9 THE FUTURE: COASTAL ENVIRONMENTS

Introduction: rising sea levels

There is a strong likelihood that if temperatures climb in coming decades, so will sea levels. Indeed, on a global basis, sea levels are rising at the present time, and always have in warmer periods in the past. The reasons why sea levels will rise include the thermal expansion of seawater (the **steric effect**), the melting of the cryosphere (glaciers, ice sheets, and permafrost) and miscellaneous anthropogenic impacts on the hydrologic cycle (which will modify how much water is stored on land). Rising sea levels will have substantial geomorphologic consequences for the world's coastlines and these in turn will have an impact on a surprisingly large proportion of Earth's human population (Leatherman, 2001). As Viles and Spencer (1995) have pointed out, about 50% of the population in the industrialized world lives within 1 km of a coast, about 60% of the world's population live in the coastal zone, and two-thirds of the world's cities with populations of over 2.5 million people are near estuaries. Thirteen of the world's twenty largest cities are located on coasts.

The steric effect

As the oceans warm up, their density decreases and their volume increases. This thermal expansion, the steric effect, causes sea levels to rise. Uncertainties arise as to the rate at which different parts of the oceans will warm up in response to increased atmospheric temperatures (Gregory et al., 2001). Long-term records of ocean warming and sea-level change are also sparse.

Church et al. (2001, table 11.2) present data on global average sea-level rise due to thermal expansion during the twentieth century. Rates between 1910 and 1990 ranged between 0.25 and 0.75 mm per year while those between 1960 and 1990 ranged between 0.60 and 1.09 mm per year. The steric effect over the period 1910 to 1990 accounts for at least one-third of the observed eustatic change over that period.

With regard to the future, sea-level rise caused by thermal expansion between 1990 and 2100 is thought likely to be around 0.28 m, out of a total predicted sea-level rise due to climate change of 0.49 m (Church et al., 2001, table 11.14). In other words, it will exceed in importance the role of ice-cap and glacier melting.

Anthropogenic contributions to sea-level change

Human activities have an impact upon the hydrologic cycle in a wide range of ways, so that changes in sea level could result from such processes as groundwater exploitation, drainage of lakes and wetlands, the construction of reservoirs behind dams, and modification of runoff and evapotranspiration rates from different types of land cover. Some of these processes could accelerate sea-level rise by adding more water to the oceans, whereas others would decelerate the rate of sea-level rise by impounding water on land. Unfortunately, quantitative data are sparse and sometimes contradictory and so it is difficult to assess the importance of anthropogenic contributions to sea-level change.

Reduction in lake-water volumes

Increased use of irrigation and interbasin water transfers have contributed to reduced volumes of water being stored in lake basins. Classic examples of this include the Aral and Caspian Seas of Central Asia and Owens Lake in the southwest USA. Unfortunately, not all lake level declines are due to human actions and it is not always easy to disentangle these from the effects of natural decadal-scale climate fluctuations. Equally it is difficult to calculate the proportion of the extracted water that reaches the world ocean by runoff and evapotranspiration. Some of it may enter groundwater stores, causing groundwater levels to rise and waterlogging of the land to occur. Controversy has thus arisen on this topic. On the one hand Sahagian (2000) has argued that this could be a process of some significance, perhaps contributing around 0.2 mm per year to sea-level rise, whereas Gornitz et al. (1997) believe that the net effect of the drying up of interior lakes is probably only small and indirect.

Water impoundment in reservoirs

Recent decades have seen the construction of large numbers of major dams and reservoirs. Today, nearly 500,000 km^2 of land worldwide are inundated by reservoirs that are capable of storing 6000 km^3 of water (Gleick, 2002).

Newman and Fairbridge (1986) calculated that between 1957 and 1982 as much as 0.75 mm per year of sea-level rise potential was stored in reservoirs and irrigation projects. More recently, Gornitz et al. (1997) have estimated that 13.6 mm of sea-level rise potential has been impounded in reservoirs, equivalent since the 1950s to a potential average reduction in sea-level rise of 0.34 mm per year. However, as Sahagian (2000: 43) reported, 'the total amount of impounded water is not known because there have been no comprehensive inventories of the millions of small reservoirs such as farm ponds and rice paddies. As a result, the total contribution of impounded water to global hydrological balance has remained unclear and most likely severely underestimated.'

Groundwater mining

Groundwater mining is the withdrawal of groundwater in excess of natural recharge. Many large groundwater aquifers are currently being heavily mined, leading to massive falls in aquifer levels and volumes. Examples include the Ogallala aquifer of the High Plains in the USA, the Nubian Sandstone aquifers of Libya and elsewhere in North Africa, and the great aquifers of the Arabian Peninsula. Estimates of the volumes of groundwater that are being removed from storage on a global basis are around 1000 to 1300 km^3, but not all of it is transferred to the oceans. Nonetheless, Church et al. (2001: 657) believe that it is probably the largest positive anthropogenic contributor to sea-level rise (apart from anthropogenic climatic change), amounting to perhaps 0.2 to 1.00 mm per year. This source may not continue indefinitely, because groundwater contributions may become exhausted in some regions.

Urbanization and runoff

Urbanization leads to a net increase in total runoff from the land surface due to the spread of areas of impermeable ground (e.g., concrete, tile, and tarmac covered surfaces), which impede groundwater replenishment. In addition, water may be evacuated efficiently from urban areas in storm-water drains and sewers. Plausible rates of sea-level rise due to this mechanism are in the range of 0.35 to 0.41 mm per year.

Deforestation and runoff

As discussed in Chapter 5, deforestation can lead to increases in runoff from land surfaces. For example, experiments with tropical catchments have shown typical increases in streamflow of 400–450 mm per year (Anderson and Spencer, 1991). The reasons for this include changes in rainfall interception, transpiration, and soil structure. Combining data on rates of tropical deforestation with average values for increases in runoff, Gornitz et al. (1997: 153) came up with an estimate that this could cause a rate of sea-level rise of 0.13 mm per year.

Wetland losses

The reclamation or drainage of wetlands may release stored water that can then enter the oceans. Gornitz et al. (1997: 153) do not see this as a significant mechanism of sea-level change.

Irrigation

The area of irrigated land on Earth has gone up dramatically, amounting to *c.* 45×10^6 hectares in 1900 and 240×10^6 hectares in 1990. During the 1950s the irrigated area increased by over 4% annually, although the figure has now dropped to only about 1%. Gornitz et al. (1997) estimated that evapotranspiration of water from irrigated surfaces would lead to an increase in the water content of the atmosphere and so to a fall in sea level of 0.14–0.15 mm per year. Irrigation water might also infiltrate into groundwater aquifers, removing 0.40–0.48 mm per year of sea-level equivalent.

Synthesis

As we have already noted, individual mechanisms can in some cases be significant in amount, but some serve to augment sea-level rise and some to reduce it. Gornitz et al. (1997: 158) summarized their analysis thus:

> Increases in runoff from groundwater mining and impermeable urbanized surfaces are potentially important anthropogenic sources contributing to sea-level rise. Runoff from tropical deforestation and water released by oxidation of fossil fuel and biomass, including wetlands clearance, provide a smaller share of the total. Taken together, these processes could augment sea level by some 0.6–1.0 mm/year.
>
> On the other hand, storage of water behind dams, and losses of water due to infiltration beneath reservoirs and irrigated fields, along with evaporation from these surfaces could prevent the equivalent of 1.5–1.8 mm/year from reaching the ocean. The net effect of all of these anthropogenic processes is to withhold the equivalent of 0.8 ± 0.4 mm/year from the sea. This rate represents a significant fraction of the observed recent sea-level rise of 1–2 mm/year, but opposite in sign.

The IPCC (Church et al., 2001: 658) came to three conclusions.

1 The effect of changes in terrestrial water storage on sea level might be considerable.
2 The net effect on sea level could be of either sign.
3 The rate of hydrologic interference has increased over the past few decades.

Permafrost degradation

Increases in temperatures in cold regions will lead to substantial reductions in the area of permanently frozen subsoil (permafrost). In effect, ground ice will be converted to liquid water, which could in principle contribute to a rise in sea level. However, a proportion, unknown, of this water could be captured in ponds, thermokarst lakes, and marshes, rather than running off into the oceans. Another uncertainty relates to the volume of ground ice that will melt. Bearing these caveats in mind, Church et al. (2001: 658) suggest that the contribution of permafrost to sea-level rise between 1990 and 2100 will be somewhere between 0 and 25 mm (0–0.23 mm per year).

Melting of glaciers and sea-level rise

Many glaciers are expected to be reduced in area and volume as global warming occurs and their meltwater will flow into the oceans, causing sea level to rise (Gregory and Oerlemans, 1998). In some situations, on the other hand, positive changes in precipitation may nourish glaciers causing them to maintain themselves or to grow in spite of a warming tendency. Other uncertainties are produced by a lack of knowledge about how the **mass balance of glaciers** will respond to differing degrees of warming. The IPCC (2001, table 11.14) suggests that between 1990 and 2100 the melting of glaciers will contribute *c.* 0.01 to 0.23 m of sea-level change, with a best estimate of 0.16 m, second therefore only to steric effects. The problems of mass balance estimations have been well summarized by Arendt et al., 2002:

> Conventional mass balance programs are too costly and difficult to sample adequately the > 160,000 glaciers on Earth. At present, there are only but 40 glaciers worldwide with continuous balance measurements spanning more than 20 years. High-latitude glaciers, which are particularly important because predicted climate warming may be greatest there, receive even less attention because of their remote locations. Glaciers that are monitored routinely are often chosen more for their ease of access and manageable size than for how well they represent a given region or how large a contribution they might make to changing sea level. As a result, global mass balance data are biased toward small glaciers (< 20 km^2) rather than those that contain the most ice (> 100 km^2). Also, large cumulative errors can result from using only a few point measurements to estimate glacier-wide mass balances on an individual glacier.

They used laser altimetry to estimate volume changes of Alaskan glaciers from the mid-1950s to the mid-1990s and suggested that they may have contributed 0.27 ± 0.10 mm per year of sea-level change. This is considerably more than had previously been appreciated.

Ice sheets and sea-level rise

The ice sheets of Greenland and Antarctica have proved pivotal in ideas about future sea-level changes. In the 1980s there were fears that ice sheets could decay at near catastrophic rates, causing sea levels to rise rapidly and substantially, perhaps by 3 m or more by 2100. Since that time this has become thought to be less likely and that far from suffering major decay the ice sheets (particularly of Antarctica) might show some accumulation of mass as a result of increasing levels of nourishment by snow. However, considerable debates still exist on this and there remain major uncertainties about how the two great polar ice masses may respond in coming decades and whether the West Antarctic Ice Sheet (WAIS) is inherently and dangerously unstable (see, e.g., Oppenheimer, 1998; Sabadini, 2002; van der Ween, 2002). These are issues that will be addressed in Chapter 11.

Suffice it to say now that were the WAIS to collapse into the ocean, the impacts of the resulting 5 m increase in sea level would be catastrophic for many coastal lowlands. The IPCC (2001) thought that it was very unlikely that this would occur in the twenty-first century. They believed that the contribution that the Antarctica and Greenland ice caps would make to sea-level change would be modest and less than those of glaciers or thermal expansion (the steric effect). The Greenland contribution between 1990 and 2100 was thought to be in the range −0.02 to 0.09 m, and for the Antarctic −0.17 to 0.02 m.

The amount of change by 2100

Over the years there has been a considerable diversity of views about how much sea-level rise is likely to occur by 2100. In general, however, estimates have tended to be revised downward through time (French et al., 1994; Pirazzoli, 1996) (Figure 9.1). They have now settled at best estimates of just under 50 cm by 2100. This implies rates of sea-level rise of around 5 mm per year, which compares with a rate of about 1.5 to 2.0 mm during the twentieth century (Miller and Douglas, 2004).

Land subsidence

The effects of global sea-level rise will be compounded in those areas that suffer from local subsidence as a result of local tectonic movements, isostatic adjustments, and fluid abstraction. Areas where land is rising because of **isostasy** (e.g., Fennoscandia or the Canadian Shield) or because of tectonic uplift (e.g., much of the Pacific coast of the Americas) will be less at risk than subsiding regions (e.g., the deltas of the Mississippi and Nile Rivers) (Figure 9.2).

Areas of appreciable subsidence include some ocean islands. Indeed, crucial to Darwin's model of atoll

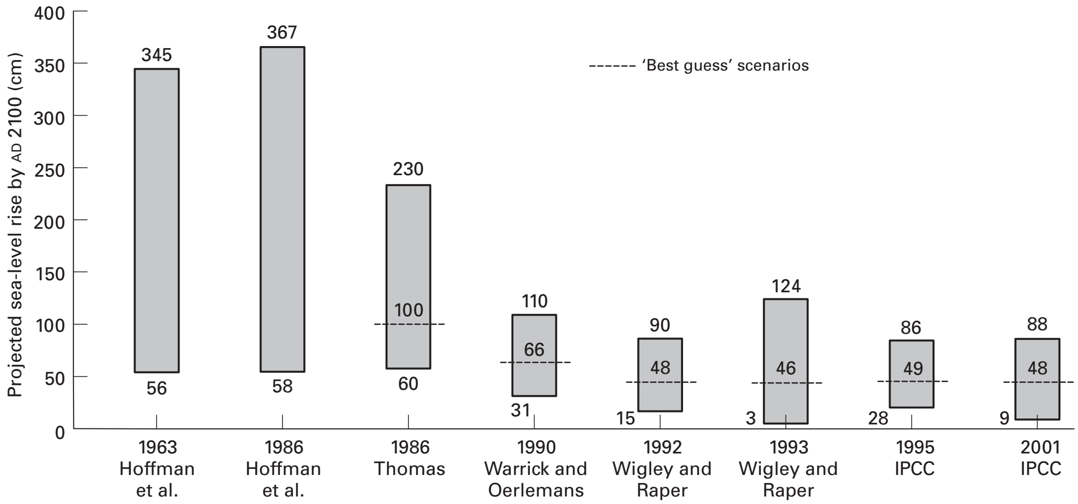

Figure 9.1 Revisions of anticipated sea-level rise by 2100 (after French et al., 1994, with modifications).

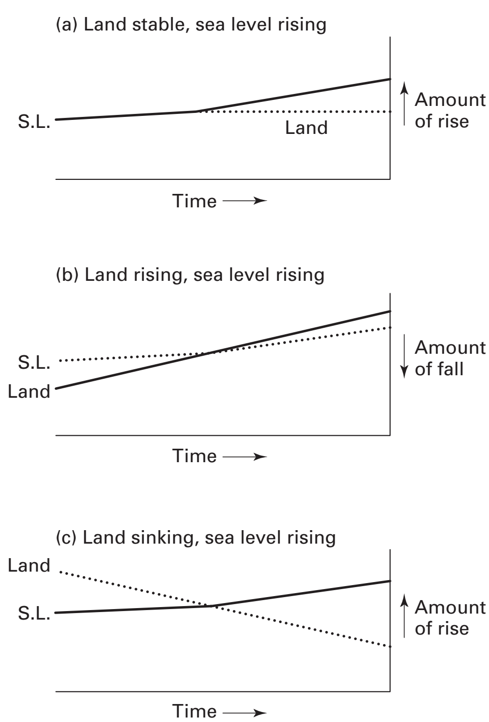

Figure 9.2 The effects of global sea-level rise will be either compounded or mitigated according to whether the local environment is one that is stable, rising, or sinking.

evolution is the idea that subsidence has occurred, and the presence of guyots and seamounts in the Pacific Ocean attest to the fact that such subsidence has been a reality over wide areas. In addition there are tide gauge records from the Hawaiian Ridge that demonstrate ongoing subsidence rates at the present day (1500 mm per 1000 years for Oahu and 3500 mm per 1000 years for Hawaii). Moreover, the occurrence of Holocene and Pleistocene submerged terraces and drowned reefs on the flanks of these two islands demonstrates the reality of this process. The thick coral accumulations that are superimposed on basaltic platforms that were once at sea level indicate that subsidence has continued on timescales of tens of millions of years. The coral cap at Eniwetok, which dates back to the early Eocene (*c.* 60 million years ago), is 1400 m thick, and that at Bikini (of Miocene age) some 1300 m thick. Differences in the depths of wave-worn platforms along a volcanic chain, if the chronology of the seamount formation can be established by potassium–argon and other dating techniques, provide estimates of subsidence rates. In the case of the Tasman Sea chains off Australia, rates of subsidence appear to have been of the order of 28 m per millions years (28 mm per 1000 years).

The causes of this subsidence are a matter of some debate (see Lambeck, 1988: 506–9). Some of it may be caused by the loading of volcanic material on to the crust, but some may be due to a gradual contraction of the seafloor as the ocean lithosphere moves away from either the ridge or the hotspot that led to the initial formation of the island volcanoes.

Another type of situation prone to subsidence is the river delta. Loading of sediment on to the crust by the river causes subsidence to occur. So, for example, Fairbridge (1983) calculated subsidence of the Mississippi at a rate of *c.* 15 mm per year during the Holocene, while Stanley and Chen (1993) calculated Holocene subsidence rates for the Yangtze delta in China as 1.6–4.4 mm per year. The Rhone delta has subsided at between 0.5 and 4.5 mm per year (L'Homer, 1992), while the Nile delta is subsiding at *c.* 4.7 mm per year (Sherif and Singh, 1999).

Some areas are prone to subsidence because of ongoing adjustment to the application and removal of ice loadings to the crust in the Pleistocene. During glacials, areas directly under the weight of ice caps were depressed, whereas areas adjacent to them bobbed up by way of compensation (the so-called peripheral bulge). Conversely, during the Holocene, following removal of the ice load, the formerly glaciated areas have rebounded whereas the marginal areas have foundered. A good example of this is the Laurentide area of North America.

Elsewhere, human actions can promote subsidence (see Chapter 6): the withdrawal of groundwater, oil, and gas (Table 6.6); the extraction of coal, salt, sulfur, and other solids, through mining; the hydrocompaction of sediments; the oxidation and shrinkage of organic deposits such as peats and humus-rich soils; the melting of permafrost; and the catastrophic development of sinkholes in karstic terrain.

Figure 9.3 (from Bird, 1993, figure 2) shows those sectors of the world's coastline that have been subsiding in recent decades, including a large tract of the

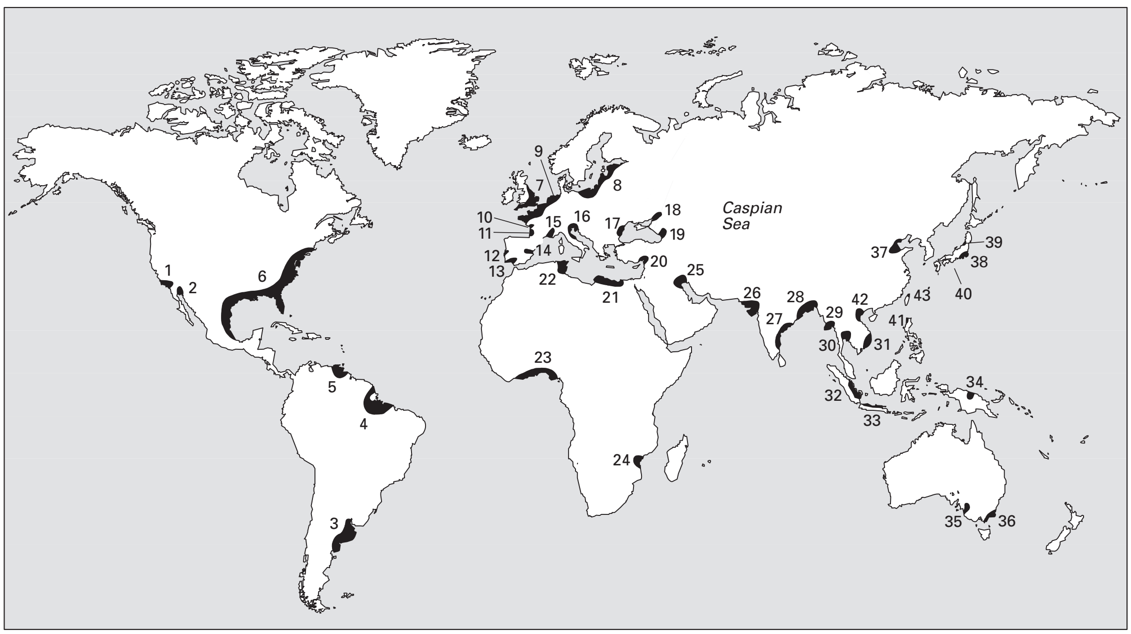

Figure 9.3 Sectors of the world's coastline that have been subsiding in recent decades, as indicated by evidence of tectonic movements, increasing marine flooding, geomorphic and ecological indications, geodetic surveys, and groups of tide gauges recording a rise of mean sea level greater than 2 mm per year over the past three decades: 1, Long Beach area, southern California; 2, Columbia River delta, head of Gulf of California; 3, Gulf of La Plata, Argentina; 4, Amazon delta; 5, Orinoco delta; 6, Gulf and Atlantic coast, Mexico and USA; 7, southern and eastern England; 8, the southern Baltic from Estonia to Poland; 9, northern Germany, The Netherlands, Belgium and northern France; 10, Loire estuary, western France; 11, Vendée, western France; 12, Lisbon region, Portugal; 13, Guadalquavir delta, Spain; 14, Ebro delta, Spain; 15, Rhone delta, France; 16, northern Adriatic from Rimini to Venice and Grado; 17, Danube delta, Romania; 18, eastern Sea of Azov; 19, Poti Swamp, Georgian Black Sea coast; 20, southeast Turkey; 21, Nile delta to Libya; 22, northeast Tunisia; 23, Nigerian coast, especially the Niger delta; 24, Zambezi delta; 25, Tigris-Euphrates delta; 26, Rann of Kutch; 27, southeastern India; 28, Ganges-Brahamputra delta; 29, Irrawaddy delta; 30, Bangkok coastal region; 31, Mekong delta; 32, eastern Sumatra; 33, northern Java deltaic coast; 34, Sepik delta; 35, Port Adelaide region; 36, Corner Inlet region; 37, Hwang-ho delta; 38, head of Tokyo Bay; 39, Niigata, Japan; 40, Maizuru, Japan; 41, Manila; 42, Red River delta, North Vietnam; 43, northern Taiwan. (Modified from Bird, 1985, figure 2.)

eastern seaboard of the USA, southeastern England, and some of the world's great river deltas (e.g., India, Ganges, Mekong, Tigris–Euphrates, and Zambezi).

How fast are sea levels rising?

Observational estimates of sea-level change are based on the short (only a decade or so) satellite record of sea-level height (Cabanes et al., 2001) and on the larger, but geographically sparse and uneven tide-gauge network. Tide gauges are rare in mid-ocean locations and very inadequate for the ocean-dominated Southern Hemisphere. In addition, difficulties in determining global (eustatic) sea-level changes are bedeviled by the need to make allowances for land motion caused by post-glacial isostatic rebound and tectonic movements (Church, 2001). In general it has been estimated that rates of sea-level rise from 1910 to 1990 average 1.0–2.0 mm per year (Church et al., 2001, table 11.10). On the other hand over the period 1993–1998, the Topex/Poseidon satellite suggested a global mean rate of sea-level rise of 3.2 ± 0.2 mm per year (Cabanes et al., 2001).

Coral reefs

Coral reefs, landforms more or less restricted to the tropics, have fascinated scientists for over 150 years. Some of the most pertinent observations of them were made in the 1830s by Charles Darwin during his voyage on *The Beagle*. He recognized that there were three major kinds: fringing reefs, barrier reefs, and atolls. These he related in a logical sequence related to progressive subsidence, and recognized the importance of sea-level change for their development.

Coral reefs are extremely important habitats. Not only do they have an estimated area of 600,000 km^2 globally, they are also the hosts to a great diversity of species, especially in the warm waters of the Indian Ocean and the western Pacific. They have been called the marine version of the tropical rainforests, rivaling their terrestrial counterparts in both richness of species and biological productivity. They are also of aesthetic importance. As Norris (2001: 37) wrote:

> Clear blue sea, brightly colored fish and amazing underwater structures built up over generations from the calcified remains of the tiny animals called coral polyps. It's a spectacular sight. No one who has ever visited one of the world's great coral reefs is ever likely to forget it. But it looks like we are one of the last generations who will have the opportunity to experience this spectacle. Reefs as we know them are on their way out.

Table 9.1 Some influences of global warming on the state of coral reefs

An increase in	*Effect*
Sea-surface temperatures	Will cause stress (bleaching, etc.) in some areas Will stimulate growth in some areas
Storm frequency and intensity	Will build up islands by throwing up coral debris Will erode reefs Will cause change in species composition
Sea levels	Will stimulate reef growth (if slow) Will cause inundation and cause corals to give up* (if fast)

*Corals will also give up if they are stressed (e.g., by pollution, increased temperatures, ultraviolet effects, etc.).

The reason for Norris's pessimism is that coral reefs are under a whole series of direct anthropogenic threats (pollution, sedimentation, dynamiting, overfishing, etc.), and that they face a suite of potential threats from climate change and sea-level rise (Table 9.1).

One potential change is that of hurricane frequency, intensity, and distribution. They might build some coral islands up, erase others, and through high levels of runoff and sediment delivery they could change the turbidity and salinity of the water in which corals grow. If hurricane frequency, intensity, and geographical spread were to change, there would be significant implications. It is, however, not entirely clear just how much these important characteristics will change. Intuitively one would expect cyclone activity to become more frequent, intense, and extensive if sea-surface temperatures (SSTs) were to rise, because SST is a clear control of where they develop. Indeed, there is a threshold at about 26.5–27.0°C below which tropical cyclones do not tend to form. Moreover, there is some evidence that increasingly low pressure centers can be maintained as SSTs rise, and Emanuel (1987) has employed a GCM which predicts that with a doubling of present atmospheric greenhouse gas levels there will be an increase of 40–50% in the destructive potential of tropical cyclones. However, the IPCC and some individual scientists are far from convinced that global warming will invariably stimulate cyclone activity. Raper (1993) has argued that there is as yet no convincing empirical evidence to support a relationship between SSTs and cyclone intensities, or that SSTs are the primary variable in whether incipient storms develop into cyclones or not.

Increased sea-surface temperature could have deleterious consequences for corals that are near their thermal maximum (Hoegh-Guldberg, 2001). Most coral species cannot tolerate temperature greater than about 30°C and even a rise in seawater temperature of 1–2°C could adversely affect many shallow-water coral species.

Increased temperatures in recent years have been identified as a cause of widespread coral bleaching (loss of symbiotic zooxanthellae). Those corals stressed by temperature or pollution might well find it more difficult to cope with rapidly rising sea levels than would healthy coral. Moreover, it is possible that increased ultraviolet radiation because of ozone layer depletion could aggravate bleaching and mortality caused by global warming. Various studies suggest that coral bleaching was a widespread feature in the warm years of the 1980s and 1990s.

Coral bleaching, which can produce mass mortality of corals in extreme cases, has been found to be strongly correlated with elevated water temperatures and high UV solar irradiance (e.g., Brown, 1997; Spencer et al., 2000). Although bleaching itself is a complex phenomenon to which corals can respond in a variety of ways (Brown et al., 2000; Fitt et al., 2001; Loya et al., 2001), ENSO-related heating, cooling, and migrations of ocean water masses have been found to be important controls of mass bleaching episodes (Spencer et al., 2000). For example, in 1998, sea-surface temperatures in the tropical Indian Ocean were as much as 3–5°C above normal, and this led to up to 90% coral mortality in shallow areas (Wilkinson et al., 1999; Edwards et al., 2001). McClanahan (2000) notes that warm conditions of between 25 and 29°C favor coral growth, survival, and species richness, and that somewhere about 30°C there are species-, environment-, or regionally specific thresholds above which many of the dominant coral species are lost. As Hoegh-Guldberg (1999), Souter and Linden (2000), Sheppard (2003), and others have suggested, continued warming trends superimposed on interannual and decadal patterns of variability are likely to increase the incidence of bleaching and coral mortality unless significant adaptation to increased temperatures occurs.

Indeed, Goreau and Hayes (1994) have produced maps of global coral bleaching episodes between 1983 and 1991 and have related them to maps of sea-surface temperatures over that period. They find that areas of severe bleaching are related to what they describe as ocean 'hotspots' where marked positive temperature anomalies exist. They argue that coral reefs are ecosystems that may be uniquely prone to the effects of global warming:

> If global warming continues, almost all ecosystems can be replaced by migration of species from lower latitudes, except for the warmest ecosystems. These have no source of immigrants already adapted to warmer conditions. Their species must evolve new environmental tolerances if their descendants are to survive, a much slower process than migrations (pp. 179–80).

However, Kinsey and Hopley (1991) believe that few of the reefs in the world are so close to the limits of temperature tolerance that they are likely to fail to adapt satisfactorily to an increase in ocean temperature of 1–2°C, provided that there are not very many more short-term temperature deviations. Indeed in general they believe that reef growth will be stimulated by the rising sea levels of a warmer world, and they predict that reef productivity could double in the next 100 years from around 900 to 1800 million tonnes per year. They do, however, point to a range of subsidiary factors that could serve to diminish the increase in productivity: increased cloud cover in a warmer world could reduce calcification because of reduced rates of photosynthesis; increased rainfall levels and hurricane activity could cause storm damage and freshwater kills; and a drop in seawater pH might adversely affect calcification.

In the 1980s there were widespread fears that if rates of sea-level rise were high (perhaps 2–3 m or more by 2100) then coral reefs would be unable to keep up and submergence of whole atolls might occur. Particular concern was expressed about the potential fate of Tokelau, the Marshall Islands, Tuvalu, the Line Islands, and Kiribati in the Pacific Ocean, and of the Maldives in the Indian Ocean. However, with the reduced expectations for the degree of sea-level rise that may occur, there has arisen a belief that coral reefs may survive and even prosper with moderate rates of sea-level rise. As is the case with coastal marshes and other wetlands, reefs are dynamic features that may be able to respond adequately to rises in sea level (Spencer, 1995). It is also important to realize that their condition depends on factors other than the rate of submergence.

An example of the pessimistic tone of opinion in the 1980s is provided by Buddemeier and Smith (1988). Employing 15 mm per year as the probable rise of sea level over the next century, they suggest that this would be (p. 51) 'five times the present modal rate of vertical accretion on coral reef flats and 50% greater than the maximum vertical accretion rates apparently attained by coral reefs'. Using a variety of techniques they believed (p. 53) 'the best overall estimate of the sustained maximum of reef growth to be 10 mm/ year . . .'. They predicted (p. 54) that

> inundated reef flats in areas of heavy seas will be subjected to progressively more destructive wave activity as larger waves move across the deepening flats . . . Reef growth on the seaward portions of inundated, wave swept reef flats may therefore be negligible compared to sea level rise over the next century, and such reef flats may become submerged by almost 1.5 m.

In addition to the potential effects of submergence, there is the possibility that higher sea levels could promote accelerated erosion of reefs (Dickinson, 1999).

In recent years, fears have been expressed that corals will suffer from the increasing levels of carbon dioxide in the atmosphere. These are expected to reach double pre-industrial levels by the year 2065. A coral reef, as Kleypas et al. (1999) have pointed out, represents the net accumulation of calcium carbonate produced by corals and other calcifying organisms. Thus if calcification was to decline, then reef-building capacity would also decline. Such a decline could occur if there were to be a change in the saturation state of aragonite (a carbonate mineral) in surface seawater. Increased concentrations of carbon dioxide decrease the aragonite saturation state.

By the middle of the twenty-first century the aragonite saturation state in the tropics could be reduced by as much as 30%. Equally, Leclercq et al. (2000) have calculated that the calcification rate of scleractinian-dominated communities could decrease by 21% between the pre-industrial period (1880) and the year (2065) at which atmospheric carbon dioxide concentrations will double. Over the same period, the pH of seawater will decline from 8.08 to 7.93. It is this that drives down the aragonite saturation state (Gattuso et al., 1998).

Salt marshes and mangrove swamps

Salt marshes, including the mangrove swamps of the tropics, are extremely valuable ecosystems that are potentially highly vulnerable in the face of sea-level rise, particularly in those circumstances where sea defenses and other barriers prevent the landward migration of marshes as sea-level rises. Sediment supply, organic and inorganic, is a crucial issue (Reed, 2002). On coasts with limited sediment supply a rise in sea level will impede the normal process of marsh progradation, and increasing wave attack will start or accelerate erosion along their seaward margins. The tidal creeks that flow across the marsh will tend to become wider, deeper, and more extended headwards as the marsh is submerged. The marsh will attempt to move landwards, and where the hinterland is low lying the salt marsh vegetation will tend to take over from freshwater or terrigenous communities. Such landward movement is impossible where seawalls or embankments have been built at the inner margins of a marsh (Figure 9.4). Equally, if there is very limited availability of sediment the marsh may not build up and inwards, so that in such circumstances the salt marsh will cease to exist (Bird, 1993). Marshes would be further threatened if climate change caused increased incidence of severe droughts (Thomson et al., 2002). Table 9.2 demonstrates the differences between marshes in terms of their sensitivity. However, salt marshes are highly dynamic features and in some situations may well be able to cope, even with quite rapid rises of sea level (Reed, 1995).

Reed (1990) suggests that salt marshes in riverine settings may receive sufficient inputs of sediment that they are able to accrete rapidly enough to keep pace with projected rises of sea level. Areas of high-tidal range, such as the marshes of the Severn Estuary in England and Wales, or the Tagus Estuary of Portugal, are also areas of high sediment-transport potential and

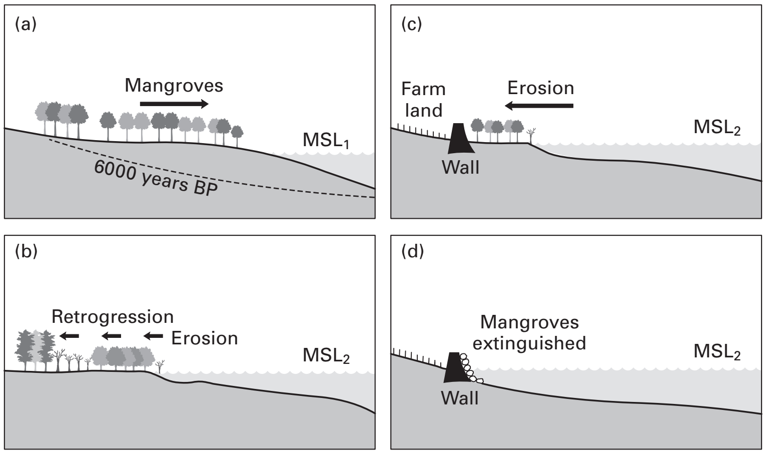

Figure 9.4 Changes on mangrove-fringed coasts as sea level rises (a and b) will be modified where they are backed by a wall built to protect farmland (c). The mangrove fringe will then be narrowed by erosion, and may eventually disappear (d), unless there is a sufficient sediment supply to maintain the substrate and enable the mangroves to persist. MSL, mean sea level. (Modified from Bird, 1993, figure 45.)

Table 9.2 Salt marsh vulnerability

Less sensitive	*More sensitive*
Areas of high sediment input	Areas of subsidence
Areas of high tidal range (high sediment transport potential)	Areas of low sediment input (e.g., cyclically abandoned delta areas)
Areas with effective organic accumulation	Mangroves (longer life cycle, therefore slower response)
	Constraint by seawalls, etc. (nowhere to go)
	Microtidal areas (rise in sea level represents a larger proportion of total tidal range)
	Reef settings (lack of allogenic sediment)

may thus be less vulnerable to sea-level rise (Simas et al., 2001). Likewise, some vegetation associations, e.g., *Spartinia swards*, may be relatively more effective than others at encouraging accretion, and organic matter accumulation may itself be significant in promoting vertical build-up of some marsh surfaces. For marshes that are dependent upon inorganic-sediment accretion, the increased storm activity and beach erosion that might be associated with the greenhouse effect could conceivably mobilize sufficient sediments in coastal areas to increase their sediment supply.

Marsh areas that may be highly prone to sea-level rise include areas of deltaic sedimentation where, because of sediment movement controls (e.g., reservoir construction) or because of cyclic changes in the location of centers of deposition, rates of sediment supply are low. Such areas may also be areas with high rates of subsidence. A classic example of this are portions of the Mississippi delta. Park et al. (1986) undertook a survey of coastal wetlands in the USA and suggested that sea-level change could, by 2100, lead to a loss of between 22 and 56% of the 1975 wetland area, according to the degree of sea-level rise that takes place.

Ray et al. (1992) have examined the past and future response of salt marshes in Virginia in the USA to changes in sea level, and have been particularly interested in the response of mid-lagoon marshes. Between 6000 and 2000 years ago, when the rate of sea-level rise was about 3 mm year, the lagoons were primarily open water environments with little evidence of mid-lagoon marshes. After AD 1200, land subsidence in the area slowed down, and sea-level rise was only about 1.3 mm per year. This permitted mid-lagoon marshes to expand and flourish. Such marshes in general lack an inorganic sediment supply so that their upward growth rate approximates only about 1.5 mm per year. Present rates of sea-level rise exceed that figure (they are about 2 mm per year) and cartographic analysis shows a 16% loss of marsh between 1852 and 1968. As sea-level rise accelerates as a result of global warming, almost all mid-lagoon marsh will be lost.

Rises in sea level will increase nearshore water depths and thereby modify wave refraction patterns. This means that wave energy amounts will also change at different points along a particular shoreline. Pethick (1993) maintains that this could be significant for the

Figure 9.5 A mangrove swamp in the Seychelles. These important wetlands are relatively slow-growing and their nature is clearly very intimately connected to the frequency and duration of inundation by tides. They may, therefore, be examples of hot spots that are especially sensitive to sea-level change.

classic Scolt Head Island salt marshes of Norfolk, eastern England, which are at present within a low to medium wave energy zone. After a 1.0 m rise in sea level these marshes will experience high wave energy because of the migration of wave foci. Pethick remarks:

> The result will be to force the long shore migration of salt marsh and mudflat systems over distances of up to ten kilometres in 50 years – a yearly migration rate of 200 metres. It is doubtful whether salt marsh vegetation could survive in such a transitory environment – although intertidal mudflat organisms may be competent to do so – and a reduction or total loss of these open coast wetlands may result (p. 166).

One particular type of marsh that may be affected by anthropogenically accelerated sea-level rise is the mangrove swamp (Figure 9.5). As with other types of marsh the exact response will depend on the local setting, sources, and rates of sediment supply, and the rate of sea-level rise itself. However, mangroves may respond rather differently from other marshes in that they are composed of relatively long-lived trees and shrubs which means that the speed of zonation change will be less (Woodroffe, 1990).

Like salt marshes, however, mangroves trap sediment to construct a depositional terrace in the upper intertidal zone. Where there is only a modest sediment supply, submergence by rising sea level may cause dieback of vegetation and erosion of their seaward margins. As their seaward margin erodes backwards, the mangroves will attempt to spread landward, displacing existing freshwater swamps or forest. As with normal salt marshes this would not be possible if they were backed by walls or embankments.

The degree of disruption is likely to be greatest in microtidal areas, where any rise in sea level represents a larger proportion of the total tidal range than in macrotidal areas. The setting of mangrove swamps will be very important in determining how they respond. River-dominated systems with large **allochthonous** sediment supply will have faster rates of shoreline progradation and deltaic plain accretion and so may be able to keep pace with relatively rapid rates of sea-level rise. By contrast in reef settings, in which sedimentation is primarily **autochthonous**, mangrove surfaces are less likely to be able to keep up with sea-level rises. This is the view of Ellison and Stoddard (1990: 161) who argued that low island mangrove ecosystems (mangals) have in the past been able to keep up with a sea-level rise of up to 8–9 cm per 100 years, but that at rates over 12 cm per 100 years they had not been able to persist.

Snedaker (1995) finds it difficult, however, to reconcile the Ellison and Stoddard view with what has happened in South Florida, where relative sea level rose by about 30 cm over 147 years (equivalent to 23 cm per 100 years). The mangrove swamps of the area did not for the most part appear to have been adversely stressed by this. Snedaker argues that precipitation and catchment runoff changes also need to be considered, as for any given sea-level elevation reduced rainfall and precipitation would result in higher salinity and greater seawater-sulfate exposure. These in turn would be associated with decreased production and increased organic matter decomposition, which would lead to subsidence. On the other hand, under conditions with higher rainfall and runoff the reverse would occur, so that mangrove production would increase and sediment elevations would be maintained.

The ability of mangrove propagules to take root and become established in intertidal areas subjected to a higher mean sea level is in part dependent on species (Ellison and Farnsworth, 1997). In general, the large propagule species (e.g., *Rhizophora* spp.) can become established in rather deeper water than can the smaller propagule species (e.g., *Avicennia* spp.). The latter has aerial roots which project only vertically above tidal muds for short distances (Snedaker, 1993).

In arid areas, such as the Arabian Gulf in the Middle East, extensive tracts of coastline are fringed by low-level salt-plains called *sabkhas*. These features are generally regarded as equilibrium forms that are produced by depositional processes (e.g., wind erosion and storm surge effects). They tend to occur at or about high tide level. Because of the range of depositional processes involved in their development they might be able to adjust to a rising sea level, but quantitative data on present and past rates of accretion are sparse. A large proportion of the industrial and urban infrastructure of the United Arab Emirates is located on or in close proximity to sabkhas.

Salt marsh, swamp, and sabkha regression caused by climatic and sea-level changes will compound the problems of wetland loss and degradation caused by other human activities (Wells, 1996; Kennish, 2001), including reclamation, ditching, diking, dredging, pollution, and sediment starvation. More than half the original salt marsh habitat in the USA has already been lost, and Shriner and Street (1998: 298) suggest that a 50 cm rise in sea level would inundate approximately 50% of North American coastal wetlands in the twenty-first century. On a global basis Nicholls et al. (1999) suggest that by the 2080s sea-level rise could cause the loss of up to 22% of the world's coastal wetlands.

River deltas

Deltaic coasts and their environs are home to large numbers of people. They are likely to be threatened by submergence as sea levels rise, especially where prospects of compensating sediment accretion are not evident. Many deltas are currently zones of subsidence because of the isostatic effects of the sedimentation that caused them to form. This will compound the effects of eustatic sea-level rise (Milliman and Haq, 1996).

It needs to be remembered, however, that deltas will not solely be affected by sea-level changes. The delta lands of Bangladesh (Warrick and Ahmad, 1996), for example, receive very heavy sediment loads from the rivers that feed them so that it is the relative rates of accretion and inundation that will be crucial (Milliman et al., 1989). Land-use changes upstream, such as deforestation, could increase rates of sediment accumulation. Deltas could also be affected by changing tropical cyclone activity.

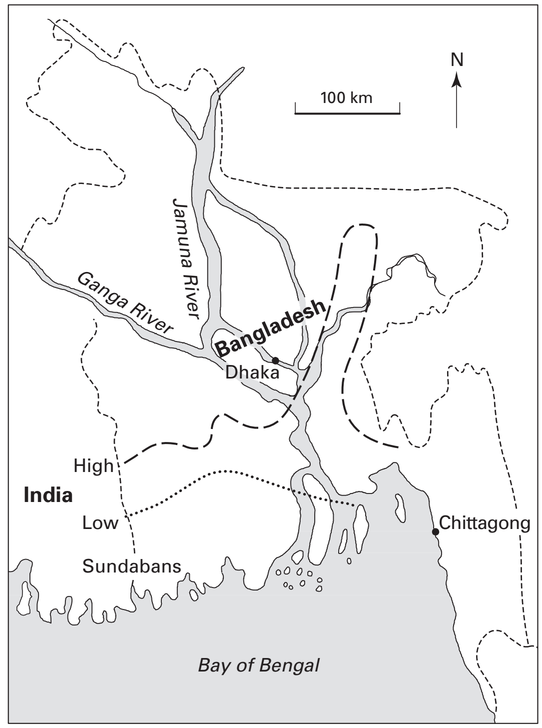

Figure 9.6 Projected areas of flooding as a result of sea-level change in Bangladesh, for two scenarios (low = 1 m and high = 3 m) (modified after Broadus et al., 1986, figure 7).

Broadus et al. (1986), for example, calculate that were the sea level to rise by just 1 m in 100 years, 12–15% of Egypt's arable land would be lost and 16% of the population would have to be relocated. With a 3 m rise the figures would be a 20% loss of arable land and a need to relocate 21% of the population. The cities of Alexandria, Rosetta, and Port Said are at particular risk and even a sea-level rise of 50 cm could mean that 2 million people would have to abandon their homes (El-Raey, 1997). In Bangladesh (Figure 9.6) a 1 m rise would inundate 11.5% of the total land area of the state and affect 9% of the population directly, while a 3 m rise in sea level would inundate 29% of the land area and affect 21% of the population. It is sobering to remember that at the present time approximately one-half of Bangladesh's rice production is in the area that is less than 1 m above sea level. Many of the world's major conurbations might be flooded in whole or in

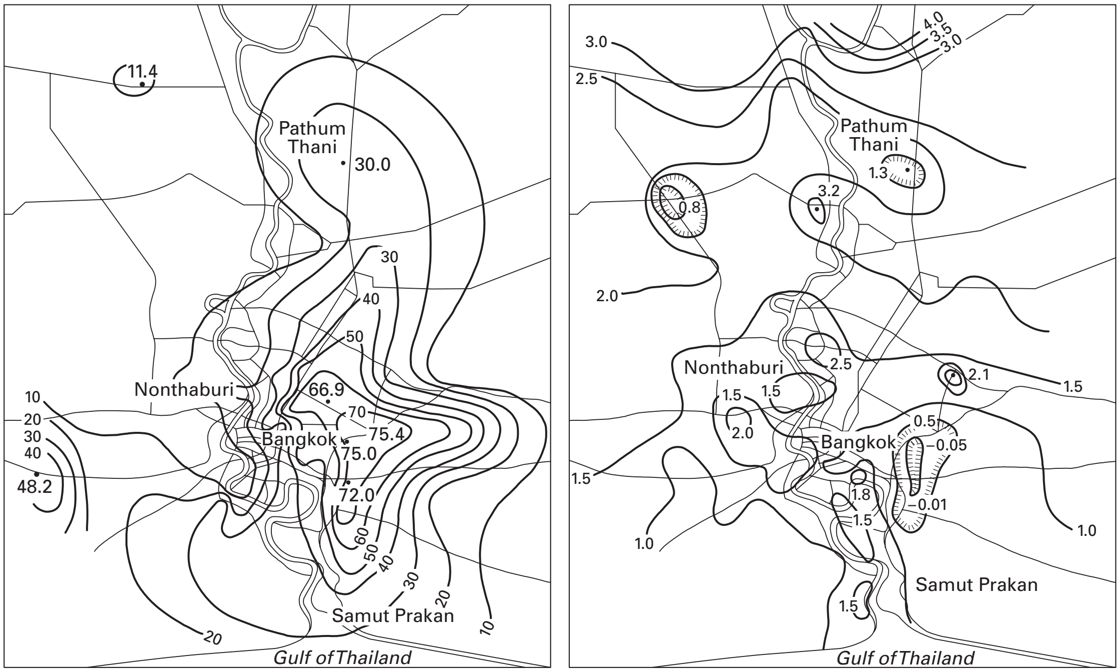

Figure 9.7 Total subsidence (in cm), 1978–87 (left) and ground elevation of Bangkok, 1987 (right). (Modified from Nutalaya et al., 1996, figures 3 and 9.)

part, sewers and drains rendered inoperative (Kuo, 1986), and peri-urban agricultural productivity reduced by saltwater incursion (Chen and Yong, 1999).

Even without accelerating sea-level rise, the Nile delta has been suffering accelerated recession because of sediment retention by dams. The Nile sediments, on reaching the sea, used to move eastward with the general anticlockwise direction of water movements in that part of the eastern Mediterranean, generating sandbars and dunes, which contributed to delta accretion. About a century ago an inverse process was initiated and the delta began to retreat. For example, the Rosetta mouth of the Nile lost about 1.6 km of its length from 1898 to 1954. The imbalance between sedimentation and erosion appears to have started with the delta barrages (1861) and then been continued by later works, including the Sennar Dam (1925), Gebel Aulia Dam (1937), Khasm el Girba Dam (1966), Roseires Dam (1966), and the Aswan High Dam itself. In addition, large amounts of sediment are retained in an extremely dense network of irrigation and drainage channels that has been developed in the Nile delta itself (Stanley, 1996). Much of the Egyptian coast is now 'undernourished' with sediment and, as a result of this overall erosion of the shoreline, the sandbars bordering Lake Manzala and Lake Burullus on the seaward side are eroded and likely to collapse. If this were to happen, the lakes would be converted into marine bays, so that saline water would come into direct contact with low lying cultivated land and freshwater aquifers.

Bangkok is an example of a city that is being threatened by a combination of accelerated subsidence and accelerated sea-level rise (Nutalaya et al., 1996). More than 4550 km^2 of the city was affected by land subsidence between 1960 and 1988, and 20 to 160 cm of depression of the land surface occurred. The situation is critical because Bangkok is situated on a very flat, low lying area, where the ground level elevations range from only 0 to 1.5 m above mean sea level (Figure 9.7).

Estuaries

Estuaries, located between the land and the sea, are unique and sensitive ecosystems that are locations of many major ports and industrial concerns (Dyer, 1995). They could be impacted upon by a range of climate-related variables, including the amount of freshwater and sediment coming from the land, the temperature and salinity of the estuarine water, and tidal range.

The overall effect of sea-level rise has been summarized by Chappell et al. (1996: 224):

In the earliest stages of sea-level rise, the extent of tidal flooding on high tide flats will increase and networks of small tidal creeks will expand into estuarine floodplains. Levels of tidal rivers will be breached and brackish water will invade freshwater floodplains. Eventually, tidal flow in main channels will increase to carry the enlarged tidal prism, and channels are likely to widen. In systems with extensive tidal floodplains, tidal regimes will shift towards ebb dominance and net sediment flux will tend to be offshore, except in macrotidal systems. Salt-water intrusion could be a major change, with saline water extending far inland as sea level rose. This in turn could affect swamp and marsh vegetation (Mulrennan and Woodroffe, 1998).

Attempts have been made to model the transgression rate for landward extending estuaries. For the Humber estuary in northeast England, Pethick (2001: 34) reports that the mean migration rate of the estuary as a whole would be 1.3 m per millimeter rise in sea level (or 8 m per year assuming a 6 mm per year rise in sea level).

Cliffed coasts

When sea level rises, nearshore waters deepen, shore platforms become submerged, deeper water allows larger waves to reach the bases of cliffs, and the cliffs suffer accelerated rates of retreat, especially if they are made of susceptible materials. In addition, sea-level rise is likely to cause increased frequencies of coastal landslides. Accelerated cliff retreat and land sliding may cause an augmented supply of sediment to adjacent and down-drift beaches. This could be beneficial. However, as Bray et al. (1992: 86) remarked:

The location of the benefit will depend upon sediment transport conditions and their relation to sediment supply. Where littoral transport is poorly developed, beaches will accrete in front of cliffs as sea level rises, and tend to offset any trend for increased retreat. Where littoral transport is efficient, cliff retreat is likely to increase significantly as sea level rises because extra sediments yielded are rapidly removed from the eroding cliffs. Coasts of this type are likely to become increasingly valuable in the future because they are sensitive to sea-level rise and can supply large quantities of sediment to downdrift beaches. A further complicating factor is that cliff erosion products differ with respect to their mobility and coarse durable materials may be retained on the upper shoreface and provide natural armoring. Soft-rock cliffs yielding a high proportion of coarse durable products are therefore likely to be less sensitive to sea-level rise.

Cliffs on high latitude coasts (e.g., in Siberia, Alaska, and Canada) might be especially seriously affected by global change. On the one hand, coasts formed of weak sediments that are currently cemented by permafrost would lose strength if warming caused the permafrost to melt. On the other hand, melting of sea ice would expose them to greater wave affects from open water.

Sandy beaches

Bruun (1962) developed a widely cited model of the response of a sandy beach to sea-level rise in a situation where the beach was initially in equilibrium, neither gaining nor losing sediment (Figure 9.8). As Bird (1993: 56) has explained:

Erosion of the upper beach would then occur, with removal of sand to the nearshore zone in such a way as to restore the previous transverse profile. In effect, there would be an upward and landward migration of the transverse profile, so that the coastline would recede beyond the limits of submergence. This restoration would be completed when the sea became stable at a higher level, and coastline recession

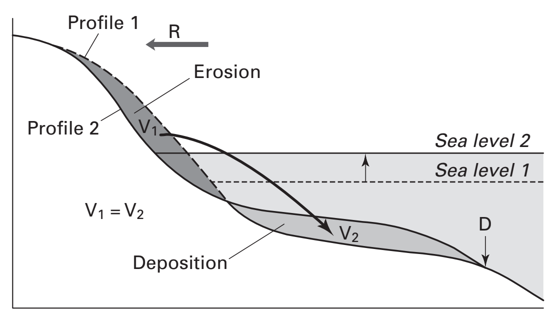

Figure 9.8 Sea-level rise and coastline changes. The Bruun Rule states that a sea-level rise will lead to erosion of the beach and removal of a volume of sand (V_1) seaward to be deposited (V_2) in such a way as to restore the initial transverse profile landward of *D*, the outer boundary of nearshore sand deposits. The coastline will retreat (*R*) until stability is restored after the sea-level rise comes to an end. The coastline thus recedes further than it would if submergence were not accompanied by erosion.

would come to an end after a new equilibrium was achieved. The extent of recession was predicted by using a formula that translates into a 'rule of thumb' whereby the coastline retreats 50–100 times the dimensions of the rise in sea level: a 1 m rise would cause the beach to retreat by 50–100 m. Since many seaside resort beaches are no more than 30 m wide, the implication is that these beaches will have disappeared by the time the sea has risen 15–30 cm (i.e., by the year 2030), unless they are artificially replaced.

This influential model, often called the Bruun Rule, has been widely used. It is important to recognize, however, that there are some constraints on its applicability (Wells, 1995; Healy, 1996). The rule assumes that no sand is lost to long shore transport and that an offshore 'closure depth' exists beyond which there is no sediment exchange. Moreover, the rule does not allow for shoreward transport of sediment as overwash or for those situations where the slope of the coastal plain is too gentle for sufficient sand to be available as a source for supplying the offshore. In addition, the rule was originally proposed for beaches that were initially in equilibrium. However, as Bird (1993: 58) has pointed out, only a small proportion of the world's sandy beaches can in fact be considered to be in equilibrium. Beach erosion is widespread. Furthermore, there is a lack of established criteria to ascertain whether the original shoreline really is in dynamic equilibrium (Healy, 1991).

In spite of these limitations the Bruun Rule remains widely used, and can be stated mathematically as follows (Gornitz et al., 2002: 68):

$$\mathrm{S} = (A \times B)/d$$

where S is shoreline movement, A is sea-level rise, d is maximum depth of beach profile, measured from the berm elevation for each project location to the estimated depth of closure, and B is the horizontal length of the profile, measured from the beginning of the berm to the intersection with the estimated depth of closure. The depth of closure is taken as the minimum water depth at which no significantly measurable change occurs in bottom depth. It is often erroneously interpreted to mean the depth at which no sediment moves in deeper water. 'Closure' is a somewhat ambiguous term in that it can vary, depending on waves and other hydrodynamic forces.

An example of the testing of the validity of the Bruun Rule is the study by List et al. (1997) of the barrier islands of Louisiana in the USA. Using bathymetric surveys over about a century, they found that only a portion of their studied profiles met the equilibrium criterion of the Bruun Rule. Furthermore, using those shore profiles that did meet the equilibrium criterion, they determined measured rates of relative sea-level rise so that they could hindcast shoreline retreat rates using the Bruun Rule formula. They found that the modeled and observed shoreline retreat rates showed no significant correlation. They suggested that if the Bruun Rule is inadequate for hindcasting it would also be inadequate for forecasting future rates of beach retreat.

There are various other techniques that can be used to predict rates of shoreline retreat. One of these is historical trend analysis, which is based upon extrapolating the trend of shoreline change with respect to recorded sea-level rise over a given historical period:

$$R_2 = (R_1/S_1) \times S_2$$

where S_1 is historical sea-level rise, S_2 is future sea-level rise, R_1 is historical retreat rate, and R_2 is future retreat rate.

Barrier islands, such as those that line the eastern seaboard of the USA and the southern North Sea, are dynamic landforms that will tend to migrate inland with rising sea levels and increased intensity of overtopping by waves (Eitner, 1996). If sea-level rise is not too rapid, and if they are not constrained by human activities (e.g., engineering structures and erosion control measures), they are moved inland by washover – a process similar to rolling up a rug (Titus, 1990); as the island rolls landward, it builds landwards and remains above sea level (Figure 9.9). As sea level rises, they will be exposed to higher storm surges and greater flooding. In the New York area, Gornitz et al. (2002) have calculated that by the 2080s the return period of the 100-year storm flood could be reduced to between 4 and 60 years (depending on location).

The role of sediment starvation

Many beaches are currently eroding because they are no longer being replenished with sediment by rivers. This is because of the construction of dams, which trap much of the sediment that would otherwise go to the sea.

The River Nile now transports only about 8% of its natural load below the Aswan Dam. Even more

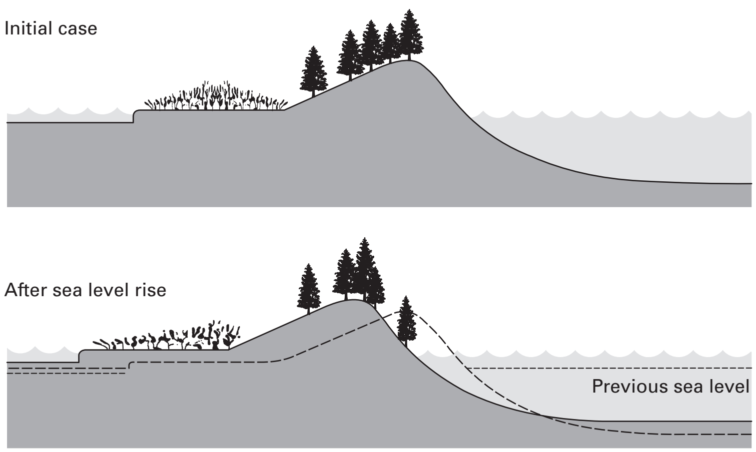

Figure 9.9 Overwash: natural response of undeveloped barrier islands to sea-level rise. (Source: Titus, 1990, figure 2.)

dramatic is the picture for the Colorado River in the USA. Prior to 1930 it carried around 125–150 million tonnes of suspended sediment to its delta at the head of the Gulf of California. Following the construction of a series of dams, the Colorado now discharges neither sediment nor water into the ocean (Schwarz et al., 1991). Similarly, rivers on the eastern seaboard of the USA draining into the Gulf of Mexico or the Atlantic have shown marked falls in sediment loadings, and four major Texan rivers carried in 1961–70 on average only about one-fifth of what they carried in 1931–40. Likewise, in France the Rhône only carries about 5% of the load that it did in the nineteenth century, while in Asia the Indus discharges less than 20% of the load if did before the construction of large barrages over the past half century (Milliman, 1990).

Conclusions

Coastal environments are already suffering from many human impacts, including sediment starvation. However, these dynamic regions, at the interface between land and sea and the home to many people, will also be subjected in the future to the combined effects of climate change and sea-level rise. Plainly coastlines will be directly impacted by climatic change through such mechanisms as changes in storm surges and hurricanes, but they will also be impacted by sea-level changes resulting from global climatic change, local subsidence, and miscellaneous modifications of the hydrologic cycle.

If, on average, sea level rises by around half a meter during this century, it will greatly modify susceptible and sensitive coastlines, including marshes and swamps, estuaries, soft cliffs, barrier islands, and sandy beaches.

Points for review

In what ways may human activities influence global sea levels?
How may coral reefs respond to global warming?
Which coastal environments may be particularly susceptible to sea-level rise?
What do you understand by the Bruun Rule? Is it a rule that has wide applicability?

Guide to reading

Bird, E. C. F., 1993, *Submerging coasts.* Chichester: Wiley. A geomorphologic overview of how coasts will respond to sea-level rise.
Douglas, B. C., Kearney, M. S. and Leatherman, S. P. (eds), 2000, *Sea level rise: history and consequences.* San Diego: Academic Press. An edited volume, with a long time perspective, which considers some of the consequences of sea-level change.
Milliman, J. D., and Haq, B. V. (eds), 1996, *Sea-level rise and coastal subsidence.* Dordrecht: Kluwer. An edited discussion of the ways in which sea-level rise and land subsidence may impact upon coastal environments.