Chapter 2

Dynamics of the Coastal Zone

James P. M. Syvitski, Nick Harvey, Eric Wolanski, William C. Burnett, Gerardo M. E. Perillo, Vivien Gornitz
Contributors: *Russell K. Arthurton, Henry Bokuniewicz, Janet W. Campbell, Lee Cooper, Kenneth Dunton, Shu Gao, Patrick P. Hesp, Yoshiki Saito, Joe Salisbury, Maria Snoussi, Wyss W.-S. Yim*

2.1 Introduction

Earth's coastline has evolved for many thousands of years, experiencing changes to habitat, coastal dynamics and the supply of sediment from the continental interior. Relative sea level has risen in some areas, but fallen elsewhere. There is an acknowledged range in natural variability within a given region of the global coastal zone, within a context of longer-term geological processes.

Many of the regional controls on sea level involve long-term geological processes (e.g., subsidence, isostasy), and have a profound influence on controlling short-term dynamics. As sea levels fluctuate, the morphology of a coastal zone will further evolve, changing the boundary conditions of other coastal processes: circulation, waves, tides and the storage of sediment on flood plains.

Human development of coastal regions has modified pristine coastlines around the globe, by deforestation, cultivation, changes in habitat, urbanisation, agricultural impoundment and upstream changes to river flow. Humans can also influence changes in relative sea level at the local scale. For example, removal of groundwater and hydrocarbons from subterranean reservoirs may cause subsidence in nearby areas, with a concomitant rise in relative sea level. Our concern in LOICZ is not just in the magnitude of change, but also in the recent and accelerated rate of change. Our interests extend to whether alterations on the local level can cumulatively give rise to coastal zone changes of global significance.

Climate warming may also contribute significantly to sea level fluctuations. Predictions by the International Panel on Climate Change (IPCC) suggest that sea level is rising globally (15 to 95 cm by 2100) as a result of the recent warming of the ocean and the melting of ice caps (Houghton et al. 2001). As sea levels rise, coastal destabilisation may occur due to accelerated beach erosion, trapping of river sediment on flood plains and increasing water residence during floods. The predicted IPCC climate-warming scenario will undoubtedly impact some regions more than others. The Siberian coast is experiencing a reduction in offshore sea-ice cover, with a associated increase in ocean fetch, leading to higher sea levels during the open-water summer and acceleration of coastal erosion. Recent studies also suggest that tropical and temperate coastal environments are experiencing stormier conditions (i.e., increased numbers and severity of hurricanes). Will local storm surges magnify the impact of a global sea-level rise, increasing risks to humans and their infrastructure? Are there negative feedbacks to engineering options for the protection of coastal settlements?

Perhaps the largest impact on coastal stability is due to modification to the global flux of sediment to the coastal zone. Changes in global hydrology have modified the timing and intensity of floods, and therefore the effective discharge available for sediment transport. Climate shifts have varied the contributions from meltwater (snow, ice), altered the intensity of rainfall, changed drainage basin water-storage capacity, and altered precipitation and evaporation rates. Human influences have also greatly modified downstream flow. Over half of the world's rivers have seen stream-flow modification through the construction of large reservoirs. These and other rivers have also been impacted by water withdrawal for agriculture, industry and settlements.

Our understanding of the importance of submarine groundwater discharge in the coastal zone and of its processes has improved markedly in recent years; a significant impetus has been given to this understanding by the LOICZ-associated SCOR Working Group 112. The outcomes of its work are summarised in this chapter.

Human migration to the coastal zone and consequent land-use changes have also greatly impacted the stability of our coastal areas. Human impacts on the coastal zone ranges from massive (e.g., reduction in wetlands, urbanisation) to non-existent (e.g., many polar coastlines). This chapter synthesises how climate shifts and humans can affect and have affected our coasts on a global scale.

2.2 Impacts of Local, Regional and Global Sea-level Fluctuations

2.2.1 Processes and Mechanisms: Coastal Dynamics

2.2.1.1 Impacts of Local, Regional and Global Sea-level Fluctuations

In 1993 the LOICZ Science Plan (Holligan and de Boois 1993) commented on the problems of accurate measurement of sea-level change using tide-gauge data, and the need for more sensitive measurement techniques. The Plan also stated that "*... spatial variations in local sea level must be known in order to predict large-scale changes to coastal systems.*" (Holligan and de Boois 1993, p. 24), noting that local sea-level changes are affected by:

- tectonics,
- climatic change impacts on winds, waves and ocean circulation, and
- coastal subsidence, often exacerbated by human impacts, which can produce land/sea-level change rates higher than those of global sea-level rise.

Despite the use of more accurate global sea-level measurement techniques over the last ten years, such as satellite altimetry and geodetic levelling (see Nerem and Mitchum 2001) and the refined global viscoelastic analysis of glacio-isostatic adjustment (GIA; see Peltier 2001, 2002), there remains a need for better understanding of existing local sea-level change data and acquisition of new data.

The following is an overview of the current state of knowledge on the coastal impacts of sea-level fluctuations.

2.2.1.2 Reconstruction of Past Changes in the Coastal Zone

Current and predicted sea-level changes need to be considered in their geological context. The LOICZ Science Plan notes the necessity of developing historical reconstructions of sea-level change "*... in order to examine the responses of particular coastal systems to relatively large changes in external forcing that might occur in the future ...*" and that "*... many of these analyses directly complement studies of the IGBP project on Past Global Changes (PAGES).*" (Holligan and de Boois 1993).

Sea-level studies from a LOICZ perspective need to be placed in the context of PAGES sea-level studies for three reasons:

1. Past sea-level changes provide a perspective on the cyclical nature of sea levels and the extent to which current and predicted sea-level changes are perturbations from natural cycles.
2. Sea-level response following the last glacial maximum of 20 000 years ago (20 ka) varied around the globe according to the coastal adjustment to the post-glacial redistribution of water and ice, and differential loading of the lithosphere.
3. Very recent geological evidence of sea levels during the last 2 000 years provides insights into subtle regional climate change linked to oceanic fluctuations.

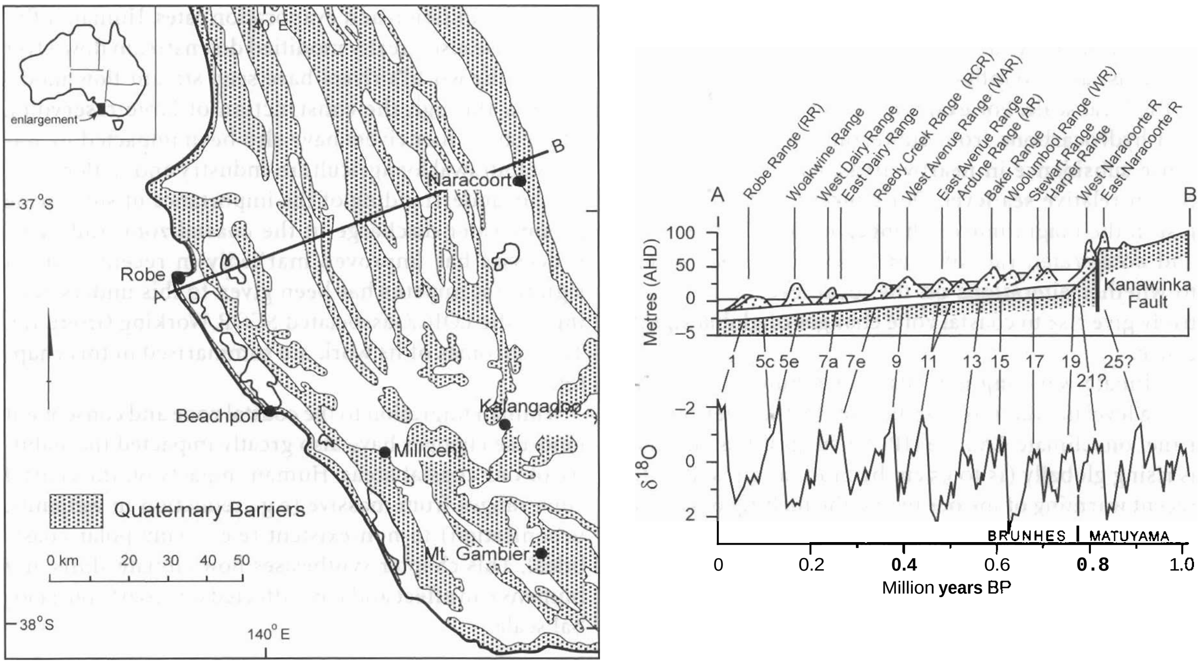


Fig. 2.1. Sea-level rise. Coastal barriers correlated with oxygen isotope changes – 800 000 years BP (from Belperio 1995)

2.2.1.2.1 *Cycles of Sea-level Change – Glacial-interglacial Cycles over 70 000 years BP*

Evidence of the cyclical nature of sea-level changes was discovered following the identification of global temperature fluctuations deduced from oxygen isotope ($^{18}O/^{16}O$) ratios of planktonic foraminifera from deep-sea sediments (Emiliani 1955). These ratios allowed interpretation of global temperature changes over time and demonstrated that a quasi-periodic cycling of climate recurred more frequently than the four major glaciations recognised from previous northern hemisphere stratigraphic studies. One good example of coastal change correlated with the oxygen isotope record across a number of glacial-interglacial cycles is from south-eastern Australia, where a series of stranded dune barriers associated with high sea-level stands has been preserved across a slowly uplifting coastal plain. The barriers, dating back to 800 000 years BP (Huntley et al. 1993, 1994), record coastal sedimentation linked to at least 10 interglacial high stands of sea level (Belperio 1995; Fig. 2.1).

Detailed climatic reconstructions for the last four major glacial cycles come from the Vostok ice-core (Petit et al. 1999), which has provided new data on the cyclical nature of global climate for glacial-interglacial cycles. These permit extrapolation of the timing and magnitude of sea-level fluctuations over the last 400 000 years, from the cyclical nature in the concentrations of CO, and CH_4 "greenhouse" gases, which oscillated within a well-defined range between glacial and interglacial periods. The modern elevated concentrations of CO, and CH_4 in the current inter-glacial high sea-level period (Raynaud et al. 2000; Fig. 2.2) are greatly elevated above the upper boundary of global gas composition for the interglacial periods as derived from the Vostok ice-core.

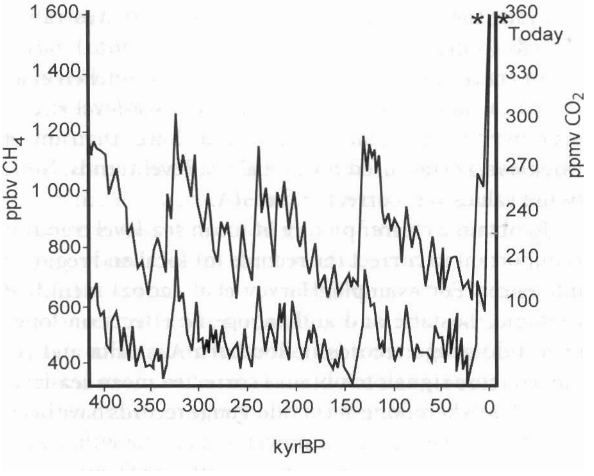


Fig. 2.2. Sea-level rise. Global greenhouse-gas concentrations from the Vostok ice-core – 400 000 years BP (from Alverson et al. 2001 adapted from Raynaud et al. 2000, after Petit et al. 1999)

This perturbation from normal cycles confirms the need for predictions of global warming and associated sea-level rise estimates as outlined by the IPCC (Houghton et al. 1996, 2001). Thus the geological evidence demonstrates a well-defined periodicity in climate-related sea-level changes. The enhanced greenhouse gas concentrations in the current interglacial indicate the potential for a sea-level response different from the previous interglacial periods. These findings have contributed to the IGBP findings that Earth systems are currently operating in a "non-analogue" state (Steffen et al. 2003).

Further geological evidence of sea-level changes across an entire glacial-interglacial cycle comes from the coral record from Barbados (Bender et al. 1979, Broecker 1979, Gallup et al. 1994, Bard et al. 1990, Blanchon and Eisenhauer 2001) and from the detailed record of coral terraces preserved on the rapidly uplifting coast of the Huon Peninsula in New Guinea. Reworking of the Huon Peninsula data has provided a close correlation with the oxygen isotope records (Chappell et al. 1996; Fig. 2.3).

2.2.1.2.2 *Postglacial Sea-level Response – the Last 20 000 Years*

Geophysical modelling of the surface of Earth has demonstrated the variable regional response of the world's coastlines to the postglacial global redistribution of ice and water (see Houghton et al. 1996). Postglacial sea-level changes are still impacting on the coast in many parts of the world, as has been shown in models of local or regional glacio-isostatic movements in the northern hemisphere and the global pattern of hydro-isostatic coastal adjustment (Houghton et al. 1996, 2001; Peltier 2001, 2002). The impact of water redistribution throughout the world's oceans has consequences for the magnitude of landlsea movements that are predicted to occur in response to the altered isostatic loads, particularly around continental margins. The results from these geophysical models are refined by geological studies.

2.2.1.2.3 *Recent Sea-level Changes – the Last 2 000 Years*

Recent studies demonstrate the potential for fine-scale resolution in reconstructing climate change during the late Holocene period. Developments with Thermal Ionisation Mass Spectrometry (TIMS) in conjunction with the measurement of Thorium/Uranium (Th/U) isotopes could provide dating resolution of 10–15 years at 2000 years BP (Goodwin 2002). Fixed biological indicators such as coral microatolls (Smithers and Woodroffe 2000) and encrusting tubeworms (Baker and Haworth 2000) can yield high-resolution paleo sea-level datums. In particular, the microtopography of the upper surface of microatolls together with X-radiography of the annual growth banding can reflect sea-level fluctuations and even

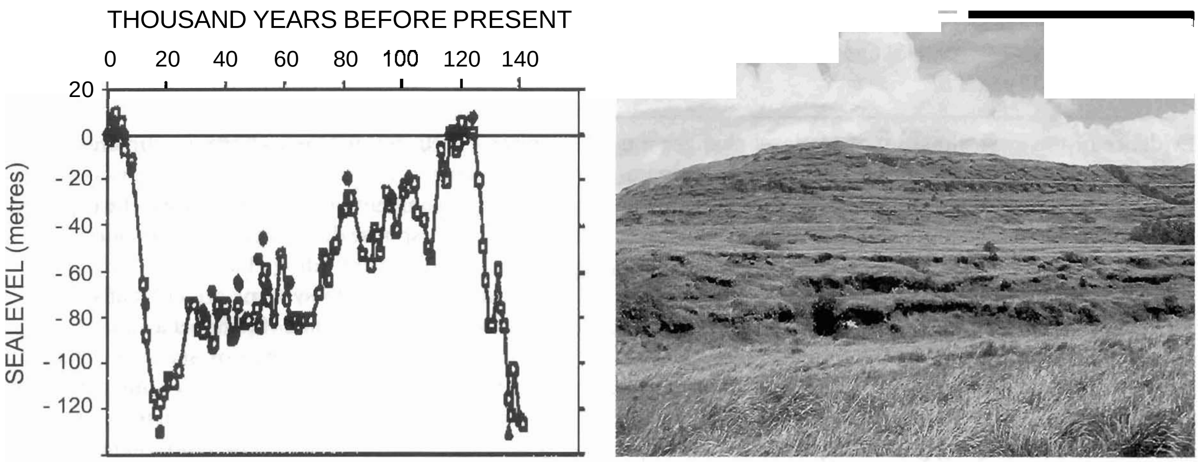


Fig. 2.3. Sea-level rise. Record of sea-level change from the Huon Peninsula, New Guinea – 125 000 years BP (from Chappell et al. 1996; photograph B. Pillans)

record major regional climate-induced sea-level changes, such as El Niño events in both the Pacific and Indian oceans (Woodroffe and McLean 1990, Spencer et al. 1997, Smithers and Woodroffe 2000, Woodroffe and Gagan 2000).

The paleo-climatic record is a vital tool in resolution of the debate over the global proliferation of the Mediaeval warm period (1050–690 years BP) and the Little Ice Age (575–150 years BP), especially in relation to regional variability in sea surface temperatures (SST) and ocean circulation during these periods (Goodwin 2002). Multi-decadal sea-level changes of up to 1 m, but typically ±0.5 m or less, can be detected in the recent coastal geological record (Goodwin et al. 2000, Goodwin 2002). Given that the sea-level fluctuations over the last few millennia are of a similar magnitude to those predicted for greenhouse-induced sea-level rises over the next 100 years, it is important to use the geological record to understand coastal response to sea-level changes of this magnitude. There are long-term proxy SST records for the tropical Atlantic, Pacific and Indian oceans (Dunbar and Cole 1999); Goodwin (2002) suggests that the way to link high-resolution sea-level changes with SST, and ultimately with climate models, lies in a coupled approach that derives both proxy SST and sea-level chronology in regional coral paleo-climatic studies.

2.2.1.3 *Historical Records of Sea-level Change – the Last 720 Years*

The IPCC Third Assessment Report (2001 IPPC: Houghton et al. 2001) concluded from analysis of global average sea-level rise derived from tide-gauge records that:

- the average rate of sea-level rise was less in the 19^{th} century than in the 20^{th} century, based on tide-gauge records spanning many decades;
- mean sea-level rise was in the range 1–2 mm yr^{-1} with a central value of 1.5 mm yr^{-1} during the 20^{th} century; and
- there was no widespread increase in extremes apart from that associated with a change in the mean, despite a decadal variability in extremes.

Tide-gauge records measure only relative sea level, so while it is important to have long-term reliable records, it is equally important to have sites free of vertical crustal movements due to plate tectonics (to be able to correct records for glacial rebound) and to be either insensitive to small oceanographic changes or capable of accounting for these (Douglas 2001). Douglas (2001) noted the very large location bias of the northern hemisphere for long-term (> 20 years) tide-gauge sites and selected only 27 sites (with records > 70 years) from 10 regions of the globe as suitable for establishing a 20^{th} century global rate of sea-level rise. However, sites from the relatively stable Australian continent were excluded from that analysis, thereby omitting the two longest Australian records (both > 80 years) (Houghton et al. 2001). Based on an array of long-term reliable records, Mitchell et al. (2001) calculated an average Australian sea-level rise of 0.3 mm yr^{-1}. This value is significantly lower than any of Douglas's 27 sites used for global sea-level trends. None of the values was corrected for GIA.

- To obtain a clearer picture of mean sea-level trends it is important to correct the records for local and regional influences. For example, Harvey et al. (2002) identified geologic, isostatic and anthropogenic effects on long-term tide-gauge records in southern Australia and removed these signals to obtain a corrected mean sea-level trend. Elsewhere, long-term tide-gauge records have been adjusted for vertical land movements using either geological methods (Gornitz and Lebedeff 1987, Shennan and Woodworth 1992, Gornitz 1995, Peltier and Jiang 1997,

Table 2.1.
Sea-level rise. Estimates of trends from tide-gauge data (after Houghton et al. 2001)

Source	Rate (mm yr^{-1})	Vertical land movement (method)	Region
Peltier and Tushingham (1989,1991)	2.4	ICE-3G/M1	Global
Peltier and Jiang (1997)	1.8	Geological	Global
Douglas (1997)	1.8	ICE-3G/M1	Global
Douglas (1991)	1.8	ICE-3G/M1	Global
Trupin and Wahr (1990)	1.75	ICE-3G/M1	Global
Mitrovica and Davis (1995)	1.4	PGR Model	Global
Gornitz and Lebedeff (1987)	1.2	Geological	Global
Peltier and Jiang (1997)	2.0	ICE-4G/M2	US East Coast
Peltier (1996)	1.9	ICE-4G/M2	US East Coast
Gornitz (1995)	1.5	Geological	US East Coast
Davis and Mitrovica (1996)	1.5	PGR Model	US East Coast
Larnbeck et al. (1998)	1.1	PGR Model	Fenno-scandia
Shennan and Woodworth (1992)	1.0	Geological	NW Europe
Woodworth et al. (1999)	1.0	Geological	British Isles
Houghton et al. (2001)	1.07	PGR Model	East coast Australia
Houghton et al. (2001)	1.55	PGR Model	West coast Australia
Harvey et al. (2002)	0.46	Geological	South coast Australia

Woodworth et al. 1999) or post-glacial rebound models (Peltier and Tushingham 1989, 1991, Trupin and Wahr 1990, Douglas 1991, Mitrovica and Davis 1995, Peltier 1996, Peltier and Jiang 1997, Lambeck et al. 1998). The resulting estimates, reviewed by the 2001 IPCC, excluded a wide range of rates reflecting, in part, the different assumptions and methods used for estimating vertical land movement, as well as the different criteria used in selection of the tidal data (Houghton et al. 2001).

Examples of sea-level rise estimates from various regions (Table 2.1) show that:

- The North American east coast has had significantly higher rates according to Peltier (1996) than those calculated by both Gornitz (1995) and Mitrovica and Davis (1995). Houghton et al. (2001) obtained slightly higher rates using the ICE-4G/M2 model to adjust the US east coast estimates.
- European rates have been relatively lower than those of North America. This may reveal a real regional difference in sea level because of higher rates of sea-level rise for the sub-tropical gyres of the North Atlantic in recent decades (Houghton et al. 2001).
- Australian data from two long-term sites, Sydney (82-year record) and Fremantle (91-year record), included GIA corrections from the Australian-based rebound calculations of Lambeck and Nakada (1990), yielding rates of 1.07 mm yr^{-1} and 1.55 mm yr^{-1}.

Sea-level rise estimates from southern Australia, which are significantly lower than other estimates (Harvey et al. 2002), were derived from a number of tide-gauge sites with records up to 60 years and individual rates between 0.14 and 0.87 mm yr^{-1}, following correction by geological methods. The far-field data from the relatively stable Australian continent result in lower rates than the central value of 1.5 mm yr^{-1} for global average sea-level rise trend accepted by Houghton et al. (2001) in the 2001 IPCC. Douglas and Peltier (2002) also argue against the 2001 IPCC central value, suggesting that a global average should be closer to 2 mm yr^{-1}.

As noted in the LOICZ Science Plan (Holligan and de Boois 1993), there are problems of accurate measurement of sea-level change using tide-gauge data, and there is a need for more sensitive measurement techniques. Detailed analysis and careful selection of tide-gauge sites coupled with the correction of long-term tide-gauge data (particularly for GIA) has considerably improved the accuracy of sea-level trends derived from tidal data over the last ten years. The introduction of satellite altimetry following the launch of the TOPEX/POSEIDON project in 1992 provided a new tool for measuring sea surface height from space (Nerem and Mitchum 2001). Recent installation of tide gauges, such as the SEAFRAME array in the south-western Pacific (Mitchell et al. 2001), has included GPS-positioned gauges, which will soon be capable of accurate geodetic survey. With two to three decades of precision monitoring data from this type of recording instrument collected from a global array of similar sites, it should be possible to measure the acceleration in sea-level rise and to help corroborate climate models (Nerem and Mitchum 2001).

2.2.2 Evolving Morphology and Boundary Conditions

2.2.2.1 *Sea-level Changes and Impacts on Coastal Dynamics*

Geological evidence of past sea-level changes has demonstrated that localised land/sea movements are affecting the current sea-level record at different rates, over different time-scales and at different spatial scales. A major UK-based land-ocean interaction study (LOIS, Huntley et al. 2001) examined coastal response to sea-level change over the postglacial sea-level transgression. This study produced detailed simulation modelling of coastal response for a section of the east coast of the UK, based on a vast array of paleo-climate and sedimentology data over the last 7000 years, to predict coastal erosion and accretion with the rising sea and allied changes to the tidal pattern.

In some parts of the world, geological studies have been linked to present-day measurements of sea level using tide-gauge data. Harvey et al. (2002) demonstrated geologic, isostatic and anthropogenic influences on southern Australian tide-gauge records occurring at time-scales of 10^6, 10^4 and 10^2 years, and rates of 0.07 mm yr^{-1}, 0.4 mm yr^{-1} and 2.0 mm yr^{-1}. While the impact of longer-term geological uplift rates in the south-eastern region of South Australia were clearly visible in the stranded shorelines at approximate 100000 year intervals (Harvey et al. 2001), the impacts of these low rates of relative sea-level fall at the modern coast are harder to identify. In contrast, the human impact on subsidence rates and consequent rapid relative sea-level rise near Port Adelaide were demonstrated by landward migration of mangroves, loss of sand from metropolitan beaches and increasing problems of flooding at high tides.

Given the differential global response to the postglacial sea-level rise, some coasts have established a morphological equilibrium to a sea level that has been relatively stable or slightly higher over the last 6000–7000 years while other coasts are still adjusting to the postglacial sea-level rise. There are numerous studies of coastal processes including a number of texts, but the most relevant are those that examine the impact of a rising sea, such as Bird (1993) who produced an overview of submerging coasts as a sequel to his earlier work on global shoreline changes. However, before discussing coastal impacts from rising sea level, it is pertinent to examine the latest IPCC predictions for sea-level rise.

2.2.2.2 *Sea-level Rise Projections – IPCC*

The most comprehensive scientific assessment of sea-level rise projections associated with global warming comes from the IPCC in its various assessment reports. Over the last ten years, these projections have been revised downwards from the initial best-estimate in 1991 of sea-level rise as 0.65 m to the year 2100 (Houghton et al. 1991, Houghton et al. 1992). Subsequent calculations have resulted from either more qualitative expert analysis (0.61 m by 2087: Woodworth 1993), or detailed re-calculation (0.46 m by 2100; Wigley and Raper 1993). The 1996 IPCC report considered that the major conclusions reached by the 1991 Panel remained qualitatively un-

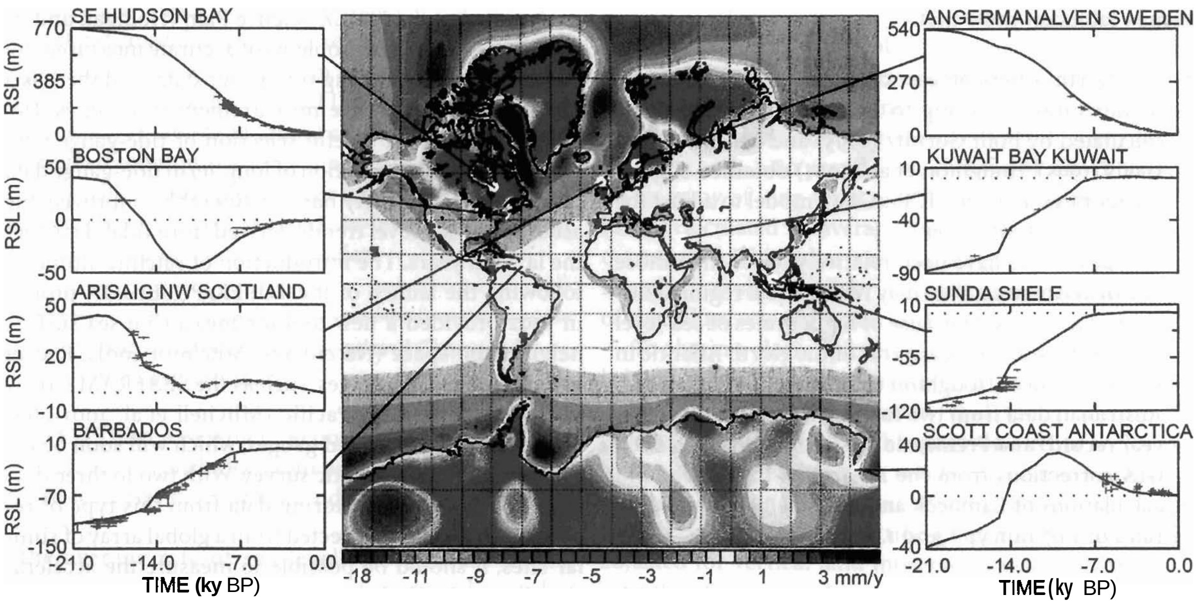


Fig. 2.4. Sea-level rise. Global isostatic response estimated from modelling (from Douglas and Peltier 2002)

changed (Houghton et al. 1996). However, all subsequent estimates of predicted sea-level rise are significantly lower than that estimated in 1991, although the 1995 evaluation warned that direct comparisons could not be made due to differences in factors such as emission scenarios and changes in radiative forcing and that the understanding of climatelsea-level relationships has not changed. The best-estimate by the 1995 Panel was that sea level would rise 49 cm by the year 2100, with a range of uncertainty of 20–86 cm. This projection is lower than that presented by the 1991 Panel due mainly to the lower temperature projections, the inclusion of a slow-down of the thermohaline circulation and changes to the glacier model (Houghton et al. 1996).

The 2001 IPCC scientific assessment (Houghton et al. 2001) reached a number of conclusions about the factors affecting current sea-level change. Over the last century, ocean thermal expansion is estimated to have contributed 0–0.7 mm yr^{-1} based on Atmosphere-Ocean General Circulation Models. The contribution from the melting of glaciers and ice-caps is estimated to range from 0.2 to 0.4 mm yr^{-1}, from observational and modelling studies which include contributions from the Greenland ice sheet (0.0–0.1 mm yr^{-1}) and from the Antarctic ice sheet (–0.2–0.0 mm yr^{-1}). Thus, the eustatic sea-level rise for the last century is estimated as between –0.8 and +2.2 mm yr^{-1}, with a central value lower than expected from the observational records (Houghton et al. 2001).

The 2001 IPCC used 35 SRES (Special Report on Emission Scenarios; Naki-envoi et al. 2000) to project a sea-level rise of 0.09–0.88 m (central value 0.48 m) for the period 1990 to 2100 (see Fig. 2.4). While this central value (not referred to as a best estimate) is similar to the best estimate of the 1995 IPCC, its range of uncertainty is larger. If present rates of terrestrial storage continue then the projections could vary by as much as –0.21 to +0.11 m, and the achievement of the central value by 2100 would require a rate of sea-level rise between 2.2 and 4.4 times the rate for the last century (Houghton et al. 2001).

2.2.2.3 *Coastal Response to a Rising Sea*

A number of coasts around the world have been subsiding in recent decades (Bird 1993, 1996, 2000), including high-latitude coasts in Siberia, Canada and Alaska. The 1995 and 2001 IPCC reports (Houghton et al. 1996, 2001) together with the relevant contributions from the IPCC Working Group II (Watson et al. 1998, McCarthy et al. 2001) outlined the key coastal areas that will be subject to the greatest impact from accelerated sea-level rise (low-lying coral islands, deltaic and coastal plains, sand beaches, barrier coasts, coastal wetlands and lagoons), and further noted the potential impact on gravel beaches and barriers, unlithified cliff coasts and ice-rich cliff coasts. A number of researchers have examined the impact of a rising sea on various types of coast (e.g., Restrepo et al. 2002, Leatherman 2001, Nicholls and Leatherman 1995, Woodroffe 1990, Ellison and Stoddart 1991, Milliman and Haq 1996), and there has been considerable research on coastal vulnerability assessment (see Sect. 2.3.3). The IPCC Working Group II in its contribution to the 2001 IPCC (McCarthy et al. 2001) identified the key potential impacts of climate change and sea-level rise on coastal systems as:

- increased coastal erosion
- inhibition of primary production processes
- more extensive coastal inundation
- higher storm-surge flooding
- landward intrusion of seawater in estuaries and aquifers
- changes in surface water quality and groundwater characteristics
- changes in the distribution of pathogenic micro-organisms
- higher sea surface temperatures
- reduced sea ice cover.

2.2.2.3.1 *Beaches, Barriers and Cliff Coasts*

About 70% of the world's sandy coasts, which occupy about 20% of the global coastline, have been retreating over the last century; 20–30% have been stable and less than 10% have been advancing (Bird 1993). Bird (1993) further argued that sea-level rise will begin to erode the stable coasts and stabilise the accreting coasts. Leatherman (2001) suggested that the figure for eroding sandy coasts is closer to 80–90% for the better-studied and better-documented US sandy coasts.

The impact of sea-level rise on sandy coasts is usually discussed in terms of the simple two-dimensional Bruun (1962) rule which asserts that sandy coasts will adjust and maintain their equilibrium in response to sea-level rise. The Bruun rule has been criticised because its assumptions rarely apply in the real world. However, Leatherman et al. (2000) computed shoreline change rates along the US east coast, producing a good correlation between sea-level rise and long-term erosion on eroding beaches. The results gave ratios of shoreline change to sea-level rise ranging from 110 to 181, compared with the ratios of 50 to 200 according to Bruun's calculations. Leatherman (2001) asserted that this confirms that the lateral beach erosion rate is always two orders of magnitude greater than the rate of sea-level rise.

While the Bruun rule may apply to locations with sufficient sediment supply, uninhibited equilibrium profile development and minimal longshore drift, these conditions are rarely all achieved in one location. Cowell and Thom (1994) attempted a model for shoreline response

Text Box 2.1. Coastal dune response to sea-level rise

Patrick P. Hesp

Sand dune systems may respond to sea-level change in a variety of ways depending on the prevailing wind and wave energy, surf zone-beach type, sediment store and supply, biogeographical province and rate of sea-level rise. The nature of the sea-level change may also be important, as a smoothly rising curve may produce significantly different results from those of a fluctuating curve that displays minor regressions or still-stands.

Sea-level rise does not necessarily equate with shoreline erosion and retreat. Foredunes have been formed and prograded to form foredune plains (including some so-called beach ridge plains) and prograded barriers in sites where sea level is rising relatively slowly and there is a significant sediment supply. Many of the US east coast and Gulf coast barriers have foredune plains that prograded from 4 000 years BP as the rate of sea-level rise slowed and sediment was delivered from the shelf and nearshore (Hesp and Short 1999). However, once the sediment supply diminished or stopped but sea level continued to rise, shoreline erosion and barrier retreat became common. Low foredunes are overwashed, high foredunes are scarped and either removed to be reformed landwards or "roll over" and gradually retreat landwards (Psuty 1992, Ritchie and Penland 1990).

Parabolic dune-fields and transgressive dune-fields may form, or have formed, as a response to rising sea or lake levels. This occurred particularly in the period 10 000 to 7 000 years BP (e.g., Pye and Bowman 1984, Thom et al. 1994, Hesp 1993, Shulmeister et al. 1993, Young et al. 1993, Pye 1983), but also at various stages throughout the mid- to late Holocene (Arbogast 2000, Loope and Arbogast 2000). The seaward edges of parabolic and transgressive dune-fields may be eroded directly by rising sea level, destabilising dunes and releasing sediment for downwind transport. This would occur particularly in regions of high alongshore and onshore sediment supply and/or in regions of high wind and wave energy (e.g., Illenberger and Rust 1988, Fryberger et al. 1990). Semi-arid and arid climates would enhance the development of transgressive dune-fields due to the lack or paucity of vegetation cover. Coastal erosion in the last 100 years has led to the initiation of a contemporary transgressive dune-field phase in some coastal sites (e.g., Froidefond and Legigan 1985, Bressolier et al. 1990).

Given that some dune-fields were initiated by sea-level transgression (Thom et al. 1981, 1992, Short 1987, 1988a, 1988b, Jelgersma et al. 1970, Klijn 1990, Hesp and Thom 1990), it is possible that changes in sea level could also be responsible for the initiation of dune phases, although there is little proof to substantiate this statement. A rise in sea level may initiate dune erosion, destabilisation and the formation of a new dune phase which largely or entirely results from the cannibalisation of older dunes. Alternatively, a few meters of sea-level regression would lead to stranding of lower beach and surf zone sands and activation of sands formerly below wave base. Such changes would be most dramatic on fine-sand, low-gradient, high-energy dissipative beaches. Christiansen and Bowman (1986) and Christiansen et al. (1990) reviewed two scenarios in which dunes may be formed by either rising (transgressive) or falling (regressive) sea levels, following Pye (1983) and after Jennings (1967). They tentatively linked a major phase of Danish dune development with falling or lower sea levels (the Little Ice Age). Sea-level regression on Lake Ontario led to the evolution of dune phases according to Davidson-Arnott and Pyskir (1988).

Since the foredune is either under immediate physical threat or can potentially become relict during sea-level rise or sea-level fall, it is often the response of the foredune that predicates to the fate of the dune-field as a whole. Some possible scenarios include:

1. ***Sea-level fall:*** Beach width increases from a small to a significant amount depending on beach slope (and therefore surf zone-beach type), and the amount and rate of fall. Generally, slowly falling sea level would not result in a wider beach as pioneer vegetation would colonise newly available sand.
   a The foredune stoss slope is supplied with sediment and grows seawards and upwards. A new incipient foredune develops and eventually becomes established (Saunders and Davidson-Arnott 1990). The critical factor here is that pioneer vegetation species are able to keep up with beach width increases or progradation, and therefore the sediment is locked up in foredune development rather than being released landwards to initiate a dune-field phase. This is similar to minor to major beach progradation without any change in sea level (e.g., Hesp 1984a, 1984b). Increased storminess has little effect because the foredune is able to recover by scarp filling and revegetation.
   b The foredune stoss slope and crest is supplied with an excess of sediment off the wider foreshore, which leads to either (i) local and/or more widespread death of pioneer and intermediate species, or (ii) sediment bypassing across the foredune. Subsequent blowout development (or supply of sediment through blowouts already present) and initiation of a phase of foredune instability and breakdown may occur.
   c The foredune is severely scarped at the time sea/lake level starts to fall. Pioneer vegetation is absent (Saunders and Davidson-Arnott 1990) and the additional supply of sediment from the widening beach is transported over the foredune; a phase of dune instability begins. Blowout, parabolic and transgressive dune-field development is possible. A period of increased storminess would aid this process despite water-level fall.
2. ***Sea-level rise:*** Beach width decreases from a small to a significant amount depending on the amount of sea-level rise and surf zone beach type.
   a The foredune stoss slope erodes, blowout development occurs, the crest increases in height and the foredune gradually retreats landwards or "rolls over" (e.g., Saunders and Davidson-Arnott 1990, Psuty 1992, Ritchie and Penland 1990, Giles and McCann 1997).
   b The foredune may be completely destroyed, and either reformed (as a new dune) some distance landwards, or be absent for some period. Landward dunes may be re-activated or may evolve into more erosional dune types.
   c The foredune is scarped, and is subsequently destabilised to form a foredune/blowout complex. This may gradually retreat landwards. Alternatively, parabolic, sand sheet or transgressive dune-fields of various sizes may evolve. Increased storminess merely quickens the process.
   d Sediment supply to the system is still significant and progradation takes place despite sea-level rise. A series of foredunes is built over time to form a foredune plain (e.g., the distal ends of some barriers and spits: Hesp and Short 1999).

Some of these scenarios have been observed in the short-term (10 to 60 years), while others are merely inferred but do explain some observed dune-field patterns. Overall, a relatively wide range of medium- to long-term responses is possible for a relatively small range of initial foredune types and sea-level rise conditions. None of these scenarios for foredunes accounts for environments and coastal dune development where foredunes are absent or very poorly developed. A significant additional complication not examined here is regional wind velocities (and general climatic factors). If regional wind velocities are high, for example, foredunes may always be erosional due to regular blowout development even where sea level is rising, stable or falling, or the coast is prograding.

allowing for sea-level rise and variation in sediment availability within sandy barrier-dune complexes. Hesp and Short (1999) discussed the impact of a rising sea on dune coasts (see Text Box 2.1). There have also been attempts to model coastal responses for other types of coasts, for example, the gravel barrier and cliffed coasts of eastern Canada (Forbes et al. 1995), cliffed coasts in southern England (Bray and Hooke 1997) and cliff erosion associated with El Nifio events impacting on the Californian coast (Komar 1998).

2.2.2.3.2 *Deltaic Coasts*

Deltaic coasts have a particular significance in terms of potential impact from accelerated sea-level rise because many are heavily populated and are already susceptible to inundation, subsidence, shoreline recession and sediment starvation (McCarthy et al. 2001). According to Alam (1996) subsidence rates can reach up to 20 mm yr^{-1} when compaction is combined with other tectonic effects or isostatic loading. Increased subsidence due to groundwater withdrawal alone was attributed to a 17 mm yr^{-1} rise in the relative sea level over a 35-year period for the Bangkok area of the Chao Phraya delta (Sabhasri and Suwarnarat 1996). Impacts of sediment starvation on eroding deltaic coasts have been well-documented for the Nile, Indus, Ebro and Mississippi rivers (Day et al. 1997) and the Rhone and Ebro rivers (Jiménez and Sánchez-Arcilla 1997).

River regulation and management are likely to have greater impacts than climate change for highly-regulated river deltas (Sánchez-Arcilla et al. 1998) (see Text Box 2.2). In the regulated Nile, Mackenzie and Ganges rivers land has been lost due to wave erosion on the outer deltas (McCarthy et al. 2001). Many other deltas could be affected if the rate of sediment accumulation does not match the rate of relative sea-level rise (including that due to subsidence and compaction) and the consequent increased rate of sediment removal.

2.2.2.3.3 *Coastal Wetlands*

The impact of a rising sea on coastal wetlands needs to be placed in the context of the already significant human impact on the coastal zone. For example, large areas of mangroves have been cleared globally for firewood, charcoal or to allow coastal developments such as aquaculture. Thailand alone has lost 50% of its mangroves due to clearance activities over the last 35 years (Aksornkoae 1993). Sediment flux is a key determinant of mangrove response to sea-level rise.

There are various predictions about the impact on different coasts such as low islands, high islands or protected coastal settings (McCarthy et al. 2001) dependent on factors such as sediment supply, mangrove stand composition and tidal range. However, subsiding mangrove coasts provide a good analogue for response to a rising sea, with evidence of mangrove advance inland unless locally impeded by artificial structures (see Harvey and Caton 2003). This impediment to wetland advance caused by coastal defence structures has been referred to as coastal squeeze (Nicholls and Branson 1998).

Coastal marshes respond to sea-level rise by landward horizontal colonisation in a similar manner to mangroves. They also accrete vertically and can be a good geological indicator of paleo sea levels provided caution is exercised with the interpretation, particularly in relation to sediment compaction and spatial aspects of marsh species composition (Kearney 2001). The temporal variability in marsh accretion rates is also important, with examples from Nova Scotia and the Gulf of Mexico showing localised relative sea-level rise of up to 10 mm yr^{-1} (McCarthy et al. 2001). Elsewhere, coastal marshes have difficulty keeping up with sea-level rise on subsiding coasts, as observed in southern England, north-western France and in the Venice Lagoon (Bird 1996), while rapid recession of seaward margins of marshes and mangrove swamps can occur unless there is sufficient peat accumulation or sedimentation rates. On a global scale, almost one quarter of the world's wetlands could be lost by the 2080s as a result of sea-level rise (Nicholls et al. 1999).

2.2.2.3.4 *Tropical* Reef *Coasts*

Over geological time, coral reef coasts have shown an ability to respond to a rising sea-level. Vertical Holocene accretion of up to 26 m has been recorded on the Great Barrier Reef, with growth rates of 6 mm yr^{-1} (Harvey 1986), and 33 m of vertical Holocene accretion have been determined for Atlantic reefs (MacIntyre et al. 1977). Consequently, it has been suggested that healthy reefs with an upper growth limit of 10 mm yr^{-1} will be able to keep up with projected rates of sea-level rise (Buddemeier and Smith 1988, Schlager 1999). However, over half of the world's coral reefs are estimated to be already at risk from human activities and many are degraded due to human impact such as pollution and increased sedimentation (Wilkinson 2000, McCarthy et al. 2001; see Text Box 2.3).

Significant climate change impact on coral reefs is likely to result from increased sea surface temperatures (SSTs) and a reduction in reef calcification. Coral bleaching resulting from loss of symbiotic algae has been predicted to become a more frequent occurrence due to increased SSTs associated with global warming (McCarthy et al. 2001, Walther et al. 2002). While some coral bleaching occurs with an annual frequency (Brown et al. 2000), large-scale periods of coral bleaching over the last 20 years have been associated with El Nifio events and

Text Box 2.2. Deltaic coast dynamics

Yoshiki Saito

Deltas are one of the principal coastal landforms and an important area for human activities. Deltaic coasts are affected by changes both on the land (in the drainage basin) and in the ocean. Sea-level rise is a typical example of a change in the ocean that may severely affect the coastline. Decreases in sediment and water discharge caused by dam construction, sand dredging in river channels and water usage in drainage basins are examples of changes on the land that also impact coasts.

Fig. TB2.2.1. Coastal erosion along a muddy mangrove coast, Chao Phraya River delta. Electric poles still stand in the sea after the shoreline has retreated (photo Yoshiki Saito)

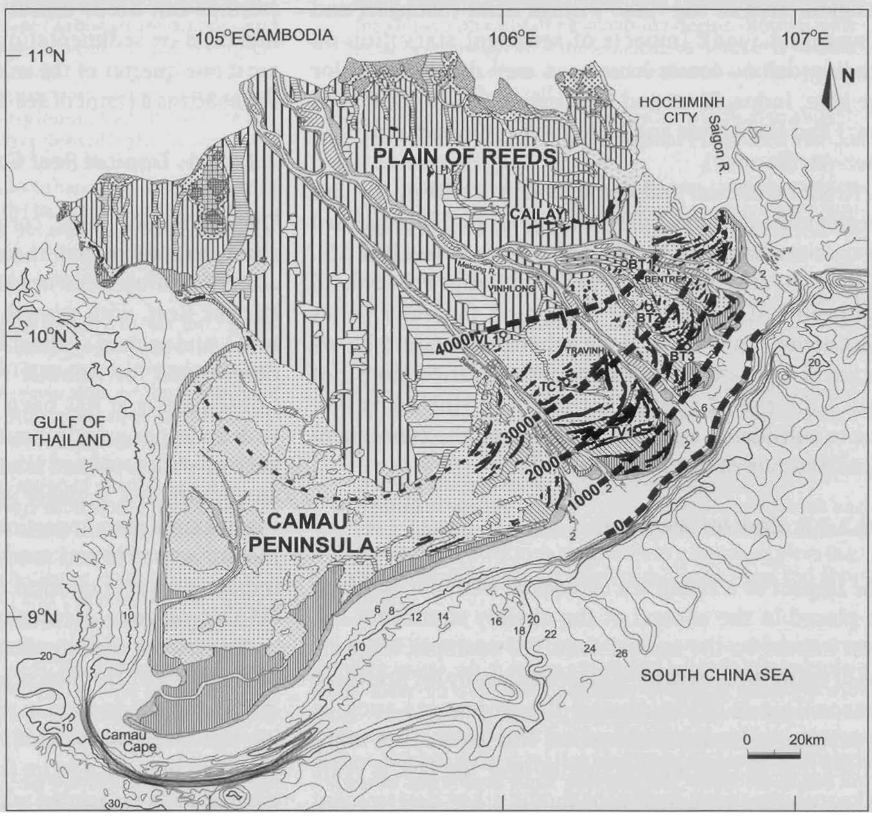


Fig. TB2.2.2. Millennial-scale paleotopography of the Mekong River delta (from Ta et al. 2002)

A delta is an irregular progradation of the shoreline directly fed by a river. Therefore, seaward shoreline migration is an essential feature. Most deltas in the world today were initiated about 8000 years ago. After the last glacial maximum, about 20000 years BP, melting water from glaciers gave rise globally to a rapid sea-level rise except in glaciated areas, where relative sea level fell because of isostatic effects. The rising sea level resulted in the landward migration of the shoreline and the retreat of river mouths, which led globally to the formation of estuaries. The decelerated sea-level rise or slightly falling sea levels over the last 8000 years has allowed the seaward migration of the shoreline at river mouths and the formation of deltas. Sea-level changes during the last 6000 years varied from region to region, controlled mainly by eustatic changes (global sea-level changes), glacio-isostasy, hydro-isostasy and local tectonics, which operate even today.

At present, the formerly glaciated lands are uplifting and surrounding areas, in general, are subsiding. Typical subsiding areas are middle to southern North America, middle to southern Europe, northern Africa and the western and northern parts of the Middle East. Along these shores are found barrier islands and estuaries, which are characteristic features of coastal topography during transgression (landward shoreline migration). Exceptions are those areas where rivers with high sediment supply allow seaward shoreline migration at their mouths (e.g., Mississippi and Nile deltas). Regions far from those glaciated experienced a sea-level highstand, above the present level, about 4000–6000 BP, but since then relative sea level has fallen in such regions, mainly as a result of hydro-isostasy. Deltas are well-developed in these regions, in particular in Asia, middle Africa and middle to northern South America. The coasts of Australia have also experienced a stable or slightly falling sea level over the last 6000 years, but the low sediment supply from the dry continent has allowed unfilled estuaries to persist (Saito 2001).

Wide coastal delta plains are important areas in which people live, carry on economic activities, grow or collect food. However, deltaic coasts around the world have recently become sites of serious environmental problems. One outstanding problem is erosion. A major cause of deltaic erosion is a decrease in sediment discharged from delta-forming rivers because of the construction of dams. About 30% of the global sediments previously discharged to the oceans are now trapped in reservoirs behind dams. Because of the construction of dams (including the Aswan and Aswan High dams), the Nile River mouth has experienced serious erosion, retreating more than 4 km during the last century. Over the last 30 years, dam construction and increased water consumption have caused a dramatic decrease in the amount of sediment and water discharged from the Yellow (Huanghe) River in China, which previously was the second largest river in the world after the Ganges-Brahmaputra in terms of sediment discharge. During the 1980s, sediment discharge from the Yellow River dropped to 70% of that of the 1960s–1970s and to 30% during the 1990s. Moreover, since the Xiaolangzi dam began operation in 1999, discharge has diminished to <10% of that of the 1960s–1970s, resulting in serious coastal erosion. Another related problem is the drying up of the lower reaches of the river. In the worst years of the 1990s, the Yellow River was dry for more than 200 days over more than 500 km from the river mouth because of increased water consumption (Yang et al. 2002).

Another cause of coastal erosion is the relative sea-level rise associated with land subsidence caused by groundwater pumping. The Chao Phraya River delta of Thailand has experienced serious coastal erosion since 1970. Excess groundwater pumping in the Bangkok area has caused rapid subsidence, to more than 10 cm yr^{-1}, and the ground level has subsided more than 2 m, not only in the Bangkok metropolitan area, but also in the coastal region south of Bangkok. Relatively, sea level rose more than 50 cm during 1970–1990 in the coastal region and the shoreline has retreated up to 700 m (Saito 2001). Electric poles can be seen still standing in the sea as a result of the shoreline retreat (Fig. TB2.2.1). The increase in water depth in the near-shore zone caused by subsidence allowed increased wave energy that resulted in coastal erosion, while the cutting of mangroves along the coast accelerated the shoreline retreat. The shoreline retreat occurred even during the early stage of land subsidence, indicating that even a 10 cm subsidence (equivalent to a 10 cm relative sea-level rise) can induce serious coastal erosion. However, since 1992, the shoreline has been stable because of the regulation of groundwater pumping.

Coastal erosion occurs not only at the shoreline (intertidal zonelbeaches), but also in nearshore zones. On wave-dominated coasts, waves can erode substrata to –50 m water depth and strongly influence shoreface topography to 15 m water depth. If there is no sediment supply to the nearshore zone, coastal erosion occurs naturally. Sub-aqueous parts of deltaic coasts are occasionally eroded by waves in advance of the retreat of the shoreline, even if sediments are supplied. Even offshore of a stable shoreline, subtidal flats, the delta front platform and the upper shoreface may be gradually eroded beneath the water.

Since seaward shoreline migration occurs naturally along deltaic coasts, a stable shoreline is not a normal deltaic feature. It is important to understand the natural state of deltaic progradation. For example, the Mekong River delta has prograded steadily for the last 3000 years at a rate of 10–20 km yr^{-1} (Ta et al. 2002, Fig. TB2.2.2). Time-series analyses, using a combination of millennial, centennial and decadal time-scales, give a clearer understanding of the nature of deltas and allow prediction or modelling future delta evolution. The evaluation of deltaic coasts and human impacts on the shoreline requires knowledge of the natural state of deltas and the natural changes that they undergo.

To prevent the erosion of present shorelines, appropriate quantities of sediments are needed above a threshold value. If the sediment supplied from rivers decreases below this value, which is different for each river and delta, deltaic coasts experience serious coastal erosion problems (Fig. TB2.2.3). The sediment discharge levels of the Nile and Yellow rivers are below this level at present.

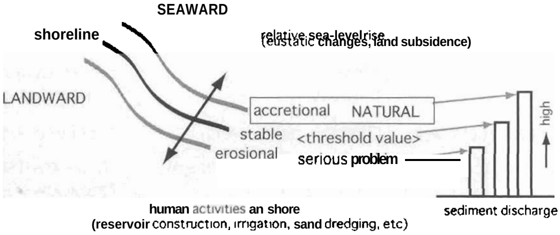


Fig.TB2.2.3. River sediment discharge and shoreline migration of deltas (from Y. Saito 2003, unpublished)

Text Box 2.3. The carbonate coasts of eastern Africa

Russell K. Arthurton

The coasts of Kenya, Tanzania and northern Mozambique have long been a locus of marine-sourced biogenic calcium carbonate sedimentation (UNEP 1998, 2001). They are part of a tectonically passive continental margin on which reefs and backreef sediments – occurring now as limestones – have formed during previous episodes of high global sea level. Their geomorphology includes terraces, cliffs and intertidal platforms, fashioned by sediment accretion, erosion and lithification in response to sea-level changes (Arthurton 2003). The limestones are important aquifers and give rise to freshwater springs in the intertidal zone.

Today's reefs date from about 6 000 BP, when global sea level had risen to near its present position from its low-stand at the last glacial maximum. They are accreting on the ocean-facing rims of limestone terraces, formerly sub-aerial but now forming intertidal to shallow subtidal patches and fringing platforms (Table TB2.3.1). Much of the biogenic carbonate debris (including broken coral) produced by wave impact on the reef front is transported landwards onto a reef bar, where the ocean swell breaks during most stages of the tide. Here the sediment becomes entrapped and enveloped by a tough, coralline algal mat, gradually raising the level of the bar, or it is carried across the bar to contribute to the sediment veneer of the backreef platform. Other biogenic contributions to the platform sediment budget are sourced on the platform itself. Besides molluscan shell debris, these include carbonate sand that has been derived as fragments from the calcified thalli of the alga ***Halimeda,*** which thrives in low intertidal to shallow subtidal thickets. Accumulations of these fragments form the substrata of lagoonal coral mounds and, further inshore, seagrass meadows. Landward parts of the platforms may be rock surfaces free from sediments other than beach sands or beach rocks (lithified former beach sands).

While a few island beaches protected from terrigenous sediment influx may consist exclusively of reef- and platform-derived carbonate sand, the beaches of the mainland fringing- and patch-reef coasts are formed largely of quartz sand – not carbonate, "coral" sand as is commonly supposed. The quartz sand is derived from river discharge, as well as by erosion of beach heads – usually in beach-plain sands – by extreme wave impacts and by exchanges with coastal dune systems.

During the last 6000 years the reef bars have grown upwards to enclose lagoons, while patchy veneers of sandy carbonate debris have accumulated on the platforms (Fig. TB2.3.1). As well, beach sands have accreted over the landward parts of platforms to form beach-plains, and dune systems have developed around the mouths of sediment-charged rivers by transfer of sand from the backshore. Erosion of limestone during this period has been largely limited to the undercutting of exposed cliffs by wave impact (Fig. TB2.3.2).

Table TB2.3.1. Eastern African carbonate coast types and their components in relation to resources and susceptibility to physical change (from Kairu and Nyandwi 2000)

Primary coast type	Geomorphological components	Resources (excluding fisheries)	Susceptibility to physical change
Fringing reef coasts	Forereefs and reef aprons	Reef ecosystem, ecotourism	Dynamite fishing, bleaching, pollution and siltation affecting coral growth, storm damage
	Reef bars and backreef lagoons	Reef ecosystem, ecotourism and coastal defence	Tourism-related damage, sea-level rise
	Backreef platforms with sediment veneer	Halimeda thickets, seagrass meadows, seaweed culture	Sediments may be ephemeral, especially in landward parts; pollution, eutrophication
	Backreef rock platforms		Resistant to erosion
	Beach-rocks	Coastal defence	May be resistant to erosion
	Sand beaches	Tourism, recreation, coastal defence	Shoreface erosion and accretion
	Sand dunes	Coastal defence, groundwater	Beach-head erosion and accretion, aeolian deflation and accretion
	Beach plains	Agriculture, settlements, tourism	Beach-head erosion and accretion
	Rock cliffs	Coastal defence	Resistant except where soft or weathered
	Hinterland, limestone terraces	Groundwater, tourism infrastructure	Resistant exceot where soft or weathered
Patch reef coasts	Offshore patch reefs	Reef ecosystem, eco-tourism	Dynamite fishing, bleaching, pollution
	Intertidal flats (sediments)	Mangrove stands, seagrass meadows	Sediments may be ephemeral; erosion exacerbated by clear felling; pollution, eutrophication
	Rock platforms	Coastal defence	Resistant to erosion
	Beach-rocks	Tourism, recreation, coastal defence	May be resistant to erosion
	Sand beaches including spits	Agriculture, settlements, tourism	Shoreface erosion and accretion
	Beach plains, delta plains	Coastal defence	Beach-head erosion and accretion
	Rock cliffs	Groundwater, tourism infrastructure	Resistant except where soft or weathered
	Hinterland, limestone terraces		Resistant except where soft or weathered

Text Box 2.4. Feedbacks associated with sea-level rise along Arctic coasts

Kenneth *Dunton* and *Lee* Cooper

Arctic shelves constitute 25% of the Arctic Ocean surface area and are the largest continental shelves in the world ocean. They are heavily influenced by coastal erosion, runoff from the large rivers and sea ice, which acts as a major geological agent. Increased rates of erosion, changes in river outflow and varying ice conditions have a direct influence on the input of nutrients, organic carbon and sediments (Rouse 2000, Chapin et al. 2000, Lammers et al. 2001). Such changes have important biological and biogeochemical implications (Fig. TB2.4.1), with respect to feedbacks to the ocean-ice-atmosphere system (Moritz et al. 1990, Tynan and Demaster 1997, Aagaard et al. 1999, Morison et al. 2000, Johnson and Polyakov 2001).

Permafrost on land and in undersea deposits currently sequesters large amounts of radiatively active gases such as methane (Anisimov and Nelson 1996, Danilov 2000, Lee and Holder 2001). However, since many Arctic shorelines are erosional in nature, significant amounts of this methane and the oxidisable organic carbon stored in peat are available for release to the atmosphere. There is also a substantial dissolved organic matter component associated with the retreating shoreline, and allochthonous contributions appear to be of relatively greater importance in the Arctic than in other oceans (Wheeler et al. 1996, Wheeler et al. 1997, Guay et al. 1999, Opsahl et al. 1999). Release of greenhouse gases would have a positive feedback on the degradation of additional permafrost and release of more gas. Continuation of the current apparent decline in sea ice spatial extent and thickness (e.g., Rothrock et al. 1999) could increase water column productivity over the continental shelves, while retreat of sea ice beyond the continental shelf could lead to the disappearance of habitat for ice-associated organisms (e.g., gray whales, walruses, diving ducks and bearded seals). Shoreline erosion rates are also likely to increase with longer open-water periods without protection of sea ice from storm and wave damage. Many of these projected changes are likely to impact the communities that live near the land-sea boundary of the Arctic region.

Off-shore transport of organic matter is significant over the wide and shallow Siberian shelves (Semiletov 1999, Romankevich et al. 2000, Semiletov et al. 2001). Additional climate warming, increased precipitation and increased ultraviolet radiation fluxes could lead to higher remineralisation rates in oxidised waters and sediments on Arctic shelves (Dixon et al. 1994, Freeman et al. 2001). In the Laptev Sea, the supply of sediment and organic carbon from coastal erosion appears to exceed that from riverine input (Rachold et al. 2000), despite the outflow of the Lena River into the Laptev Sea, its mouth one of the world's largest deltas (Fig. TB2.4.2). The lengthening of the ice-free season in summer and the retreat of the summer minimum ice edge further away from the coasts during the past decade is likely to increase the transfer of wave and thermal energy to the coasts, potentially accelerating rates of coastal retreat in the future. Changes in the

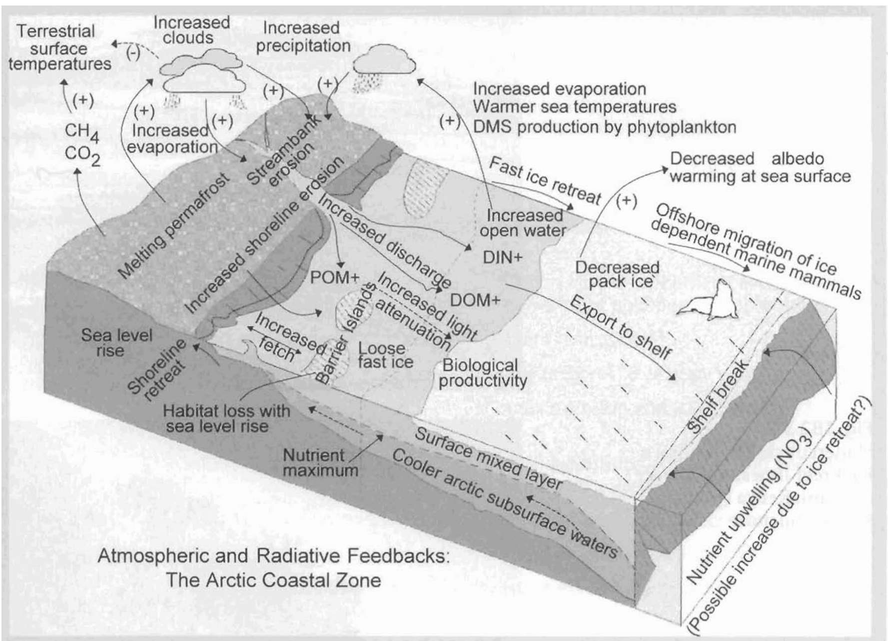


Fig. TB2.4.1.
Major processes and feedbacks associated with climate change in high Arctic coastal ecosystems (from http://arctic.bio.utk.edu/#raise)

increased seasonal maximum temperatures (Hoegh-Guldberg 1999). In recent years, satellite-based SST anomalies have been used to predict and describe the spatial extent of mass coral bleaching events (Skirving and Guinotte 2001, see also http://www.osdpd.noaa.gov//*PSB/EPS/SST/climohot.html*), but less severe events have been more difficult to correlate with coral bleaching in other studies. Coral reefs may be differentially affected by bleaching events and in response show reduced coral and habitat diversity (McCarthy et al. 2001). In addition to bleaching in response to increasing temperatures, increased CO, has been predicted to reduce reef calcification rates to such an extent that its effect should be clearly manifested later in the 21st century (Kleypas et al. 1999, see Text Box 3.2 Chap. 3). A greater frequency of coral bleaching events together with a general reduction in coral reef calcification is likely to produce a geographically variable response to climate change and sea-level rise.

Arctic atmosphere-ice-ocean system recorded during the past decade (Serreze et al. 2000, Huntington 2000) and the reduction in the ice season, combined with a shrinking and thinning of the Arctic sea-ice cover (Rothrock et al. 1999, Serreze et al. 2000), are likely to have profound impacts on the life cycle of marine mammals and impacts on infrastructure and development in the circum-Arctic.

Rates of coastal erosion in the Alaskan and central and eastern Siberian Arctic have been estimated as several meters to tens of meters per year (Are 1999). Paradoxically, a shortened ice season not only results in a loss of protection of the coastline but also increases the action of ice as both an erosional and transport agent. Sea ice plays a major role in the transport of eroded terrigenous sediments onto the Arctic shelf (Stierle and Eicken 2002, Reimnitz et al. 1993, Pfirman et al. 1997, Eicken 2004). The importance of ice transport processes is likely to grow with increases in wind fetch due to reduced ice cover and more frequent and stronger storm events (Proshutinsky et al. 1999). The increase in coastal erosion will be offset locally around the mouths of Arctic rivers that are expected to deliver more sediment with the warming of the hinterland (Syvitski 2002).

Arctic coastal communities depend on access to the sea and to sea ice, but are vulnerable to flooding and erosion. Key human impacts identified by arctic residents include coastal erosion, recent declines in ice extent and thickness, less stable shore-fast ice, changes in permafrost depth, gouging of shelves and coast by sea ice, pile-up of ice on shore, sea-level rise and storm hazards, including flooding. Because of the ice content of coastal sediments, rapid coastal erosion is highly variable and will not be uniform in terms of how it affects individual settlements.

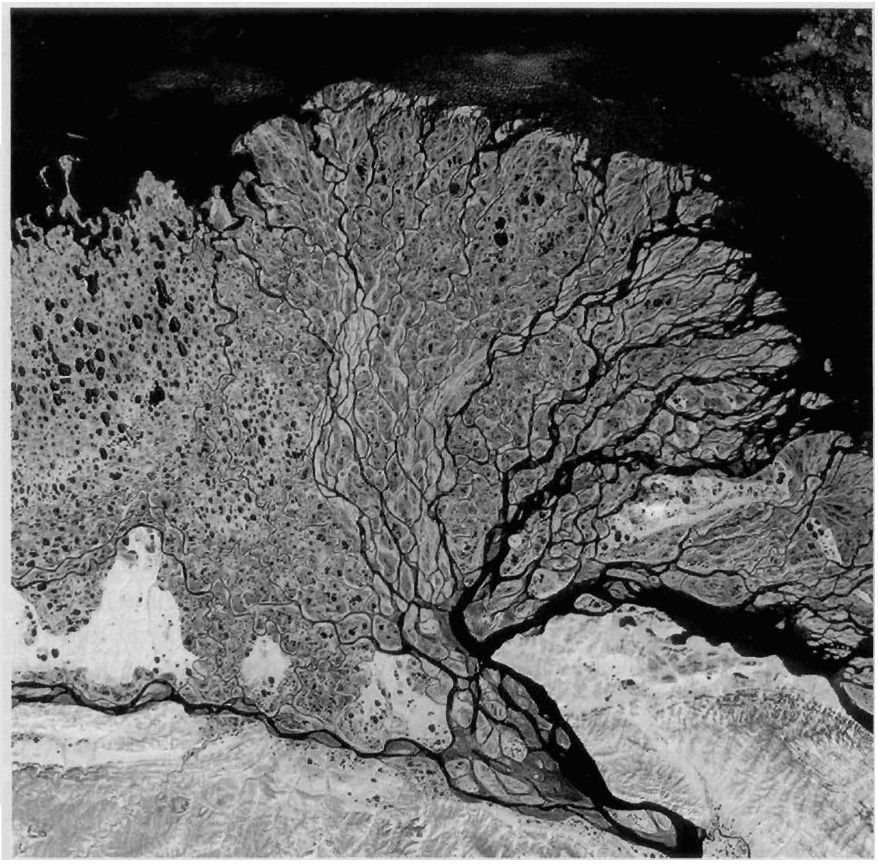


Fig. TB2.4.2.
LandSat 7 image of the Lena River delta, northeastern Siberia (from *http://www.visibleearth.nasa.gov/cgi-bin/viewrecord/18024*)

2.2.2.3.5 *High-latitude Coasts*

Climate models predict a higher than average temperature rise in high latitudes. High-latitude coasts are particularly susceptible to increased periods of ice thaw leading to a reduction of sea ice, creating greater wave exposure and exposing unlithified coastal sediments (see Text Box 2.4). There is evidence of rapid cliff recession in Siberian glacial and peri-glacial deposits in response to ice thaw (Bird 1996) and a seasonally determined active-thaw layer in high-latitude beaches and nearshore zones (Nairn et al. 1998). Any increase in global temperatures will have the effect of extending the periods of thaw and consequently increasing coastal vulnerability. Evidence of this is beginning to emerge from the rapidly eroding sandy coasts in the Gulf of St Lawrence where severe erosion in recent years has been linked to warmer winters (Forbes et al. 1997).

Another impact of global warming is the potential for the thawing of sea ice to increase areas of open water in high latitudes and create a longer fetch for wave generation. Solomon et al. (1994) demonstrated increased erosion rates from model studies on the Canadian coast using global warming predictions of the reduction of sea ice in the area. This type of impact will be exacerbated for high-latitude, low-energy mud and sand coasts such as in the Canadian Arctic Archipelago (Forbes and Taylor 1994).

2.2.3 *Coastal Storms and Coastal Protection*

2.2.3.1 *Increased Frequency and Intensity of Coastal Storms*

Concern has been expressed about the possible increase in the frequency and intensity of coastal storms (e.g., WASA 1998). Observed climate variability and change records show that variations in tropical and extra-tropical storm frequency and intensity are linked to inter-decadal to multi-decadal variations (Houghton et al. 2001) and that there are no significant trends over the 20th century. However, changes have been identified in both the northern hemisphere circulation, linked to the North Atlantic Oscillation, and the variability of the El Niiio-Southern Oscillation (ENSO) with a move toward more frequent El Niño events since 1976. At present there is insufficient evidence to link these changes to anthropogenic forcing (Houghton et al. 2001).

There is conflicting opinion about the projected changes to storm intensity and frequency with increased warming. For example, on the issue of frequency of extra-tropical cyclones, Zhang and Wang (1997) suggested there will be fewer storms with a moister atmosphere, whereas Simmonds and Keay (2000) proposed that an increase in moisture should increase them. A number of theoretical and model-based studies predict increases of 5–10% in peak wind intensities of tropical cyclones/hurricanes and 20–30% increase of mean precipitation in some regions, although there is no evidence of projected changes in the frequency or areas of formation (Houghton et al. 2001). The climate models generally all indicate that the intensity of rainfall events will increase. Some model results presented in the 2001 IPCC report suggested that "*... precipitation extremes increase more than does the mean and the return period for extreme precipitation events decreases almost everywhere ...*" (Houghton et al. 2001). There is also agreement from most models showing a mean El Niiio-like response for the tropical Pacific with a shift of the mean precipitation to the east. A rise in mean sea level alone will increase flood risk.

Although there is no consensus at a global level for changes in coastal storm frequency within climate models, regional studies have predicted changes. Two model studies for north-western Europe examined the sea-level impacts of predicted changes in storm climatology and found significant increases in five-year extreme water levels (Houghton et al. 2001). Similarly, a regional study for southern Australia (Hubbert and McInnes 1999) demonstrated that increased wind speeds with cold front-associated storm surges could increase flood events in coastal localities.

The predicted sea-level rise associated with global warming must be considered in conjunction with the prospect of increased storm intensity. Raised water levels alone will allow higher energy waves to reach the coast and consequently reduce the average recurrence interval (ARI) of major storm damage events. Increased storm intensity will further reduce the ARI of storm events used for coastal planning and management purposes, such as storms with an ARI of 50 to 100 years. This collectively increases the risk to coastal populations, although locally it will depend on coastal resilience, which in turn depends on ecological, geomorphic and socio-economic variables.

The notion of the physical susceptibility to sea-level rise of natural systems was discussed by McCarthy et al. (2001), who defined the capacity of the system to respond as its resilience and resistance. While the physical concepts refer to the natural coastal system, resilience is affected by human activities. Socio-economic resilience is the capacity of a society (including its technical, institutional, economic and cultural abilities) to cope with the impacts of sea-level rise and climate change. This may involve any of the IPCC coastal response options of protection, accommodation and retreat (see Sect. 2.2.3.4). A range of technological adaptation strategies exists (Klein et al. 2000), although the use of protective works, for example, needs to be considered within the planning context of a particular country. Differences between countries in terms of physical susceptibility to sea-level rise and socio-economic resilience are highlighted by the vulnerability of small island states, which are located mainly in the tropics and subtropics. For this reason the 2001 IPCC report devoted a separate chapter to small island states in its discussion on climate change impacts, adaptation and vulnerability (McCarthy et al. 2001).

2.2.3.2 *Coastal Vulnerability Assessment*

The concept of coastal vulnerability differences between countries that have varying capacities to respond led to the notion of coastal vulnerability assessment. The drive for a global approach to coastal vulnerability assessment from the IPCC through its "Common Methodology" was intended to assess the implications and costs of human-induced climatic change on coastal systems. Vulnerability was defined as "*... a nation's ability to cope with the consequences of an acceleration in sea-level rise and other coastal impacts of global climate change.*" (IPCC 1992) and included the impact that these changes may have on

socio-economic and ecological systems. The aim was to use cost-benefit analysis of vulnerable areas to assess the best response option, such as protection of the coast by defence works, accommodation of changes, retreating from vulnerable areas, or to do nothing (IPCC 1992). Results were elaborated in a number of tables to produce vulnerability classes of low, medium, high and critical, based on relative or absolute quantities. The LOICZ Implementation Plan recognised the need for detailed research in this area with a specific long-term objective for "*... improved methodologies for vulnerability assessment at the regional and global scales ...*" (Pernetta and Milliman 1995).

Eastern hemisphere vulnerability studies revealed a lack of data on basic coastal topography and a lack of operational technical capacity for describing the complicated non-linear geomorphological and ecological impacts of climate change. The spatial distribution of relative sea-level rise and other coastal implications were ignored due to a lack of regional climate scenarios, as were the other potential impacts of climate change such as extreme events (McLean and Mimura 1993). It was concluded that more work was needed on broader socio-economic needs, including traditional aesthetic and cultural values such as those of subsistence economies and traditional land tenure systems. A further criticism was that the concept of vulnerability did not take into account the concept of resilience of coastal systems to the various stresses (McLean and Mimura 1993).

Text Box 2.5. Coastal Vulnerability Index (CVI)

Vivien Gornitz

Future increases in global mean sea level associated with global warming will exacerbate existing coastal hazards. The relative vulnerability of different coastal environments to sea-level rise can be compared on regional to national scales by means of a Coastal Vulnerability Index (CVI), which incorporates basic information on a number of geophysical and oceanographic factors. In its original form, the CVI was based on seven coastal "risk" variables - mean elevation, shoreline displacement, relative sea-level rise rates, mean tidal range, significant wave height, geology and geomorphology. Each risk variable was ranked on a linear scale of 1 to 5 ("very low risk" to "very high risk"), where 5 represents the highest risk (Gornitz and White 1992). A number of meteorological variables were later added and several formulae for a CVI were proposed and tested (Gornitz et al. 1994). Since they showed a high degree of correlation (Gornitz et al. 1994), the following form of the CVI was adopted:

$$CVI = [(x_1 \cdot x_2 \cdot x_3 \cdot \ldots x_n)/n]^{1/2}$$

where, n = number of variables present, x = the risk classification of each variable on a scale of 1 to 5 (with 5 representing the highest risk), x_1 = mean elevation, x_2 = mean shoreline accretion/erosion rate, $x_3 \ldots x_n$ = relative sea-level change rate. The square root was introduced to compress the extreme range of scores.

The US Geological Survey (Thieler and Hammar-Klose 1999, Hammar-Klose and Thieler 2001) has adopted this basic approach with several modifications, in addition to updating and refining the original data sets. The number of variables in the CVI was reduced to six, omitting the geology variable on the grounds that it is highly correlated with geomorphology. A regional slope extending 50 km landward and seaward of the shoreline (rather than the mean elevation) was obtained from topographic and bathymetric data. Microtidal (< 2 m) coasts were assigned a high risk rating whereas macrotidal coasts (> 4 m) received a low rating, reversing the scheme originally used by Gornitz and White (1992). The US Geological Survey based their change on the influence of storms on the coast. A microtidal shoreline is always "near" high tide, and thus at greatest risk to flooding by storms.

Figure TB2.5.1 shows the CVI index applied to the United States Gulf of Mexico coast, a region at very high risk to sea-level rise, because of its generally low topography, low resistance to erosion and very high rates of relative sea-level rise, particularly in Louisiana (see Sect. 2.6.1). Very high risk areas along the US Atlantic Coast include portions of the south shore of Long Island, New York, New Jersey, the Chesapeake Bay area, Cape Hatteras, North Carolina and sections of South Carolina and Florida. On the US Pacific Coast, very high risk areas are more scattered along the southern California shore (e.g., near San Diego), near San Francisco and north of Cape Blanco, Oregon.

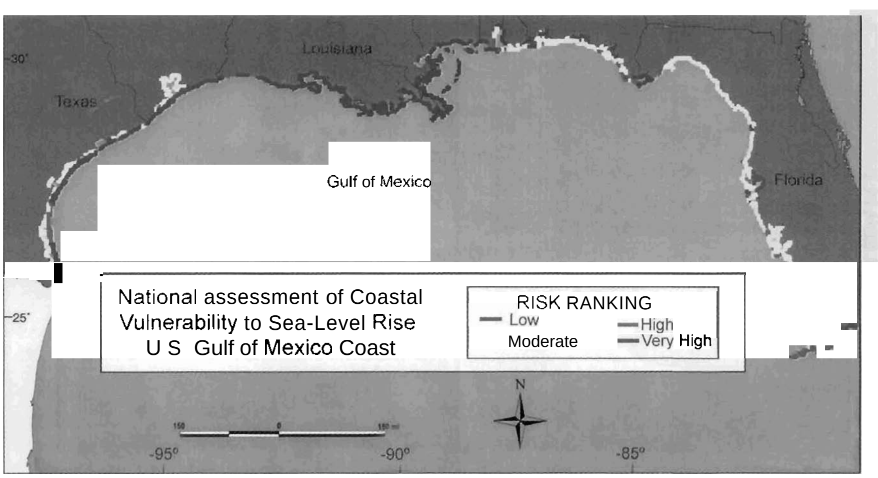


Fig. TB2.5.1. Coastal Vulnerability Index (CVI) applied to the United States Gulf of Mexico coast, a region at very high risk to sea-level rise because of its generally low topography, low resistance to erosion and very high rates of relative sea-level rise (from Thieler and Hammar-Klose 2000)

The IPCC Working Group II to the 1995 IPCC contributed studies from 23 countries that provided an estimate of land threatened by sea-level rise (Watson et al. 1996). The IPCC highlighted the vulnerability of small island states to climate change in a report on regional impacts of climate change (Watson et al. 1998) but stressed the need to consider other factors that contribute to their overall vulnerability. Subsequently, Klein and Nicholls (1999) and Harvey et al. (1999) critically analysed the "Common Methodology" approach and both evaluations proposed alternative methodologies, the latter of which has been applied in Australia. McCarthy et al. (2001) provided a review of the socio-economic impacts of vulnerability studies and studies on the economic costs of sea-level rise.

More recent attempts to use a global vulnerability approach have focused on a methodology for the synthesis and upscaling of vulnerability assessment studies (SURVAS, ***http://www.survas.mdx.ac.uk)*** in order to develop improved regional and global perspectives on accelerated sea-level rise and associated impacts (Nicholls and de la Vega-Leinert 2001, Nicholls 2002). The SURVAS methodology has been used in regional workshops in Europe, Africa and the Asia/Pacific region with outputs in the form of databases, workshop reports and scientific papers including a special issue of the Journal of Coastal Management (2002). Notwithstanding attempts to: *(i)* conceptualise the framework for coastal vulnerability assessment (McCarthy et al. 2001), (ii) modify methodologies to incorporate different socio-economic, cultural and planning contexts, or (iii) increase their usefulness with synthesis methodologies, it is clear that there has been a mixed success for the various global and regional approaches to coastal vulnerability assessment. One current initiative is a European Union-sponsored project DINAS COAST (*http://www.dinas-coast.net*) that is developing a simulation model and CD-based tool (DIVA) with application at local, regional and global scales, which integrates both environmental and social information to allow users to produce quantitative information on a range of coastal vulnerability indicators. Titus and Richman (2001) prepared vulnerability maps for the US Atlantic and Gulf coasts, based on elevation and not incorporating existing or projected coastal processes and climate change. The US Geological Survey has developed a critical vulnerability index for the US coastline (see Text Box 2.5).

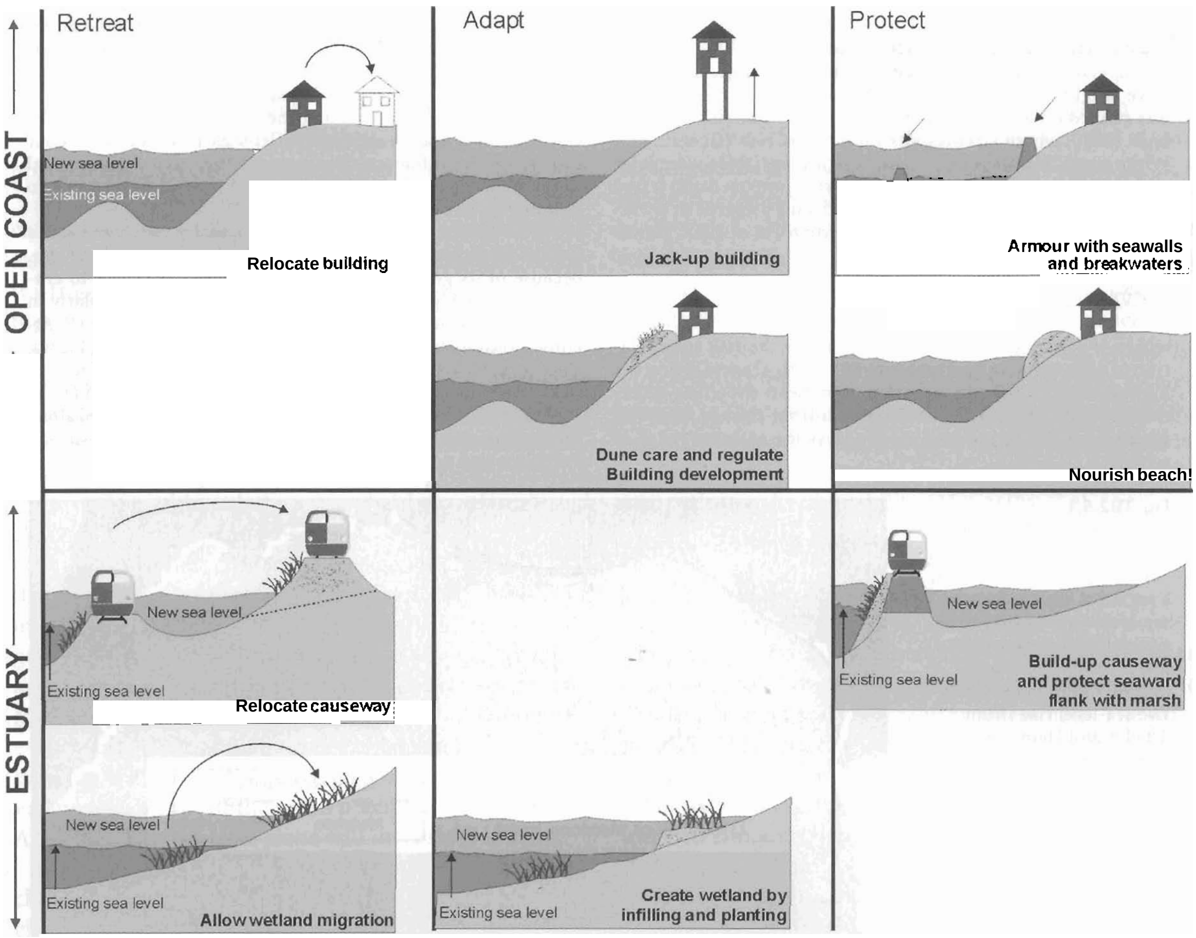


Fig. 2.5. Sea-level rise. Coastal response options for New Zealand (from Bell et al. 2001)

2.2.3.3 Management Response to Coastal Vulnerability

The IPCC approach to coastal adaptation (IPCC 1992) outlines three societal response options:

a *Retreat* – either a forced or a managed retreat where no attempt is made to protect the coast or property. This could involve the relocation of houses, for example, or the abandonment of various coastal land uses.
b *Accommodate* – an adjustment to the higher sea level, such as elevating coastal properties or changing land use to more compatible industries such as aquaculture.
c *Protect* – maintain the current land use with some form of engineering response involving either hard protection (e.g., seawalls) or soft protection (e.g., beach nourishment).

A summary of the various response options for open coasts and estuaries has been prepared for New Zealand coastal managers (Bell et al. 2001; Fig. 2.5). Various countries have adaptation strategies through managed retreat, including setback distances, no-build zones, rolling easements and managed realignments (McCarthy et al. 2001). South Australia has a policy based on earlier IPCC sea-level rise predictions (IPCC 1991, Harvey and Belperio 1994), i.e., that new coastal development should be capable of being reasonably protected from a 1 m sea-level rise by 2100. It recommends that site and building levels should be 0.3 m above the loo-year ARI water level and adjusted to allow for localised subsidence or uplift. Building floor levels should be an additional 0.25 m above this level, and approvals should depend on the capability to protect from a further 0.7 m of sea-level rise (e.g., by means of a bund wall or raising the building). In the case of flood-protected areas, the 100-year ARI design flood level for the development area must incorporate the extreme tide (plus surge) and storm-water events, together with wave effects. The policy also makes a general recommendation for an erosion setback distance based on 100 years of erosion (or 200 years for major development) at a site, allowing for local coastal processes, a sea-level rise of 0.3 m by 2050 and storm erosion from a series of severe storms. The evolving circumstances and management approaches for New York City, USA (see Text Box 2.6) add emphasis to the awareness about sea-level rise required in urban planning.

A distinction needs to be drawn between natural coastal vulnerability and the vulnerability of human lives and property to the effects of climate change and sea-level rise. Some coasts such as crystalline cliffed coasts may be resistant to these impacts, while coral reef or wetland coasts may respond naturally by accretion. Others such as barrier coasts may naturally migrate inland, but where humans or their property are at risk (e.g., on the US east coast) it is necessary to have societally-focused adaptation strategies (Titus 1998). In many cases, attempts to use hard protection have exacerbated the problem because of a lack of understanding of coastal processes (Doornkamp 1998, Pethick 2001) but there appears to be the beginning of a renaissance in concepts and policies for coastal management (for example, de Vries 2001).

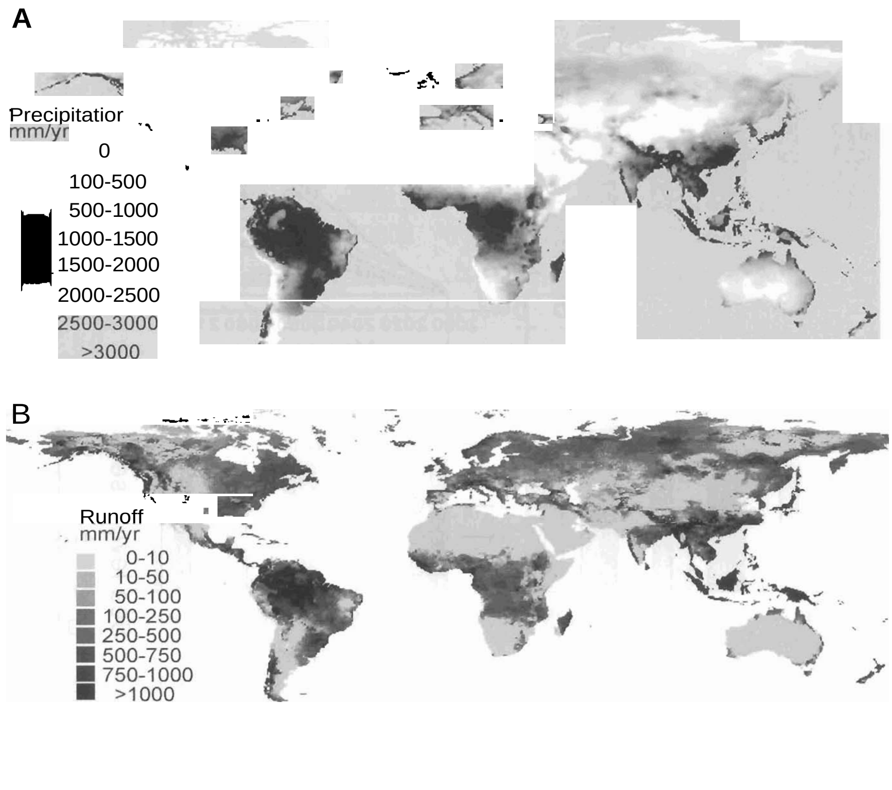


Fig. 2.6. Water flux. a Global precipitation (mm yr^{-1}) (Syvitski et al. 2003). b Hydrological runoff (mm yr^{-1}) after accounting for all forms of evapo-transpiration and human-induced consumption. The hydrological runoff divided by the drainage area equals the water discharge ($km^3 yr^{-1}$) (http://www.bafg.de/grdc.htm)

Text Box 2.6. Coastal storms and coastal protection: New York City, USA case study

Vivien Gornitz

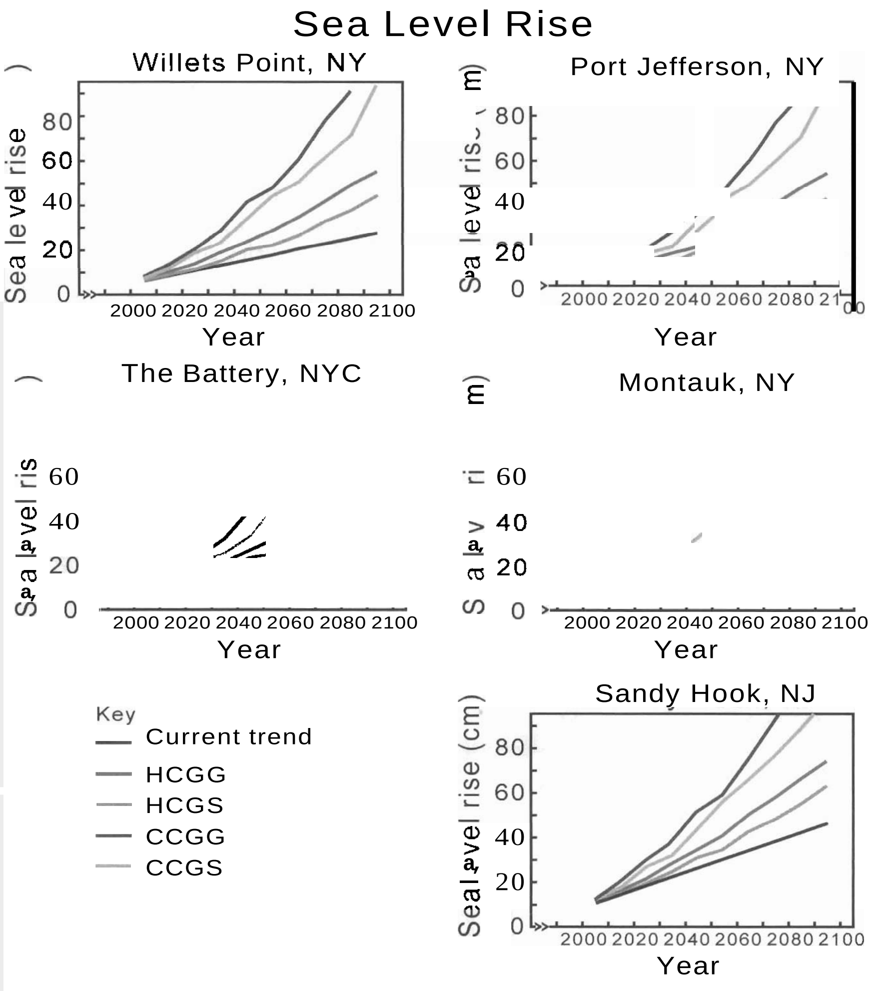
The New York City area will be increasingly vulnerable to storm surges, beach erosion and loss of wetlands as sea level rises. Four out of five of the New York City boroughs are located on islands that are interconnected by a large network of bridges and tunnels. Major portions of transportation arteries in the New York metropolitan area lie at elevations of three meters or less, and have been flooded by severe storms in the past. Within the past 40 years, at least three coastal storms have generated flood levels > 2 m above mean sea level, causing serious inundation and disruption to area transportation systems. Although there has been considerable interdecadal variability in storm activity during this period, no statistically significant secular trend has emerged. However, in recent years, new residential and commercial complexes have risen on landfill in lower Manhattan and have replaced decaying piers and factories at water's edge across the Hudson River in New Jersey. Ironically, the revitalisation of the New York City waterfront comes at a time when climate change may begin to threaten the shoreline.

Beaches and coastal wetlands provide the large regional urban population with recreational opportunities and also act as buffer zones against destructive storm surges. However, beaches of Long Island and New Jersey have been eroding historically, in part due to ongoing sea-level rise and in part due to construction of hard structures such as jetties and groynes. Coastal erosion of regional beaches, particularly after major storm events, is periodically reversed by expensive beach replenishment projects undertaken by the US Army Corps of Engineers. Since the 1920s, US$ 250 million has been spent on beach nourishment for just six case study sites in northern New Jersey and Long Island, New York (Gornitz et al. 2002).

Potential consequences of sea-level rise and storm hazards for the New York metropolitan area have been evaluated recently (Gornitz et al. 2002, Hartig et al. 2002). Regional sea level has climbed steadily by 22–39 cm during the 20th century. These values exceed the global mean rise of the last century (10–20 cm, Houghton et al. 2001) because of local subsidence due to ongoing glacial isostatic adjustments (Peltier 2001). Projections based on tide-gauge data and climate-model simulations with increasing levels of greenhouse gases ± sulfate aerosols suggest that regional sea level could move upward by another 24 to 108 cm by the 2080s (Fig. TB2.6.1).

Because of the increase in mean ocean levels, today even minor storms would produce coastal flooding equivalent to that produced earlier by a major storm. Floods from a 100-year storm (having a probability of recurring once in 100 years) could reach 3.1 to 3.8 m by the 2050s, and between 3.2 to 4.2 m by the 2080s. More importantly, flood return periods due to both mid-latitude and tropical cyclones (hurricanes) would shorten dramatically throughout the region. The current 100-year flood height would have an average recurrence period of 68 to 19 years by the 2050s

Fig. TB2.6.1.
Projections for the New York City metropolitan region based on historic tide gauge data and climate model simulations (United Kingdom Hadley Centre and Canadian Centre for Climate Modelling and Analysis) with increasing levels of greenhouse gases ± sulfate aerosols (from Gornitz et al. 2002)



and 60 to 4 years by the 2080s (Fig. TB2.6.2). Beach erosion rates could increase several-fold, requiring up to 26% more sand replenishment by volume (Gornitz et al. 2002).

The salt marshes of Jamaica Bay, Gateway National Recreation Area, New York, constitute an important regional ecological resource, providing prime habitat for migratory birds and other wildlife. Initial aerial photograph analysis of several tidal salt marshes in Jamaica Bay over a 39-year period (1959–1998) showed about 12% reduction in land area. A more comprehensive remote sensing survey has revealed even more extensive marsh losses over the entire bay. Around 51% (by area) of mapped island salt marshes vanished between 1924 and 1999, with 38 percent of the losses occurring between 1974 and 1999 (Hartig et al. 2002). Although the historical rise in sea level is probably a contributing factor, no recent acceleration in this trend has been detected. Observed losses may be related to low marsh accretion rates, stemming from a sediment deficiency caused by previous dredging for navigation, armoring of the shoreline and inland development (Hartig et al. 2002). Other possible factors include erosion by storms and nutrient enrichment of the bay. Unless restoration efforts are initiated soon, the very survival of Jamaica Bay salt marshes and their wildlife populations may become increasingly precarious in the face of future sea-level rise. Buffer zones should be designated to allow for landward migration of coastal salt marshes. However, landward movement of the Jamaica Bay salt marshes could be impeded by existing seawalls and other "hard" structures, as well as by future inland development.

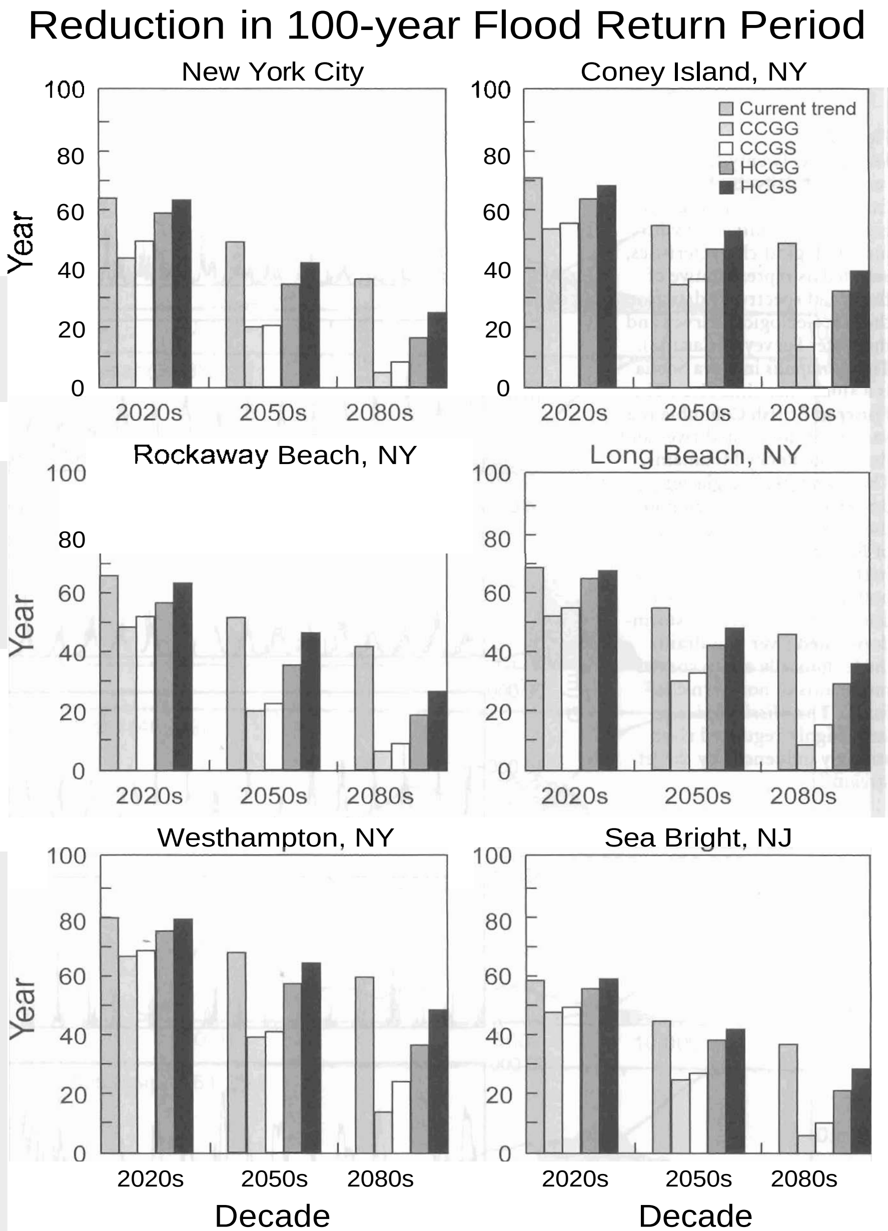


Fig. TB2.6.2. Reductions in recurrence intervals of the 100-year flood for the New York City metropolitan region due to sea-level rise based on historical data and climate model simulations (from Gornitz et al. 2002)

2.3 Changes in the Flux of Water and Sediment

2.3.1 Processes and Mechanisms

The discharge of river water to the coastline reflects the global distribution of precipitation, drainage basin area and relief, the loss of moisture back to the atmosphere through various mechanisms of evaporation and sublimation, and the time-dependent release of stored water to drainage channels (Fig. 2.6). Discharge to the coastal zone reflects this global variability.

In the sub-polar to polar regions, as well as in mountainous regions, precipitation may be initially stored in snow or ice fields and subsequently released across warmer seasons. One consequence of this initial storage is to modulate the intensity of flood waves. In less pristine regions, river flow is modulated by reservoirs and agricultural/industrial diversion schemes. Therefore, there is a wide spectrum of river mouth hydrographs (Fig. 2.7). Small maritime rivers (e.g., Annapolis) are flashy with little flow modulation. Snowmelt-dominated rivers (e.g., Fraser and Liard) show freshet flood waves that rise sharply in the spring and progressively decrease into the early summer. Glacier-dominated rivers (e.g., Homathko) discharge much of their water in the warm summer season, punctuated by outbursts and storm events. Storm-dominated rivers (e.g., Eel) reflect the duration and intensity of climatological events (e.g., El Niño storms, hurricanes). Larger rivers (e.g., Mississippi) reflect continental-scale influences such as the behaviour of the jet stream.

Rivers are the chief mechanism for the delivery of terrestrial sediment to the ocean (> 95%, Table 2.2). Sediment is delivered as bedload (sediment moved along the river bed by rolling, skipping or sliding) and suspended

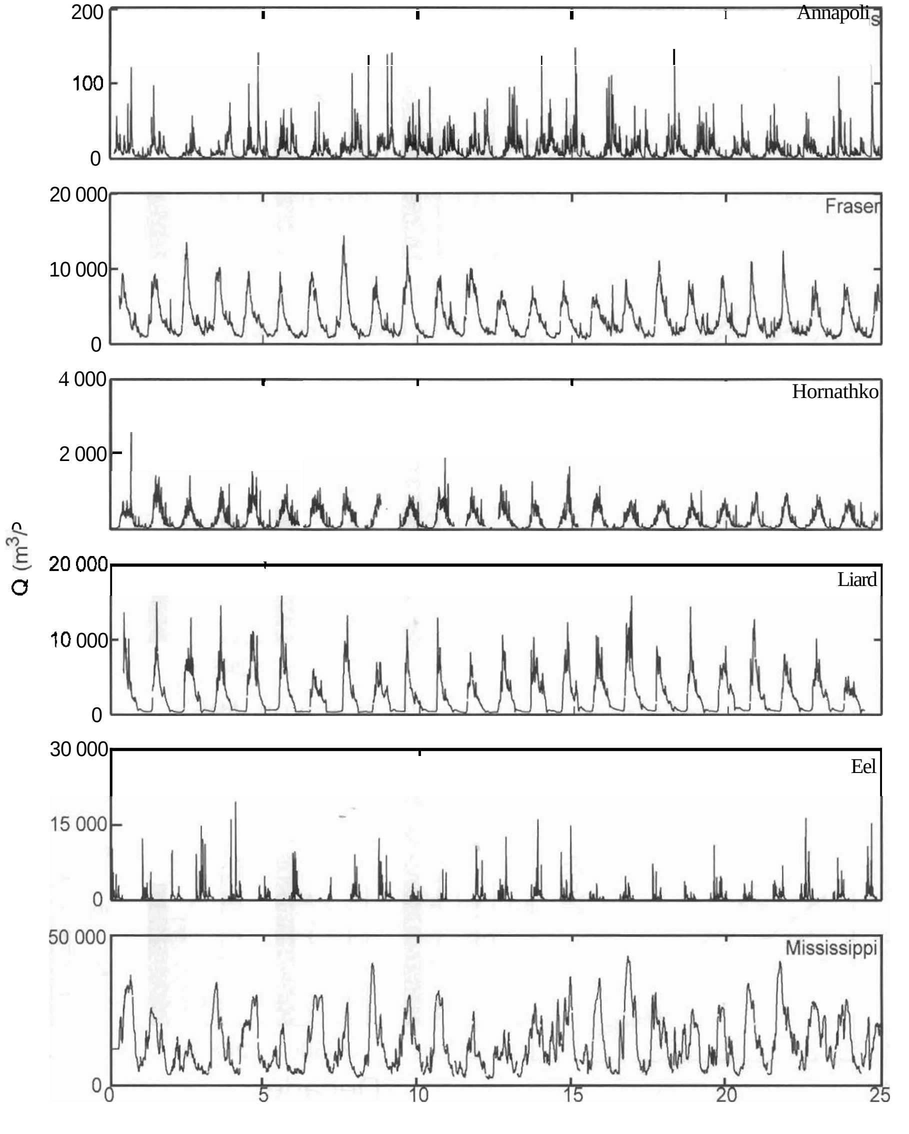


Fig. 2.7. Water flux. Twenty-five year records of daily discharge from six North American rivers with very different hydromorphological characteristics, selected as representative of the broad spectrum (data from the U.S. Geological Survey and the Water Survey of Canada). The ***Annapolis*** in Nova Scotia is a small, maritime river. The ***Fraser*** in British Columbia is a snowmelt-dominated river and drains the Rocky Mountains. The ***Homathko*** is a glacier-dominated river incised into the rugged coastal mountains of British Columbia. The ***Liard*** in the Northwest Territories is both glacier- and snowmelt-dominated. The ***Eel*** is a storm-dominated river that drains the tectonically active coastal mountains of northern California. The ***Mississippi*** is a large highly-regulated river, strongly influenced by the jet stream

load (sediment that is fully supported by fluid flow and maintained by the upward component of fluid turbulence). Bedload is flow-dependent, difficult to measure but easier to predict; suspended load is source-dependent, relatively easy to measure but hard to predict. The proportion of suspended load to bedload increases with the size of the drainage basin, and is typically in the range of 10:1 to 100:1 of the total sediment flux to the continental margins (Meade et al. 1990).

Most rivers transport 90% of their load during 6 to 60% of their time (Fig. 2.8). At the extreme, the Eel River, which drains the steep-sloped coastal mountains of California, delivered more sediment in three days than it had carried in total over the prior 7 years. Seasonality is important when examining sediment flux patterns. Rivers fed by ice fields carry most of their sediment in mid- to late summer. Rivers dominated by snow discharge carry most sediment in early spring. Some river basins have limited and seasonal sediment sources such as snowmelt, where once the seasonal sediment is flushed out of the river, the river remains clean for the rest of the year.

A river's total sediment load depends largely on its drainage area, relief and average basin temperature: the larger the basin, the higher the mountains and the

Table 2.2. Sediment flux. Global estimates of the flux of sediment from land to the ocean (Syvitski 2003a)

Transport mechanism	Global flux (Gt yr^{-1})
Rivers:	
suspended load	18
bed load	2
dissolved load	5
Glaciers, sea ice, icebergs	2
Wind	0.7
Coastal erosion	0.4

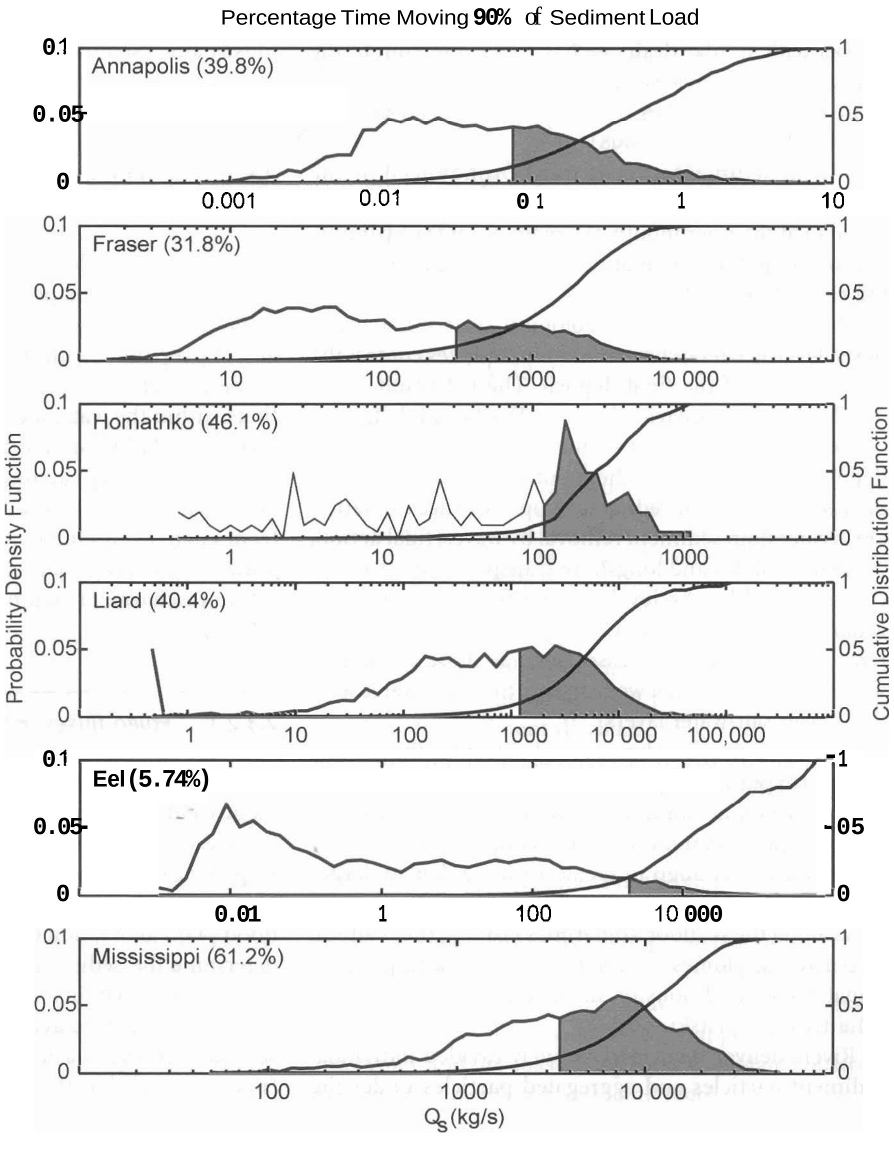


Fig. 2.8. Sediment flux. Six selected North American rivers having different hydro-morphological characteristics (see Fig. 2.7) displaying their sediment discharge to the coast in terms of their probability frequency distribution and cumulative distribution. Displayed in red are the events that together contribute 90% of the load discharged each year. Inset boxes show the amount of time needed to carry 90% of the sediment load; for example, the Eel River will discharge 90% of its load in 5.7% of the year (21 days)

warmer the basin, the greater the load (Milliman and Syvitski 1992, Morehead et al. 2003). Regional features and geological structure also substantially influence the sediment load of rivers (Wang et al. 1998). Geological factors on sediment yield include the bedrock type being eroded (sedimentary material is more easily eroded than crystalline material), the development of extensive soils and the impact of the last ice age. For example, the runoff of the Yellow River is about 1/15 of that of the Mississippi River, 1/183 of Amazon River, and 1/2 of Nile River, but the sediment load of the Yellow River is three times the load of the Mississippi River, twice that of the Amazon River and nine times more than the Nile River. The difference in sediment concentrations is even higher. The reason is that the Yellow River flows through the Loess Plateau, and the unconsolidated loess is easily eroded. The major load of silt carried by the Yellow River is eroded mainly from its middle and lower reaches, i.e., the Loess Plateau and ancient fluvial fan (Wang et al. 1998). The effect of glaciation is seen by comparing two rivers from Alaska with similar drainage basin size and runoff: the glacial-fed Cooper River has a yield (1200 t $km^{-2} yr^{-1}$) more than ten times larger than the non-glacial Kuskokwim River (Wang et al. 1998).

Fluvial sedimentation is most pronounced at the mouths of rivers and is dominated by the deposition of bedload at the river mouth, the main point for hydraulic transition and sedimentation under the seaward-flowing river plume. Tides and wave action rework this sediment along the coastline. Geological-scale fluctuations in sea level, glaciers, climate and earthquakes control the development of the final deposit. The net result is the development of estuaries, lagoons, fjords and deltas.

Deltas have nearly flat surface expressions of alluvial deposits formed around the mouth of a river. A marine delta survives only if sediment supply and accumulation are greater than sediment removal by waves, tidal action, submarine slides and longshore transport. Estuaries are characterised by the level of stratification within the water column: a balance of buoyancy forces set up by river discharge and processes such as those associated with tidal action which work to mix the fresh water with the denser salt-water layers.

Much of a river's sediment load will enter the sea as a surface (hypopycnal) plume: fresh water flowing into an ambient basin of saline water. A bottom (hyperpycnal) plume occurs when the sediment concentration in river water is heavy enough to create a plume density larger than the seawater density. The plume then plunges and flows along the seafloor, sometimes eroding the seafloor. Hyperpycnal plumes are often generated during exceptional flood conditions of small to medium-size rivers (Mulder and Syvitski 1995).

Rivers deliver their load as finely divided individual sediment particles and aggregated particles under the influence of hydrous oxides, organic coatings and microflora. As the river water mixes with the marine water, flocculation of silts and clays rapidly occurs. In the flocculated state, particles settle faster (by orders of magnitude) than if they were to settle individually. The influence of organic matter dominates particle dynamics in the ocean.

Seasonality in hydrological and biological events can affect the agglomeration of marine particulate matter. Zooplankton graze on river-transported flocs, producing mineral-bearing faecal pellets that rapidly settle to the seafloor. The spatial distribution and density of zooplankton is highly seasonal, often reaching maximum populations in the late spring and early fall, coincident with floods and maximum sedimentation. In temperate regions of southern Canada, the peak runoff of lowland rivers is synchronous with winter rainstorms, while the peak runoff of glacier- or snow-fed rivers is during the spring-summer period. Since the maximum biological productivity occurs during the latter period, organisms would mostly affect sedimentation of particles discharged from glacier- and snow-fed rivers (Wang et al. 1998).

2.3.2 Sediment Flux to the Coast: Climate versus Humans

Anthropogenic influences and changing climate can affect the "natural" supply and flux of sediment along hydrological pathways. Rivers and their drainage basins evolve over time, with the discharge to the ocean from modern rivers strongly influenced both by paleo-conditions within the watershed and by human perturbations (Hay 1994, Milliman and Syvitski 1992). Understanding sediment discharge across a broad time-scale allows for better predictions of the impact of humans as distinct from changes in climate. For example, the trapping of sediment by terrestrial reservoirs is fundamental to the future discharge of sediment to the coastal oceans (Hu et al. 1998).

2.3.2.1 Paleo-fluxes before Human Influence

Fluvial systems evolve with the landscape, so that sediment yield observed today is influenced by the geological history of the paleo-systems. Rivers like the Mississippi, for example, were fed by ice-sheet fed rivers (Ohio, Missouri) during the Pleistocene, resulting in extensive flood plain alluviation (Hay 1994). It remains difficult to determine the sediment flux from pristine rivers, given the natural variability within river systems.

While there is no accepted value for the paleo-flux of sediment to the coastal oceans, Milliman and Syvitski (1992) argued that the modern –20 Gt yr^{-1} global flux

Fig. 2.9. Sediment flux. Predictions of the sediment load of rivers with basins larger than 25 000 km^2 (Vorosmarty and Syvitski, unpubl. data 2002). Much of the world sediment is shed from the rivers that drain the Himalayas and the Tibetan Plateau

estimate (bedload plus suspended load) might have been 50% smaller about 2 000 years ago, when human impact was minimal. Changes due to humans and/or climate affect small river basins more dramatically than larger river basins, due to the modulating ability of large rivers. The predominance of studies in larger basins may therefore skew our view on paleo-flux estimates.

2.3.2.2 *Present Flux*

Milliman and Syvitski (1992) estimated the global flux of suspended sediment to the coastal zone as 18 Gt yr^{-1} using a typological approach based on 280 rivers that emphasised the importance of fluxes from small mountainous rivers. Even now and relative to their number, very few small rivers have been monitored worldwide. Small rivers are highly impacted by rare events (e.g., landslides, floods), and lack of data on these events remains a fundamental problem in determining a global flux estimate. Using digital elevation data covering the global landmass to constrain the upscaling exercise, the annual sediment flux to the global ocean is determined to be 24 Gt yr^{-1} (Syvitski 2003c) (Fig. 2.9).

Meade (1996), however, notes that any global estimates are not the sediment flux to the coastal ocean, but flux estimates to the most seaward gauging stations in the river basins. These stations are often located well inland, and more seaward filters such as deltaic and tidal flats often influence the magnitude of the sediment load reaching the coastal ocean. Examples include:

- in the Amazon River, 20% of the annually delivered load (1 Gt yr^{-1}) is retained by its delta; the remaining 80% is deposited on the continental shelf and coast.
- in the Ganges and Brahmaputra rivers, 55% of their combined annual sediment load (1.1 Gt yr^{-1}) is retained by their delta, with 36% reaching the shelf and 9% reaching the deep sea.
- in the Yellow River, 82% of the annual load (1.1 Gt yr^{-1}) is retained by its delta; the remaining 18% is deposited on the shelf and coast.

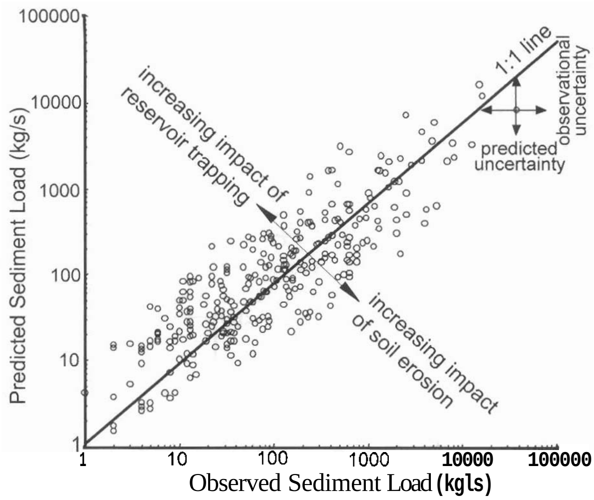


Fig. 2.10. Sediment flux. The sediment load of global rivers, based on the typological model of Syvitski et al. (2003), with

$$\overline{Q_s} = \alpha_3 A^{\alpha_4} R^{\alpha_5} e^{kT}$$

where Q_s is long-term sediment load, A is basin area, R is river basin relief, and $\bar{T}$ is basin averaged temperature. Observational errors are globally variable and large, and relate to the short duration and quality of measurement. The scatter around the 1:1 lines is largely an indication of the impact of human activities (e.g., soil erosion or impoundment by reservoirs)

Another problem facing all coastal flux estimates is that most of the observational data covers only a few years. This leads us to question the usefulness of mean numbers for sediment discharge (Fig. 2.10). Both inter- and intra-annual variations within river basins need to be considered (Fig. 2.11); inter-annual variability can exceed an order of magnitude. Another problem with observational datasets is that they are some decades old (1960s and 1970s), with most of the global observational network now abandoned. Global estimates will need continual re-examination following construction of more dams and other engineering projects. Vorosmarty et al. (1997) estimated that approximately 30% of the global sediment flux to the coastal zone is trapped behind large reservoirs, and the coastal impact of such dam construction can be great (Hart and Long 1990).

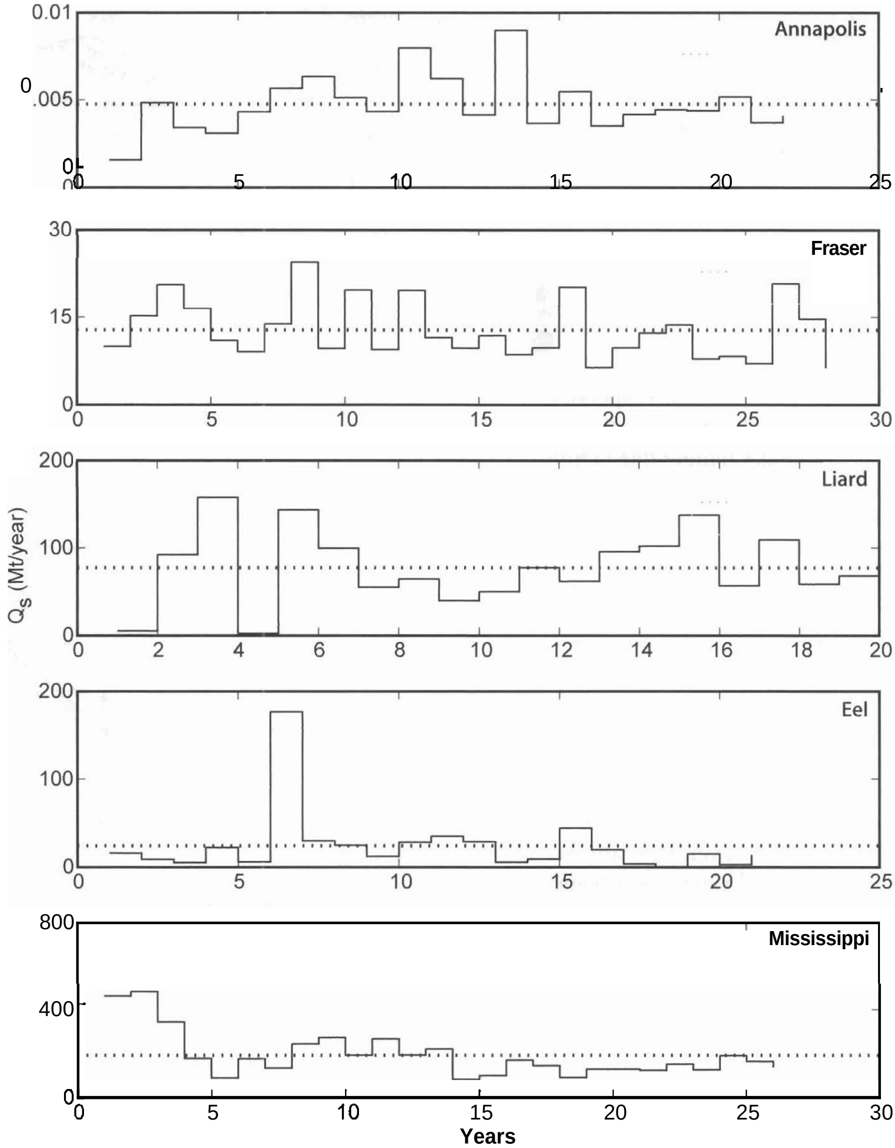


Fig. 2.11.
Sediment flux. The inter-annual variability in sediment delivery to the coast of five selected rivers (see Fig. 2.7). Annual variability can range from an order of magnitude (e.g., Eel and Liard rivers) to a factor of two (e.g., Annapolis and Fraser rivers). The decreasing annual load of the Mississippi River relates to the increasing impact of sediment impoundment

Sediment is stored in large river systems across many different time-scales with little linkage between the original erosion of uplands and subsequent sediment discharge at river mouths. Ninety percent of the sediment eroded off the land surface is stored somewhere between the uplands and the sea (Meade 1996). Twentieth century erosion across the conterminous USA (5.3 Gt yr^{-1}) (Holeman 1980) was an order of magnitude higher than the fluvial discharge of sediment (0.445 Gt yr^{-1}) (Curtis et al. 1973, also see Smith et al. 2001). At the seasonal scale, sediment is stored in riverbeds and along their banks at low or falling discharges, but re-suspends at high or rising discharges. Temporarily stored sediment may be washed out of the system before peak discharge is reached. At the decade to century time-scale, sediment moves down the drainage system in a series of sediment pulses, and may take decades to gain the lowest reaches of the flood basin (Madej and Ozaki 1996). Understanding these scales will play an important role in deciphering the difference between soil erosion and sediment discharge by rivers (Meade 1996).

2.3.2.3 *Sediment Flux and Climate Change*

Recent studies point to a strong coupling of river discharge and climate oscillations. The El Niiio-Southern Oscillation-induced climate changes recur on a multi-decadal timescale in general agreement with the Pacific/North American climate pattern (Inman and Jenkins 1999). A dry climate was observed in southern California from 1944 to about 1968 and a wet climate from about 1969 to the present. The dry period was characterised by consistently low sediment flux out of southern Califor-

nian rivers. The wet period has an annual suspended sediment flux about five times greater than the dry period, caused by strong El Niño events that produce floods with an average recurrence of about 5 years. The average sediment flux out of southern Californian rivers during the three major flood years was 27 times greater than the flux during the 1944–68 dry climate period. Similar trends were observed for the Eel River of northern California (Syvitski and Morehead 1999). In contrast, sediment loads are low for the Yellow River during El NiAo events when the Southern Oscillation Index is negative (Hu et al. 1998), i.e., an opposite response to the impact of southern California rivers.

There is a close interdependence between climate, land use, vegetation cover density and erosion rates. After human settlement effects, climate shifts are often the major driving factor on sediment flux. For example, given a sharp rise in precipitation following a decade of relatively low rainfall in East Africa, sediment yields greatly increased (Wasson 1996). One of the largest impacts of climate change is through changes in the overall water balance with subsequent impacts on land cover density and thus erosion rates. Relatively modest shifts in average climate conditions (i.e., 1 to 2 °C, < 20% precipitation) have large impacts on the behaviour of a river's flood response and thus sediment yield (Knox 1993).

Large continents are influenced by a number of climatic phenomena over different time periods. Individual regions may respond differently to climate forcing, yielding a varied response of changes in sediment flux for a given climatic event. The response will depend on the duration of the climate fluctuation and the variability in spatial properties of such parameters as relief, geology and hydrological processes.

2.3.2.4 Anthropogenic Influences

Human impact on the flux of sediment to the global ocean is well recognised (Berner and Berner 1987, Milliman et al. 1987, Hu et al. 1998, Saito et al. 2001). Land use is probably the dominant control on particulate fluxes in areas of low relief and large-scale urbanisation, in contrast with mountainous regions where natural processes are likely to still dominate (Wasson 1996). Land clearing in low relief areas increases sediment yields by more than an order of magnitude (Douglas 1993), an effect that increases with decreasing drainage area.

In the wet tropics, intense rainfall coupled with deforestation, overgrazing and other poor farming practices considerably increases soil erosion (Wolanski and Spagnol 2000). In some cases the effects are catastrophic (e.g., mud slides). These practices, together with the discharge of mine tailings to rivers, have increased the sediment loads carried by rivers many times above natural backgrounds. Except where dams capture the sediment, the future for tropical estuaries and coasts is increased muddiness and increased flooding. This in turn reduces primary productivity and impacts the tourism industry with the inherent loss of aesthetics. Even in less extreme cases, increased water turbidity still leads to environmental degradation from the smothering of coral reef organisms and seagrasses (McLaughlin et al. 2003). The mud also affects the biological properties of the water and the benthic food chains in tropical river deltas, which economic planners have generally chosen to ignore. Clearly, in such systems maintenance of a healthy marine environment requires concomitant management of the land. This situation contrasts with that in developed countries in middle latitudes, where estuaries are generally suffering from sediment starvation due to extensive damming and river flow regulation (Wolanski et al. 2003a).

The Mediterranean landscape may be the most human-impacted terrain on earth. Around 75% of the average sediment yield (1100 t $km^{-2} yr^{-1}$) of Mediterranean headwater river basins may be attributed to human activity (Dedkov and Mozzherin 1992). The terrain is naturally vulnerable to processes of erosion with its steep slopes, high relative relief, fissile sedimentary rocks, thin erodible soil covers and active tectonic settings (Woodward 1995). Severe land degradation has taken place including badland formation, representative of acute land deterioration. Rising human population exacerbates these conditions. In tropical areas of the globe, deforested land is increasing at 1×10^5 $km^2 yr^{-1}$ (Hu et al. 1998) and may account for the extraordinary sediment loads from Oceania discussed by Milliman and Syvitski (1992).

Where the natural sediment balance of estuaries is disturbed by an increase in a river's sediment load and/or change in the intensity of its annual flood wave, changes in bathymetry as a result of net siltation or erosion will ensue. The extent and speed of an estuarine response will depend on bathymetry, intensity of floods, tidal range and riverine sediment inflow. An estuary can change markedly in a few decades.

Impoundments provide important benefits to society through flood control, power generation, water storage and release for agriculture, industry and municipalities. Impoundments also provide a unique recreational resource (Vorosmarty et al. 1997). Negative environmental impacts include dislocation of human populations, siltation of reservoirs, downstream scouring of channels, interference with migration, life cycle and habitat of aquatic organisms, eutrophication and anoxia, increases in the occurrence and severity of stagnant-water diseases, and irretrievable water loss through reservoir evaporation and groundwater seepage. There is also the decrease in sediment supply to the ocean. Between 1951 and 1982 dams were being constructed at a rate of 900 per year. Prior to 1950 there were only eight dams in China; by 1982 the number had increased to 18 600, or 55% of the world total (with the US at 16% and Japan at 6%; Vörös-

marty et al. 1997). A case study for Morocco (Text Box 2.7) highlights the effects of damming on sediment supply to the coastal zone.

The Colorado River once supplied about 150×10^6 t yr^{-1} of sediment to the Gulf of California. Sediment trapping by dams has starved the Colorado delta and this has resulted in coastal recession (Walling and Fan 2003). The Ribarroja-Mequinenza Dam on the Ebro River in Spain traps 96% of the river sediment, which has led to coastal recession at the river mouth and cessation of the sea-

Text Box 2.7. Role of reservoirs and sediments: Moroccan case study

Maria Snoussi

The suspended sediment fluxes of the main Moroccan rivers flowing into the Atlantic Ocean (Fig. TB2.7.1) were estimated to be 40×10^6 t yr^{-1}, i.e., a yield of 750 t km^{-2} yr^{-1}. This rate, which is one of the highest in Africa in terms of specific sediment yield, is probably due to the drainage basins being characterised by young mountains, extensive sedimentary rocks, irregular and often stormy precipitation and scarce vegetation. The Sebou River's exceptionally high sediment yield (995 t km^{-2} yr^{-1}) reflects the erodible rocks of its drainage area and steep slopes that cause landslides and mudflows.

In these semi-arid conditions, where rainfall occurs mostly as short and heavy storms, flood events dominate the water and sediment fluxes. For example, during the 1963 floods, the maximum water discharge of the Sebou River was 8 000 $m^3 s^{-1}$ (about 60 times greater than the mean annual value) and that of the Moulouya River 5 200 $m^3 s^{-1}$ (nearly 240 times greater than the mean annual discharge). Suspended matter reached several hundred g l^{-1} during the flood events. While land erosion and sediment delivery in this semi-arid region are products of physical factors (e.g., climate, topography, soil erodability and the local vegetation status), human activities (deforestation, overgrazing and damming) are exacerbating these processes.

Damming has increased over recent decades to better manage the water shortages in Morocco caused by recurrent droughts. However, the reservoirs have experienced siltation from the high rates of natural and accelerated erosion in the hinterland. The annual sedimentation rate in the main reservoirs averages 50×10^6 $m^3 yr^{-1}$. This high siltation rate has serious environmental and socio-economic impacts, because it reduces the reservoir capacity and may be affecting the morphological equilibrium of the coastline.

The Sebou River basin is a clear example of the effects of the construction of reservoirs on river sediment fluxes. In the period 1940–1972, before dam construction, the average suspended sediment input to the Atlantic Ocean was about 34×10^6 t yr^{-1}. Since construction of the reservoirs, > 95% of the total sediment load has been trapped (Haida 2000). The calculated trap efficiency is 85–99%, and the lifespan of the reservoirs has diminished from 540 years to as little as 42 years. Similarly, the Moulouya River now delivers to the Mediterranean Sea only 7% of the sediment load transported before it was dammed (Snoussi et al. 2002); the Mohamed V reservoir will fill with sediment within 59 years, having lost 35% of its storage capacity between 1967 and 1991. It is estimated that by 2030, 70 000 ha of irrigated land and 300 MW of electricity will be lost as a consequence of the high rates of dam siltation.

Considering all the dams, sediment trapping has reduced the catchment areas as effective sources of coastal sediment by 60–70%, which may have a profound effect on the morphological evolution of the coastline. However, it is difficult to isolate the effects of damming on the shoreline evolution, since influences other than fluvial discharge (e.g., human activities at the coast) affect the coast at different rates and times.

The potential effects of the construction of the Mohamed V dam on the Moulouya shoreline morphology were examined by comparison of two sets of aerial photographs (1958 and 1988). Before construction of the dam (1958), the lower Moulouya River was sinuous to meandering and the mouth was much wider than it is now. Fluvial sediment discharge was sufficient to support progradation of deltaic deposits in the eastern part of the river mouth. Since construction of the Mohamed V dam, the river mouth and coastline have changed markedly. The influence of marine waves has increased because of weaker fluvial hydraulic power, leading to reworking of shoreline sediments, narrowing of the river mouth and accumulation of mouth bars. Net littoral transport is about 165 000 $m^3 yr^{-1}$. Wave-induced sand transport is directed westwards and has led to the accretion of the west coast, while the east coast, no longer fed by fluvial inputs, has retreated. Many other inlets, apart from those that are stabilised (Loukkos, Sebou and BouRegreg) have spits directed westward along the coastline. In some cases marine storms transport enough sand to fill the inlets periodically (Moulay Bouselham) or completely (Oued Massa). These changes in coastal geomorphology are probably largely due to the reduction of flow discharges by damming. In the Sebou River, fluvial competence has become very weak and now the estuary is not flushed as frequently.

In summary, the semi-arid fluvial fluxes to the coast depend strongly on climatic variations. At the seasonal scale, fluvial input occurs only during flood events (a few days per year); in non-dammed rivers, a large amount of suspended sediment reaches the coast. At other times, the rivers have weak flow or become dry, reinforcing the marine influence at the inlets. The construction of dams during recent decades has exacerbated sediment and water impoundment, while the land-ocean interface has become dominated by marine forcing. The projections of climate change trends over the next 20 years, according to IPCC methodology, predict for Morocco a general decrease of about 4% annual precipitation, a 15% decrease of surface water discharge and an increase in the frequency of extreme events (floods and droughts). In such conditions, the construction of more dams is likely to have serious environmental and related societal impacts on the coastal zone.

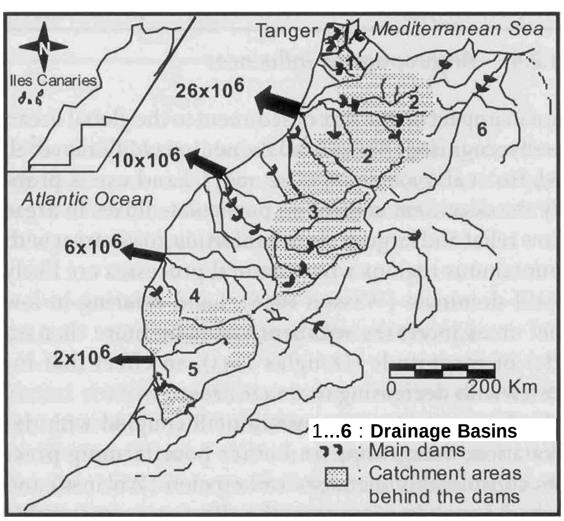


Fig. TB2.7.1. Suspended sediment discharges (t yr^{-1}) of the main Moroccan rivers into the Atlantic Ocean before the construction of dams. The ***shaded areas*** represent catchments for which sediment has been impounded following the construction of dams

ward progradation of the delta (Guillen and Palanques 1997; Jimenez and Sanchez-Arcilla 1993). The 40% decrease in suspended sediment load between 1963 and 1989 in the Mississippi River may be the major cause for the recession of the Mississippi deltaic coast (Walling and Fan 2003).

Water diversion can also decrease sediment flow and generate coastal erosion. For instance, water diversion from China's Luanhe River has decreased the riverine sediment load by 95% and has resulted in the delta's recession at a rate of 17.4 m yr^{-1} (Qian 1994).

This story is repeated in the Nile River, which carried about 135×10^6 t yr^{-1} of sediment before construction of the Aswan High Dam (Stanley and Warne 1993). For millennia, annual floods provided Egypt with much-needed water to irrigate farmers' fields. The historic annual sediment input to the Nile River delta also helped offset geologic subsidence rates that range from < 1 mm yr^{-1} to > 4 mm yr^{-1} in the north-eastern delta region (Stanley and Warne 1993). Before completion of the High Aswan Dam in 1964, the Nile River delivered an annual average of –84 km^3 of water and $\sim 124 \times 10^6$ tonnes of sediment to the coast, plus an additional 9.5 x 10^6 tonnes of suspended sediments deposited on the Nile floodplains. Subsequently, water reaching the coast has been reduced by 80% and sediment loads by over 98% (Stanley and Warne 1993, El-Sayed 1996). The only fresh sediment now reaching the coast comes via longshore transport and aeolian activity. As a consequence, erosion along parts of the shoreline has intensified and salinisation of cultivated land has increased. With 98% of this sediment now trapped in the reservoirs, coastal erosion is intense – the Rosetta and Arietta promontories are eroding at the rates of 106 m yr^{-1} and 10 m yr^{-1}.

In semi-arid regions, river floods can be suppressed as a result of large dams. The balance between the scouring of the estuary during occasional river floods and the regular import of coastal zone sediment into the estuary by tidal pumping is then disturbed. The estuary may subsequently silt up, as occurred with the Ord River estuary in north-western tropical Australia (Wolanski et al. 2001). Macro-tides drive rapid tidal pumping and the estuarine bathymetry can change rapidly. Hence, the Ord River estuary has silted measurably by 30×10^6 m^3 over the last 30 years, and the estuarine cross-sectional area has decreased by about 50%. The Ord River estuary-may have been made geomorphologically unstable by dams suppressing large river floods. Such effects are likely to be more prevalent for large dams in arid areas because they are designed to store water for several years and consequently capture most of the water from rare, large flood events; in humid areas dams are filled annually or seasonally.

When sediment from human-induced soil erosion exceeds the trapping capacity of the estuary, mud deposits in coastal waters. This process is exacerbated by reclamation of estuarine wetlands reclaimed for farming and for settlement developments (Wolanski et al. 2004). Here, the change in coastal properties may be very rapid, occurring within a few decades. Examples of this are found in bays along the Queensland coast adjacent to the Great Barrier Reef, which have become permanently muddy in only a few decades following land clearing and accelerated soil erosion (Wolanski and Duke 2002).

Reduced sediment loads to rivers through damming increases coastal erosion and deterioration of coastal marine ecosystems. Following completion of the Aswan Dam in 1964, the sardine fish catch in waters adjacent to the Nile River delta was reduced by 95% in response to reduced nutrient discharge and the delta shrank rapidly (Hu et al. 1998). Recently, with the increased use of artificial fertilisers and the expanded croplands in its delta region, nutrient loads to the Nile have increased and some fisheries stocks have improved or been replenished (Nixon 2003). After the US catchment of the Colorado River was dammed, sediment and nutrient discharge plummeted and the shrimp catch in Baja California collapsed (Hu et al. 1998). Completion of the Kotri Barrage on the Indus River in 1956 resulted in fish catches decreasing by a factor of three (Hu et al. 1998). Similar conditions occurred in the Bohai Sea when the sediment discharge of the Yellow River was reduced along with water and nutrient discharge; the shrimp fishery has decreased by 85% and the percentage of high-quality fish catch has declined by an order of magnitude (Hu et al. 1998). About 2 500 years ago the Yellow River was not muddy and its sediment discharge was one-tenth that of 30 years ago (Milliman et al. 1987, Saito et al. 2001) when it peaked in response to rapid cultivation of the Loess Plateau. A combination of soil preservation practices in the 1980s and dam construction (3380 reservoirs and another 30 000 diversion works of various scales) have reduced both water and sediment discharge. The Yellow River now runs dry for many months of the year; in contrast, the river flooded on average every three years over the preceding 4 000 years (Saito et al. 1994).

2.3.2.5 Near-future Sediment Flux

The future flux of sediment to the coastal oceans will continue to be influenced by humans and/or climate change. Determining the balance between increasing sediment loads (due to land use, engineering, climate change and climate variability) and decreasing sediment loads (due to reservoirs, engineering, climate change and climate variability) is of utmost importance for sound coastal zone and resource management. In general, the future load of rivers should be less than the present estimates, mostly because of the construction of large dams. This projection may be in error, as we do not fully comprehend the balance between sediment retention

Text Box 2.8. Satellite monitoring of water turbidity in a coastal system: northern Gulf of Mexico[1]

Joe Salisbury and Janet W. Campbell

Fluvial input and processes of wind-driven re-suspension and transport control the distribution of suspended sediment in river-dominated coastal regions. The influences of discharge and winds on surface sediment concentration can be explored by analysing the covariance between these variables. Recently these relationships were investigated in the northern Gulf of Mexico using satellite-derived wind and sediment concentration data (Salisbury et al. 2001). Temporal correlation coefficients were mapped at pixel level for time-series of wind stress and sediment concentration, and Mississippi River discharge and sediment concentrations.

Figure TB2.8.1 shows the time-series at two locations to illustrate the degree of coherence between signals where there is a high positive correlation between discharge and sediment concentration ($r > 0.7$, see Fig. TB2.8.2a), and between wind stress and sediment ($r > 0.6$, see Fig. TB2.8.3b). Correlation maps based on the time series of 8-day averages for the period 20 September 1997 to 31 December 2000 reveal the long-term patterns (Figs. TB2.8.2 and TB2.8.3). A region of high correlation ($r > 0.7$) between the Mississippi River discharge and sediment concentrations (Fig. TB 2.8.2) was located near the delta and a region of significant but lower correlations extended eastward toward the Alabama coast. This region of fluvial influence was spatially separate from the regions where wind stress and sediment correlations were significant (Fig. TB2.8.3). The wind-influenced regions are associated with shallow shelf areas, as one might expect. The boundary of the wind-influenced region off the Louisiana–Texas coast is aligned with the 100 m isobath and the waters with highest correlation ($r > 0.6$) had depths < 50 m.

Maps of the correlation between suspended sediment concentration and river discharge indicate regions that are fluvially influenced, whereas maps of the correlation between wind stress and sediment indicate regions where wind-mixing accounts for sediment re-suspension and subsequent transport. The regions of positive wind and sediment correlation are spatially disjointed from regions of positive discharge and sediment correlation. This indicates dominance by one of the two sediment mobilisation and/or transport processes (i.e., sediment delivery by rivers, wind-driven re-suspension and subsequent transport). It is probable that influences of winds and discharge on sediment distributions can be investigated independently. Using these methods, investigators interested in ecosystems dominated by one process or the other can confine the focus of their work. Further, spatio-temporal correlation methods represent an opportunity to study the transformation and fate of river- and wind-influenced constituents.

[1] Figures TB2.8.1–TB2.8.3 are reprinted from Deep Sea Research II, vol 51, No. 10–11, pp 1187–1203, Salisbury et al.: "On the seasonal correlation of surface particle fields with wind stress and Mississippi discharge in the northern Gulf of Mexico." Copyright 2004, with permission from Elsevier.

schemes and soil erosion perturbations. Time-series data are needed to determine trends, with a focus on the last 20 years. These data are often lacking and new methods need to be developed (or old ones reassessed) to effectively utilise available data for water discharge. Globally, we need to determine:

a How long until we fill the terrestrial sediment sinks (natural and artificial)?
b What effect will the resulting reservoir state have on the coastal zone and the global sediment flux?
c What are the sensitivities of local and regional coastal settings to erosion on a global scale?

When modelling the possible future sediment flux, economics needs to be considered. The effect of human development and land use is vital in understanding the global sediment flux and regional variations. The global volume of sediment entering the coastal zone may not be the important variable; the change in sediment yield on a regional basis may be of much more importance. For example, sediment-starved regions may undergo erosion, while sediment-inundated regions may experience biological and ecosystem consequences such as burial of benthic biota.

Historical land-use and sediment discharge response has come a full circle for the eastern seaboard of the USA (Pasternack et al. 2001). After European settlement (post-1740) sedimentation rates increased eight-fold through early deforestation and agriculture (1750–1820), then increased another three-fold during the period of peak deforestation and intensive agriculture (1820–1920), and finally were reduced by an order of magnitude (i.e., close to pre-colonisation values) during the period of dam-building and urbanisation (1920 to present). A similar set of histories for other regions must be completed before we can understand and predict the global flux of sediment.

The combination of climate (drying) and human water utilisation drive many hydrological and erosional systems. Two situations might be considered:

1. regions that become moister and start generating runoff, thereby initiating sediment transport to the ocean, and, more probably,
2. regions that cease water flowing to the ocean because of a combination of drying and water utilisation by growing populations.

- Important regions of runoff change include the Mediterranean basin, sub-Saharan Africa, southwestern North America, central Asia and eastern South America. These regions similarly witness the impact of humans on scarce water resources. Research must identify the population thresholds and behaviours that have strong hydrological and erosional effects. Some coastal issues include the erosion or subsidence of sediment-starved deltas and delayed responses. The coupling of increased nutrient inputs and decreased sediment loads may promote coastal-zone eutrophication and hypoxia.

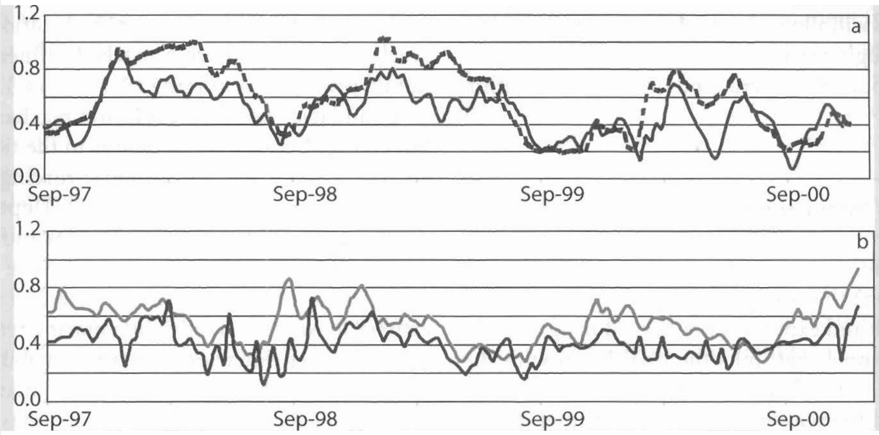


Fig. TB2.8.1. Time-series of the Mississippi River discharge (*blue*), wind stress (green) (derived from the National Center for Environmental Predictions, NCEP gridded data) and average sediment concentrations derived from SeaWiFS data (red) for two locations. **a** 4 × 4 pixel box located near the mouth of the Mississippi River (see point *A* in Fig. TB2.8.2) where there is a high correlation between discharge and sediment concentration. **b** 4 × 4 pixel box located in a region where there is a high correlation between wind stress and sediment concentration (see point B in Fig. TB2.8.3)

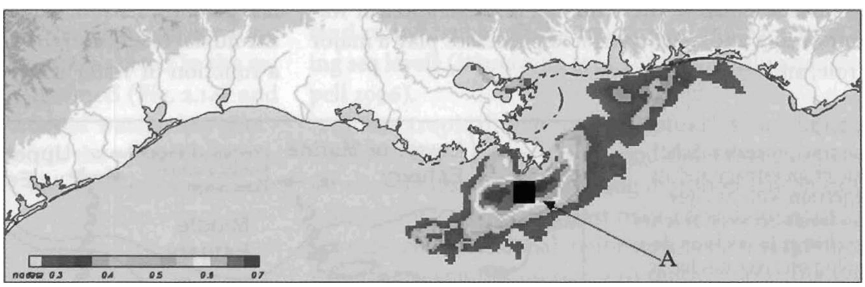


Fig. TB2.8.2. Correlation maps for the period 1997–2000 based on 8-day averages of Mississippi River discharge and SeaWiFS-derived sediment data showing region of significant correlation between the Mississippi discharge and the sediment concentration

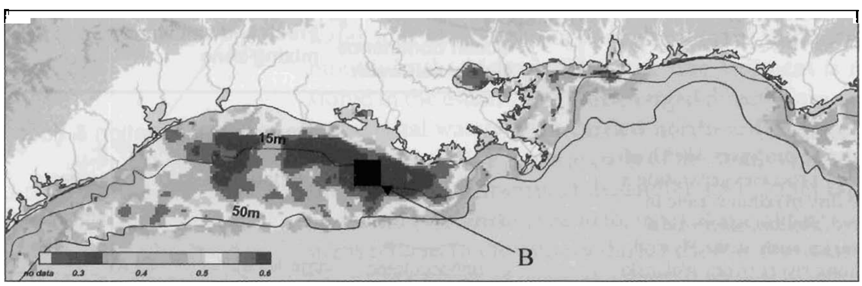


Fig. TB2.8.3. Correlation maps for the period 1997–2000 based on 8-day averages of NCEP-derived wind stress and SeaWiFS-derived sediment data showing a region of significant correlation between the wind stress and the sediment concentration

For society to understand and better manage sediment fluxes a more systematic approach is required (for example, see Text Box 2.8) using new technologies and improving general information. Clearly, there is a need to:

- Assemble existing maps and databases for coastal zone morphology and sediment situations at the global scale. Observations made from this activity could be linked to up-river processes and information about rates of change of documented human impact.
- Establish global maps delineating sediment sources and/or sensitivity to disturbance. This would allow for a better understanding of the effect of change on the system.
- Create an index to encapsulate sediment transit times within basins. This index must be scale-independent as transit times in small river basins are expected to be much shorter than those for larger river basins. This infers that changes occur much more rapidly in smaller basins than in larger ones.
- Determine how long before river loads fill up the terrestrial sediment traps, and what the subsequent impacts will be downstream in the coastal zone. Effort is needed to establish the links between land and ocean.

- Delineate the balance between increasing and decreasing sediment loads due to humans and/or climate change, and the actual and potential impacts.
- Link coastal sediment budgets to terrestrial sediment budgets. This will allow a bridge between the data from upstream gauging stations and the coastal ocean, taking into account the interaction and filtering within estuaries.

2.4 Estuarine Interactions

A tidal inlet system consists of a tidal basin (often in the form of a lagoon) that stores seawater during the flood phase of the tide and exports water mass during the ebb, an entrance channel that links the tidal basin with the open sea, and flood and ebb tidal deltas that are located in the basin and the seaward side of the channel (Hayes 1980, see Text Box 2.9). An estuary typically has three major reaches (Dionne 1963; Fig. 2.12):

a a marine or lower estuary, where processes are dominated by oceanic influence,
b a middle estuary, where mixing of fresh and salt water occurs and density-related processes play a major role, and
c an upper or fluvial estuary, where there is no salt intrusion but tidal effects impact circulation inducing changes in water level and reversing currents.

The "effective tidal limit" is the location in the fluvial reach where tides have a marked influence on the dynamic processes occurring (Perillo 1995).

Boundaries between estuarine reaches are not fixed but change because of variations in the tidal range (e.g., spring-neap cycles, atmospheric forcing) and in river discharge (e.g., high/low runoff). Depending on the dominant factor, the boundaries may move either landward or seaward. Not all estuaries have three reaches. Some estuaries may have only one (e.g., Bahia Blanca estuary in Argentina only has a marine reach as the middle and upper estuaries are within a tributary; the Amazon River in Brazil is mostly a fluvial estuary) or two (e.g., the Negro River estuary in Argentina has only the middle and upper estuaries). Of course the presence/absence of any reach varies with time, most particularly in relation to season.

Most studies of river dynamics and sediment delivery to the coastal zone provide information down to the last gauging station, which is usually located well above the fluvial reach, and the water discharge is calculated as a function of water level. However, there is a relatively

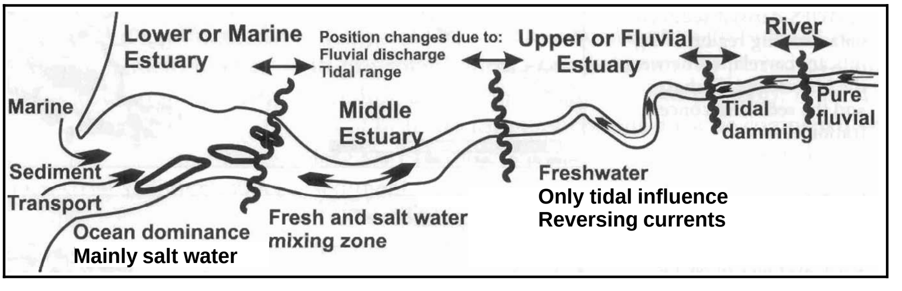


Fig. 2.12. Estuarine processes. Schematic of an estuary and its integration with its river. Boundaries between reaches may change in position depending on river discharge and tidal range (modified from Perillo 1995)

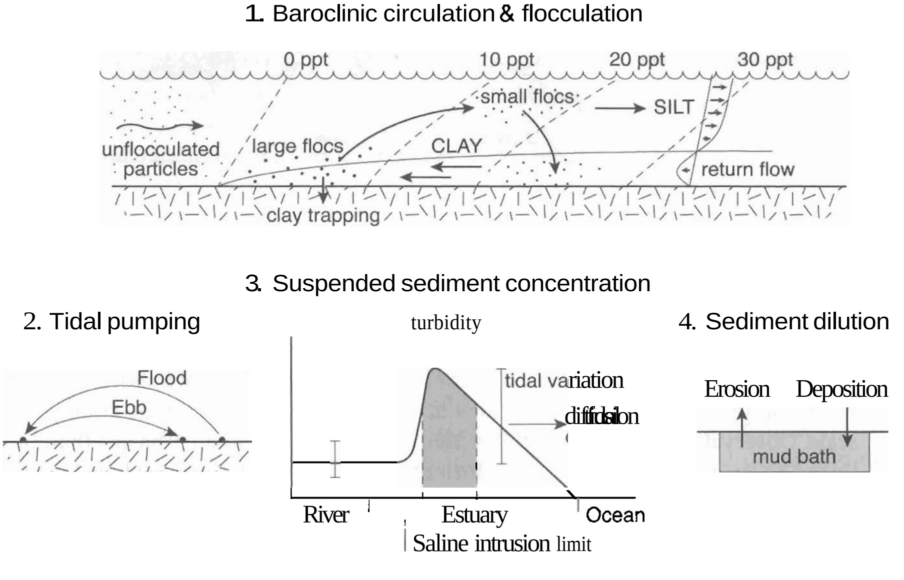


Fig. 2.13. Estuarine processes. Sketch of the key processes generating a turbidity maximum zone in turbid, shallow, macro-tidal estuaries, such as the Fly and Mekong rivers (from Wolanski et al. 1995)

small reach between the upper estuary and the pure fluvial conditions in which changes in water level do occur, but there are no reversing currents: river discharge actually stops for a short period of time. Perillo (2000) defined this reach as *the zone of tidal damming*.

Although the interaction between freshwater and saltwater plays a major role in estuarine dynamics and sediment transport, the geomorphology (shape) of the estuary provides an important control on the interaction between tides and freshwater. Regardless of runoff volume and the tidal range at the mouth, what makes each estuary truly unique is its own particular geomorphology. As the shape of the estuary changes in response to the influences of sediment supply and deposition, current and wave erosion, so will the tidal propagation change with concomitant feedback on the geomorphology.

The Fly River estuary is a prime example of an estuary without river floods that can overwhelm the tidal system and yield freshwater to the mouth of the estuary. In the absence of such floods, the estuary will filter, trap and process the suspended, fine sediment particles. The particles move with the water currents but also have internal motions (Fig. 2.13) that include flocculation, sorting between clay and silt, erosion and settling at tidal frequency, and interaction with the plankton forming muddy marine snow. The particles are sorted in the estuary, clay being preferentially retained (Fig. 2.14) and silt preferentially exported to coastal waters; key processes are the internal circulation in the estuary and the asymmetry between flood and ebb tidal currents. Both

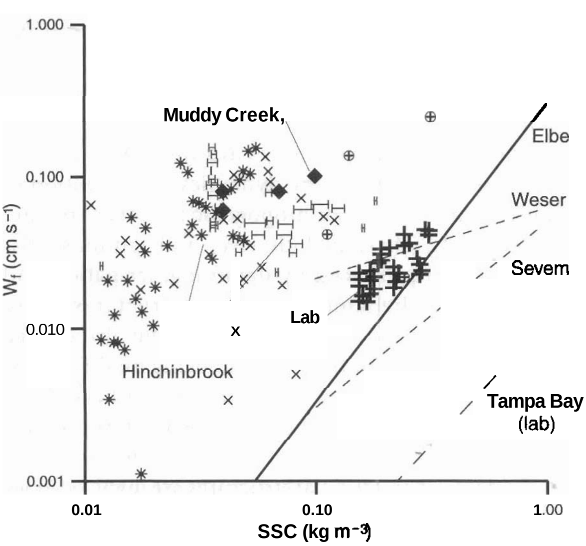


Fig. 2.14. Estuarine processes. Relationship between the suspended sediment concentration (SSC) and the settling velocity (w_f) of fine suspended sediment. Settling velocity values are much higher in tropical (e.g., Hinchinbrook River and Muddy Creek) than in temperate waters (Elbe, Severn and Weser rivers). Laboratory results underestimate the settling velocity because they neglect the formation of muddy marine snow, a dominant process in the field (from Wolanski et al. 2001, adapted from Dronkers and van Leussen 1988)

processes lead to the formation of a turbidity maximum (Wolanski et al. 1995).

Some estuaries are filling with sediment that is imported from coastal waters. The Fly River estuary may receive about ten times more sediment from coastal waters than from the river. The imported sediment may be derived from other large rivers nearby which enter the estuary by littoral drift, a process very much dependent on the local, coastal oceanographic conditions and the wind in coastal waters. Such infilling is commonly observed in small estuaries near large rivers. The small Jiaojiang River estuary facing the South China Sea is rapidly infilling (at about 0.13 m yr^{-1}) with fine sediment discharged into coastal waters by the Yangtze River 200 km away (Guan et al. 1998). This infilling requires continuous dredging to maintain navigability. Estuaries along the tropical coast of South America north of the Amazon River mouth are infilling with fine Amazon sediment. In the South Alligator River in tropical northern Australia, much of the infilling material may be pre-Holocene sediment originally deposited along the coast and drowned as sea level rose and the coast retreated at the end of the last ice age. Here, sediment is imported into the estuary ten times faster from the sea than from land runoff, enabling the estuary to keep pace with rising sea levels (Woodroffe et al. 1986, Wolanski and Chappell 1996).

Some tropical, macro-tidal estuaries can be overwhelmed during river flood conditions that can persist for several months where a long monsoon season exists. Alternatively, floods may last only a few days in arid areas subject to occasional major storms such as typhoons. Estuaries respond differently to these contrasting flood regimes. When floods persist for several months, freshwater may extend throughout the estuary to the river mouth. In the Mekong River, the fine sediment is not stored in the estuary, but is discharged directly into shallow, coastal waters, then carried northward and southward along the Vietnamese coast of the South China Sea, the direction of transport depending on oceanography and wind (Wolanski et al. 1996, 1997). Some of that sediment returns to the estuary during the low-flow season. Over time-scales of years, the natural system may be at quasi-equilibrium in the sense that the size, shape and depth of the system may change only very slowly, evolving at time-scales of decades. This is important because it gives humans the time to adapt to changing estuaries and coasts.

In some rivers there is practically no estuary. This can occur when the river discharge is high and the river is sufficiently shallow so that the water is fresh at the mouth. All the riverine sediment is then exported offshore. Examples include the Sepik River in Papua New Guinea, where there is no continental shelf and sediment settles to abyssal depths (Chappell 1993); the Amazon River in

small reach between the upper estuary and the pure fluvial conditions in which changes in water level do occur, but there are no reversing currents: river discharge actually stops for a short period of time. Perillo (2000) defined this reach as *the* zone of tidal **damming.**

Although the interaction between freshwater and saltwater plays a major role in estuarine dynamics and sediment transport, the geomorphology (shape) of the estuary provides an important control on the interaction between tides and freshwater. Regardless of runoff volume and the tidal range at the mouth, what makes each estuary truly unique is its own particular geomorphology. As the shape of the estuary changes in response to the influences of sediment supply and deposition, current and wave erosion, so will the tidal propagation change with concomitant feedback on the geomorphology.

The Fly River estuary is a prime example of an estuary without river floods that can overwhelm the tidal system and yield freshwater to the mouth of the estuary. In the absence of such floods, the estuary will filter, trap and process the suspended, fine sediment particles. The particles move with the water currents but also have internal motions (Fig. 2.13) that include flocculation, sorting between clay and silt, erosion and settling at tidal frequency, and interaction with the plankton forming muddy marine snow. The particles are sorted in the estuary, clay being preferentially retained (Fig. 2.14) and silt preferentially exported to coastal waters; key processes are the internal circulation in the estuary and the asymmetry between flood and ebb tidal currents. Both

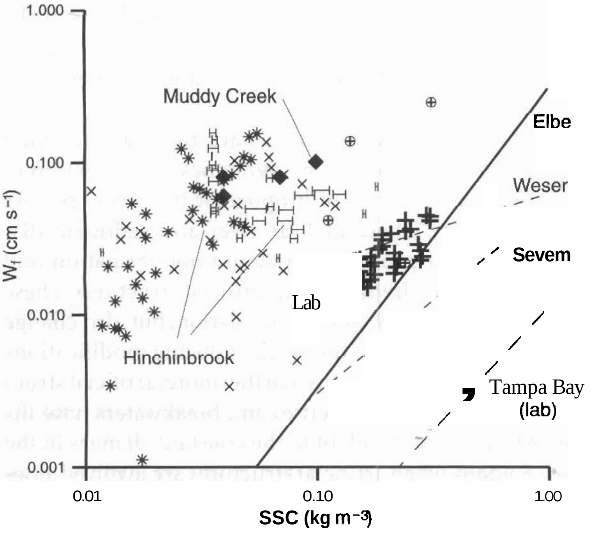


Fig. 2.14. Estuarine processes. Relationship between the suspended sediment concentration (SSC) and the settling velocity (w_f) of fine suspended sediment. Settling velocity values are much higher in tropical (e.g., Hinchinbrook River and Muddy Creek) than in temperate waters (Elbe, Severn and Weser rivers). Laboratory results underestimate the settling velocity because they neglect the formation of muddy marine snow, a dominant process in the field (from Wolanski et al. 2001, adapted from Dronkers and van Leussen 1988)

processes lead to the formation of a turbidity maximum (Wolanski et al. 1995).

Some estuaries are filling with sediment that is imported from coastal waters. The Fly River estuary may receive about ten times more sediment from coastal waters than from the river. The imported sediment may be derived from other large rivers nearby which enter the estuary by littoral drift, a process very much dependent on the local, coastal oceanographic conditions and the wind in coastal waters. Such infilling is commonly observed in small estuaries near large rivers. The small Jiaojiang River estuary facing the South China Sea is rapidly infilling (at about 0.13 m yr^{-1}) with fine sediment discharged into coastal waters by the Yangtze River 200 km away (Guan et al. 1998). This infilling requires continuous dredging to maintain navigability. Estuaries along the tropical coast of South America north of the Amazon River mouth are infilling with fine Amazon sediment. In the South Alligator River in tropical northern Australia, much of the infilling material may be pre-Holocene sediment originally deposited along the coast and drowned as sea level rose and the coast retreated at the end of the last ice age. Here, sediment is imported into the estuary ten times faster from the sea than from land runoff, enabling the estuary to keep pace with rising sea levels (Woodroffe et al. 1986, Wolanski and Chappell 1996).

Some tropical, macro-tidal estuaries can be overwhelmed during river flood conditions that can persist for several months where a long monsoon season exists. Alternatively, floods may last only a few days in arid areas subject to occasional major storms such as typhoons. Estuaries respond differently to these contrasting flood regimes. When floods persist for several months, freshwater may extend throughout the estuary to the river mouth. In the Mekong River, the fine sediment is not stored in the estuary, but is discharged directly into shallow, coastal waters, then carried northward and southward along the Vietnamese coast of the South China Sea, the direction of transport depending on oceanography and wind (Wolanski et al. 1996, 1997). Some of that sediment returns to the estuary during the low-flow season. Over time-scales of years, the natural system may be at quasi-equilibrium in the sense that the size, shape and depth of the system may change only very slowly, evolving at time-scales of decades. This is important because it gives humans the time to adapt to changing estuaries and coasts.

In some rivers there is practically no estuary. This can occur when the river discharge is high and the river is sufficiently shallow so that the water is fresh at the mouth. All the riverine sediment is then exported offshore. Examples include the Sepik River in Papua New Guinea, where there is no continental shelf and sediment settles to abyssal depths (Chappell 1993); the Amazon River in

Brazil, where the freshwater discharge and shallow water maintain freshwater to the river mouth, so that sediment is deposited in coastal waters from where it is transported as sand and mud waves along the coast (Nittrouer and DeMaster 1986); and the La Sa Fua River in Guam, where sediment settles over and smothers coral reefs, accumulating for several months (occasionally for one or even two years), before being stirred and flushed out by typhoon-generated waves (Wolanski et al. 2003b).

A sediment equilibrium over time-scales of years may also exist in macro-tidal, semi-arid tropical estuaries. In the Fitzroy River in Western Australia, runoff is very brief and major river floods lasting only a few days are a result of the passage of tropical cyclones (typhoons, hurricanes) (Wolanski and Spagnol 2003). During these few days, the estuary is scoured and fine sediment is exported offshore where it is deposited in coastal waters, to return to the estuary over the rest of the (dry) year as a result of tidal pumping at a rate of 3 000 to 30 000 t d^{-1} (Wolanski and Spagnol 2003). This results in an accumulation of about 0.2–2.0 cm yr^{-1} in the estuary. Estuaries of this type are thus in balance over a time-scale of years, between scouring during rare floods and infilling the rest of year with riverine sediment deposited offshore and entrained back into the estuary.

In the coastal zone, sediment-trapping mechanisms may act individually and collectively, reflecting the interplay of the major dynamic processes (tides, waves, river and groundwater discharge) with the coastal geomorphology (see Text Box 2.9). The shape of the coast determines how tides propagate along the estuary, and the relationship between convergence and friction was used by Le Floch (1961) to classify estuaries. A funnel-shape versus a constant or landward-increasing cross-section (although quite rare) determines whether tidal energy is concentrated or diffused along its pathway. Boundary and bottom roughness elements (e.g., tributaries, intertidal zones, bedforms, sinuosity) dissipate tidal energy. The degree of tidal wave asymmetry and the resulting asymmetry in tidal currents is strongly related to the geomorphological influence on tidal propagation.

Sediment is mostly delivered to the coastal zone by rivers, although the effects of coastal erosion, longshore transport and shore-normal wave and tidal action can be locally important. A large portion of the sedimentary material reaching the coast is retained in estuaries and deltas (see Sect. 2.3.1). The development of an estuary rather than a delta is a function of the relative energy of the river discharge and the dissipative energy of the marine forces (tides and waves) acting on the discharge. This balance seldom reaches equilibrium as both mechanisms are continuously changing across temporal and spatial scales.

A major question presently unanswered for most estuarine systems is their sediment retention. Perillo (2000) proposed a Retention Index (RI) as *the ratio between the sediment permanently retained into the coastal zone to the total sediment input provided by fluvial, marine, atmospheric and even the coastal zone proper.* This retention index considers the long-term residence period, ranging from several years to infinity, and is directly related to the long-term geomorphologic evolution of the coastal zone. Different reaches of an estuary may have different retention indices.

Most estuarine process research is focused on the formation of the turbidity maximum that forms at the landward end of the salt wedge in response to the strong density gradient (Fig. 2.13). Few investigations have addressed the influence of estuarine geomorphology (e.g., presence or absence of tidal flats/marshes/mangroves, tidal channel morphology) and the propagation of the tidal wave along the estuary (e.g., asymmetry of the wave and currents, changes in tidal range) or the interactions between the turbidity maximum and sediment retention in estuaries. Seldom are changes in sediment transport and retention measured along the whole estuary and into the tidal-influenced reaches of the river, nor is their time evolution assessed. Interactions between geomorphology and advection processes are highly non-linear, making difficult the prediction of the rate of sediment transport and the fate of sediment input into coastal areas.

Flow regulation structures may trap sediment and the estuary may be starved of sediment. The location of the turbidity maximum may then migrate seasonally back and forth in the estuary in response to the seasonally varying freshwater discharge. In many cases, however, the estuary responds by not just a movement in the location of the turbidity maximum, but by changing bathymetry and the location of the coast.

Although natural processes are the major driving mechanisms controlling the dynamics and retention of sediments in estuaries, anthropogenic influences require detailed study. These include river and sediment discharge control produced by damming, irrigation and water pumping, dredging and artificial structures. These procedures not only control circulation, but also change the geomorphology of the reach, inducing modifications in all trapping mechanisms. Furthermore, artificial structures such as harbours, jetties and breakwaters have little or no capacity to adapt to the constant changes in the flow. Seldom, when artificial structures are involved, does the system reach any equilibrium.

2.5 Groundwater Inputs to the Coastal Zone

Although not as obvious as river discharge, continental groundwater also discharges directly into the ocean wherever a coastal aquifer is connected to the sea. Artesian

Text Box 2.9. Physical changes of tidal inlet systems

Shu Gao

Coastal embayments form an important part of the coastal zone, with tidal inlets present in many places along the world coastlines. These systems were formed during the Holocene when sea level rose to inundate low-lying coastal areas. Subsequently, these systems have been affected by wave- and tidally-induced sediment transport and river discharges, and many tidal basins have disappeared due to sediment infilling. For example, 4400 years BP the Bergen Inlet in the Netherlands was one of the largest inlet systems of the region (Beets et al. 1996). Sediment infilling within the tidal basin ensured that the entire basin was filled by sediment within 1200 years. Thus, tidal inlets are short-lived features on a geological time-scale.

Infilling of a tidal basin does not always take place at a constant rate. Negative feedbacks, for example changes in time-velocity asymmetry patterns, play an important role in extending the lifetime of a tidal inlet. The tidal currents within the entrance channel are influenced by, among other factors, flood and ebb durations. Generally, a longer flood duration is associated with stronger ebb currents, and a shorter flood duration is related to weaker ebb currents. The former situation is referred to as "ebb dominance" and the latter as "flood dominance". What factors control these difference patterns? Speer and Aubrey (1985) identified two parameters responsible for the difference: the tidal range (R) to water depth (H) ratio for the entrance channel, and the intertidal zone area (A,) to the total area (A) ratio for the tidal basin. A large RIH ratio favours flood dominance, while a large A_0/A ratio favours ebb dominance. From this it may be inferred that at an early stage of tidal inlet evolution, flood dominance is most probable because the R/H ratio tends to be large and the A_0/A ratio small, resulting in intense sediment transport into the tidal basin. Infilling of sediment increases the intertidal zone area and scouring of the channel increases the channel depth. Thus, the RIH ratio is reduced and the A_0/A ratio is enhanced. Eventually, ebb dominance will replace flood dominance, with a consequent reduction in the rate of sediment infilling. This negative feedback mechanism explains why most present-day inlet systems are ebb-dominated.

More complicated inlet behaviours with regard to time-velocity asymmetry patterns have been observed (Jia et al. 2003). Yuehu is a small inlet system with a tidal basin 5 km^2 in area, lying in the eastern part of Shandong Peninsula, China. The inlet system was largely un-impacted before the late 1970s when, for aquaculture purposes, the entrance to the lagoon was artificially closed and some of the intertidal areas were reclaimed. As a result, the inlet system now experiences a smaller flood duration. However, the tidal currents show complex patterns (Fig. TB2.9.1), which may be attributed to regional tidal characteristics and local entrance channel geometry. For example, the shape of the entrance channel is such that the cross-sectional area changes rapidly at certain water levels, altering the tidal currents. Because changes in tidal-inlet geomorphology will cause the coastal zone to change in terms of material cycling and ecosystem evolution, it is important to understand the processes and behaviours of tidal inlet systems. In particular, the negative feedbacks described above may be used to prolong the lifespan of these coastal systems.

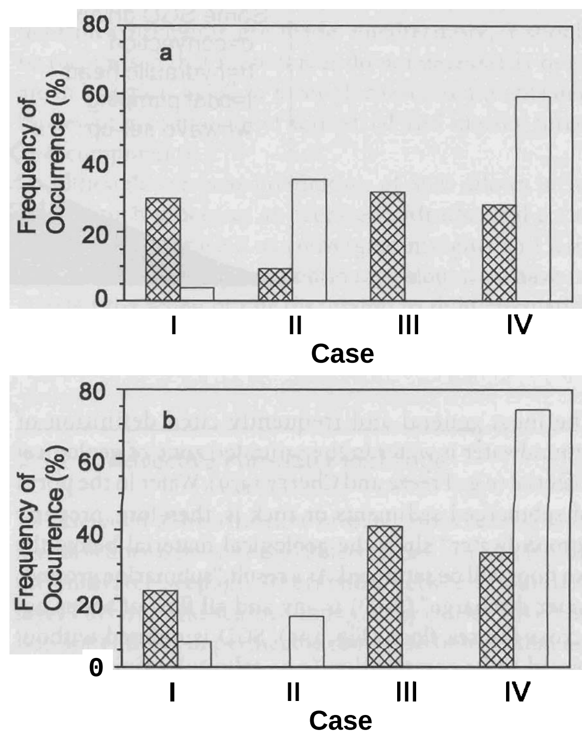


Fig. TB2.9.1. Frequency of occurrence for different time-velocity asymmetry patterns associated with the Yuehu inlet system, eastern China: (a) winter 1998 and **(b)** summer 1999. Case I – flood duration is shorter than ebb, and flood current velocity is larger than ebb. Case II – flood duration is longer than ebb, and flood current velocity is smaller than ebb. Case III – flood duration is longer than ebb, and flood current velocity is larger than ebb. Case IV – flood duration is shorter than ebb, and flood current velocity is smaller than ebb. The ***shaded bars*** denote cross-sectional mean currents, the ***blank bars*** vertically averaged tidal currents at the entrance center

aquifers can extend for considerable distances from shore underneath the continental shelf, with discharge to the ocean at their points of outcrop. In some cases, these deeper aquifers may have fractures or other breaches in the overlying confining layers, allowing groundwater to flow into the sea. Although submarine springs and seeps have been known for many years (written accounts exist from at least the Roman period), these features have traditionally been perceived as hydrological "curiosities" rather than objects for serious scientific investigation (Kohout 1966).

Within the last few decades, recognition has emerged that, at least in some cases, groundwater discharge into the sea may be both volumetrically and chemically important (Johannes 1980). It is now widely recognised that there are several oceanic processes that drive advective flow or re-circulated seawater through permeable sediments in addition to fresh groundwater flow driven by hydraulic gradients on land. The terrestrial-driven and ocean-derived flows grade into each other, especially near the coast. It is important to have a nomenclature that is compatible to both types of flow.

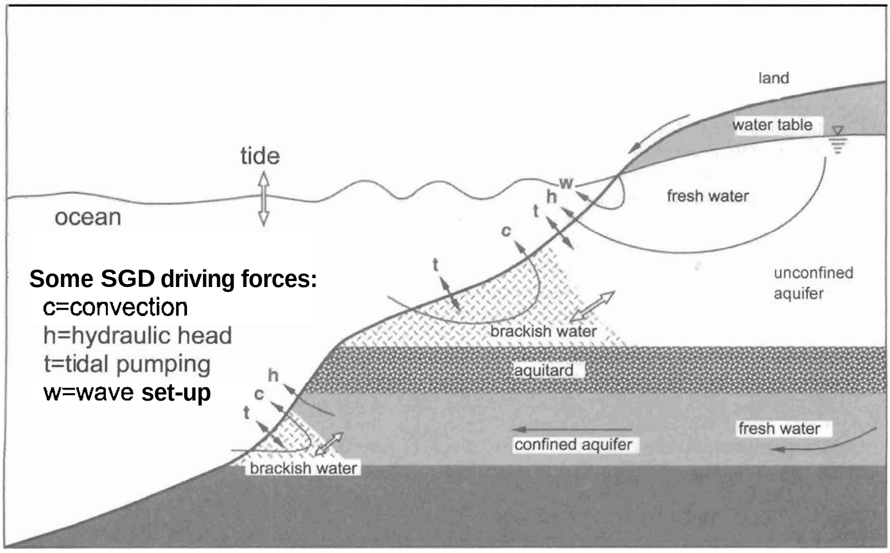


Fig. 2.15.
Submarine groundwater flux. Nomenclature of fluid exchange and schematic depiction (no scale) of processes associated with submarine groundwater discharge. Arrows indicate fluid movement (from Burnett et al. 2003, modified from Thibodeaux and Boyle 1987)

2.5.1 A New Understanding

The most general and frequently cited definition of groundwater is water in the saturated zone of geological material (e.g., Freeze and Cherry 1979). Water in the pores of submerged sediments or rock is, therefore, properly "groundwater" since the geological material below the sea floor will be saturated. As a result, "submarine groundwater discharge" (SGD) is any and all flow of water out across the sea floor (Fig. 2.15). SGD is defined without regard to its composition (e.g., salinity), origin or phenomena driving the flow. Where the sediments are saturated, as they are expected to be in all submerged materials, "groundwater" is synonymous with "porewater."

Traditional hydrology has been concerned with terrestrial groundwater. As a result, groundwater has been defined as rainwater that has infiltrated and percolated to the water table, or similar definitions consistent with the applications to freshwater, terrestrial systems (e.g., Considine 1995). Such qualifications on the definition of groundwater lead to conceptual problems when dealing with submarine discharges. SGD does not have to be terrestrially derived, although it can be and is in many prominent situations. When SGD is measured, there is seldom a way to evaluate its source. While it may be legitimate to require water classified as "groundwater" to move according to Darcy's Law, even that may be too restrictive in some highly channelised (e.g., karst) situations. At least one definition of groundwater explicitly excludes underground streams (Wyatt 1986) while another specifically includes them (Bates and Jackson 1984; Jackson 1997). Since karst is such an important setting for SGD, we include such features.

In the marine environment, submarine groundwater recharge (SGR) also occurs as tides, waves, currents, sea-level fluctuations and density differences force seawater into the sea floor. This water eventually must leave the aquifer. Recycling of ocean water has been referred to, in various articles, as "irrigation" or "ventilation" or "transportation:' terms usually applied to the surficial layers within a meter or so of the sea floor (e.g., Conley and Inman 1994). In some cases, the recharged waters are discharged locally. In other cases, waters can emerge far from the source, even sub-aerially as in saline springs in Hawaii (Cooper et al. 1964).

SGD can vary widely over time and space. Stirring or agitating of porewater by short-period water-waves, without necessarily producing any net flow has been referred to as "wave pumping" or "wave stirring" (e.g., Harrison et al. 1983, Riedl et al. 1972). "Floating" (Thorstenson and MacKenzie 1974) or "salt fingering" (Gorsink and Baker 1990) describe the situation when, if the density of the ocean water increases above that of the porewater for any reason, porewater can float out of the sediment by gravitational convection in an exchange with denser seawater, again without a net discharge. SGD has also been referred to as "flushing." Flushing generally involves a continuous replacement of porewater driven by the hydraulic gradients ashore or pressure gradients in the coastal ocean. Gradients may be due to wave set-up at the shore (Li et al. 1999), tidal pumping at the shore (Nielsen 1990), er differences in tidal elevations across narrow reefs or barrier islands (Bokuniewicz and Pavlik 1990, Reich et al. 2002).

The system of terminology developed by the SCOR Working Group 112 (Burnett et al. 2003) is illustrated in Fig. 2.16. The flow of water across the sea floor can be divided into SGD, a discharging flow out across the sea floor, or SGR, a recharging flow in across the sea floor. The two terms do not have to balance because SGD can, and often will, include a component of terrestrially re-

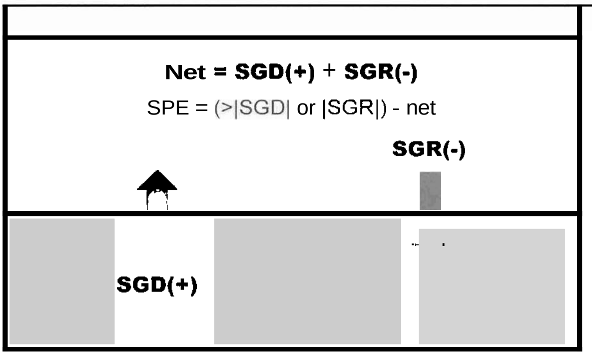


Fig. 2.16. Submarine groundwater flux. Submarine groundwater discharge (SGD) comprises any fluid flow upward (+) across the sea floor while submarine groundwater recharge (SGR) is fluid flow into (–) the seabed. The "net" flow is the mathematical sum of these two components that do not necessarily balance. Submarine porewater exchange (SPE), analogous to the mixing parameter in the LOICZ biogeochemical budget terminology, is the larger absolute value of SGD or SGR minus the net flow. This assumes that the domain of the SPE remains constant

charged water. Alternatively, some or all of the SGR can penetrate the sub-aerial aquifer, raising the water table or discharging as terrestrial surface waters (e.g., saline springs) rather than discharging out across the sea floor. The net discharge is the difference between these two components. The "submarine porewater exchange" (SPE) is the difference between the larger of the two (SGD or SGR) volume fluxes and the net. In other words, SPE is the smaller absolute value of either SGD or SGR. SPE is equivalent in magnitude to V_x (a mixing term) in the LOICZ biogeochemical budget convention (Gordon et al. 1996, see Chap. 3). By that convention, V_x is positive for flows into the sediment.

SGD may consist of multiple components. One is meteoric water that had fallen on dry land as atmospheric precipitation, infiltrated the soil on rock, and percolated to the water table. It can be driven across the sea floor by the onshore hydraulic gradients although it is possible that this exchange takes place as gravitational convection, i.e., buoyant fresh porewater "floating" across the sea floor into the open, salt water. Another important component of SGD may be re-circulated seawater that can also be driven in part by hydraulic gradients on land as well as by various oceanic forces. In some cases, SGD can also contain saline connate groundwater or groundwater whose salinity has been raised by dissolution of salt within the aquifer itself. While it is tempting to subdivide SGD in the terminology according to its principal components, i.e., fresh water and saline water, we have elected not to do this, as it could lead to somewhat arbitrary decisions when mixtures of fresh and salt waters are encountered.

SGD could be classified based on the driving forces. Terrestrial hydraulic gradients and the consequent motion of the meteoric groundwater can drive the seepage of infiltrated seawater. SGD could be driven by any of a number of oceanic processes, such as wave pumping, wave set-up or set-down and tidal pumping. A third class of endogenic drivers such as thermal gradients, osmotic pressures, inverted density stratification or consolidation must also be considered. Although such divisions are useful in our efforts to understand the mechanisms involved, they cannot serve as a primary basis for classification and terminology because we rarely have sufficient information to define the driving forces when field measurements are made. Furthermore, it should be recognised that these forces do not necessarily operate in isolation, i.e., flow through coastal sediments may represent a composite of terrestrial and marine forces and components.

Although our broad definition of SGD allows inclusion of such processes as deep-sea hydrothermal circulation, fluid expulsion at convergent margins and density-driven cold seeps on continental slopes, we have restricted the scope of our discussion to fluid circulation through continental shelf sediments with emphasis on the coastal zone.

2.5.2 Advective Porewater Exchange

As discussed above, the flow of water from coastal sediments is not exclusively tied to terrestrially driven (fresh) groundwater seepage. Interaction between boundary layer currents and sea-bed topography causes advective porewater flows in permeable coastal sediments that are not as conspicuous as submarine springs but can be an important mechanism controlling the geochemical characteristics of the water column in the coastal zone (Shum and Sundby 1996, Boudreau et al. 2001).

The lower permeability limit for porewater transport that significantly exceeds diffusive transport is approximately 10^{-12} m^2 (Huettel et al. 1996). In general, the mean grain size and the permeabilities of the sediment surface layers increase from the continental rise towards the coast (Emery and Uchupi 1972). The decrease in water depth amplifies the effect of bottom currents; at water depths < 100 m, the wave orbital motion reaches the sea bed and tidal current speeds increase, producing strong bottom shear and sediment erosion (Nittrouer and Wright 1994, Jing et al. 1996). In this zone, frequent resuspension of terrigenous shelf deposits and removal of the fine material by cross-shelf currents result in well-sorted sand beds that are characterised by high permeabilities that permit measurable porewater flows.

In the upper part of the continental slope, most of the 10^{-12} m^2 permeability isolines turn parallel to the 200 to 500 m isobaths (Riedl et al. 1972). The contribution of the porous component to the surface sediments of the shelf of the eastern USA is about 92% (gravel 11%, shell 14%, sand 67%; Hayes 1967, Stoddart 1969, Riedl et al. 1972).

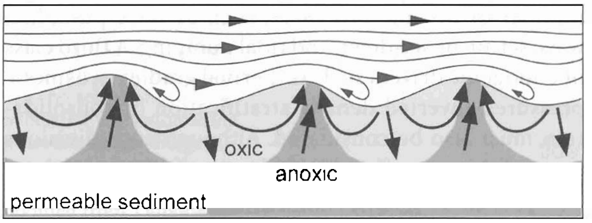


Fig. 2.17. Submarine groundwater flux. Schematic view of porewater flow directions under ripples exposed to unidirectional flow. Water intrudes the sediment at the upstream face of the ripple and moves on a curved path to the downstream slope where it is released from the sediment (from *http://www.scor-wg114.de/*)

The area of higher permeability sand sediments on an average shelf may be about 65%, in agreement with the estimates of Emery (1968).

Where bottom currents are sufficiently strong to resuspend and winnow sediment, the seabed surface is structured by ripples (Wiberg and Harris 1994) that cause the advective porewater exchange. The deflection of the unidirectional or oscillating boundary-layer currents at sediment surface structures (e.g., ripples, biogenic topography), produces local pressure gradients that drive porewater flows and interfacial fluid exchange (Webb and Theodor 1968, Thibodeaux and Boyle 1987).

In surface depressions (ripple troughs) water penetrates into the sediment and flows on a curved path towards protruding surface structures (the ripple crests), where the porewater is released (Huettel and Gust 1992; Fig. 2.17). For reasons of mass balance, the same volume of water that is forced into the sediment also flows from the bed. The resulting porewater circulation carries organic matter and oxygen to the sediment, creates horizontal concentration gradients that can be as strong as the vertical gradients, and increases the flux of porewater constituents across the sediment-water interface. Due to the continuously changing sediment topography and boundary-layer flow characteristics, advective porewater circulation and ensuing biogeochemical zonation are highly variable in space and time (Huettel et al. 1996).

The depth of the sediment layer that is affected by this advective exchange is related to the size and spacing of the ripples and, in homogeneous sediment, reaches down to approximately two times the ripple wavelength (Hutchinson and Webster 1998). Under calm hydrodynamic conditions (friction velocity, $u^* < 5$ cm s-'), the porewater velocities in typical shelf sands (200 ym, permeability, $k = 10^{-11}$ m^2) may reach vertical velocities of 2 to 3 cm h^{-1} in the uppermost centimeter of a rippled bed (5 cm amplitude, 30 cm wavelength; Huettel et al. 1996). With these upwelling velocities, the sediment exposed to unidirectional flow (e.g., during receding tide) releases approximately 75 to 100 L m^{-2} d^{-1}. In flume experiments, surface gravity waves caused porewater release rates up to 222 L m^{-2} d^{-1} in similar sediments, demonstrating that oscillating boundary flows can effectively enhance the fluid exchange between the seabed and the water column.

Riedl et al. (1972) measured interfacial porewater exchange caused by surface gravity waves using thermistor flow sensors embedded in permeable shelf sands. They calculated that worldwide the waves filter a volume of 97×10^3 km^3 yr^{-1} through the permeable shelf sediments. The intertidal pump, driven by swash and tidal water level changes, moves another 1.2×10^3 km^3 yr^{-1} through the sandy beaches of the world. These estimates suggest that wave action filters the total ocean volume through permeable sediments within 14 000 years.

2.5.3 Magnitude of Submarine Groundwater Discharge

There are many factors that affect rates of fresh groundwater flow into the coastal zone, either directly or indirectly. Driving force and transmissivity are the main factors that determine the flux of terrestrially-derived SGD. The driving force is a function of the hydraulic gradient (influenced by topography) and the terrestrial groundwater recharge rate (affected by precipitation and evapotranspiration). The types and extent of vegetation as well as climate will determine evapo-transpiration rates. Transmissivity may be controlled by permeability (geology) and development of river systems (geomorphology). Thus, parameters related to geology, precipitation, vegetation (land use) and topography are all contributing factors in determining rates of fresh groundwater flow to the sea.

Without the benefit of measurements, one may predict that land-derived SGD fluxes would be high in areas of high permeability (karst), high relief near the coast, areas without well-developed river systems (some large oceanic islands), and regions with high groundwater recharge rates (humid tropics). To evaluate the importance of groundwater pathways to the coastal zone, direct assessments are required by modelling, direct measurements, geochemical tracers or other approaches (Burnett et al. 2001, Burnett et al. 2002, Burnett et al. 2003).

SGD estimates have been made for many independent studies performed on the east coast of the United States, in Europe, Japan and Oceania (Taniguchi et al. 2002). Some studies have been done on the west coast of the US and in Hawaii. A summary of estimated and measured fluid discharges to the ocean and across the seabed shows that while unit fluxes are small, total discharge values can be huge (Table 2.3). Many SGD measurements have been made in karst areas where the hydraulic conductivity of the aquifers is large, and thus significant amounts of fresh groundwater discharge are expected under reasonable hydraulic gradients.

Table 2.3. Submarine groundwater flux. Fluxes of terrestrially-derived water and seawater to the coastal zone and ocean expressed as volumetric discharge per unit time, flux per unit length of shoreline, and unit area fluxes

Flux	Fluid composition/origin[a]	Reference
Global discharge ($km^3\ yr^{-1}$)		
Rivers = 37 400	Fresh/terrestrial	Berner and Berner (1987)
Fresh groundwater seepage = 2 200[b]	Fresh/terrestrial	Zekster (2000)
"Intertidal pump" = 1 170	Composite/mixed	Riedl et al. (1972)
"Subtidal pump" = 95 700	Seawater/marine	Riedl et al. (1972)
Shoreline fluxes ($m^3\ m^{-1}\ day^{-1}$)		
Rivers = 170	Fresh/terrestrial	Calculated using shoreline length[c] = 600 000 km
Groundwater seepage = 10	Fresh/terrestrial	Calculated from Zekster (2000)
Measured (Florida) = 3–35	Composite/mixed	Cable et al. (1997), Burnett et al. (2002)
Measured (Perth) = 2–8	Composite/mixed	Burnett and Turner (2001)
Unit fluxes ($m^3\ m^{-2}\ yr^{-1}$)		
Seepage meters = 5–100	Composite/mixed	Calculated from Taniguchi et al. (2002)[d]
Subtidal pump = 3.5	Seawater/marine	Riedl et al. (1972)

[a] Composition refers to fresh (or meteoric) water, seawater or a mixture; "origin" refers to driving force, either terrestrial hydraulic gradients or marine forcing (tidal pumping, wave set-up) or a mixture of terrestrial and marine.

[b] Estimates for the freshwater component of groundwater discharge to the ocean vary tremendously, from values of 100 $km^3\ yr^{-1}$ (COSOD II 1987) to ~4 000 $km^3\ yr^{-1}$ (Garrels and MacKenzie 1971). The value cited here represents ~6% of the river flow.

[c] The length of a shoreline depends upon the resolution of the measurement interval so there are no definitive values (see Text Box 1.1, Chap. 1). Estimates between about 500 000 and 600 000 km appear reasonable for global shoreline lengths. Zekster (2000) stated that "... a shoreline length of 600 000 km excludes Antarctica, Greenland, the Arctic and some permafrost regions."

[d] Approximately 85% of the measured seepage values from the coastal zone in this compilation fall in this range (1–30 cm day^{-1}).

2.5.4 Biogeochemical Implications

The advective flux of terrestrially-driven groundwater through coastal sediments is becoming recognised as an important mechanism for transferring material from the land to the ocean (Cathles 1990, Valiela et al. 1990, Moore 1996, Jickells 1998). Flow may occur through the surficial aquifer or through breaches in deeper semi-confined coastal aquifers (Moore 1999). This process may affect the biogeochemistry of estuaries and the coastal ocean through the addition of nutrients, metals and carbon (Moore 1996).

Most geochemical studies of sediments have concentrated on muddy sites where diffusion and biological mixing drive exchange with the overlying ocean. The techniques and models used in those studies are not applicable to sites where advection through permeable sediments is the primary exchange agent. Results based only on muddy areas may seriously underestimate the fluxes of biogeochemically important materials in the coastal ocean.

During the passage of terrestrially-derived fluids through the sediments, mixing of seawater with fresh groundwater and chemical reactions of the fluids with solid phases may occur. The emerging fluid is chemically distinct from the groundwater and seawater end-members. Concentrations of nutrients, trace metals, organic carbon, methane and CO_2 may be considerably higher than surface ocean waters (Simmons 1992, Bugna et al. 1996, Paerl 1997, Cai and Wang 1998). Major ions may be affected by diagenesis of solid phases (Burt 1993). Because SGD may bypass the estuary filter, this source term may affect the coastal ocean quite differently from river discharge.

In these coastal aquifers or "subterranean estuaries" (Moore 1999), chemical reactions between the mixed waters and aquifer solids modify the fluid composition much as riverine particles and suspended sediments modify the composition of surface estuarine waters. The importance of chemical reactions between aquifer solids and a mixture of seawater and fresh groundwater is well-recognised by geochemists (Runnels 1969, Back et al. 1979). For example, mixing of seawater supersaturated with calcite and fresh groundwater saturated with calcite can result in solutions that are either supersaturated or undersaturated (Plummer 1975). This mechanism explains the massive dissolution of limestone along the northern Yucatan Peninsula (Back et al. 1979). Dissolution of submarine limestone by groundwater flow creates distinctive canyons and escarpments on continental margins (Paull et al. 1990).

Calcite dissolution may also be driven by addition of CO_2 to fluids in the subterranean estuary. Salt water penetrating the Floridian aquifer near Savannah, Georgia, is enriched in inorganic carbon and calcium, as well as

ammonium and phosphate, relative to seawater and fresh groundwater end members, due to oxidation of organic carbon within the aquifer or CO_2 infiltration from shallower aquifers (Burt 1993). Shallow coastal groundwaters are highly supersaturated with respect to CO_2 (Cai and Wang 1998).

Groundwater may be an important source of nutrients for coral reefs (Marsh 1977, D'Elia et al. 1981, Crossland 1982, Umezawa et al. 2002) or other communities on the continental shelf (Johannes 1980, Simmons 1992, Jahnke et al. 2000). The fluxes of nitrogen and phosphorus to the Georgia and South Carolina shelf from SGD have been estimated to exceed fluxes from local rivers (Simmons 1992; Krest et al. 2000).

Groundwater-borne nutrients can have significant effects on water quality in surface estuaries (Reay et al. 1992). Groundwater may have nutrient concentrations several orders of magnitude greater than surface waters either via contamination (e.g., from septic systems) or natural processes. Thus, nutrient concentrations in coastal groundwater, which are modified by man-made changes to coastal regions, may be a significant factor in the eutrophication of nearshore waters (Valiela et al. 1990).

The ecological significance of the advective porewater flow beyond the zone of influence of SGD derived from land is fundamentally different from that of the terrestrially-derived groundwater inputs, as there is no import of allochthonous substances into the coastal zone. Nevertheless, advective porewater transport may be important for the coastal cycles of matter through the acceleration of the deposition and mineralisation process (Marinelli et al. 1998). Due to advective porewater exchange, coastal permeable sands function like expansive biocatalytic filter systems and may in part be responsible for the tight cycling of matter in shelf waters, reducing the export to the deep ocean (Huettel et al. 1998).

Flume studies have shown that with interfacial water flows, suspended particles and phytoplankton are filtered from the water column, thereby increasing the deposition rate (Huettel and Rusch 2000). Within the sediment, the organic particles are exposed to higher mechanical stress, higher bacterial abundances and higher exo-enzyme concentrations, which accelerates their decomposition. Directed advective transport of oxygen and other electron acceptors into the sediment (Lohse et al. 1996, Ziebis et al. 1996) and simultaneous transport of decomposition products (e.g., HCO_3^-, inorganic nutrients) out of the sands (Gehlen et al. 1995, Huettel et al. 1998) further enhance the sedimentary degradation and convert the permeable sands into efficient bioreactors (Huettel and Rusch 2000).

The effect of advective porewater flow from the biocatalytic sand filter is documented by high benthic primary production on coarse organic-poor sands reaching 800 mg C m^{-2} d^{-1}. About 30% of the continental shelf sea floor (3.4×10^8 km^2) receives sufficient light to support significant rates of benthic primary production that would result in an estimated production of 2.9×10^{14} g C yr^{-1} (~0.3 Gt C yr^{-1}; Nelson et al. 1999, Jahnke et al. 2000). However, this indirect evidence for advective nutrient release may be strongly affected by the impact of nutrient-rich terrestrially-derived groundwater through the sandy seabed. Investigations using isotopes and other novel techniques are needed to assess the contribution of the two processes to benthic primary production. How these advective processes affect the biogeochemistry of estuaries and the continental shelf is only beginning to be appreciated.

Much work remains before SGD can be evaluated relative to more conventional processes. For example, nutrients may enter the coastal ocean through rivers, the atmosphere and upwelling at the shelf break, as well as in SGD. The biological effects of these inputs depend not only on the magnitude of the input but how and where the nutrients are delivered. A relatively small input to an isolated estuary may have an effect much different than a more substantial input spread over a large fraction of the shelf. Differing amounts of delivered nitrogen, phosphorus and silica may also create distinct responses. To achieve a more complete understanding of the role of advective processes in sediments, studies over a range of scales and environments are required.

2.6 Influence of Human Activities on Material Fluxes

The stability of the coastal zone is affected by the influence of humans on upland water resources through marked changes in timing, flux and dispersal of water, sediments and nutrients. Although there are numerous large construction projects that exemplify the impact of humans at the local scale, one example is highlighted here: the building of an artificial island off Hong Kong (see Text Box 2.10). Below we limit our discussion to more global issues.

2.6.1 The Role of Dams and Other Land Transformations

Massive anthropogenic transformations of the Earth's surface during the 20th century have begun to impact continental-scale patterns of river runoff, sedimentation and coastal erosion, ultimately affecting global sea-level rise (Gornitz 2001). Sequestration of water in reservoirs and artificial lakes diminishes the outflow of water to the sea. Conversely, groundwater mining, deforestation and urbanisation increase the volume of water delivered to the oceans. Groundwater mining or overdraft – the withdrawal of groundwater in excess of natural recharge

Text Box 2.10. Construction of the new Hong Kong International Airport

Wyss W.-S. Yim

Anthropogenic activity in the coastal zone is accelerating land-ocean interaction at an unprecedented rate through land reclamation from the sea for major infrastructures. One example is the recently completed Hong Kong International Airport with a total land area of 12.48 km^2 located just south of the Tropic of Cancer on the inner continental shelf of the northern South China Sea (Fig. TB2.10.1). Table TB2.10.1 shows some statistics for the airport.

The site of the new Hong Kong International Airport lies immediately south of the Pearl River delta front within the Pearl River Estuary (Fig. TB2.10.1). An island platform was created for the airport by levelling two islands (Chek Lap Kok and Lam Chau) with a total area of 3.1 km^2, and reclaiming a further 9.38 km^2 from the sea. For investigating the ground conditions, about 2 000 boreholes and over 5 000 cone penetration tests were made, providing a vast geo-technical database (Anon 1996). Offshore site preparation involved dredging the seabed to completely remove the soft marine clay of Holocene age and to partially remove the comparatively stiff marine clay of Last Interglacial age, to depths ranging approximately from 4 to 30 m below mean sea level. A complete and thick sequence of unconsolidated Quaternary deposits occurs to the west of Lam Chau, and include:

1. Marine unit 1 (M1) of Holocene or oxygen-isotope (OI) stage 1 age with abundant shells and shell fragments (Yim and Li 1983) radiocarbon dated at no older than 8 200 years BP (Yim 1999) post-dating the meltwater pulse of the Laurentide icesheet.
2. Terrestrial unit 1 (T1) of last glacial to pre-last interglacial or OI stage 2–4 age.
3. Marine unit 2 (M2) of last interglacial or OI stage 5 age with rare shell fragments because of post-depositional groundwater dissolution (Yim and Li 1983). Oyster shells found elsewhere in Hong Kong from the same unit have yielded a uranium-series age of about 130 000 years BP (Yim et al. 1990).
4. Terrestrial unit 2 (T2) of second last glacial or OI stage 6 age.

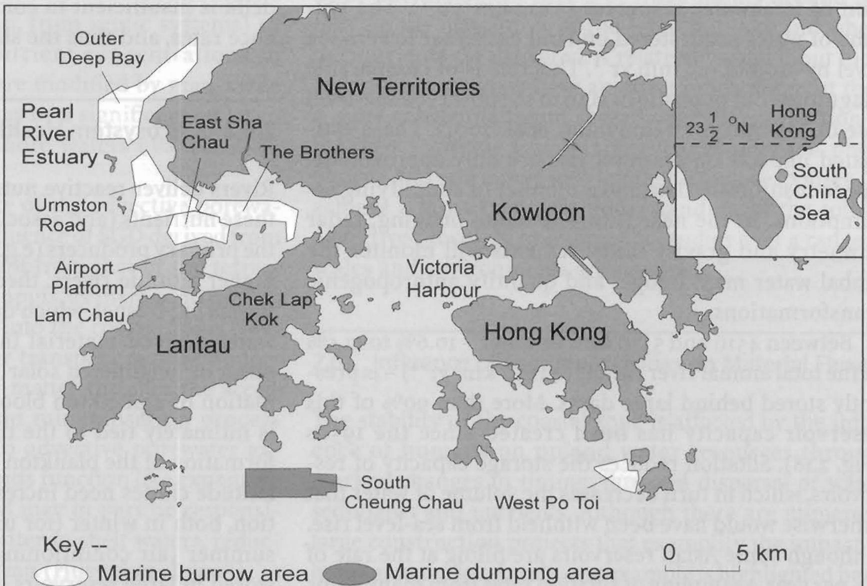


Fig. TB2.10.1. Location map of the Hong Kong International Airport, marine borrow areas and mud disposal areas

Table TB2.10.1. Some statistics of the new Hong Kong International Airport

Date of opening of airport	July 6 1998
Construction time of airport island	42 months
Total cost of site preparation	US$1.16 billion
Total area of airport	12.48 km^2
Total area reclaimed from the sea	9.38 km^2
Dimensions of two runways	Length 3.8 km, width 60 m, 1.525 km apart
Volume of marine clay dredged	68.8 million m^3
World's dredging fleet	75%
Volume of overburden removed from marine burrow areas to exploit sand	40 million m^3
Volume of sand brought in from marine burrow areas for reclamation	76 million m^3
Total volume of fill	197 million m^3
Minimum dredged level	3 m below Principal Datum
Maximum dredged level	29 m below Principal Datum
Total length of seawall	13 km

5. Marine unit 3 (M3) of second last interglacial or OI stage 7 age.
6. Terrestrial unit 3 (T3) of third last glacial or OI stage 8 age, thermo-luminescence dated at about 250000 yearsBP (Yim et al. 2002).
7. Marine unit 4 (M4) of third last interglacial or OI stage 9 age. Oyster shells found elsewhere in Hong Kong from the same unit have yielded a uranium-series age of 308 000 years BP (Yim and Choy 2000).
8. Terrestrial unit 4 (T4) of fourth last glacial or OI stage 10 age.
9. Residual soil pre-dating fourth last glacial or OI stage 10 age.

In other Hong Kong waters, a maximum of 5 marine units and 5 terrestrial units has been observed, in agreement with the 5 interglacial-glacial cycles of the Vostok ice core in Antarctica (Petit et al. 1999).

Between the islands of Chek Lap Kok and Lam Chau (Fig. TB2.10.1), a paleosol referred to as a paleo-desiccated crust by Tovey and Yim (2002) is well developed on top of the M2 unit. This crust was formed by acid-sulfate soil development through the oxidation of pyrite present. Because the crust was indurated by iron-oxide cementation, its removal by dredging was found to be difficult. West of Lam Chau, older palaeosols with crusts showing greater induration than the top of the M2 unit are found on top of the M3 and M4 units reflecting the ageing effect under terrestrial conditions. Sea-level and paleo-environmental changes determined the engineering properties of the "M" and "T" units (Yim and Choy 2000). Within the reclamation, the greatest ground settlement was found in areas where the cumulative thickness of "M" units left in place is the greatest. Figure TB2.10.2 shows a three-dimensional model of the post-dredging ground surface within the area of the airport platform.

For land reclamation, sand was imported from six marine borrow areas within Hong Kong waters, while muds dredged both from the airport site and the marine borrow areas were disposed in an offshore dumping area south of Cheung Chau (Fig. TB2.10.1). The bulk of the sand mined from the marine borrow areas for reclamation pre-date the OI stage 5 and were formed during low sea-level stands under terrestrial conditions (Yim 2001). Figure TB2.10.3 shows a trailing suction hopper dredger similar to the type used for the reclamation of the airport and Fig. TB2.10.4 shows the partially completed airport platform in 1995.

Construction of the Hong Kong International Airport has had considerable impact on the coastal environment (ERL Asia Limited 1982). The rate of change during the 42-month construction period of the airport platform exceeded that of any natural processes. Of the impacts, the most important ones included:

- sediment removal from the seabed of the inner continental shelf and their redistribution,
- destruction of the natural shoreline and its replacement with a seawall,
- loss of coastal habitats,
- changed coastal hydrology through the creation of the airport platform, and
- increased loading on the earth's crust through the placement of construction materials on the platform.

If the present-day sea level is maintained over the next 1000 years, the Pearl River delta is expected to continue to migrate southwards and should merge with Lantau Island. However, the design life of the airport is 100–150 years. Additional impacts not considered here are the road and rail links between the airport and the urban areas of Hong Kong.

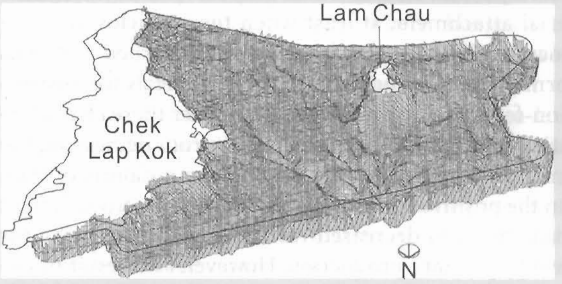


Fig. TB2.10.2. A three-dimensional model of the ground surface after dredging within the area of the airport platform (from Plant et al. 1998)

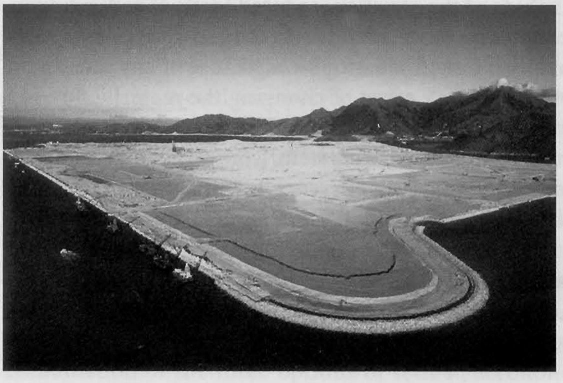


Fig. TB2.10.3. View of the Hong Kong International Airport construction site during 1995 (© K. Bartlett, Pacific Century Publishers Limited)

Fig. TB2.10.4. A trailing suction hopper dredger from the Netherlands similar to the type used for the reclamation of the new Hong Kong International Airport in action in Hong Kong. Sand is placed hydraulically using a technique known as rainbowing (© W. W.-S. Yim)

as trace metals and nutrients with a resultant change in the concentrations of the dissolved and particulate fractions, particularly in regions of low salinity; (*b*) they provide a particle surface which may serve as a site for bacterial attachment, at least when the particles occur in flocs; and (*c*) the particles and their absorbed materials form sources of nutrients and trace metals for suspension-feeding organisms. Taken together these characteristics infer effects that might result from decreasing the amount of sediment discharged from a dammed river. On the positive side of the equation, decreased sediment loads point to decreased turbidity levels necessary for healthy primary production. However, as a result of decreased suspended sediment delivered offshore of the Mississippi River in addition to increased nutrient loads, the water clarity has decreased due to the increase in chlorophyll biomass (Rabalais et al. 2002). On the negative side of the equation, reduced loads of sediment, nutrients and trace metals discharged from a river will greatly influence the trophic structure of any ecosystem. One secondary effect of changing the material flux of a river would be to change the diversity and structure of the benthic coastal ecosystem, from one of bacterial preference and sediment tolerance, to one with plankton preference and sediment intolerance (Farrow et al. 1983).

2.6.3 The Vitality of Coastal Wetlands, Mangroves and Reefs

Many coastal ecosystems depend on the seasonal variability of water levels, water salinity and water clarity. A reduced seasonal flood-wave limits the flooded area of a delta and the health of the wetland community. Examples abound, from breeding habitats of birds and coastal mammals to the algal and seagrass communities of coastal areas. River diversion is the most extreme example of an anthropogenic influence on the coastal zone. In Canada, diversion schemes include the wholesale diversion of rivers, with massive effects on the coastal zone (D'Anglejan 1994). In the residual rivers, lower water levels are predicted to increase the susceptibility of river-banks cut into sensitive clays, leading to mass wasting and increased estuarine sedimentation. Higher rates of sedimentation in the estuary, coupled with increased intertidal vegetation, further alter the estuarine circulation with augmented salt intrusion and increased particle retention. Brackish swamps will invade with time.

Estuaries marked with increased and regulated flows are expected to rapidly prograde their offshore delta, reduce any winter ice cover and modify their shoreline profile over time. The increased size of the plume would disperse fine-grained sediment over a larger area and probably change the offshore benthic ecosystem. Together, these conditions are expected to lead to changes in offshore circulation patterns.

Increased siltation in some areas can stimulate the growth of mangroves just as decreased coastal siltation can decrease the growth of mangroves (Wang et al. 1998). Increased siltation in the coastal zone could hinder the growth of offshore coral reefs.

In tropical systems, the major polluter is considered to be river-borne mud. Most tropical estuaries are now extremely shallow and turbid due to a combination of intense tropical rainfall and the lack of effective restrictions on human activities in the river catchment that lead to increased erosion. Dams and water diversion schemes in tropical catchments (e.g., Mekong) are predicted to result in:

- increased salinity intrusion up the estuary;
- increased wave-induced coastal erosion reflecting a decrease (expected) in coastal sedimentation (during the high flow season), that would otherwise provide protection for the coast;
- reduced annual flood wave, which would increase the likelihood of estuarine siltation; and
- decreased wet-season flushing of acid sulfates formed in the dry season in the top layer of the soils in the upper delta (acid sulfate being a significant hindrance to farming in the dry season).

2.6.4 Sediment Dispersion and Grain Size Effects

Hinterland changes in a river's hydrography will have profound changes on the hydrodynamics that control the energy level within a delta system and the flux of a material load across a delta system. This truism leads to new understandings of how anthropogenic changes on a river system fundamentally change the residence time of sediment within a delta. Changes will lead to differences in the residence time proportioned to different sediment grain sizes. As the sediment load within a river system is artificially reduced, concomitant with reduction of the flood wave and river-bank control (for example, the Ebro River, Spain; Sanchez-Arcilla et al. 1998), wave energy begins to reshape the delta deposits. This reshaping increases coastal erosion in certain areas while longshore transport increases the deposition rate in other areas. As a consequence, clayey material has a decreased residence time on the delta flood plain, whereas sand material has an increased residence time along the exposed coast.

The dispersion of river-borne particulate load away from the river mouth depends on coastal circulation (i.e., upwelling, downwelling, longshore currents), and on the character of the emanating plume (i.e., surface or subsurface). Road construction and newly introduced river-valley farming along Taiwanese rivers, for example, have led to highly elevated sediment loads and an increased likelihood of subsurface (hyperpycnal) flows (Syvitski 2003b). These hyperpycnal plumes flow along the seafloor

and travel directly down the continental slope, often within submarine canyons. Such currents will reduce the flux of sediment to the shelf environment and reduce the availability of freshwater and nutrients to the coastal zone ecosystems. In contrast, reservoir control will reduce the flood wave and the concomitant delivery of sediment to the coastal zone. In affected coastal environments the dispersal of sedimentary material is greatly reduced (Syvitski et al. 1998).

2.7 Summary

2.7.1 Impacts of Local, Regional and Global Sea-level Fluctuations

Scientific estimates of sea-level rise associated with global change have often dominated the political debate on climate change impacts in the coastal zone. Over the last decade there has been an increase in the accuracy of global sea-level measurement from techniques such as satellite altimetry and geodetic levelling together with the refined global viscoelastic analysis of glacio-isostatic adjustment. However, there is still a need for detailed research at local and regional scales into sea-level change and more accurate prediction of the impact of these changes.

Placing present and predicted sea-level changes within a geological context is important, so as to provide a perspective on the cyclical nature of sea-level and the extent to which current and predicted sea-level changes are perturbations beyond natural cycles, particularly the 100 000-year glacial/interglacial cycles. The geological record also shows that there has been a globally variable but predictable coastal response to the sea-level rise that followed the last glacial maximum. In addition, recent geological research shows prospects for linking recent geological records (the last 2 000 years) with sea surface temperature measurements and regional climate change.

The historical record of sea-level change interpreted from tide-gauge data demonstrates that the average rate of sea-level rise was less in the 19th century than in the 20th century. However, the raw tide-gauge data need to be corrected for local and regional influences either with modelling calculations or directly by geological investigations near tide-gauge sites. Resultant mean sea-level trends for the 20th century show a mean sea-level rise in the range of 1–2 mm yr^{-1} with a central value of 1.5 mm yr^{-1}. There remains some debate about these results, which in part relates to the different correction methods.

An important element of sea-level change research is the prediction of impacts on coastal dynamics. This is underpinned by intensive studies, such as LOIS, the UK program that detailed the evolving coastal dynamics following the postglacial sea-level rise for a section of the UK coast. Examination of the impacts of a rising sea level on different types of coast (sandy coasts, deltaic coasts, tropical coasts, low-latitude coasts) demonstrated that many are already eroding and that new threats are appearing in areas such as the Arctic, where changes in the extent of sea ice have in places produced a different wave climate with consequent coastal impacts.

The most recent scientific projections of a sea-level rise are within the range of 0.09 to 0.88 m for the period 1990 to 2100 with a central value of 0.48 m and, if current rates of terrestrial storage continue, the projections could vary by as much as −0.21 to +0.11 m. Even the achievement of the central value by 2100 would require a rate of sea-level rise between 2.2 and 4.4 times the rate for the 20th century.

In response to these predictions, various coastal vulnerability assessments have been undertaken. Initially a common global methodology was developed by the IPCC but this proved either too difficult or inappropriate for a number of countries. Regional vulnerability assessments have been conducted for specific areas such as the Pacific, where there are many low-lying developing countries perceived to be at risk. Alternative approaches to the IPCC common methodology include attempts to refine and upscale regional vulnerability data (as in the SURVAS project) or to develop specific coastal vulnerability indices (as used in the United States of America).

Although the management response to sea-level rise varies around the world, generally, it comprises some combination of the three options of retreat, accommodate or protect. A distinction needs to be made between natural coastal vulnerability and the vulnerability of human lives and property that may be put at risk by the effects of climate change and sea-level rise. In many cases, poor coastal planning has resulted in the need for rapid and expensive adjustment to the consequences of sea-level change. However, there is a clear need for better understanding of local sea-level change in addition to adopting a precautionary approach in planning to meet the consequences of predicted global sea-level rise.

2.7.2 Sediment Flux to the Coast

River systems evolve across time, influenced by paleo-conditions within the watershed, shorter period fluctuations in climate and, most recently, perturbations by humans. The sediment load delivered by world rivers is a determining factor for the long-term stability of our coastal zones. Ocean energy (tides, waves, currents) reworks the river-supplied coastal sediment to form and maintain our varied coastlines: estuaries, beaches and deltas. If the sediment supply from the land is reduced, then ocean energy will begin to relentlessly attack and erode shores.

Apart from humans, climate shifts are often the major driving factor on sediment flux. Large continents are

Text Box 2.11. Mar del Plata: A cautionary tale

Gerardo M. E. Perillo

Mar del Plata is the most important tourist city of Argentina. It has about 500 000 permanent inhabitants but, during the summer period (December-February), the population may increase 6-fold. Founded in 1874, the extended bay beaches between points of Devonian orthoquartzitic rocks were the main features that attracted visitors. At both sides of the city, Pleistocene loess cliffs (up to 25 m high) are a significant source of sediment for littoral transport.

The coast at Mar del Plata has a general trend SW-NE (Fig. TB2.11.1) and the coastline receives waves from the south and southeast for about 80% of the time, resulting in net northward littoral transport of about 390 000 $m^3 yr^{-1}$ (Caviglia et al. 1991). The coast is microtidal, having a semidiurnal tide with mean spring and neap ranges of 0.91 and 0.61 m, and wave height is between 0.5 to 5.5 m with periods between 6 and 16 s (Sunrise Technical Consultants 1968).

Mar del Plata was also a fishing village. As the population grew and agricultural development on the adjacent Pampas increased, the Mar del Plata harbour was constructed between 1914 and 1919 by partly closing a small bay with two large jetties. Construction proceeded in the absence of oceanographic evaluation.

Shortly after harbour construction, considerable erosion occurred on most beaches north of the harbor and a dissipative beach started to build to the south (Punta Mogotes Beach). Sand retention on the beaches became a priority for local and provincial authorities, with construction of a series of wooden groynes in the 1930s, replaced by cement in the 1950s. As more sand was retained by the urban beaches, erosion extended to other beaches, requiring more groyne construction.

As beach erosion grew more extensive north of the city, the associated Pleistocene loess cliffs were exposed to wave attack and started to retreat. By the 1960s, cliff erosion endangered houses, highway infrastructure and beach tourism, and in 1972–74 a 5 m high wall was constructed along 800 m of the coast in addition to new groynes.

By this time, marked erosion processes had reached the next town (Santa Clara del Mar) located 20 km to the north; between both towns beaches had disappeared and the cliffs were fully exposed and actively retreating. Nevertheless, groynes were still being built along Santa Clara del Mar, which by 1977 had lost practically all its beaches.

Between 1978 and 1984, a completely new plan for building groynes and breakwaters was developed, and groynes (Fig. TB2.11.2) were built from Mar del Plata to the mouth of Mar Chiquita coastal lagoon (about 40 km to the north); the coast of this area was retreating at an average rate of 5 $m\,yr^{-1}$ (Schanck et al. 1990). Cliffs were then protected by placing riprap stone along their base (Fig. TB2.11.2c).

Some of the semi-enclosed groynes and breakwaters built were very closely spaced, so that small artificial bays developed (Fig. TB2.11.2a) with a very restricted circulation. As sand filled these bays, the beach profile became very steep (> 30°) with water depths of 3–4 m attained in a few meters. These were not only dangerous for swimming, but the resulting embayments in northern locations became polluted by Mar del Plata sewage discharged from offshore (5 km) outflows.

In 1998, Mar del Plata city and harbour authorities accepted that groynes and breakwaters were not a solution. Major dredging of the harbour mouth produced coarse to very coarse sand that was pumped to Grande, Varese and Bristol beaches (Fig. TB2.11.2). The Grande and Bristol beaches were then about 250 m wide and elevated by 1.5 m above the previous beach level. All the groynes at Bristol beach were fully covered by the sand.

A few months after the nourishment, a major storm produced flooding in the city and river flows from the hills that severely eroded the beaches and moved sand offshore. Wave action has now eroded 50% of the original nourishment material leaving beaches 170–190 m wide, and all groynes have outcropped and are now active in trapping sediment.

Although the solution of nourishing the beaches from the harbour mouth shoal sand may have been a good solution, it was made without a clear understanding of sand grain-sizes required to produce an equilibrium beach. Neither was the solution a "one-time" nourishment operation. The probable "best solution" to preserve beaches to the north of the harbour is to dredge or pump sand that accumulates along the South Jetty and at Mogotes Beach and bypass it to the side of the North Jetty, thus allowing the local dynamics to return the littoral transport lost some 90 years ago.

influenced by a variety of climatic phenomena over different time periods. Individual regions may respond differently to climatic forcing, resulting in variable changes in sediment flux for a given climatic event. The response will depend on the duration of the climate fluctuation and the variability in spatial properties of such parameters as relief, geology and hydrological processes.

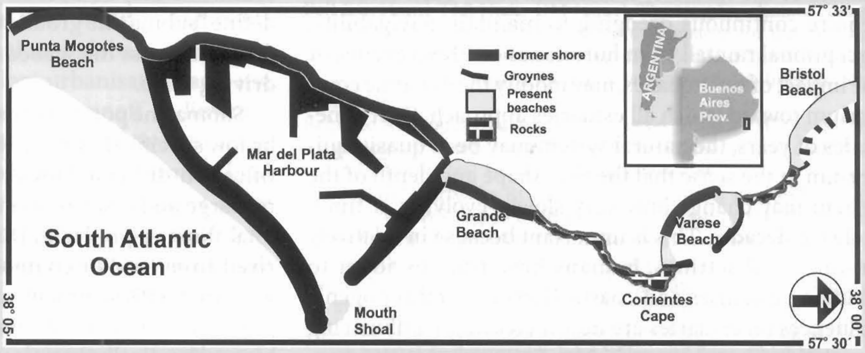
The modern global flux of sediment to the coastal zone (20 Gt yr^{-1}) has two competing Anthropocene forces:

a land-use practices that increase soil erosion (agriculture, deforestation, industrialisation, mining), and
b practices that decrease soil erosion (engineering of waterways including the trapping of sediment by reservoirs).

While humans have substantively increased the continental flux of sediment during the last two millennia, most of the land-derived erosional sediment remains stored someplace between the uplands and the sea. Storage of sediment in large reservoirs constructed during the last 50 years has decreased the global flux of sediment to the coastal zone by 30%. By decreasing sediment loads, coastal erosion is accelerated and coastal ecosystems deteriorate. There is now good documentation on the reduction of local fisheries (e.g., sardine and shrimp catch) as a consequence of hinterland reservoir construction. The coupling of increased nutrient inputs and decreased sediment loads may promote coastal-zone eutrophication and hypoxia.

Future research must identify population thresholds and behaviours that have strong hydrological and sediment flux consequences and lag-times in coastal response. The influence of humans and/or climate affects smaller river basins more dramatically than larger river basins, due to the modulating ability of large rivers. New techniques must be developed to address the coastal response to these sensitive smaller systems. Scaling techniques must also be employed to address the quality of our global databases as: (*a*) most of the observational data was determined across only a few years and interannual variability can exceed mean values by an order of

Fig. TB2.11.1. Schematic map of the harbour and adjacent beach areas of Mar del Plata, Argentina



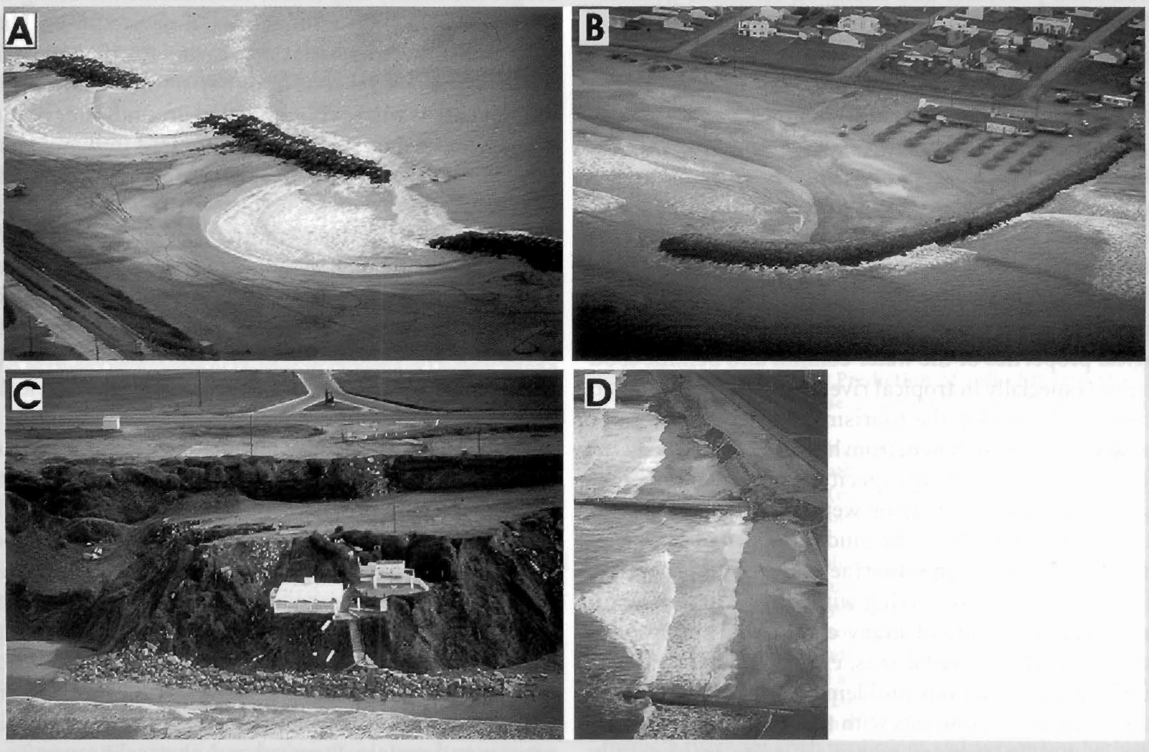


Fig. TB2.11.2. Some of the mechanisms of beach and cliff protection constructed along the Mar del Plata coast, Argentina

magnitude, and (*b*) observational datasets are already a few decades old during a time of rapid change resulting from increased human impacts.

2.7.3 Dynamics at the Estuarine Interface

A major portion of the sedimentary material reaching the coast may be retained in estuaries and deltas, but for most coastal systems the sediment retention level is largely unknown. Geomorphology is probably the most important single factor controlling sediment retention, as the degree of tidal-wave asymmetry is strongly related to the geomorphological influence on tidal propagation. Interactions between geomorphology and advection processes are highly non-linear, making the fate of sediment input into coastal areas difficult to predict, across both short and long time periods.

Even though smaller rivers have relatively large sediment loads, they do not have the strength to drive sediment through the estuary, especially on meso- and macro-tidal coasts where most of the material is retained within the estuary. In the absence of river floods that can overwhelm the system and flush freshwater all the way

to the mouth of the estuary, the filtering, trapping and processing of suspended particles in the estuary will be a function of internal circulation and the form and dynamics of the turbidity maximum. Many such estuaries require continuous dredging to maintain navigability. Exceptional rainfall from hurricanes or ENSO events, or the impact of earthquakes, may modify the dynamic equilibrium toward which all estuaries approach. Over timescales of years, the natural system may be at quasi-equilibrium in the sense that the size, shape and depth of the system may change only very slowly, evolving at timescales of decades. This is important because in relatively un-impacted settings, humans have time to adapt to changing estuaries and coasts. However, anthropogenic influences on estuaries are now increasingly influencing circulation processes and changing estuarine geomorphology and thereby modifying the trapping mechanisms. Artificial structures such as harbours, jetties and breakwaters have little or no capability to adapt to the constant changes in estuarine dynamics.

In the wet tropics, intense rainfall coupled with deforestation, land clearing, overgrazing and other poor farming practices increase soil erosion and sediment flux many times above natural backgrounds. Siltation results in decreased water visibility and increased muddiness, often causing environmental degradation by the smothering of benthic organisms, driving changes in the biological properties of the water column and benthic food chains (especially in tropical river deltas), and leading to economic losses for the tourism industry due to poor aesthetics. When sediment from human-induced soil erosion exceeds the trapping capacity of the estuary (a process exacerbated by estuarine wetlands reclamation and other human activities), the mud is deposited in coastal waters. This change in estuarine and coastal properties may be very rapid, occurring within a few decades, and may threaten the life of many coral reefs. For tropical estuaries and the coastal seas, environmental degradation by mud is a serious problem.

This situation contrasts with that in developed countries in middle latitudes, where estuaries are generally suffering from sediment starvation due to extensive damming and river flow regulation. Sediment retention by dams leads to accelerated coastal recession (e.g., the deltas of the Colorado, Nile, Ebro, Mississippi and Volta rivers). Water diversion schemes also accelerate coastal erosion.

2.7.4 Groundwater Inputs

Both terrestrial and marine forces drive underground fluid flows in the coastal zone. Hydraulic gradients on land result in groundwater seepage near shore and may contribute to flows further out on the shelf from confined aquifers. Marine processes such as tidal pumping and current-induced topographic flow may occur anywhere on the shelf where permeable sediments are present. The terrestrial and oceanic forces overlap spatially and thus measured fluid flow through coastal sediments may be a result of composite forcing. We thus define "submarine groundwater discharge" (SGD) as any flow out across the seabed, regardless of composition or driving force.

Submarine porewater exchange (SPE) is characterised by low specific flow rates that make detection and quantification difficult. However, because such flows (both recharge and discharge) may occur over large areas, the total flux is significant. Discharging fluids, whether derived from land or composed of re-circulated seawater, will react with sediment components changing its composition. These fluids may thus become a source of biogeochemically important constituents including nutrients, metals and radionuclides. If derived from land, such fluids will represent a pathway for new material fluxes to the coastal zone and may result in diffuse pollution in areas where contaminated groundwater occurs.

Additional data collection is required in many areas, especially in South America, Africa and southern Asia where, to our knowledge, no or few assessments are currently described in the literature. We recommend an approach that targets representative types of coastal aquifers based on geology (e.g., karst, coastal plains, deltaic) and environmental parameters (e.g., precipitation, temperature). The production of an SGD database and globalisation efforts are necessary to integrate SGD on a global scale.

Improvements must also be made to techniques used for measurements of SGD. The sensitivity of porewater exchange to local pressure perturbations necessitates the development of new non-invasive methods, and we need to revise our monitoring, measuring and sampling strategies in permeable seabeds. The calculation of realistic estimates of advective porewater exchange in the coastal zone and on the shelf requires concurrent data on bottom current characteristics, sediment topography and sediment permeability. Long-term studies, isotopic signatures and certain chemical and physical properties of the pore fluids may provide indications of the origin and physical forcings of the pore-fluid flows. However, in many cases, several of these processes will be active simultaneously and their separation based on measured data may be very difficult if not impossible. An indispensable tool for overcoming this problem is the modelling of the various transport processes and their effects on the biogeochemical cycles. A new generation of dynamic models is needed to explain the porewater flow observed in natural environments.

The implication of SGD for coastal area management requires assessment of the importance of hydrologic flow in a particular region in a much more time-efficient manner than currently exists. This may require expanding our use of remote sensing, geophysical tools and typology.

2.7.5 The Human Dynamic

Changes to the natural environment due to human intervention impact the flux and flow of water, sediments and nutrients on local, regional and global scales. Construction of dams, deforestation, urbanisation and use of groundwater have modified river runoff, sedimentation patterns, coastline structure and relative sea level. Changes to timing of nutrient fluxes and sediment load have affected trophic structures within ecosystems. River diversion and modified seasonal variability of water levels have affected the health of coastal wetlands, mangrove and coral reef communities as well as leading to changes in shoreline profile and ecosystem composition of rivers, estuaries and coastal seas. Awareness of anthropogenic impacts needs to lead to technological innovations to manage the system as it exists, and to minimise further degradation of coastal systems (see Text Box 2.11).

References

Aagaard K, Darby DA, Falkner K, Flato G, Grebemeier J, Measures C, Walsh JE (1999) Marine science in the Arctic: a strategy. Arctic Research Consortium of the United States (ARCUS), Fairbanks, AK

Aksornkoae S (1993) Ecology and Management of Mangroves. IUCN, Bangkok, Thailand

Alam M (1996) Subsidence of Ganges-Brahmaputra Delta of Bangladesh and associated drainage, sedimentation and salinity problems. In Milliman JD and Haq BU (eds) Sea-level rise and coastal subsidence: causes, consequences and strategies. Kluwer Academic Publishers, Dordrecht, The Netherlands, pp 169–192

Alverson K, Oldfield F, Bradley RS (eds) (2001) Past global changes and their significance for the future. IGBP Science Series No. 3: Environmental variability and climate change (PAGES), Pergamon

Anisimov OA, Nelson FE (1996) Permafrost distribution in the northern hemisphere under scenarios of climatic change. Global and Planetary Change 14:59–72

Anon (1996) Geotechnical aspects of Hong Kong's New Airport. Ground Engineering Special Supplement December 1996

Arbogast AF (2000) Estimating the time since final stabilization of a perched dune field along Lake Superior. Professional Geographer 52(4):594–606

Are FE (1999) The role of coastal retreat for sedimentation in the Laptev Sea. In: Kassen H, Bauch HA, Dmitrenko IA, Eichen H, Hubberten H-W, Melles M Thiede J, Timokhov LA (eds) Land-Ocean Systems in the Siberian Arctic: Dynamics and History, Springer, Berlin, pp 287–295

Arthurton RS (2003) The fringing reef coasts of eastern Africa – present processes in their long-term context. Western Indian Ocean Journal of Marine Science 2(1):1–13

Back W, Hanshaw BB, Pyler TE, Plummer LN, Weide AE (1979) Geochemical significance of groundwater discharge in Caleta Xel Ha, Quintana Roo, Mexico. Water Resources Research 15: 1521–1535

Baker RGV, Haworth RJ (2000) Smooth or oscillating late Holocene sea-level curve? Evidence from cross-regional statistical regressions of fixed biological indicators. Marine Geology 163(1–4): 353–365

Bard E, Hamelin B, Fairbanks RG (1990) U/Th ages obtained by mass spectrometry in corals from Barbados: sea level during the past 130 000 years. Nature 346:456

Bates RL, Jackson JA (1984) Dictionary of Geological Terms. American Geological Institute, NY

Beets, DJ, Roep TB, Westenrhoff WE (1996) The Holocene Bergen Inlet: closing history and related barrier progradation. In: Beets DJ, Fischer MM, Gans W de (editors) Coastal Studies on the Holocene of the Netherlands. Boom-Planeta, The Netherlands, pp 97–131

Bell RG, Hume TM, Hicks DM (2001) Planning for climate change effects on coastal margins. A report prepared for the Ministry for the Environment as part of the New Zealand Climate Change Programme, Ministry for the Environment, Wellington, New Zealand

Belperio AP (1995) Quaternary. In: Drexel JF, Preiss WV (eds) The geology of South Australia, vol 2: the Phanerozoic. South Australia Geological Survey Bulletin 54:219–281

Bender ML, Fairbanks RG, Taylor FW, Matthews RK, Goddard JG, Broecker WS (1979) Uranium-series dating of the Pleistocene reef tracts of Barbados, West Indies. Geological Society of America Bulletin, part I, vol 90, pp 577–594

Berner EK, Berner RA (1987) The global water cycle: geochemistry and environment. Prentice-Hall, Englewood Cliffs, NJ

Bird ECF (1993) Submerging coasts: the effects of a rising sea level on coastal environments. John Wiley and Sons, Chichester

Bird ECF (1996) Coastal erosion and rising sea level. In Milliman JD, Haq BU (eds) (1996) Sea-level rise and coastal subsidence: causes, consequences and strategies. Kluwer Academic Publishers, Dordrecht, The Netherlands, pp 87–103

Bird ECF (2000) Coastal geomorphology: an introduction. Wiley, Chichester

Blanchon P, Eisenhauer A (2001) Multi-stage reef development on Barbados during the last interglaciation. Quaternary Science Reviews 20:1093–1112

Bokuniewicz HJ, Pavlik B (1990) Groundwater seepage along a barrier island. Biogeochemistry 10:257–276

Boudreau BP, Huettel M, Froster S, Jahnke RA, McLachlan A, Middelburg JJ, Nielsen P, Sansone F, Taghon G, Van Raaphorst W, Webster I, Weslawski JM, Wiberg P, Sundby B (2001) Permeable marine sediments: overturning an old paradigm. EOS 82: 133–136

Bray MJ, Hooke JM (1997) Prediction of soft-cliff retreat with accelerating sea level rise. Journal of Coastal Research 13(2): 453–467

Bressolier C, Froidefond J-M, Thomas Y-F (1990) Chronology of coastal dunes in the south-west of France. In: Bakker Th W, Jungerius PD, Klijn JA (eds) Dunes of the European coasts. Geomorphology – hydrology – soils. Catena Supplement 18: 101–107

Broecker WS (1979) Thermohaline circulation, the Achilles heel of our climate system: will man-made CO_2 upset the current balance? Science 278:1582–1588

Brown BE, Dunne RP, Goodson MS, Douglas AE (2000) Bleaching patterns in reef corals. Nature 404:99–105

Bruun P (1962) Sea-level rise as a cause of shore erosion. Journal of Waterways and Harbors Division (ASCE) 88:116–130

Buddemeier RW, Smith SV (1988) Coral reef research in an era of rapidly rising sea level: predictions and suggestions for long-term research. Coral Reefs 7:51–56

Bugna GC, Chanton JP, Cable JE, Burnett WC, Cable PH (1996) The importance of groundwater discharge to the methane budgets of nearshore and continental shelf waters of the northeastern Gulf of Mexico. Geochimica Cosmochimica Acta 60:4735–4746

Burnett WC, Turner J (2001) LOICZ group investigates groundwater discharge in Australia. LOICZ Newsletter 18:1–4

Burnett WC, Taniguchi M, Oberdorfer J (2001) Measurement and significance of the direct discharge of groundwater into the coastal zone. Journal of Sea Research 46(2):109–116

Burnett WC, Chanton J, Christoff J, Kontar E, Krupa S, Lambert M, Moore W, O'Rourke D, Paulsen R, Smith C, Smith L, Taniguchi M (2002) Assessing methodologies for measuring groundwater discharge to the ocean. EOS 83:117–123

Burnett WC, Chanton JP, Kontar E (2003) Submarine groundwater discharge. Biogeochemistry 66:1–202

Burt RA (1993) Groundwater chemical evolution and diagenetic processes in the Upper Floridan aquifer, southern South Carolina and northeastern Georgia. US Geological Survey Water-Supply Paper 2392

Cable JE, Burnett WC, Chanton JP, Corbett DR, Cable PH (1997) Field evaluation of seepage meters in the coastal marine environment. Estuarine, Coastal and Shelf Science 45:367–375
Cai WJ, Wang Y (1998) The chemistry, fluxes and sources of carbon dioxide in the estuarine waters of the Satilla and Altamaha rivers, Georgia. Limnology and Oceanography 43:657–668
Cathles LM (1990) Scales and effects of fluid flow in the upper crust. Science 248:323–330
Caviglia FJ, Pousa JL, Lanfredi NW (1991) A determination of the energy flux constant from dredge records. Journal of Coastal Research 7:543–550
Chapin FS, McGuire AD, Randerson J, Pielke R, Baldocchi D, Hobbie SE, Roulet N, Eugster W, Kasischke E, Rastetter EB, Zimov SA, Running SW (2000) Arctic and boreal ecosystems of western North America as components of the climate system. Global Change Biology 6:211–223
Chappell, J (1993) Contrasting Holocene sedimentary geologies of the lower Daly River, northern Australia and lower Sepik-Ramu, Papua New Guinea. Sedimentary Geology 83:339–358.
Chappell, J, Omura A, Esat T, McCulloch M, Pandolfi J, Ota Y, Pillans B (1996) Reconciliation of Late Quaternary sea levels derived from coral terraces at Huon Peninsula with deep sea oxygen isotope records. Earth and Planetary Science Letters 141:227–236
Christiansen C, Bowman D (1986) Sea-level changes, coastal dune building and sand drift, NW Jutland, Denmark. Geografisk Tidsskrift 86:28–31
Christiansen C, Dalsgaard K, Moller JT, Bowman D (1990) Coastal dunes in Denmark. Chronology in relation to sea level. In: Bakker Th W, Jungerius PD, Klijn JA (eds) Dunes of the European coasts. Geomorphology – hydrology – soils. Catena Supplement 18:61–70
Conley DC, Inman DL (1994) Ventilated oscillatory boundary layers. Journal of Fluid Mechanics 273:261–284
Considine DM (ed) (1995) Van Nostrand's scientific encyclopedia, 8th edn. NY, Van Nostrand, Reinhold
Cooper HH, Kohout FA, Henry HR, Glover RE (1964) Sea water in coastal aquifers. US Geological Survey Water Supply Paper 1613-C:84
Cowell PJ, Thom BG (1994) Morphodynamics of coastal evolution. In: Carter RWG, Woodroffe CD (eds) Coastal evolution: late Quaternary shoreline morphodynamics. Cambridge University Press, Cambridge, United Kingdom and New York, NY, USA, pp 33–86
Crossland CJ (1982) Dissolved nutrients in reef waters of Sesoko Island, Okinawa: a preliminary study. Galaxea 1:47–54
Curtis WF, Culbertson K, Chase EB (1973) Fluvial-sediment discharge to the oceans from the conterminous United States. US Geological Survey Circular 670:1–17
D'Anglejan B (1994) Potential sedimentological impacts of hydroelectric developments in James Bay and Hudson Bay. James Bay Publication Series, Paper No. 5
Danilov ID (2000) Submarine permafrost of the Arctic Shelf and related processes of cryogenic lithomorphogenesis. Oceanology 40:714–722
Davidson-Arnott RGD, Pyskir NM (1988) Morphology and formation of a Holocene coastal dune field, Bruce Peninsula, Ontario. Geographie Physique et Quaternaire 42(2):163–170
Davis JL, Mitrovica JX (1996) Glacial isostatic adjustment and the anomalous tide gauge record of eastern North America. Nature 379:331–333
Day JW, Martin JF, Cordoch L, Templet PH (1997) System functioning as a basis for sustainable management of deltaic ecosystems. Coastal Management 25:115–153
Dedkov AP, Mozzherin VI (1992) Erosion and sediment yield in mountain regions of the world. In: Walling DE, Davies TR, Hasholt B (eds) Erosion, debris flows and environment in mountain regions (Proceedings of the Chengdu Symposium, July 1992). IAHS Publication 209:29–36
D'Elia CF, Webb KL, Porter JW (1981) Nitrate-rich groundwater inputs to Discovery Bay, Jamaica: a significant source of N to local reefs? Bulletin of Marine Science 31:903–910
de Vries I (2001) Shoreline and land-use development perspectives in the context of long-term decreasing functional diversity in the Delta Region, SW Netherlands. In: von Bodungen B and Turner RK (eds) Science and Integrated Coastal Management, Dahlem Workshop Report 85, Dahlem University Press, Berlin, pp 113–120
Dionne JC (1963) Towards a more adequate definition of the St Lawrence estuary. Journal of Geomorphology 7:36–44
Dixon RK, Brown S, Houghton RA, Solomon AM, Trexler MC, Wisniewski J (1994) Carbon pools and flux of global forest ecosystems. Science 263:185–190
Doornkamp JC (1998) Coastal flooding, global warming and environmental management. Journal of Environmental Management 52:327–333
Douglas BC (1991) Global sea level rise. Journal of Geophysical Research 96(C4):6981–6992
Douglas BC (1997) Global sea rise: a redetermination. Surveys in Geophysics 18:279–292
Douglas BC (2001) Sea level change in the era of the recording tide gauge. In: Douglas BC, Kearney MS, Leatherman SP (eds) Sea-level rise: history and consequences. Academic Press, International Geophysics Series vol 75, pp 37–64
Douglas, BC, Peltier, WR (2002) The puzzle of global sea-level rise. Physics Today, March 2002, pp 35–40
Douglas I (1993) Sediment transfer and siltation. In: Turner BL, Clark WC, Kates RW, Richards JF, Mathews JT, Meyer WB (eds) The earth as transformed by human action. Cambridge University Press, Cambridge, pp 215–234
Dronkers J, van Leussen W (1988) Physical processes in estuaries. Springer-Verlag, Berlin
Dunbar R, Cole J (1999) Annual records of tropical systems (ARTS). Recommendations for research. PAGES Workshop Report, Series 99-1
Eicken H (2004) The role of Arctic sea ice in transporting and cycling terrestrial organic matter. In: Stein R, Macdonald RW (eds) The organic carbon cycle in the Arctic Ocean: present and past. Springer-Verlag, Berlin
Ellison JC, Stoddart DR (1991) Mangrove ecosystem collapse with predicted sea-level rise: Holocene analogues and implications. Journal of Coastal Research 7:151–165
El-Sayed M Kh (1996) Rising sea level and subsidence of the northern Nile Delta. In: Milliman JD, Haq BU Sea-level Rise and Coastal Subsidence. Kluwer Academic Publishers, Dordrecht, the Netherlands, pp 215–233
Emery KO (1968) Relict sediments on continental shelves of the world. American Association of Petroleum Geologists 52:445–464
Emery KO, Uchupi E (1972) Western North Atlantic Ocean: topography, rocks, structures, water, life and sediments. American Association of Petroleum Geologists Memoir 17
Emiliani C (1955) Pleistocene temperatures. Journal of Geology 63: 539–578
ERL (Asia) Limited (1982) Proposed replacement airport at Chek Lap Kok – potential impact on the marine environment. Prepared for the Civil Aviation Department, Hong Kong Government, Hong Kong
Farrow GE, Syvitski JPM, Tunnicliffe V (1983) Suspended particulate loading on the macrobenthos in a highly turbid fjord: Knight Inlet, British Columbia. Canadian Journal of Fisheries and Aquatic Sciences 40:100–116
Forbes DL, Taylor RB (1994) Ice in the shore zone and the geomorphology of cold coasts. Progress in Physical Geography 18:59–89
Forbes DL, Orford JD, Carter RWG, Shaw J, Jennings SC (1995) Morphodynamic evolution, self-organization and instability of coarse-clastic barriers on paraglacial coasts. Marine Geology 126:63–85
Forbes DL, Orford JD, Taylor RB, Shaw J (1997) Interdecadal variation in shoreline recession on the Atlantic Coast of Nova Scotia. In: Proceedings of the Canadian Coastal Conference '97, Guelph, Ontario, Canadian Coastal Science and Engineering Association, Ottowa, Canada, pp 360–374
Freeman C, Evans CD, Monteith DT (2001) Export of organic carbon from peat soils. Nature 412:785
Freeze RA, Cherry JA (1979) Groundwater. Prentice-Hall Inc. Englewood Cliffs, NJ
Froidefond J-M, Legigan P (1985) La grande dune du Pilat et la progression des dunes sur la littoral Aquitaine. Bulletin Institut Geologique du Bassin d'Aquitaine 38:69–79
Fryberger S, Krystinik LF, Schenk CJ (1990) Tidally flooded back-barrier dunefield, Guerrero Negro area, Baja California, Mexico. Sedimentology 37: 1–23

Gallup CD, Edwards RL, Johnson RG (1994) The timing of high sea levels over the past 200 000 years. Science 263:796–800

Garrels RM, MacKenzie FT (1971) Evolution of sedimentary rocks. Norton & Co, New York

Gehlen M, Malschaert H, van Raaphorst WR (1995) Spatial and temporal variability of benthic silica fluxes in the southeastern North Sea. Continental Shelf Research 15:1675–1696

Giles PT, McCann SB (1997) Foredune development on Iles de la Madeleine (Quebec), Atlantic Canada. Canadian Journal of Earth Sciences 34:1467–1476

Goodwin I (2002) Unravelling climate influences on late Holocene sea-level variability. In: Anson M, Battarbee R, Birks J, Oldfield F (eds) Global change in the Holocene. Edward Arnold, pp 406–421

Goodwin I, Harvey N, van der Plassche O, Oldfield F and Oglesby R (2000) Prospects of resolving climate-induced sea-level fluctuations over the last 2000 years. Transactions of the American Geophysical Union (EOS) 81(28):311–314

Gordon DC Jr, Boudreau PR, Mann KH, Ong JE, Silvert WL, Smith SV, Wattayakorn G, Wulff F, Yanagi T (1996) LOICZ biogeochemical modelling guidelines. LOICZ Reports & Studies No. 5, LOICZ, Texel, The Netherlands

Gornitz V (1995) A comparison of differences between recent and late Holocene sea level trends from eastern North America and other selected regions. In: Finkl CW Jr (ed) Holocene cycles: climate, sea levels and sedimentation. Journal of Coastal Research, Special Issue No. 17:287–297

Gornitz V (2001) Impoundment, groundwater mining, and other hydrologic transformations: impacts on global sea level rise. In: Douglas BC, Kearney M.S, Leatherman SP (eds) Sea level rise: history and consequences. Academic Press, San Diego, CA, pp 97–119

Gornitz V, Lebedeff S (1987) Global sea level changes during the past century. In: Nummedal D, Pilkey OH, Howard JD (eds) Sea level fluctuation and coastal evolution. Society for Economic Paleontologists and Mineralogists Special Publication 41:3–16

Gornitz V, White TW (1992) A coastal hazards database for the US east coast. ORNL/CDIAC-45, NDP-043A. Oak Ridge National Laboratory, Oak Ridge, TN

Gornitz V, Daniels RC, White TW, Birdwell KR (1994) The development of a coastal risk assessment database: vulnerability to sea level rise in the US southeast. Journal of Coastal Research, Special Issue No. 12:327–338

Gornitz V, Couch S, Hartig EK (2002) Impacts of sea level rise in the New York metropolitan area. Global and Planetary Change 32: 61–88

Gorsink JP, Baker GC (1990) Salt fingering in subsea permafrost: some stability and energy considerations. Journal of Geophysical Research 95:9575–9584

Guan WB, Wolanski E, Dong LX (1998) Cohesive sediment transport in the Jiaojiang River estuary, China. Estuarine, Coastal and Shelf Science 46:861–871

Guay CK, Klinkhammer GP, Falkner KK, Benner R, Coble PG, Whitledge TE, Black B, Bussell FJ, Wagner TA (1999) High-resolution measurements of dissolved organic carbon in the Arctic Ocean by in situ fiber-optic spectrometry. Geophysical Research Letters 26:1007–1010

Guillen J, Palanques A (1997) A historical perspective of the morphological evolutions in the lower Ebro River. Environmental Geology 30:174–180

Haida S (2000) Transport de matières et bilan de l'érosion mécanique et de l'altération chimique dans un bassin versant de la zone semi-aride: Impact des variations climatiques et des activités humaines. Thèse Doctorat d'Etat, Universite Ibn Tofail, Kénitra

Hammar-Klose E, Thieler ER (2001) Coastal vulnerability to sea-level rise: a preliminary database for the US Atlantic, Pacific and Gulf of Mexico coasts. US Geological Survey Digital Data Series DDS-68

Harrison WD, Musgrave D, Reeburgh WS (1983) A wave-induced transport process in marine sediments. Journal of Geophysical Research 88:7617–7622

Hart BS, Long BF (1990) Recent evolution of the Outardes Estuary, Quebec, Canada: consequences of dam construction on the river. Sedimentology 37:495–507

Hartig EK, Gornitz V, Kolker A, Mushacke F, Fallon D (2002) Anthropogenic and climate-change impacts on salt marshes of Jamaica Bay, New York City. Wetlands 22:71–89

Harvey N (1986) The Great Barrier Reef: shallow seismic investigations. Department of Geography, James Cook University of North Queensland, Monograph Series No. 14

Harvey N, Belperio AP (1994) Implications of climatic change for the South Australian coast. Transactions of the Royal Society of South Australia 118(1):45–52

Harvey N, Caton B (2003) Coastal management in Australia. Oxford University Press

Harvey N, Clouston B, Carvalho P (1999) Improving coastal vulnerability assessment methodologies for integrated coastal management: an approach from South Australia. Australian Geographical Studies 37(1):50–69

Harvey N, Belperio AP, Bourman R, Mitchell W (2002) Geologic, isostatic and anthropogenic signals affecting sea level records at tide gauge sites in southern Australia. Global and Planetary Change 32:1–11

Hay WH (1994) Pleistocene-Holocene fluxes are not the earth's norm. In: Hay WH (ed) Global sedimentary geofluxes. National Academy of Sciences Press, Washington, pp 15–27

Hayes MO (1967) Relationship between coastal climate and bottom sediment type on the inner continental shelf. Marine Geology 5: 111–132

Hayes, MO (1980) General morphology and sediment patterns in tidal inlets. Sedimentary Geology 26:139–156

Hesp PA (1984a) The formation of sand "beach ridges" and foredunes. Search 15(9–10):289–291

Hesp PA (1984b) Aspects of the geomorphology of south-western Australian estuaries. In: Hodgkin EP (Ed) Estuarine environments of the Southern Hemisphere. WA Dept of Conservation and Environment Bulletin 161

Hesp PA (1993) Late Quaternary coastal geomorphology of the South Sydney-Illawarra coast. In: Thoms M (ed) Catchments and coasts. Dept of Geography, University of Sydney Research Monograph No. 5:72–87

Hesp PA, Short AD (1999) Barrier morphodynamics. In: Short AD (ed) Handbook of beach and shoreface morphodynamics. John Wiley, pp 307–333

Hesp PA, Thom BG (1990) Geomorphology and evolution of active transgressive dunefields. In: Nordstrom KF, Psuty NP, Carter RWG (eds) Coastal dunes: form and process. John Wiley and Sons, London, pp 253–288

Hoegh-Guldberg O (1999) Climate change, coral bleaching and the future of the world's coral reefs. Journal of Marine and Freshwater Research 50:839–866

Holeman JN (1980) Erosion rates in the US estimated from the Soil Conservation Services inventory. EOS (Transactions of the AGU) 61:954

Holligan PM, de Boois H (1993) Land-Ocean Interactions in the Coastal Zone, Science Plan, International Geosphere-Biosphere Programme, Report No. 25

Houghton JT, Jenkins GJ, Ephraums JJ (1991) Climatic change. The IPCC scientific assessment. Cambridge University Press

Houghton JT, Callender BA, Varney SK (1992) Climatic change 1992. The supplementary report to the IPCC assessment. Cambridge University Press

Houghton JT, Meira Filho LG, Callander BA, Harris N, Kattenburg A, Maskell K (eds) (1996) Climatic change 1995: The science of climate change. Published for the IPCC by Cambridge University Press

Houghton JT, Ding Y, Griggs DJ, Noguer M, van der Linden PJ, Xiaosu D (eds) (2001) Climate change 2001: the scientific basis. Contribution of Working Group 1 to the Third Assessment Report of the Intergovernmental Panel on Climate Change (IPCC). New York, Cambridge University Press

Hu D, Saito Y, Kempe S (1998) Sediment and nutrient transport to the coastal zone. In: Galloway JN, Melillo JM (eds) Asian change in the context of global climate change: Impact of natural and anthropogenic changes in Asia on global biogeochemical cycles. Cambridge University Press, IGBP Publication Series 3, 245–270

Hubbert GD, McInnes K (1999) A storm surge inundation model for coastal planning and impact studies. Journal of Coastal Research 15:168–185

Huettel M, Gust G (1992) Solute release mechanisms from confined sediment cores in stirred benthic chambers and flume flows. Marine Ecology Progress Series 82:187–197
Huettel M, Rusch A (2000) Transport and degradation of phytoplankton in permeable sediment. Limnology and Oceanography 45:534–549
Huettel M, Ziebis W, Forster S (1996) Flow-induced uptake of particulate matter in permeable sediments. Limnology and Oceanography 41:309–322
Huettel M, Ziebis W, Forster S, Luther G (1998) Advective transport affecting metal and nutrient distribution and interfacial fluxes in permeable sediments. Geochimica y Cosmochimica Acta 62: 613–631
Huntington HP (2000) Native observations capture impacts of sea ice changes. Witness the Arctic 8:1–2
Huntley DJ, Hutton JT, Prescott JR (1993) The stranded beach-dune sequence of south-east South Australia: A test of thermoluminescence dating, 0–800 ka. Quaternary Science Reviews 12:1–20
Huntley DJ, Hutton JT, Prescott JR (1994) Further thermoluminescence dates for the dune sequence in the south-east of South Australia. Quaternary Science Reviews 13:201–207
Huntley D, Leeks G, Walling D (eds) (2001) Land ocean interaction measuring and modelling fluxes from river basins to coastal seas. IWA Publishing
Hutchinson PA, Webster IT (1998) Solute uptake in aquatic sediments due to current-obstacle interactions. Journal of Environmental Engineering 124:419–426
Illenburger K, Rust I (1988) A sand budget for the Alexandria coastal dunefield, South Africa. Sedimentology 35:513–521
Inman DL, Jenkins SA (1999) Climate change and the episodicity of sediment flux of small California rivers. Journal of Geology 107: 251–270
Intergovernmental Panel on Climate Change (IPCC) (1991) Common methodology for assessing vulnerability to sea-level rise. IPCC Coastal Zone Management Subgroup, Ministry of Transport and Public Works, The Hague, The Netherlands
Intergovernmental Panel on Climate Change (IPCC) (1992) Global climate change and the rising challenge of the sea. Report of the Coastal Zone Management Subgroup, Ministry of Transport, Public Works and Water Management, The Hague, The Netherlands
Intergovernmental Panel on Climate Change (IPCC) (1994) Preparing to meet the coastal challenges of the 21st century. Report of the world coast conference, 1–5 November 1993. Ministry of Transport, Public Works and Water Management, The Hague, The Netherlands
Jackson JA (ed) (1997) Glossary of Geology, 4th edn. American Geological Institute, Alexandria, VA
Jahnke R, Nelson JR, Marinelli RL, Eckman JE (2000) Benthic flux of biogenic elements on the southeastern US continental shelf: influence of pore water advective transport and benthic microalgae. Continental Shelf Research 20:109–127
Jelgersma S, De Jong JD, Zagwijn WH, Van Regteren Altena JF (1970) The coastal dunes of the western Netherlands: geology, vegetational history and archaeology. In: Mededelingen Rijks Geologische Dienst, 21, pp 93–167. Rijks Geologische Dienst, Haarlem.
Jennings JN (1967) Cliff-top dunes. Australian Geographical Studies 5(1):40–49
Jia JJ, Gao S, Xue YC (2003) Sediment dynamics of small tidal inlets: an example from Yuehu Inlet, Shandong Peninsula, China. Estuarine, Coastal and Shelf Science 57(5–6):783–801
Jickells TD (1998) Nutrient biogeochemistry of the coastal zone. Science 281:217–222
Jiménez JA, Sánchez-Arcilla A (1993) Medium-term coastal response at the Ebro Delta, Spain. Marine Geology 114:105–118
Jiménez JA, Sánchez-Arcilla A (1997) Physical impacts of climatic change on deltaic coastal systems (II): driving terms. Climatic Change 35(1):95–118
Jing L, Ridd PV, Mayocchi CL, Heron ML (1996) Wave-induced benthic velocity variations in shallow waters. Estuarine and Coastal Shelf Science 42:787–802
Johannes RE (1980) The ecological significance of the submarine discharge of groundwater. Marine Ecology Progress Series 3: 365–373
Johnson MA, Polyakov IV (2001) The Laptev Sea as a source for recent Arctic Ocean salinity changes. Geophysical Research Letters 28:2017–2020
Kairu K, Nyandwi N (eds) (2000) Guidelines for the study of shoreline change in the Western Indian Ocean region. IOC Manuals and Guides No. 40. UNESCO, Paris
Kearney MS (2001) Late Holocene sea level variation. In: Douglas BC, Kearney MS, Leatherman SP (eds) Sea level rise: history and consequences. Academic Press International Geophysics Series vol 75, pp 13–36
Klein RJT, Nicholls RJ (1999) Assessment of coastal vulnerability to climate change. Ambio 28:182–187
Klein RJT, Buckley EN, Nicholls RJ, Ragoonaden S, Aston J, Capobianco M, Mizutani M, Nunn PD (2000) Coastal adaptation. In: Metz B, Davidson OR, Martens JW, van Rooijen SN, van Wie McGrory L (eds) Methodologies and technological issues in technology transfer. A special report of IPCC Working Group III, Cambridge University Press, Cambridge, UK, pp 349–372
Kleypas JA, Buddemeier RW, Archer D, Gattuso J, Landon C, Opdyke BN (1999) Geochemical consequences of increased carbon dioxide in coral reefs. Science 284:118–120
Klijn JA (1990) The younger dunes in the Netherlands; chronology and causation. In: Bakker ThW, Jungerius PD, Klijn JA (ed) Dunes of the European coasts: geomorphology – hydrology – soils. Catena Supplement 18:89–100
Knox JC (1993) Large increase in flood magnitude in response to modest changes in climate. Nature 361:430–432
Kohout FA (1966) Submarine springs: a neglected phenomenon of coastal hydrology. Hydrology 26:391–413
Komar PD (1998b) The 1997–98 El Niño and erosion on the Oregon coast. Shore and Beach 66(3):33–41
Krest JM, Moore WS, Gardner LR, Morris J (2000) Marsh nutrient export supplied by groundwater discharge: evidence from Ra measurements. Global Biogeochemical Cycles 14:167–176
Lambeck K, Nakada M (1990) Late Pleistocene and Holocene sea-level change along the Australian coast. Palaeogeography, Palaeoclimatology, Palaeoecology (Global and Planetary Change Section) 89:143–176
Lambeck K, Smither C, Ekman M (1998) Tests of glacial rebound models for Fennoscandinavia based on instrumented sea- and lake-level records. Geophysical Journal International 135: 375–387
Lammers RB, Shiklomanov AI, Vorosmarty CJ, Fekete BM, Peterson BJ (2001) Assessment of contemporary Arctic river runoff based on observational discharge records. Journal of Geophysical Research-Atmospheres 106:3321–3334
Le Floch J (1961) Propagation de la marée dans l'estuaire de la Seine et en Seine Maritime, DSc thesis, Université de Paris
Leatherman SP (2001) Social and economic costs of sea level rise. In: Douglas BC, Kearney MS, Leatherman SP (eds) Sea level rise: history and consequences. International Geophysics Series, vol 75. Academic Press, San Diego, London
Leatherman SP, Zhang K, Douglas BC (2000) Sea level rise shown to drive coastal erosion. EOS 81(6):55–57
Lee SY, Holder GD (2001) Methane hydrates potential as a future energy source. Fuel Processing Technology 71:181–186
Lewis AG, Syvitski JPM (1983) Interaction of plankton and suspended sediment in fjords: Sedimentary Geology 36:81–92
Li L, Barry DA, Stagnitti F, Parlange JY (1999) Submarine groundwater discharge and associated chemical input to a coastal sea. Water Resources Research 35:3253–3259
Lohse L, Epping EHG, Helder W, van Raaphorst WR (1996) Oxygen porewater profiles in continental shelf sediments of the North Sea – turbulent versus molecular diffusion. Marine Ecology Progress Series 145:63–75
Loope WL, Arbogast AF (2000) Dominance of a ~150 year cycle of sand supply change in late Holocene dune-building along the eastern shore of Lake Michigan. Quaternary Research 54: 414–422
MacIntyre IB, Nurke RB, Stuckenrath R (1977) Thickest recorded Holocene reef section, Isla Periez core hole, Alacran Reef, Mexico. Geology 5:749–754

Madej MA, Ozaki V (1996) Channel response to sediment wave propagation and movement, Redwood Creek, California, USA. Earth Surface Processes and Landforms 21(10):911–927

Marinelli RL, Jahnke RA, Craven DB, Nelson JR, Eckman JE (1998) Sediment nutrient dynamics on the South Atlantic Bight continental shelf. Limnology and Oceanography 43: 1305–1320

Marsh JA (1977) Terrestrial inputs of nitrogen and phosphates on fringing reefs on Guam. Proceedings Third International Coral Reef Symposium, pp 331–336

McCarthy JJ, Canziani OF, Leary NA, Dokken DJ, White KS (eds) (2001) Climate change 2001: impacts, adaptation and vulnerability – contribution of Working Group II to the Third Assessment Report of the Intergovernmental Panel on Climate Change (IPCC). New York, Cambridge University Press

McLaughlin CJ, Smith, CA, Buddemeier RW, Bartley JD, Maxwell BA (2003) Rivers, runoff and reefs. Global and Planetary Change 39:191–199

McLean RF, Mimura N (eds) (1993) Vulnerability assessment to sea-level rise and coastal zone management. Proceedings of the IPCC Eastern Hemisphere Workshop, Tokyo

Meade RH (1996) River-sediment inputs to major deltas. In: Milliman JD, Haq BU (eds) Sea-level rise and coastal subsidence. Kluwer Academic Publishers, Dordrecht, The Netherlands, pp 63–85

Meade RH, Yuzyk TR, Day TJ (1990) Movement and storage of sediment in rivers of the United States and Canada. In: Wolman MG, Riggs HC (eds) Surface water hydrology. The geology of North America, v. O-1, Geological Society of America, Boulder CO, pp 255–280

Milliman JD, Syvitski JPM (1992) Geomorphic/tectonic control of sediment discharge to the ocean: the importance of small mountainous rivers. Journal of Geology 100(5):525–544

Milliman JD, Qin YS, Ren MEY (1987) Man's influence on the erosion and transport of sediment by Asian rivers: The Yellow River (Huanghe) example. Journal of Geology 95:751–762

Milliman JD, Haq BU (eds) (1996) Sea-level rise and coastal subsidence: causes, consequences and strategies. Kluwer Academic Publishers, Dordrecht, The Netherlands

Mitchell W, Chittleborough J, Ronai B, Lennon GW (2001) Sea level rise in Australia and the Pacific. In: Grzechnik M, Chittleborough J (eds) Proceedings of the Pacific Islands conference on climate change, climate variability and sea level rise: linking science and policy, Rarotonga, Cook Islands, 3–7 April 2000, National Tidal Facility Australia, Adelaide

Mitrovica JX, Davis JL (1995) Present-day post-glacial sea level change far from the late Pleistocene ice sheets: implications for recent analyses of tide-gauge records. Geophysical Research Letters 22:2529–2532

Moore WS (1996) Large groundwater inputs to coastal waters revealed by ^{226}Ra enrichments. Nature 380:612–614

Moore WS (1999) The subterranean estuary: a reaction zone of groundwater and seawater. Marine Chemistry 65:111–126

Morehead MD, Syvitski JPM, Hutton EWH, Peckham SD (2003) Modeling the inter-annual and intra-annual variability in the flux of sediment in ungauged river basins. Global and Planetary Change 39:95–110

Morison J, Aagaard K, Steele M (2000) Recent environmental changes in the Arctic: a review. Arctic 53:359–371

Moritz RE, Aagaard K, Baker DJ, Codispoti LA, Smith SL, Smith WO, Tipper RC, Walsh JE (1990) Arctic system science: ocean-atmosphere-ice interactions. Joint Oceanographic Institutions Inc., Washington DC

Mulder T, Syvitski JPM (1995) Turbidity currents generated at river mouths during exceptional discharge to the world oceans. Journal of Geology 103:285–298

Nairn RB, Solomon S, Kobayashi N, Virdrine J (1998) Development and testing of a thermal-mechanical model for predicting Arctic shore erosion processes. In: Lewkowicz AG, Allard M (eds) Permafrost: Proceedings of the Seventh International Conference, Yellowknife, NWT, June 23–27 Centre d'études nordiquies, Universite Laval, Quebec, Canada, pp 789–795

Naki-enovi N, Alcamo J, Davis G, de Vries B, Fenhann J, Gaffin S, Gregory K, Grübler A, Jung TY, Kram T, La Rovere EL, Michaelis L, Mori S, Morita T, Pepper W, Pitcher H, Price L, Raihi K, Roehrl A, Rogner H-H, Sankovski A, Schlesinger M, Shukla P, Smith S, Swart R, van Rooijen S, Victor N, Dadi Z (2000) IPCC Special report on emissions scenarios. Cambridge University Press, Cambridge, UK and New York, NY, USA

Nelson JR, Eckman JE, Robertson CY, Marinelli RI, Jahnke RA (1999) Benthic microalgal biomass and irradiance at the sea floor on the continental shelf of the South Atlantic Bight: spatial and temporal variability and storm effects. Continental Shelf Research 19:477–505

Nerem RS, Mitchum GT (2001) Observations of sea level change from satellite altimetry. In: Douglas BC, Kearney MS, Leatherman SP (eds) Sea level rise: history and consequences. Academic Press, International Geophysics Series, vol 75, pp 121–163

Nicholls RJ (2002) Analysis of global impacts of sea-level rise: a case study of flooding. Physics and Chemistry of the Earth, Parts A/B/C 27(32–34):1455–1466

Nicholls RJ, Branson J (1998) Coastal resilience and planning for an uncertain future: an introduction. Geographical Journal 164: 255–258

Nicholls RJ, de la Vega-Leinert A (2001) The SURVAS project at the Overview Workshop. In: de la Vega-Leinert A, Nicholls RJ (eds) Proceedings of the SURVAS Overview Workshop on the future of vulnerability and adaptation studies. The Royal Chace, London

Nicholls RJ, Leatherman SP (eds) (1995) Potential impacts of accelerated sea level rise on developing countries. Journal of Coastal Research Special Issue 14

Nicholls RJ, Hoozemans FMJ, Marchand M (1999) Increasing flood risk and wetland losses due to sea-level rise: regional and global analyses. Global Environmental Change 9:S69–S87

Nielsen P (1990) Tidal dynamics in the water table in a beach. Water Resources Research 26:2127–2134

Nittrouer, CA, DeMaster, DJ (1986) Sedimentary processes on the Amazon continental shelf: past, present and future research. Continental Shelf Research 6:5–30

Nittrouer CA, Wright LD (1994) Transport of particles across continental shelves. Reviews of Geophysics 32:85–113

Nixon SW (2003) Replacing the Nile: are anthropogenic nutrients providing the fertility once brought to the Mediterranean by a great river? Ambio 32 (1):30–39

Opsahl S, Benner R, Ammon RMW (1999) Major flux of terrigenous dissolved organic matter through the Arctic Ocean. Limnology and Oceanography 44:2017–2023

Paerl HW (1997) Coastal eutrophication and harmful algal blooms: importance of atmospheric deposition and groundwater as "new" nitrogen and other nutrient sources. Limnology and Oceanography 42:1154–1165

Pasternack GB, Brush GS, Hilgartner WB (2001) Impact of historic land-use change on sediment delivery to a Chesapeake Bay subestuarine delta. Earth Surface Processes and Landforms 26: 409–427

Paull CK, Speiss F, Curry J, Twichell D (1990) Origin of Florida Canyon and the role of spring sapping on the formation of submarine box canyons. Geological Society of America Bulletin 102: 502–515

Peltier WR (1996) Global sea level rise and glacial isostatic adjustment: an analysis of data from the east coast of North America. Geophysical Research Letters 23:717–720

Peltier WR (2001) Global glacial isostatic adjustment and modern instrumental records of relative sea level history. In: Douglas BC, Kearney MS, Leatherman SP (eds) Sea level rise: history and consequences. Academic Press, San Diego, CA, pp 65–95

Peltier WR (2002) On eustatic sea level history: last glacial maximum to Holocene. Quaternary Science Reviews 21:377–396

Peltier WR, Jiang X (1997) Mantle viscosity, glacial isostatic adjustment and the eustatic level of the sea. Surveys in Geophysics 18: 239–277

Peltier WR, Tushingham AM (1989) Global sea level rise and the greenhouse effect: might they be connected? Science 244: 806–810

Peltier WR, Tushingham AM (1991) Influence of glacial isostatic adjustment on tide gauge measurements of secular sea level change. Journal of Geophysical Research 96:6779–6796

Perillo GME (1995) Definition and geomorphologic classifications of estuaries. In Perillo GME (ed) Geomorphology and sedimentology of estuaries. Developments in Sedimentology 53:17–47, Elsevier Science BV, Amsterdam

Perillo GME (2000) Sediment budgets and fluxes in estuarine and coastal areas. IGBP-LOICZ Water Workshop, Boulder, CO (oral presentation)

Pernetta JC, Milliman JD (1995) Land-Ocean Interactions in the Coastal Zone, implementation plan. International Geosphere-Biosphere Programme, Report No. 33

Pethick J (2001) The Anglian coast. In von Bodungen B and Turner RK (eds) Science and integrated coastal management. Dahlem Workshop Report 85, Dahlem University Press, Berlin, pp 121–134

Petit JR, Jouzel J, Raynaud D, Barkov NI, Barnola J-M, Basile I, Bender, M, Chappellaz J, Davis M, Delaygue G, Delmotte M, Kotlyakov VM, Legrand M, Lipenkov VY, Lorius C, Pépin L, Ritz C, Saltzman E, Stievenard M (1999) Climate and atmospheric history of the past 420 000 years from the Vostok ice core, Antarctica. Nature 399:429–436

Pfirman S, Gascard J-C, Wollenburg I, Mudie P, Abelman A (1989) Particle-laden Eurasian Arctic sea ice: observations from July and August 1987. Polar Research 7:59–66

Pfirman S, Colony R, Nürnberg D, Eicken H, Rigor I (1997) Reconstructing the origin and trajectory of drifting Arctic sea ice. Journal of Geophysical Research 102 C6):12575–12586

Plant GW, Covil CS, Hughes RA (eds) (1998). Site preparation for the New Hong Kong International Airport. Thomas Telford, London

Plummer LN (1975) Mixing of seawater with calcium carbonate ground water. In: Whitten EHT (ed) Quantitative studies in the geological sciences. Geological Society of America Memoir 85, pp 219–238

Proshutinsky A, Polyakov IV, Johnson MA (1999) Climate states and variability of Arctic ice and water dynamics during 1946–1997. Polar Research 18:1–8

Psuty NP (1992) Spatial variation in coastal foredune development. In Carter RWG, Curtis TGF, Sheehy-Skeffington MJ (eds) Coastal dunes. AA Balkema, Rotterdam, pp 3–13

Pye K (1983) Coastal dunes. Progress in Physical Geography 7(4): 531–557

Pye K, Bowman GM (1984) The Holocene marine transgression as a forcing function in episodic dune activity on the eastern Australian coast. In: Thom BG (ed) Coastal geomorphology in Australia. Academic Press, pp 179–196

Qian CL (1994) The impacts of water withdrawal from Luanhe River on the delta. Acta Geographica Sinica 49:158–166 (in Chinese)

Rabalais NN, Turner RE, Dortch Q, Justic D, Bierman VJ, Wiseman WJ (2002) Nutrient-enhanced productivity in the northern Gulf of Mexico: past, present and future. Hydrobiologia 475/476:39–63

Rachold V, Grigoriev MN, Are FE, Solomon S, Reimnitz E, Kassens H, Antonow M (2000) Coastal erosion vs riverine sediment discharge in the Arctic Shelf seas. International Journal of Earth Sciences 89:450–460

Raynaud D, Barnola J-M, Chappellaz J, Blunier T, Indermühle A, Stauffer B (2000) The ice record of greenhouse gases: a view in the context of future changes. Quaternary Science Reviews 19:9–17

Reay WG, Gallagher DL, Simmons GM (1992) Groundwater discharge and its impact on surface water quality in a Chesapeake Bay inlet. Water Research Bulletin 28:1121–1134

Reich CD, Shinn EA, Hickey TD, Tihansky AB (2002) Tidal and meteorological influences on shallow marine groundwater flow in the upper Florida Keys. In: Porter JW, Porter KG (eds) The Everglades, Florida Bay, and coral reefs of the Florida Keys, CRC Press, Boca Raton, pp 659–676

Reimnitz E, McCormick M, McDongall K, Brouwers E (1993) Sediment export by ice rafting from a coastal polynya, Arctic Alaska, USA. Arctic and Alpine Research 25:83–98

Restrepo JD, Kjerfve B, Correa ID, González JL (2002) Morphodynamics of a high discharge tropical delta, San Juan River, Pacific coast of Colombia. Marine Geology 192: 355–381

Riedl R, Huang N, Machan R (1972) The subtidal pump: a mechanism of interstitial water exchange by wave action. Marine Biology 13:210–221

Ritchie W, Penland S (1990) Aeolian sand bodies of the south Louisiana coast. In: Nordstrom KF, Psuty NP, Carter RWG (eds) Coastal dunes form and process. John Wiley and Sons, London, pp 105–127

Romankevich EA, Vetrov AA, Vinogradov ME, Vedernikov VI (2000) Carbon cycling components in the Arctic seas of Russia: carbon fluxes from land, carbon in the bottom sediments, elements of balance. Oceanology 40:335–344

Rothrock DA, Yu Y, Maykut GA (1999) Thinning of the Arctic sea-ice cover. Geophysical Research Letters 26:3469–3472

Rouse WR (2000) The energy and water balance of high-latitude wetlands: controls and extrapolation. Global Change Biology 6: 59–68

Runnels DD (1969) Diagenesis, chemical sediments, and mixing of natural waters. Journal of Sedimentary Petrology 39:1188–1201

Sabhasri S, Suwanarat K (1996) Impact of sea level rise on flood control in Bangkok and vicinity. In: Milliman JD, Haq BU (eds) Sea-level rise and coastal subsidence: causes, consequences and strategies. Kluwer Academic Publishers, Dordrecht, The Netherlands, pp 343–356

Saito Y (2001) Deltas in Southeast and East Asia: their evolution and current problems. In: Mimura N, Yokoki H (eds) Global change and Asia Pacific coast. Proceedings of APN/SURVAS/LOICZ joint conference on coastal impacts of climate change and adaptation in the Asia-Pacific Region, APN, Kobe, Japan, pp 185–191

Saito Y, Ikehara K, Katayama H, Matsumoto E, Yang Z (1994) Course shift and sediment discharge changes of the Huanghe recorded in sediments of the East China Sea. Chishitsu News 476:8–16

Saito Y, Yang ZS, Hori K (2001) The Huanghe (Yellow River) and Changjiang (Yangtze) deltas: a review on their characteristics, evolution and sediment discharge during the Holocene. Geomorphology 41:219–231

Salisbury JE, Campbell JW, Meeker LD, Vörösmarty C (2001) Ocean color and river data reveal fluvial influence in coastal waters. EOS 82(20):226–227

Salisbury JE, Campbell JW, Linder EL, Meeker D, Müller-Karger FE, Vörösmarty CJ (2004) On the seasonal correlation of surface particle fields with wind stress and Mississippi discharge in the northern Gulf of Mexico. Deep Sea Research II 51:1187–1203

Sánchez-Arcilla A, Jiménez JA, Valdemoro HI (1998) The Ebro delta: morphodynamics and vulnerability. Journal of Coastal Research 14(3):754–772

Saunders KE, Davidson-Arnott RGD (1990) Coastal dune response to natural disturbances. Canadian Symposium on Coastal Sand Dunes, pp 321–345

Schanck EJ, Fasano JL, Lanferedi NW, Pousa JL (1990) Impacts of sea level rise on the Argentine coast. In: Changing climate and the coast. IPCC Report 2, UNEP-WMOO-USACE-EPA-NOAA, pp 363–384

Schlager W (1999) Scaling of sedimentation rates and drowning of reefs and carbonate platforms. Geology 27:183–186

Semiletov IP (1999) On aquatic sources and sinks of CO_2 and CH_4 in the Polar regions. Journal of Atmospheric Sciences 56:286–306

Semiletov IP, Savelieva NI, Weller GE (2001) Cause and effect linkages between atmosphere, the Siberian rivers and conditions in the Russian Shelf seas. In: Semiletov IP (ed) Changes in the atmosphere-land-sea system in the Amerasian Arctic, pp 63–96

Serreze MC, Walsh JE, Chapin FSI, Osterkamp T, Dyurgerov M, Romanovsky V, Oechel WC, Morison J, Zhang T, Barry RG (2000) Observational evidence of recent change in the northern high-latitude environment. Climate Change 46:159–209

Shennan I, Woodworth PL (1992) A comparison of late Holocene and twentieth-century sea level trends from the UK and North Sea region. Geophysical Journal International 109:96–105

Short AD (1987) Modes, timing and volume of Holocene cross-shore and aeolian sediment transport, Southern Australia. ASCE, Proceedings of Coastal Sediments '87, pp 1925–1937

Short AD (1988a) Response of a high through low energy sandy coast to sea-level transgression and stillstand, South Australia. Geographical Review 78:119–136

Short AD (1988b) Holocene coastal dune formation in southern Australia – a case study. Sedimentary Geology 55:121–142
Shulmeister J, Short SA, Price DM, Murray AS (1993) Pedogenic uranium/thorium and thermoluminescence chronologies and evolutionary history of a coastal dunefield, Groote Eylandt, northern Australia. Geomorphology 8:47–64
Shum KT, Sundby B (1996) Organic matter processing in continental shelf sediments – the subtidal pump revisited. Marine Chemistry 53:81–87
Simmons GM Jr (1992) Importance of submarine groundwater discharge (SGWD) and seawater cycling to material flux across sediment/water interfaces in marine environments. Marine Ecology Progress Series 84:173–184
Simmonds I, Keay K (2000) Variability of southern hemisphere extratropical cyclone behavior, 1958–97. Journal of Climate 13: 550–561
Skirving W, Guinotte J (2001) The sea surface temperature story on the Great Barrier Reef during the coral bleaching event of 1998. In Wolanski E (ed) Oceanographic processes of coral reefs. Physical and biological links in the Great Barrier Reef. CRC Press LLC, Boca Raton, Florida, pp 301–309
Smith SV, Renwick WH, Buddemeier RW, Crossland CJ (2001) Budgets of soil erosion and deposition for sediment and sedimentary organic carbon across the conterminous United States. Global Biogeochemical Cycles 15(3):697–707
Smithers SG, Woodroffe CD (2000) Microatolls as sea-level indicators on a mid-ocean atoll. Marine Geology 168:61–78
Snoussi M, Haida S, Imassi S (2002) Effects of the construction of dams on the Moulouya and the Sebou rivers (Morocco). Regional Environmental Change 3:5–12
Solomon SM, Forbes DL, Kierstead B (1994) Coastal impacts of climate change: Beaufort Sea erosion study. Canadian Climate Centre Report 94-2. Environment Canada, Downsview Ontario, Canada
Speer PE, Aubrey DG. (1985) A study of non-linear tidal propagation in shallow inlet/estuarine systems, Part II: Theory. Estuarine, Coastal and Shelf Science 21:207–224
Spencer T, Tudhope AW, French JR, Scoffin TP, Utanga A (1997) Reconstructing sea level change from coral microatolls, Tongareva (Penryn) Atoll, Northern Cook Islands. Proceedings of the 8th International Coral Reef Symposium 1:489–494
Stanley DJ, Warne AG (1993) Nile Delta: recent geological evolution and human impact. Science 260:628–634
Steffen W, Jager J, Carson DJ, Bradshaw C eds. (2003) Challenges of a changing earth. Proceedings of the Global Change Open Science Conference, Amsterdam, The Netherlands, 10–13 July 2001. Springer Verlag, Heidelberg, Germany
Stierle AP, Eicken H (2002) Sedimentary inclusions in Alaskan coastal sea ice: small-scale distribution, interannual variability and entrainment requirements. Arctic Antarctic and Alpine Research 34:465–476
Stoddart DR (1969) World erosion and sedimentation. In: Chorley RJ (ed) Water, earth and man. Methuen Co. Ltd, London, pp 43–64
Sunrise Technical Consultants (1968) Estudio mediante ensayo hidráulico sobre modelo del Puerto de Mar del Plata y sus alrededores. vol I–VIII, Tokyo, Japan
Syvitski JPM (2002) Sediment transport variability in Arctic Rivers: implications for a warmer future. Polar Research 21(2):323–330
Syvitski JPM (2003a) Sediment fluxes and rates of sedimentation. In: Middleton GV (ed) Encyclopedia of sediments and sedimentary rocks. Kluwer Academic Publishers, Dordrecht, The Netherlands, pp 600–606
Syvitski, JPM (2003b) The influence of climate on the flux of sediment to the coastal ocean. Proceedings of OCEANS 2003, San Diego, Holland Publishing, pp 496–502
Syvitski, JPM (2003c) Supply and flux of sediment along hydrological pathways: research for the 21st century. Global and Planetary Change 39:1–11
Syvitski JPM, Morehead MD (1999) Estimating river-sediment discharge to the ocean: application to the Eel Margin, northern California. Marine Geology 154:13–28
Syvitski JPM, Nicholson M, Skene K, Morehead MD (1998) PLUME1.1: Deposition of sediment from a fluvial plume. Computers and Geoscience 24(2):159–171
Syvitski, JPM, Peckham SD, Hilberman RD, Mulder T (2003) Predicting the terrestrial flux of sediment to the global ocean: a planetary perspective. Sedimentary Geology 162:5–24
Ta TKO, Nguyen VL, Tateishi M, Kobayashi I, Tanabe S, Saito Y (2002) Holocene delta evolution and sediment discharge of the Mekong River, southern Vietnam. Quaternary Science Reviews 21: 1807–1819
Taniguchi M, Burnett WC, Cable JE, Turner JV (2002) Investigation of submarine groundwater discharge. Hydrological Processes 16:2115–2129
Thibodeaux LJ, Boyle JD (1987) Bedform-generated convective transport in bottom sediment. Nature 325:341–343
Thieler RE, Hammar-Klose ES (1999) National assessment of coastal vulnerability to sea-level rise: US Atlantic Coast. USGS, Woods Hole, MA
Thieler ER, Hammar-Klose ES (2000) National assessment of coastal vulnerability to sea level rise: preliminary results for the US Gulf of Mexico coast. USGS Open-File Report 00-179
Thom BG, Bowman GM, Roy PS (1981) Late Quaternary evolution of coastal sand barriers, Port Stephens-Myall Lakes area, central NSW, Australia. Quaternary Research 15:345–364
Thom BG, Shepherd MJ, Ly C, Roy P, Bowman GM, Hesp PA (1992) Coastal geomorphology and Quaternary geology of the Port Stephens- Myall Lakes area. Department of Biogeography and Geomorphology, ANU, Monograph No. 6, ANUTech PL, Canberra
Thom BG, Hesp PA, Bryant T (1994) Last glacial "coastal dunes" in eastern Australia and implications for landscape stability during the Last Glacial Maximum. Palaeogeography, Palaeoclimatology, Palaeoecology 111:229–248.
Thorstenson DC, Mackenzie FT (1974) Time variability of porewater chemistry in recent carbon sediments, Devil's Hole, Harrington, Sound Bermuda. Geochemica Cosmochimica Acta 38:1–19
Titus JG (1998) Rising seas, coastal erosion and the takings clause: how to save wetlands and beaches without hurting property owners. Maryland Law Review 57:1279–1399
Titus JG, Richman C (2001) Maps of lands vulnerable to sea level rise: modeled elevations along the US Atlantic and Gulf Coasts. Climate Research 18:205–228
Tovey NK, Yim WW-S (2002) Desiccation of late Quaternary inner shelf sediments of Hong Kong: microfabric observations. Quaternary International 92:73–87
Trupin A, Wahr J (1990) Spectroscopic analysis of global tide gauge sea level data. Geophysical Journal International 100:441–453
Tynan CT, Demaster DP (1997) Observations and predictions of Arctic climatic change: potential effects on marine mammals. Arctic 50:308–322
Umezawa Y, Miyajima T, Kayanne H, Koike I (2002) Significance of groundwater nitrogen discharge into coral reefs at Ishigaki Island, southwest of Japan. Coral Reefs 21(4):346–356
UNEP (1998) East Africa atlas of coastal resources – Kenya. UNEP, Nairobi
UNEP (2001) East Africa Atlas of Coastal Resources – Tanzania. UNEP, Nairobi
Valiela I, Costa J, Foreman K, Teal JM, Howes B, Aubrey D (1990) Transport of groundwater-borne nutrients from watersheds and their effects on coastal waters. Biogeochemistry10:177–197
Vörösmarty CJ, Meybeck M, Fekete B, Sharma K (1997) The potential impact of neo-Castorization on sediment transport by the global network of rivers. In: Walling DE, Probst J-L (eds) Human impact on erosion and sedimentation. IAHS Publication No. 245, pp 261–273
WASA (1998) Changing waves and storms in the Northeast Atlantic. Bulletin of the American Meteorological Society 79:741–760
Walling DE, Fan D (2003) Recent trends in the suspended sediment loads of the world's rivers. Global and Planetary Change 39: 111–126
Walther G-R, Post E, Convey P, Menzel A, Parmesan C, Beebee TJC, Fromentin J-M, Hoegh-Guldberg O, Bairlein F (2002) Ecological responses to recent climate change. Nature 416:389–395
Wang Y, Ren M-e, Syvitski JPM (1998) Sediment transport and terrigenous fluxes. In: Brink KH, Robinson AR (eds) The Sea; vol 10: the global coastal ocean: processes and methods. John Wiley & Sons, New York, pp 253–292

Wasson RJ (1996) Land use and climate impacts on fluvial systems during the period of agriculture. PAGES Workshop Report, Series 96-2

Watson RT, Zinyowera MC, Moss RH, Dokken DJ (1996) Technical summary: impacts, adaptations and mitigation options. In: Watson RT, Zinyowera MC, Moss RH (eds) Climatic change 1995 – impacts, adaptations and mitigation of climate change: scientific-technical analyses. Preparing for a changing climate: the potential consequences of climate variability and change – mid Atlantic foundations. January 2001. Cambridge University Press, New York, pp 19–53

Watson RT, Zinyowera MC, Moss RH, Dokken DJ (1998) The regional impacts of climate change: an assessment of vulnerability. A special report of IPCC Working Group II, Intergovernmental Panel on Climate Change, Cambridge University Press

Webb JE, Theodor J (1968) Irrigation of submerged marine sands through wave action. Nature 220:682–685

Wells JT (1996) Subsidence, sea-level rise, and wetland loss in the lower Mississippi River delta. In: Milliman JD, Haq BU (eds) Sea-level rise and coastal subsidence. Kluwer Academic Publishing, Dordrecht, The Netherlands, pp 281–311

Wheeler PA, Gosselin M, Sherr E, Thibault D, Kirchman DL, Benner R, Whitledge TE (1996) Active cycling of organic carbon in the central Arctic ocean. Nature 380:697–699

Wheeler PA, Watkins JM, Hansing RL (1997) Nutrients, organic carbon and organic nitrogen in the upper water column of the Arctic Ocean: implications for the sources of dissolved organic carbon. Deep-Sea Research (II) 44:1571–1592

Wiberg PL, Harris CK (1994) Ripple geometry in wave-dominated environments. Journal of Geophysical Research – Oceans 99:775–789

Wigley TLM, Raper SCB (1993) Future changes in global mean temperature and sea level. In: Warrick RA, Barrow EM, Wigley TML (eds) Climate and sea level change: observations, projections and implications. Cambridge University Press, pp 111–135

Wilkinson C (ed) (2000) Status of coral reefs of the world: 2000. Australian Institute of Marine Science, Townsville

Wolanski E (ed) (2001) Oceanographic processes of coral reefs: physical and biological links in the Great Barrier Reef. CRC Press, Florida

Wolanski E, Chappell J (1996) The response of tropical Australian estuaries to a sea level rise. Journal of Marine Systems 7:267–279

Wolanski E, Duke N (2002) Mud threat to the Great Barrier Reef of Australia. In: Healy T, Wang Y, Healy J-A (eds) Muddy coasts of the world: processes, deposits and function. Elsevier Science BV, Amsterdam, pp 533–542

Wolanski E, Spagnol S (2000) Environmental degradation by mud in tropical estuaries. Regional Environmental Change 1(3–4):152–162

Wolanski E, Spagnol S (2003) Dynamics of the turbidity maximum in King Sound, tropical Western Australia. Estuarine, Coastal and Shelf Science 56:877–890

Wolanski E, King B, Galloway D (1995) Dynamics of the turbidity maximum in the Fly River estuary, Papua New Guinea. Estuarine, Coastal and Shelf Science 40:321–337

Wolanski E, Huan NN, Dao LT, Nhan NH, Thuy NN (1996) Fine sediment dynamics in the Mekong River estuary, Vietnam. Estuarine, Coastal and Shelf Science 43:565–582

Wolanski E, Nhan NH, Spagnol S (1997) Sediment dynamics during low flow conditions in the Mekong River estuary, Vietnam. Journal of Coastal Research 13:1241–1251

Wolanski E, Moore K, Spagnol S, D'Adamo N, Pattiaratchi C (2001) Rapid, human-induced siltation of the macro-tidal Ord River estuary, Western Australia. Estuarine, Coastal and Shelf Science 53:717–732

Wolanski E, Richmond RH, Davis G, Bonito V (2003a) Water and fine sediment dynamics in transient river plumes in a small, reef-fringed bay, Guam. Estuarine, Coastal and Shelf Science 56:1–13

Wolanski E, Richmond R, McCook L, Sweatman H (2003b) Mud, marine snow and coral reefs. American Scientist 91(1):44–51

Wolanski E, Boorman LA, Chicharo L, Langlois-Saliou E, Lara R, Plater AJ, Uncles RJ, Zalewski M (2004) Ecohydrology as a new tool for sustainable management of estuaries and coastal waters. Wetlands Ecology and Management 12(4):235–276

Woodroffe CD (1990) The impact of sea-level rise on mangrove shorelines. Progress in Physical Geography 14:483–520

Woodroffe C, Gagan MK (2000) Coral microatolls from the Central Pacific record late Holocene El Nino. Geophysical Research Letters 27:1511–1514

Woodroffe C, Mclean R (1990) Microatolls and recent sea-level change on coral atolls. Nature 344:531–534

Woodroffe CD, Chappell JMA, Thom BG, Wallensky E (1986) Geomorphological dynamics and evolution of the South Alligator tidal rivers and plains, Northern Territory. Australian National University, Northern Australia Research Unit, Mangrove Monograph 3, Darwin

Woodward JC (1995) Patterns of erosion and suspended sediment yield in Mediterranean river basins. In: Foster IDL, Gurnell AM, Webb BW (eds) Sediment and water quality in river catchments. John Wiley & Sons Ltd, pp 365–389

Woodworth PL (1993) Sea level changes. In Warrick RA, Barrow EM, Wigley TML (eds) Climate and sea level change: observations, projections and implications. Cambridge University Press, pp 379–391

Woodworth PL, Tsimplis MN, Flather RA, Shennan I (1999) A review of the trends observed in British Isles mean sea level data measured by tide gauges. Geophysical Journal International 136: 651–670

Wyatt A (1986) Challinor's Dictionary of Geology, 6th edn. University of Wales Press, Cardiff

Yang ZS, Saito Y, Liu B, Zhang J, Wang HJ (2002) Decadal and millennial time-scale changes of water and sediment discharges of the Huanghe (Yellow River) caused by human activities. In: Hong GH, Kremer HH, Pacyna J, Chen C-TA, Behrendt H, Salomons W and Marshall Crossland JI (2002) East Asia Basins: LOICZ Global assessment and synthesis of river catchment-coastal sea interaction and human dimensions. LOICZ Reports & Studies No. 26, LOICZ, Texel, The Netherlands, pp 118–121

Yim WW-S (1999) Radiocarbon dating and the reconstruction of late Quaternary sea-level changes. Quaternary International 55: 77–91

Yim WW-S (2001) Stratigraphy of Quaternary offshore sand and gravel deposits in the Hong Kong SAR, China. Quaternary International 82:101–116

Yim WW-S, Choy AMSF (2000) Some engineering applications of Quaternary sea-level changes in Hong Kong. In: Proceedings, Conference on Engineering Geology Hong Kong 2000, Hong Kong. The Institution of Mining & Metallurgy, London – Hong Kong Branch, 10 November 2000, pp 211–218

Yim WW-S, Li Q-Y (1983) Sea-level changes and sea-floor surficial deposits off Chek Lap Kok. In: Yim WW-S, Burnett AD (eds) Geology of surficial deposits in Hong Kong. Abstracts No. 1, Geological Society of Hong Kong and Department of Geography & Geology, The University of Hong Kong, Hong Kong, pp 48–59

Yim WW-S, Ivanovich M, Yu K-F (1990) Young age bias of radiocarbon dates in pre-Holocene marine deposits of Hong Kong and implications for Pleistocene stratigraphy. Geo-Marine Letters 10: 165–172

Yim WW-S, Price DM, Choy AMSF (2002) Distribution of moisture contents and thermoluminescence ages in an inner shelf borehole from the new Hong Kong International Airport. Quaternary International 92:35–43

Young R, Bryant EA, Price DM, Wirth LM, Pease M (1993) Theoretical constraints and chronological evidence of Holocene coastal development in central and southern NSW, Australia. Geomorphology 7:317–329

Zektser IS (2000) Groundwater and the environment: applications for the global community. Lewis Publishers, Boca Raton

Zhang Y, Wang WC (1997) Model-simulated northern winter cyclone and anticyclone activity under a greenhouse warming scenario. Journal of Climate 10:1004–1020

Ziebis W, Huettel M, Forster S (1996) Impact of biogenic sediment topography on oxygen fluxes in permeable seabeds. Marine Ecology-Progress Series 140:227–237

2.2.3.3 Management Response to Coastal Vulnerability

The IPCC approach to coastal adaptation (IPCC 1992) outlines three societal response options:

a ***Retreat*** – either a forced or a managed retreat where no attempt is made to protect the coast or property. This could involve the relocation of houses, for example, or the abandonment of various coastal land uses.
b ***Accommodate*** – an adjustment to the higher sea level, such as elevating coastal properties or changing land use to more compatible industries such as aquaculture.
c ***Protect*** – maintain the current land use with some form of engineering response involving either hard protection (e.g., seawalls) or soft protection (e.g., beach nourishment).

A summary of the various response options for open coasts and estuaries has been prepared for New Zealand coastal managers (Bell et al. 2001; Fig. 2.5). Various countries have adaptation strategies through managed retreat, including setback distances, no-build zones, rolling easements and managed realignments (McCarthy et al. 2001). South Australia has a policy based on earlier IPCC sea-level rise predictions (IPCC 1991, Harvey and Belperio 1994), i.e., that new coastal development should be capable of being reasonably protected from a 1 m sea-level rise by 2100. It recommends that site and building levels should be 0.3 m above the 100-year ARI water level and adjusted to allow for localised subsidence or uplift. Building floor levels should be an additional 0.25 m above this level, and approvals should depend on the capability to protect from a further 0.7 m of sea-level rise (e.g., by means of a bund wall or raising the building). In the case of flood-protected areas, the 100-year ARI design flood level for the development area must incorporate the extreme tide (plus surge) and storm-water events, together with wave effects. The policy also makes a general recommendation for an erosion setback distance based on 100 years of erosion (or 200 years for major development) at a site, allowing for local coastal processes, a sea-level rise of 0.3 m by 2050 and storm erosion from a series of severe storms. The evolving circumstances and management approaches for New York City, USA (see Text Box 2.6) add emphasis to the awareness about sea-level rise required in urban planning.

A distinction needs to be drawn between natural coastal vulnerability and the vulnerability of human lives and property to the effects of climate change and sea-level rise. Some coasts such as crystalline cliffed coasts may be resistant to these impacts, while coral reef or wetland coasts may respond naturally by accretion. Others such as barrier coasts may naturally migrate inland, but where humans or their property are at risk (e.g., on the US east coast) it is necessary to have societally-focused adaptation strategies (Titus 1998). In many cases, attempts to use hard protection have exacerbated the problem because of a lack of understanding of coastal processes (Doornkamp 1998, Pethick 2001) but there appears to be the beginning of a renaissance in concepts and policies for coastal management (for example, de Vries 2001).

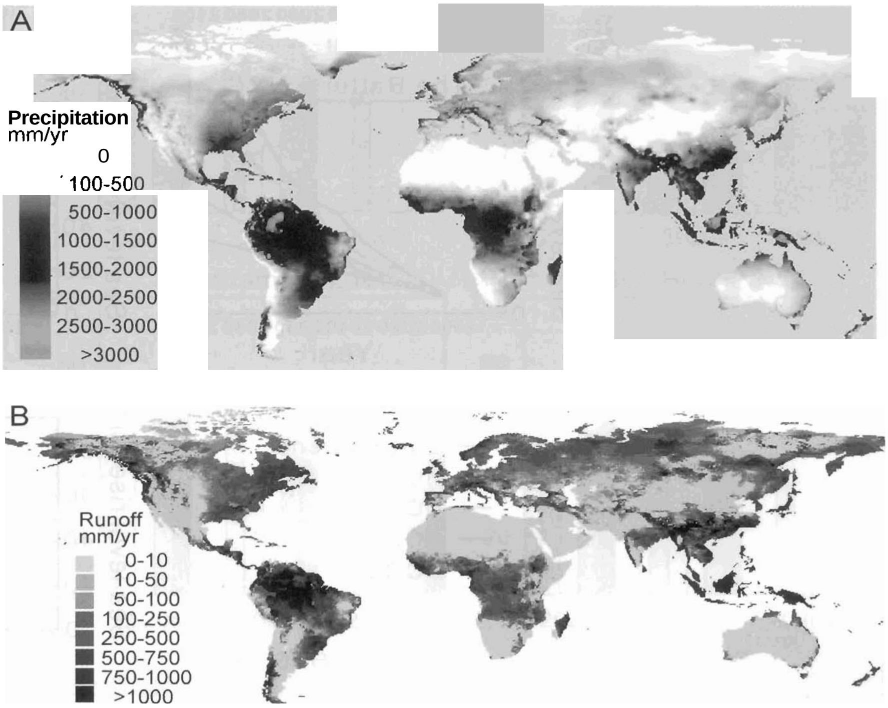


Fig. 2.6. Water flux. **a** Global precipitation (mm yr^{-1}) (Syvitski et al. 2003). **b** Hydrological runoff (mm yr^{-1}) after accounting for all forms of evapo-transpiration and human-induced consumption. The hydrological runoff divided by the drainage area equals the water discharge ($km^3 yr^{-1}$) (http://www. bafg.de/ *grdc.htm*)