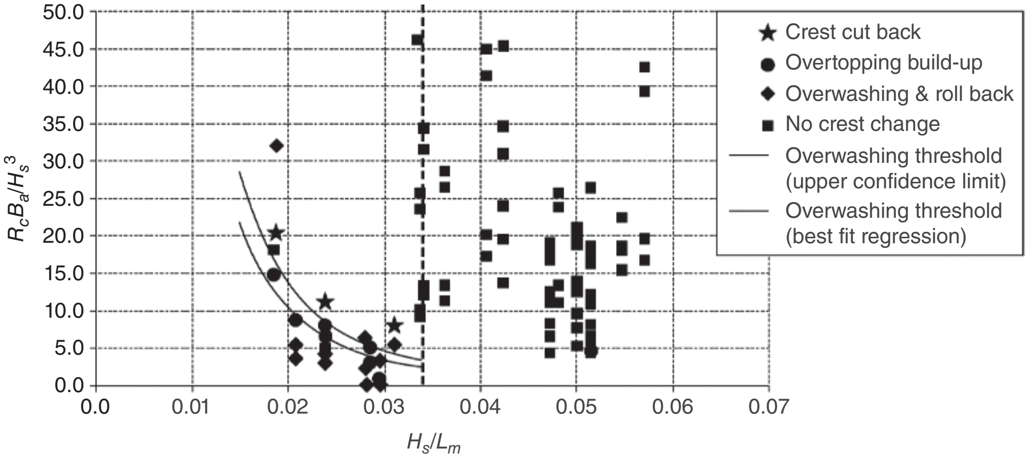

Fig. 1.17 Testing of the empirical model of Bradbury (1998) using Hurst Spit, UK. The model is based on barrier inertia parameter R_cB_a/H_s^3 and wave steepness H_s/L_m, where R_c is the barrier freeboard, B_a is the cross-sectional area of the barrier above still water level, H_s is the significant wave height, and L_m is the deep water wave steepness based on the mean wave period T_m ($L_m = gT_m^2/2\pi$, where g is gravity). The line represents the overwashing threshold, whereby conditions below the line predict barrier overwash. (Source: Data from Bradbury et al. 2005.)

1.4.4 Behaviour-oriented models

Behaviour-oriented models are realizations of coastal systems that attempt to reproduce the dominant behaviour without too much concern about the actual processes (e.g. estuarine equilibrium model of Townend and Pethick, 2002; rocky coast evolution model of Trenhaile, 2000). They tend to aggregate the complex processes into a number of simple parameterizations that can be used not only to conceptualize, but also quantify, coastal behaviour. Such models make quantitative predictions of coastal evolution and are therefore more sophisticated than conceptual models, and they may include parameterizations from empirical models. Behaviour-oriented models are often computationally efficient, because only the behaviour is modelled, rather than the detailed processes. They are therefore very useful for sensitivity analysis – for exploring 'what if' questions – thereby illuminating which aspects of the study are most in need of further study, and where more empirical data are most needed. They are most appropriate for systems that are not very well understood and/or that are very complex, because it is for these systems that sophisticated process-based models are not available or simply not good (enough). They are also useful when the input parameters are not very well constrained – there is no point in using a sophisticated model when the input data are not reliable (GIGO; garbage in, garbage out).

One of the most widely used behaviour-oriented models in coastal research, despite its shortcomings (Cooper and Pilkey, 2004), is the application of the Bruun rule to predict the effect of sea-level rise on barrier systems (Dean, 1991). According to this model, the underwater shoreface profile is described by a simple exponential profile whose overall steepness is only a function of the sediment size, and its spatial extent is determined by the closure depth, itself a function of wave conditions and sediment size. Under conditions or rising or falling sea levels, it is assumed that the profile shape is maintained, but the profile is shifted up or down, respectively, while mass is being preserved. This simple model has been used as the basis for the Shoreface Translation Model (STM) model of Cowell et al. (1995), who added a number of capabilities to the model, including back-barrier sedimentation, longshore sediment transport and dune formation. Figure 1.18 illustrates output of the STM and shows that realistic behaviour of the response of barriers to sea-level rise can be reproduced.

1.4.5 Process-based morphodynamic models

Process-based morphodynamic models include all the relevant hydrodynamic processes and link these to the morphology through sediment transport. The sediment continuity equation is used to update the evolving morphology (Box 1.3). Feedback between morphology and hydrodynamics is accounted for and such models essentially cycle through the morphodynamic loop depicted in Fig. 1.5. The models therefore include both negative and positive feedback. These models can include elements of conceptual, empirical and behaviour-oriented models, but the key element is that process-based models attempt to account for the actual processes. Figure 1.20 shows the output produced by XBeach, a process-based morphodynamic model specifically designed to predict the response

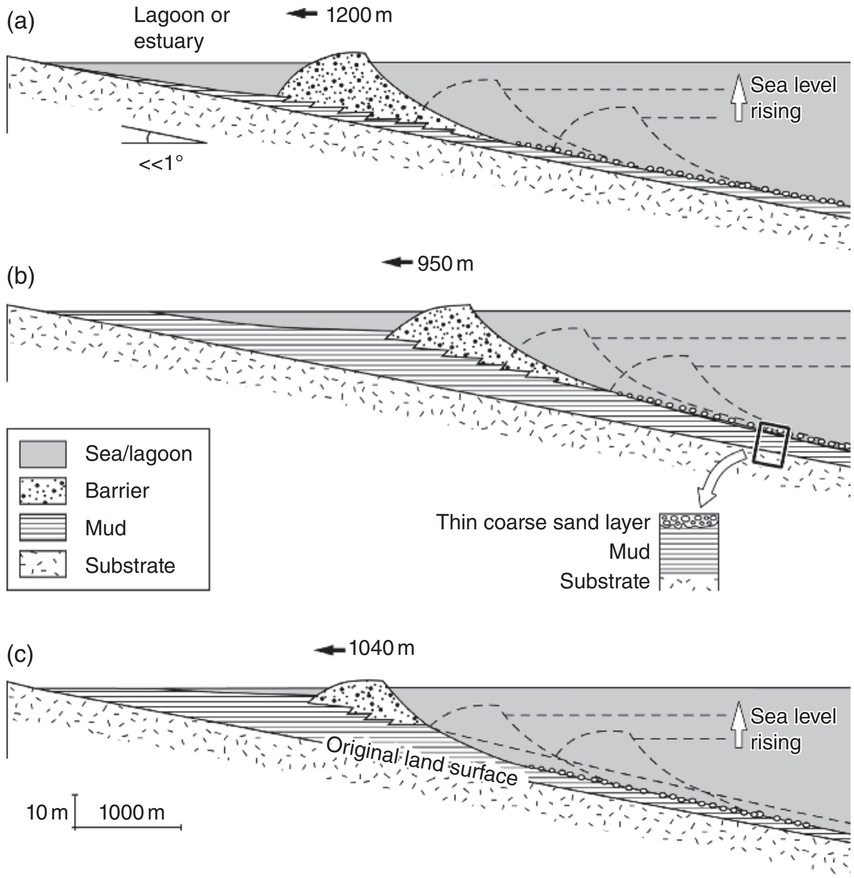

Fig. 1.18 Example of output of the Shoreline Translation Model (STM) of Cowell et al. (1995) showing the response of a barrier to sea-level rise. In Case (a), a small amount of mud deposition in the back barrier lagoon has little effect on barrier size or coastal behaviour, but with mud accumulating almost as fast as sea level is rising, as in Case (b), the barrier decreases in size and recession slows down. In Case (c), a deficit in the nearshore sediment budget, for example induced by a negative littoral drift differential, is superimposed on Case (b), resulting in a reduction in the size of the barrier and an increase in the recession rate. In addition to showing the different stages of barrier recession, the model also reproduces the stratigraphy, and in all cases lagoonal muds conformably blanket the substrate beneath and to the seaward of the barrier; after the barrier passes, wave reworking ensures that the bed is soon veneered by a coarse lag. (Source: Roy et al. 1994. Reproduced with permission of Cambridge University Press.)

of sandy barriers to hurricanes (Roelvink et al., 2009). The model includes all the key hydrodynamic processes, such as wave transformation (refraction, breaking and swash), nearshore currents and sediment transport, and is able to reproduce adequately barrier overwash and even breaching. Due to the nature of coastal morphodynamics, characterized by multiple positive and negative feedbacks, and the cumulative properties of coastal evolution, the most realistic, and therefore sophisticated, process-based models tend to work best at the shorter timescales. Over longer timescales, models' predictions rapidly diverge from observations or the model simply runs off the rails and crashes.

Whereas conceptual, empirical and behaviour-oriented models are mostly used in an explorative sense as research tools, the predictions generated by process-based numerical models are increasingly being used as a basis for policy decisions. This elevated status of this class of models, therefore, requires a lot more scrutiny with regards to the modelling results. Oreskes et al. (1994) warns of the dangers of overconfidence in modelling results due to an inappropriate consideration of what numerical models really are. It is general modelling practice to compare model predictions with observations, and when the former are consistent with the latter, usually after extensive model calibration (i.e. manipulation of the model input parameters to obtain a match between observation and model output), the model is considered verified (or validated). To claim that a model is verified is to say that its truth has been demonstrated, which implies its reliability as a basis for decision-making. Numerical models are, however, representations of the truth and can even be

METHODS BOX 1.3 Sediment continuity equation

Process-based morphodynamic models compute instantaneous flow velocities and sediment transport rates, and, because morphodynamic models take account of the morphodynamic feedback, the morphology is updated regularly to reflect the effect of the evolving morphology on the hydrodynamics. In the interest of computing efficiency, the time step for computing the hydrodynamics is generally smaller than that for updating the morphology. The ratio between the morphological step time and the hydrodynamic step time is generally $O(100–1000)$ and model output is very sensitive to this ratio.

The equation used for updating the morphology on the basis of the sediment transport rates is the sediment continuity equation, which in its differential formulation reads as:

$$\frac{dh}{dt} = (1-n)\left(\frac{dQ_x}{dx} + \frac{dQ_y}{dy}\right) + \frac{dV}{dt} \quad (1.1)$$

where dh/dt is the change in bed elevation over time, $(1–n)$ is sediment porosity, dQ_x/dx is cross-shore gradient in the cross-shore sediment flux, dQ_y/dy is longshore gradient in the longshore sediment flux, and dV/dt represents local sediment loss and gains, for example due to local sediment production, abrasion, nourishment and dredging. The sediment fluxes Q in the equation are volumetric fluxes (cubic metres per unit of time); if the fluxes are mass fluxes (kilograms per unit of time), then the sediment density will also need to be taken into account.

Process-based morphodynamic models generally operate on a rectangular grid; therefore flow velocities and sediment transport rates are computed numerically for each individual grid cell. Figure 1.19 illustrates how eqn. 1.1 is applied for a single grid cell. In the present example, the size of the grid cells are $\Delta x = 5$ m and $\Delta y = 10$ m, and the morphological time step dt is 5 minutes. It is further assumed that: cross-shore transport Q_x into and out of the grid cell is 27 m^3 and 23 m^3, respectively; the longshore transport into and out of the grid cell is 124 m^3 and 123 m^3, respectively; $n = 0.4$; and $dV/dt = 0$. In this case, the total volume of sediment entering the grid cell over dt is 5 m^3 ($\Delta Q_x = 4$ m^3 + $\Delta Q_y = 1$ m^3). This gain in sediment is distributed over the 50 m^2 grid cell; therefore, the increase in bed elevation h is 0.1 m. Since the morphological time step was 5 minutes, the rate of bed-level change dh/dt is 0.02 m per minute.

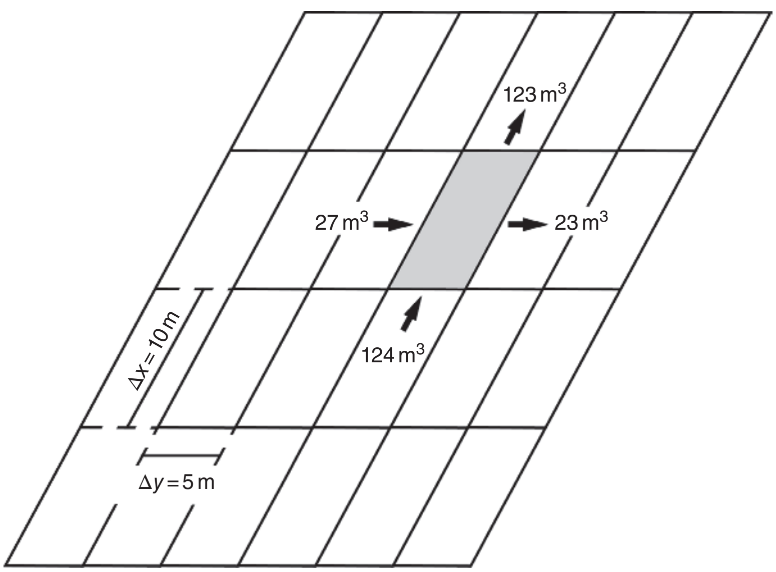

Fig. 1.19 Schematic illustrating the use of the sediment continuity equation for grid-based numerical models. See text for explanation.

considered a form of highly complex scientific hypotheses. The philosopher Karl Popper has taught us that one cannot prove a hypothesis, but one should attempt to refute it; therefore a model can never be verified. Anyone involved with modelling is aware that good agreement between model predictions and observations can be attained in more than one way (by tuning different parameters), and such good agreement does not constitute confirmation

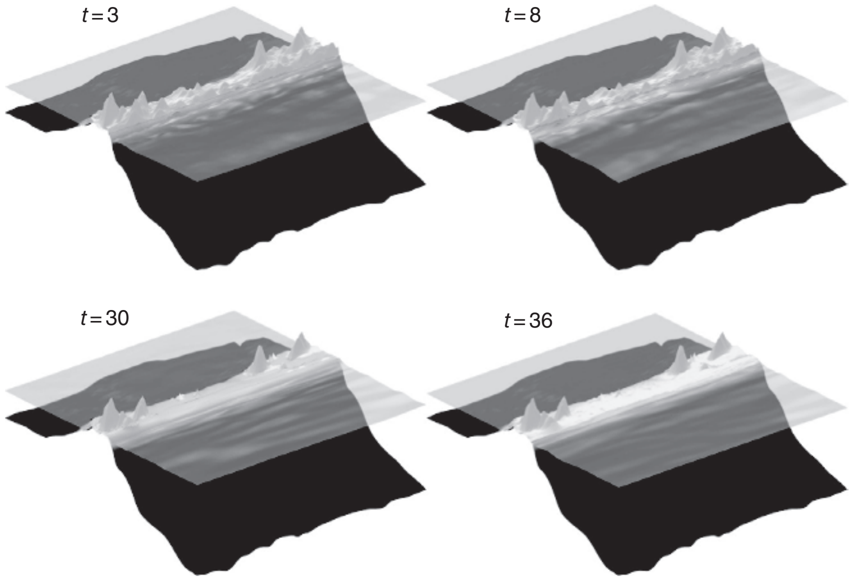

Fig. 1.20 Numerical simulation using the XBeach model of the impact of Hurricane Ivan (the largest of five hurricanes to strike to US coast in 2004) on a 2-km wide section of Santa Rosa Island, a narrow barrier island in Florida. The Gulf of Mexico is in the lower right and the back barrier bay in the upper left of all four panels. The simulation was run over 36 hours and the four panels represent the water-level variation and barrier morphology at different time steps. The maximum storm surge and significant wave height used for forcing the model were 1.75 m and 7 m, respectively, and were attained at t = 18 hours. Note the extensive overwashing at t = 30 hours and the complete destruction of the dunes in the central part of the modelled region. (Source: McCall et al. 2010. Reproduced with permission of Elsevier.) For colour details, please see Plate 4.

that the model is 'true'. At best, observations can support, or confirm, the model, and the greater the number and diversity of confirming observations, the more probable it is that the conceptualization embodied in the model is not flawed. But, whatever the level of sophistication and the agreement between predictions and observations, a model will always be a representation, and care should be taken in interpreting model results.

1.4.6 Physical models

Prior to the availability of computers, the best way to investigate the effect of coastal response to changing boundary conditions, other than making field observations, was through the use of physical modelling: the construction of physical models of the coastal system case in laboratories. Most commonly, the models are scaled-down versions of reality, but some large laboratory facilities enable the construction of models at the prototype scale (1 : 1). This methodology has been widely used in coastal engineering, for example, to investigate the effect of submerged breakwaters on shoreline stability (Ranasinghe and Turner, 2006), and still provides a good alternative to numerical process-based models.

An example of a physical model experiment designed to investigate the stability of gravel barriers is the BARDEX experiment held in the Delta Flume (length = 250 m; width = 5 m; depth = 7 m) in The Netherlands (Williams et al., 2012; Fig. 1.21). During this project, a fine-gravel barrier (D_{50} = 10 mm) was constructed in the wave flume, with a height of 4 m and a width at its base of 50 m. The barrier was heavily instrumented by a range of equipment, including bed-level sensors, current meters, water depth sensors, and was regularly profiled to measure its morphology. The barrier was constructed with a 'sea' and a 'lagoon' at either side, and by raising and lowering the water level in the lagoon (by 1 m relative to the sea level), the water table in the barrier was elevated and lowered in a controlled and reproducible manner relative to the mean sea level. At the end of the experiment, the barrier was subjected to increasingly elevated water-level conditions, eventually causing extensive overwashing of the barrier. Such measurements could not have been collected in a field setting, and the data will not only be used to increase our understanding of barrier processes, but also to help formulating and calibrating numerical models.

The main issue with physical models is scale (except for the prototype scale) and although scaling relationships are

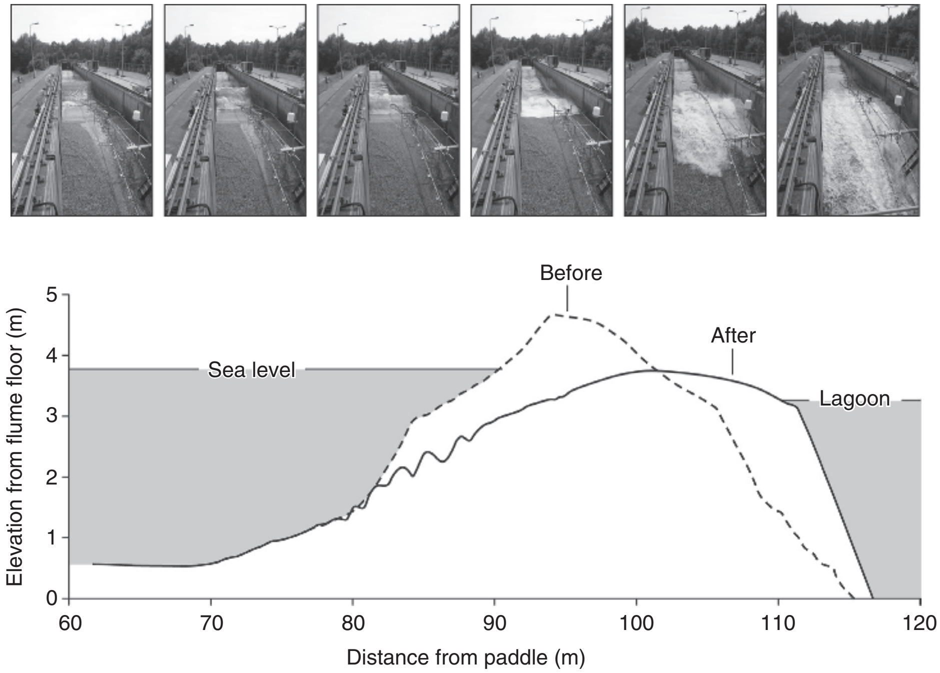

Fig. 1.21 Physical simulation of overwash on a gravel barrier constructed in the Delta Flume, The Netherlands, during the BARDEX experiment. (Source: Williams et al. 2012. Reproduced with permission of Elsevier.) The sequence of photos in the top panels represents the various stages of wave transformation of one of the overwashing waves. The bottom panel shows the morphological change after 2.5 hours of exposure to overwash, characterized by a lowering of the barrier crest by about 1 m and the development of an extensive washover deposit at the back of the barrier. The wave conditions during this test were characterized by a significant wave height of 0.8 m and a peak wave period of 8 s.

well established for interpreting the results of small-scale laboratory experiments, particularly in models that contain sediment, module scaling is never perfect. Nevertheless, physical models remain a valuable tool for coastal research.

1.5 Summary

This introductory chapter provides the theoretical framework and the scope of this book. The coast is defined as that region of the Earth's surface that has been affected by coastal processes during the Quaternary (the last 2.6M years). This definition is dynamic rather than static, and considers sediments, sea-level history and coastal processes. This book also follows the morphodynamic paradigm, according to which processes are linked with morphology through sediment transport. As the title suggests, an important focus of this book is on how global change affects coastal processes and landforms. The largest impact of future global change on coasts will be through an increase in the rate of global sea-level rise. A long-term perspective is crucial, as coastal evolution is recorded in coastal stratigraphy, which provides useful analogues for present-day and future coastal behaviour under various sea-level regimes. Models of coastal change include those based on descriptions of processes (conceptual models), those that quantify processes (empirical models) and behaviour (behaviour-oriented models), those that include morphodynamic feedback (process-based morphodynamic models), and those that are a scaled-down version of the real world (physical models). These models are useful tools for studies of coastal impact assessment and are critical to explain and forecast coastal change.

- A long-term perspective on coastal evolution is useful because the present-day coast is a product of modern and past coastal processes and is often, at least partly, controlled by inherited landforms.
- According to the coastal morphodynamic paradigm, coastal systems comprise three linked elements (morphology, processes and sediment transport) that exhibit a certain degree of autonomy in their behaviour, but are ultimately driven and controlled by environmental factors.
- Coastal systems change through positive morphodynamic feedback, and stabilize through negative morphodynamic feedback. An important property of coastal systems is self-organization, or emergence, which refers to the development of morphological features with a specific shape and/or spacing that arises from the mutual interactions between form and process.

• During the 20th century, the global temperature increased by 0.7 °C and sea level rose by 0.2 m. Currently, sea level is rising at an accelerated rate and is likely to rise about 1 m in the next 100 years.
• Models can be used to forecast coastal change due to climate change. However, regardless of the level of model complexity, it should be borne in mind that any model is a representation, and therefore a simplification, of the real world, and model predictions should be interpreted with care.

Key publications

Church, J.A., Woodworth, P.L., Aarup, T. and Wilson, W.S. (Eds), 2010. *Understanding Sea-Level Rise and Variability*. Wiley-Blackwell, Chichester, UK.
[*State-of-the-art treatise on the current understanding of sea-level changes*]

Coco, G. and Murray, A.B., 2007. Patterns in the sand: from forcing templates to self-organisation. *Geomorphology*, **91**, 271–290.
[*Very accessible discussion of self-organization applied to a range of coastal phenomena*]

Cowell, P.J. and Thom, B.G., 1994. Morphodynamics of coastal evolution. In: R.W.G. Carter and C.D. Woodroffe (Eds), *Coastal Evolution*. Cambridge University Press, Cambridge, UK, pp. 33–86.
[*Excellent account of the morphodynamic paradigm*]

Masselink, G. and Hughes, M.H., 2003. *Introduction to Coastal Processes and Geomorphology*. Arnold, London, UK.
[*Introductory book providing good framework for the current book*]

Nicholls, R.J., Wong, P.P., Burkett, V.R., et al., 2007. Coastal systems and low-lying areas. In: M.L. Parry, O.F. Canziani, J.P. Palutikof, P.J. van der Linden and C.E. Hanson (Eds), *Climate Change 2007: Impacts, Adaptation and Vulnerability*. Contribution of Working Group II to the Fourth Assessment Report of the Cambridge University Press, Cambridge, UK, pp. 315–356.
[*Comprehensive and up-to-date account of the impact of climate change on coastal systems*]

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