C

CARBON SEQUESTRATION

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Synonyms

Carbon capture; Carbon storage

Definition

Carbon sequestration is the process by which atmospheric carbon dioxide (CO_2), the most important greenhouse gas, is removed from the atmosphere and stored in the ocean, on the land surface, or in geological formations (Sundquist et al., 2008). Reservoirs that store carbon over long periods of time are called "carbon sinks." The process of carbon sequestration can occur naturally by plants via photosynthesis with subsequent storage of carbon in biomass (leaves, roots, and stems/trunks of plants) and soils. Carbon can also be sequestered by separating and capturing CO_2 emitted by industrial processes and transporting it to deep underground geological formations for permanent storage (Lal, 2008). Carbon sequestration is reported as a rate of carbon (C) storage in units of mass per time such as teragrams ($Tg = 1 \times 10^{12}$ g) C/year.

Bibliography

Lal, R., 2008. Carbon sequestration. *Philosophical Transactions of the Royal Society B Biological Sciences*, **363**, doi:10.1098/rstb.2007.2185.

Sundquist, E., Burrus, R., Faulkner, S., Gleason, R., Harden, J., Kharaka, Y., Tieszen, L., and Waldrop, M., 2008. Carbon sequestration to mitigate climate change. *U.S. Geological Survey Fact Sheet 2008–3097*. http://pubs.usgs.gov/fs/2008/3097/pdf/CarbonFS.pdf.

Cross-references

Saltmarshes

CHENIERS AND REGRESSIVE BEDFORMS

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Synonyms

Chenier plain; Sandy beach ridge

Definition

Chenier is a term devoted exclusively to linear sandy coastal ridges separated from the shoreline by muddy deposits (mudflats or marshes). The base of the cheniers should be a horizontal surface where a clear difference exists between the mud and the coarse-grained deposits. The seaward sides of the cheniers are regular and straight, while the landward side is irregular (Reineck and Singh, 1980). Cheniers do not occur alone, but in groups, with narrow spacing between individual ridges at their ends. Moreover, this arrangement can cause streams and small rivers to be deflected, creating angular or parallel drainage patterns, a feature that was noted in the Suriname chenier plain by Augustinus (1980).

Introduction

Cheniers are deposits originating from high-energy reworking of muddy coastal plains. Beach-ridge and chenier plains can be confused when it is difficult to discern if the plains are dominated by sand and mud.

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Cheniers were originally defined as shallow-based, sandy beach ridges resting on clay along a marshy or swampy, seaward facing, tidal shore, with other beach ridges stranded in a marsh behind, forming a belted marsh-and-ridge plain (Price, 1954, 1955), and usually enriched in up to pebble-size shelly material (Otvos, 2000). In other words, beach-ridge plains have local names in Louisiana (cheniers) and Suriname (ritsen). The term chenier was given to the ridges with oaks growing on them, which are called "chênes" in Southwest Louisiana. The chenier plain comprised of the vegetated marshes, water bodies (including lakes, streams, and tidal inlets), and beach ridges (Byrne et al., 1959). However, these ridges possess certain characteristics. They are less than 3 m high, 30–50 m wide, and no more than 5 m thick. The growth of these cheniers was clearly related to the mud delivered by the Atchafalaya River (Wells and Kemp, 1981).

The link between cheniers and fluvial input has been applied to other deltas or river estuaries. Since the origin of the term, they have been reported at the deltas of the Mississippi, Amazon, Orinoco, Po, and Rhone Rivers (Price, 1954). Wells and Coleman (1977) extended their studies of the suspended sediment transported from the Amazon and Orinoco deltas to the chenier plains of Suriname (Wells and Coleman, 1981). Their model related chenier growth to the number of days when the tide exceeded a certain level, the increase in sediment concentration, the decrease in sediment compaction, and the root density of mangroves at the coastline (Wells and Coleman, 1981).

In recent times, beach ridges are being used as a common term for coastal features originating from several processes: (1) swash action, (2) settling lag, (3) eolian action, (4) and storm surges (Tanner, 1995). They are also related to the episodic input of sediment, either of fluvial (Anthony, 1995) or volcanic origin (Nieuwenhuyse and Kroonenberg, 1994). They can be composed of sand, gravel, or shells (Reineck and Singh, 1980).

Sea-level trend

Although they were originally defined for low-lying deltaic plains affected by rising sea levels, as along the Mississippi River, cheniers are more abundant on plains of regressive coasts subject to periodic or episodic high-energy levels. The mid-Holocene was the time when the alluvial plains became stable (Xiqing, 1996). This occurred around the same time as the mid-Holocene sea-level maximum, which varied from less than 1 m to around 4 m above the current level in the Southern Hemisphere (Isla, 1989). However, this highstand was not uniform or stable. In the Northern Hemisphere, sea level is still rising, while in the Southern Hemisphere, it has been dropping slightly. The differences of 2–3 m in the last 5,000 years have produced significant changes in bedforms developing in estuaries. Morphodynamic models are mostly biased toward transgressive coasts (Tanner, 1995; Hesp and Short, 1999). Although progradational barriers are assumed to have a higher potential of preservation on regressive (falling sea level) coasts, not many models have been proposed for these coasts (Roy et al., 1995). Sediment availability controls the facial relationships either in transgressive or regressive sequences (Davis and Clifton, 1987; Isla, 1998).

Location

Cheniers have been observed in every estuary subject to episodic processes or where the availability of coarse material is episodic. They have been found in every continent, with the exception of Antarctica, from low to high latitudes (Figure 1). Therefore, climate only affects its composition and the type of plant communities (e.g., mangroves or salt marshes, in low or mid-high latitudes, respectively).

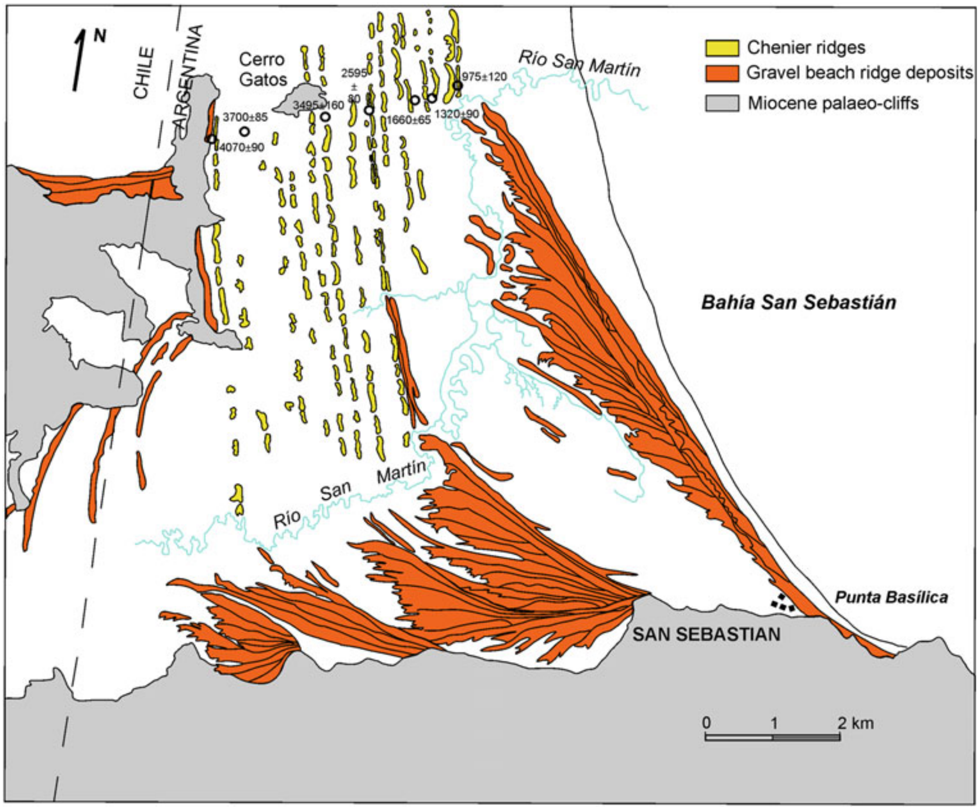
Origin and composition

As their origin implies high energy, cheniers are composed of different materials. Sandy cheniers are quite common, although they can also be composed of shells and gravel. Cheniers have been divided in medium- to coarse-sandy cheniers and fine-sandy cheniers, each one with different accumulation processes, depending on the mode of sandy supply.

Chenier ridges are formed by the interplay between washover and beach drift processes. Sand in low- to medium-energy coasts is effectively transported by constructive wave action. The grains are suspended by the turbulence of the breaking waves and transported by beach drift. During the rising tide, the sediment is stirred at the seaside of a developing chenier by approaching breakers. If the crest is low, sand will be washed over it and deposited on the lee side, causing the chenier to migrate gradually landward. Once the crest is high enough, beach drift becomes the dominant process, and the chenier begins to extend laterally, depending on the direction of the current (Augustinus, 1980).

Another mechanism of chenier formation is the switching of delta lobes (Otvos and Price, 1979; Penland and Suter, 1989). This is particularly evident in the variations of the Huang He (Yellow) River outlet to the North Jiangsu of Bohai Bay (Xitao, 1989; Yan et al., 1989).

Medium- to coarse-sandy cheniers are built up by sand delivered by longshore currents, beach drift, and washover processes. Thus, they form at or just above high-tide level. Sedimentary structures within coarse-textured cheniers include lamination in two different directions – foreslope- and backslope-parallel laminations. The foreslope-parallel lamination occupies a narrow strip on top of the chenier, with the rest of the chenier's body composed of the landward-parallel lamination. Cross-stratification sets may appear intercalated with the parallel laminate depending on the water level landward of the chenier (e.g., in mangroves or salt pans). If the chenier is sufficiently high, small washover deltas develop due to the sudden slowing down of the running water containing



Cheniers and Regressive Bedforms, Figure 1 Location of cheniers (*triangles*) and beach-ridge plains (*circles*).

sediment as it encounters stagnant waters. This process is related to spring tides and ends before the following neap tide (Augustinus, 1980; Augustinus et al., 1989).

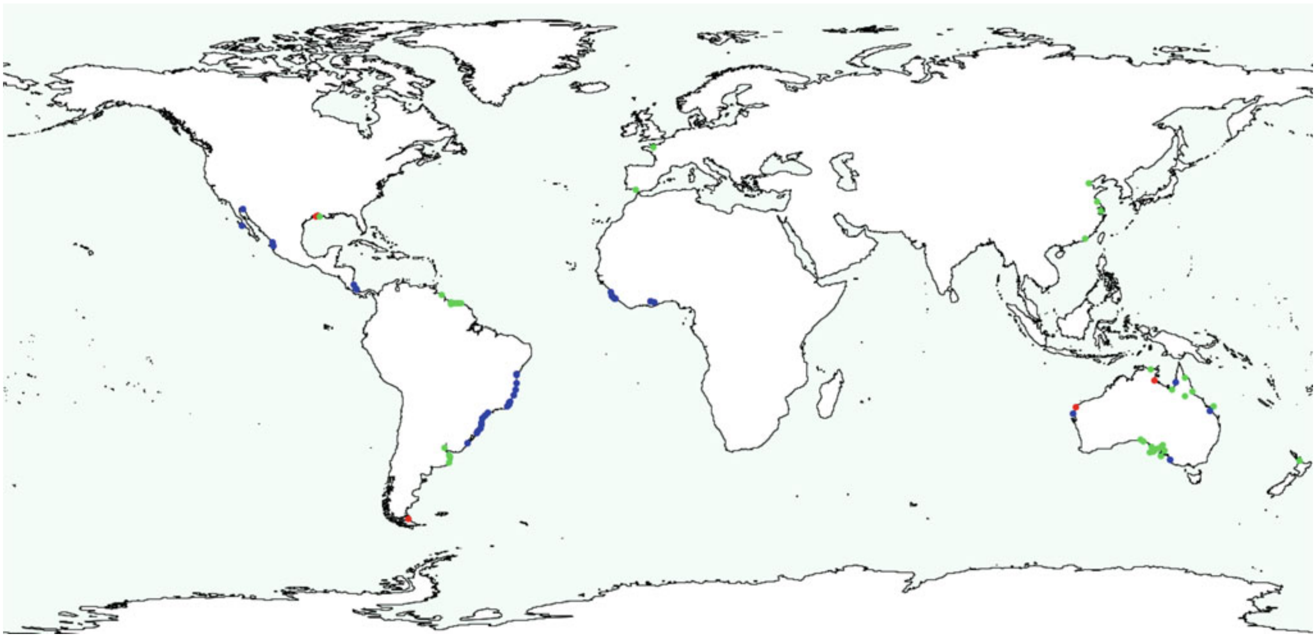
Fine-sandy cheniers originate at the mean low water level, where unsorted mudflat deposits are stirred up by waves and currents. Sand settles on the more tranquil near-shore waters, while the finer sediment is transported farther, forming a longshore bar that moves obliquely landward, gathering volume and height. When the up-current end attaches to the mainland, it appears similar to a spit, with a lagoon-like tidal flat behind it. The main body of the fine-sandy cheniers is composed of a thick bed set of gently seaward-dipping laminate with intercalated thin sets of coarser, steeply landward-dipping shelly layers. The lagoon-like tidal flat is characterized by interlayered sand/mud bedding (Augustinus, 1980; Augustinus et al., 1989).

The transition from beach ridges to cheniers is a matter of sediment availability. In San Sebastian Bay, Tierra del Fuego, a river runs between the boundary of a beach ridge composed dominantly of gravel and a chenier plain composed of shell debris in a sandy matrix and interfingered with mudflats and salt marshes (Isla et al., 1991). Although the beach-ridge plain is assumed to be older, the chenier plain developed between 4,070 and 975 years ago (Figure 2; Vilas et al., 1999). In a similar way, the beach-ridge plain expanded eastward during a period close to the maximum highstand of the headlands of the Rio de la Plata mid-Holocene embayment (Figure 3). About 2,800 years BP, the Parana Delta expanded south-eastward, restricting wave processes and increasing the mud availability. This caused a significant change, and a chenier plain that prograded until the deposition of the Parana and Uruguay Rivers generated a single lobe (Figure 3).

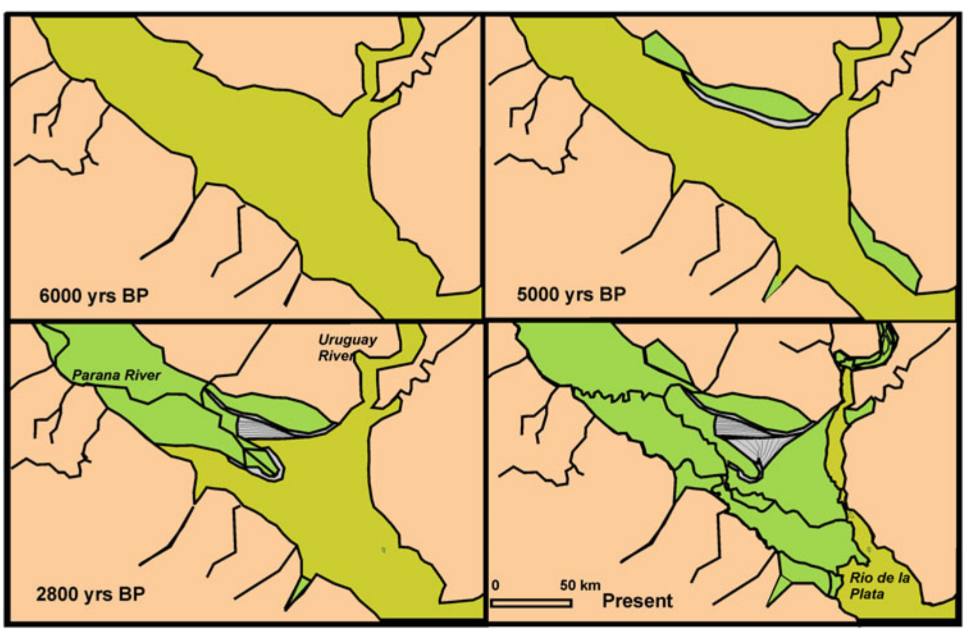
Along the Baja California Peninsula, the tide-dominated coast (macrotidal) has a chenier plain formed by deposition and starvation processes of the Colorado River Delta. On the other hand, the open-ocean coast of Vizcaino Peninsula (mesotidal) beach-ridge plains is associated with high tides and storms (Meldahl, 1995). Applying Ground Penetrating Radar (GPR) techniques, sand-dominated and mud-dominated plains have been differentiated at the Tijucas (meaning “marsh” or “swamp” in the Tupi language) River Inlet, Santa Catarina, Brazil (Buynevich et al., 2005).

Regressive bedforms

More strictly speaking, a chenier plain is composed of at least two subparallel ridges or ridge sets, “sandwiched” between tidal-subtidal mudflats. It represents multiple



Cheniers and Regressive Bedforms, Figure 2 Development of gravel and chenier plains of the San Sebastián Bay, Tierra del Fuego (Modified after Vilas et al., 1999).



Cheniers and Regressive Bedforms, Figure 3 Development of the Paraná-Uruguay composite delta (Modified from Cavalotto et al., 2005).

episodes of ridge and mudflat formation on prograding shore sectors. Mudflat progradation must bracket the ridge: a marsh-covered or barren preexisting mudflat in its rear, and a younger, possibly still active mudflat seaward (Otvos and Price, 1979; Otvos, 2000).

There are two main types of chenier plains depending on the type of coast in which they develop. Bight-coast chenier plains include those forms of the West Louisiana and Guiana coasts. In this setting, a broad indentation of the bight coast contributes to a decrease of wave energy approaching the nearshore and the settling out of muddy deposits. Bayhead chenier plains in turn are formed on bayheads of smaller dimensions than those on bight coasts. Classic bayhead chenier plains include Broad Sound and Burdekin River, Queensland (Cook and Polach, 1973), the Colorado Delta, California Gulf (Otvos and Price, 1979), Firth of Thames, New Zealand (Woodroffe et al., 1983), or at the headlands of the Rio de la Plata (Cavalotto et al., 2005). Other examples are San Sebastian (Vilas et al., 1999) and Samborombon Bays (Bértola, 1994) on the coast of Argentina. Recently, Spanish authors distinguished a third setting for chenier formation: estuarine coasts protected from the open sea by a coastal barrier (Rodríguez-Ramírez and Yáñez-Camacho, 2008). Tentatively, the ridges along the Mar Chiquita lagoon coast fall into this category.

Age

Radiocarbon dates from shells collected from cheniers do not relate directly to ridge accumulation but to the time of the organism's death before ridge formation (Woodroffe, 2002). As the maximum limit of the Holocene transgression occurred ~6,000 years ago, there are differences in the ages of the shells composing beach deposits. Shells sampled in living position from estuarine facies are better indicators of the maximum highstand (Isla, 1989; Cavalotto et al., 2005).

Schofield was the first to relate a chenier plain to the sea-level fall of ~2.6 m in 2,000 years at the North Island of New Zealand. The Miranda plain is composed of cheniers and regressive spits dated between 3,900 and 980 years BP (Schofield, 1960). Recently, the sequence of regressive spits was reanalyzed based on GPR records. Significant differences were recognized based on the architecture of the older ridges (13-6), related to mudflats, and the modern ridges (5-1) deposited on embayed tidal flats (Dougherty and Dickson, 2009).

After the maximum highstand of 5,800–5,500 years BP, a chenier plain developed as sea level dropped at the mesotidal Gulf of Carpentaria, Australia (Rhodes, 1982). At the eastern side of Queensland, another chenier plain developed in Princess Charlotte Bay. From a maximum highstand of ~6,000 years BP, beach ridges and cheniers composed of shells are interfingered (Chappell and Grindrod, 1984). Mangroves colonized cheniers in a prograding shoreline. Managing different chenier plains, significant changes in the progradation were detected over the last 6,000 years at Princess Charlotte Bay, Karumba, and South Alligator River (Woodroffe, 2002). At Southern Australia, another chenier plain composed of shells increased in size due to organic-rich mangrove facies (Short, 1988). This plain expanded to a highstand of 2–3 m above today's sea level ~6,400 years ago. The plain's growth was recorded using organic remains from the transition between *Posidonia* sea grass and sand flats and from these flats to ridges dominated by *Avicennia* mangroves (Belperio et al., 2002).

Cheniers can be used as reliable markers of paleocoastlines, since they only form on stable or slightly retreating coasts. When there is a steady sediment supply along with a gradually falling sea level, the result is an increased rate of mudflat progradation (Chappell and Grindrod, 1984), which prevents chenier formation or lowers their frequency of appearance. Moreover, an increased deposition of muds causes a reduction in shellfish communities, which in turn diminishes shell production and the quantity of the material available to form cheniers. On the northern outer shelf of the East China Sea, submerged chenier ridges were used to map a Pleistocene paleocoastline, with ages between 24,000 and 15,000 years BP. They were found at depths between 150 and 110 m (Xitao, 1989), which reflect the lowstand at the last glacial maximum. The Lelydorp member of the Coropina Formation (Suriname) consists of the deposits of an early Pleistocene chenier plain (Wong et al., 2009).

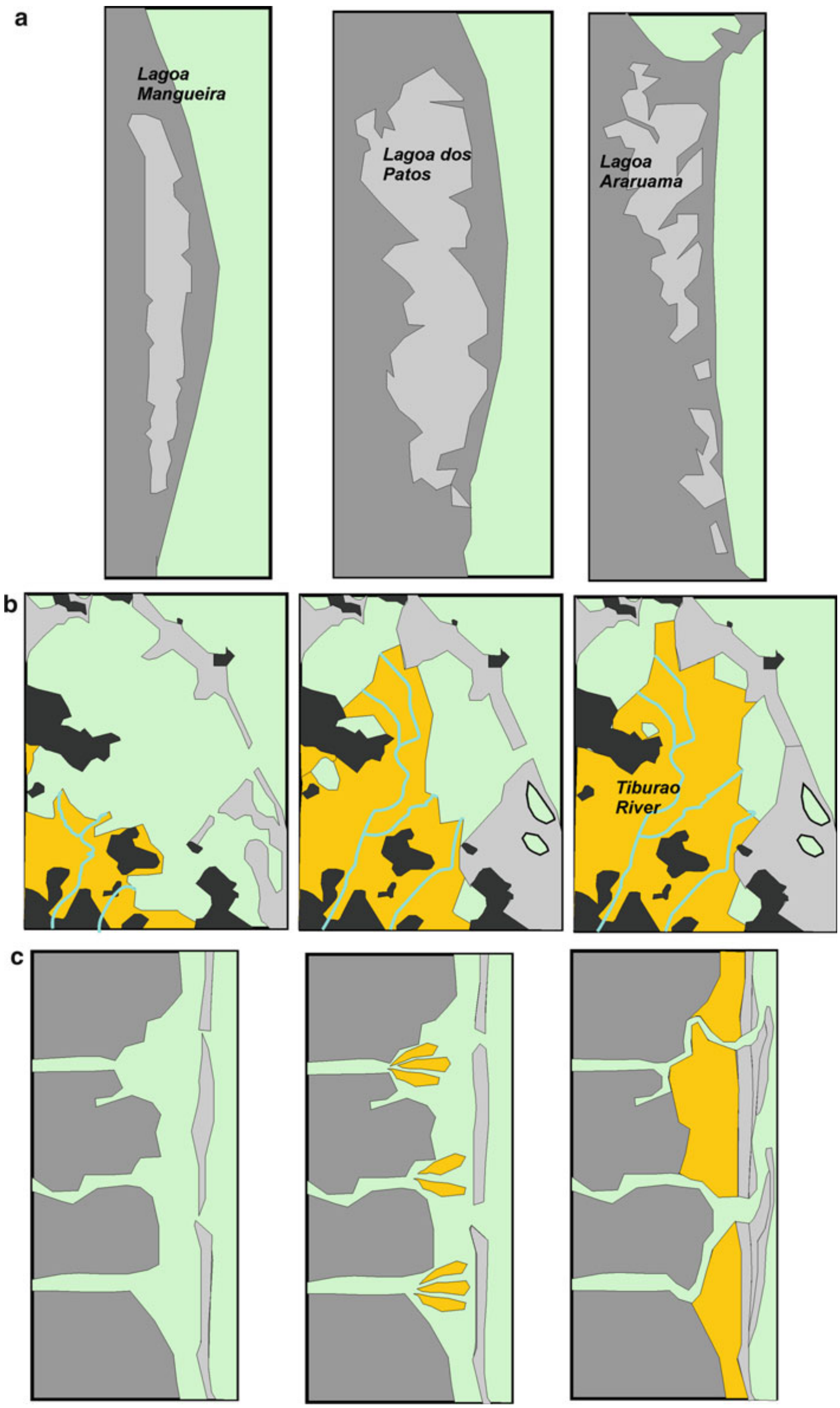
Beach ridges and delta development

Beach-ridge sequences have also been linked to delta development. The so-called wave-dominated deltas of northeastern Brazil grew in relation to beach ridges. During the maximum highstand, coastal lagoons developed and were progressively infilled by intralagoonal deltas (Figure 4; Martin and Dominguez 1994). This regressive model differs from the original view of coastal lagoon development at a coast dominated by longshore drift with a stable sea level (Lucke, 1934; Kumar and Sanders, 1974; Borrego et al., 1993).

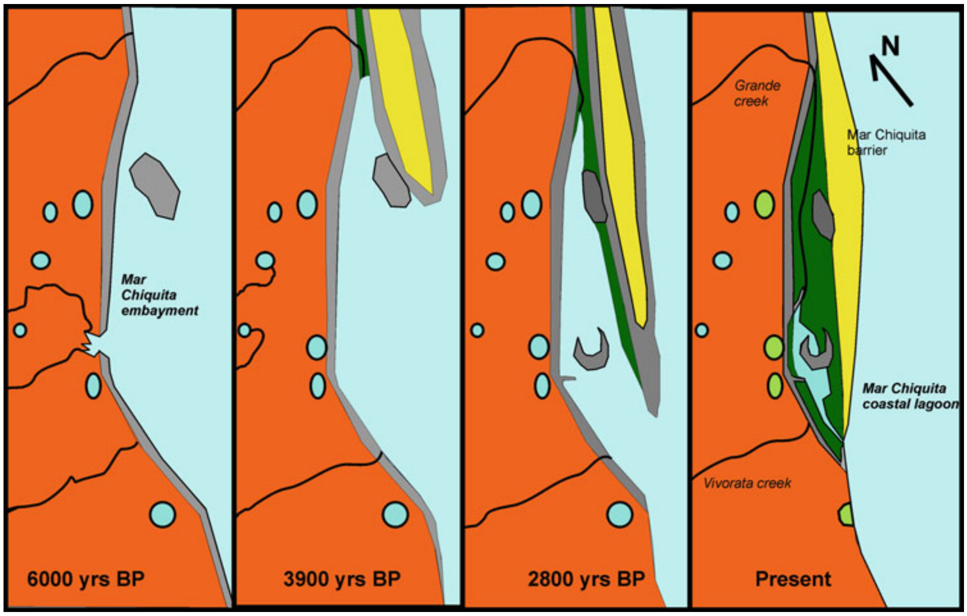
Several deltas were significantly affected in recent years by the construction of river dams or jetty improvements in coastal areas. These changes altered the sediment dynamics of their beach-ridge plains. The Volcán Dam reduced sediment transported to the Orinoco Delta (Warne et al., 2002).

The coast of Nayarit, Western Mexico, provides another good example of a regressive area where coastal lagoons and deltas interact. The region also has a good succession of beach ridges spanning in time from 4,500 years BP to present (Curray et al., 1969). Based on these progradation-aggradation examples, Curray et al. (1969) call attention to non-transgressive models. On transgressive coasts, a barrier is a single ridge transporting onshore. However, other barrier models do exist: (1) a single barrier composed of several ridges that characterize progradation in stable conditions, (2) multiple barrier systems composed of a single ridge each, and (3) multiple barriers composed of several ridges, where each characterizes a regressive coast, related to sea-level fluctuations, climate-triggered net drift changes, or variations in sediment input (deltaic variations).

The beach-ridge plains attached to the progradation of the Doce and Paraiba do Sul developed in response to a steady sea-level fall since the mid-Holocene (Martin et al., 1997). In the evolution of these intralagoonal deltas, the net littoral drift toward the north and the episodic



Cheniers and Regressive Bedforms, Figure 4 Development of Brazilian coastal lagoons. (**a**) Different stages of septation of coastal lagoons (Modified after Isla, 1995). (**b**) Development of a coastal lagoon by septation and deltaic sedimentations (Modified after Nascimento, 2010). Development of deltas filling coastal lagoons (Modified after Dominguez et al., 1987; Martin and Domínguez, 1994).



Cheniers and Regressive Bedforms, Figure 5 Development of the Mar Chiquita coastal lagoon (Modified from Schnack et al., 1982).

effects of waves from the northeast played a significant role (Martin and Suguio, 1992). It is clear that these plains were related to deltas forming within large coastal lagoons until the delta exceeded the body of these bays (Figure 4). Two generations of beach ridges were linked to fluctuations of sea level in a generalized trend of sea-level fall (Martin and Suguio, 1992). Some of these plains were originally prograding cuspate headlands (esporoes or pontales, as they are called in Brazil) which evolved later into regressive spits.

In recent years, different riprap structures have been constructed in order to control the river floods of the Magdalena River (Colombia). The modifications of the discharge channel between 1925 and 1935 produced significant morphological changes (Martínez et al., 1995). The east coast prograded 3.5 km in ~50 years (1943–1990). The two jetties constructed at the outlet (Bocas de Ceniza, 1946–1954) reduced the outlet width to 705 m (Correa et al., 2005; Alvarado Ortega, 2010). On the western coast, the man-made obstruction of longshore drift caused the disappearance of the Sabanilla, Verde, del Medio, and Arena Islands (Martínez et al., 1995).

Processes and mechanisms

Within large estuaries and open coasts, sediment can be distributed by different processes and mechanisms. As a result, several major bedforms occur in these environments that can be distinguished: (1) cuspate spits, (2) regressive spits, (3) cheniers, and (4) washovers.

Cuspate spits are common along elongated coastal lagoons where either wind or tidal currents cause resonance effects, resulting in nodal points where currents counteract and sediment is deposited. This process leads to the septation of coastal lagoons as described by Zenkovitch (1959). It should be stressed that the septation or segmentation is due to estuarine processes (Bird, 2002). The septation of coastal lagoons explains why some cuspate spits are small points, while others are very long and always perpendicular to the main flow (Figure 4).

Some of these cuspate spits can increase in size at a fast rate: the Pontal das Desertas (Lagoa dos Patos, Brazil) prograded about 3,400 m in 57 years toward the southeast (Toldo, 1991). At the region of Laguna (Santa Catarina, Brazil), two beach-ridge plains were compared in their development. The Rio do Meio and Campos Verdes plains (Laguna, Santa Catarina, Brazil) evolved from cuspate spits that divided the mid-Holocene coastal lagoon in the last 5,000 years (Fornari, 2010; Tanaka, 2010). The sediment delivered to this mid-Holocene coastal lagoon by the Tubarão River initially divided the northern lagoons (Santo Antonio and Santa Marta) and afterward the Camacho and Santa Marta lagoons (Nascimento, 2010; Figure 4).

In Mar Chiquita coastal lagoon, there is a cuspate spit 1.9 km long and 300–400 m wide, which shows incipient septation, as it enlarged in a shore-normal direction (Figure 5). The development of the coastal lagoon was influenced by the expansion of a coastal barrier southward (Schnack et al., 1982), while the waterbody was progressively reduced as regressive spits and transverse bedforms composed of shells were forming. Its profile is asymmetric with the higher slope toward the open sea. It is composed primarily by shells of the gastropod *Heleobia* (51–76 %).

Regressive spits occur where there is an interaction between episodic processes (storms) in a regressive trend dominated by longshore drift and a constant siltation rate. Regressive spits occur in sets where it is easy to recognize a different orientation between a set of spits. They can increase in size inside or outside large estuaries and can develop into recurved spits.

The Sao Francisco delta (Brazil) developed within a large coastal lagoon limited by a system of regressive spits (Figure 4; Domínguez, 2009). Fitzgerald and colleagues (1992) described the formation of ridges composed of sand (dominantly) and gravel in response to changes between storm and fair-weather conditions.

Cheniers are caused exclusively by storm effects on a coast subject to wave action or an episodic input of mud. Cheniers should be recognized in vertical sequences where the episodic deposit is overlying estuarine muds.

Washovers consist of coarse deposits on the landward side of a spit or barrier, with sets of landward-dipping lamination or stratification. Their main causes are storm surges, sometimes coinciding with spring tides, which allow the surge to exceed the barrier-spit height. Care should be taken to distinguish between operative washovers, related to present-day storms, and nonoperative washovers that are isolated by processes taking place today. Usually they lie landward of the active washovers.

Summary

Cheniers can occur in different settings where a low-lying plain is subject to periodic high-energy events or drastic fluctuations of sediment availability. The mid-Holocene sea-level fall favors the progradation of chenier plains. Large estuaries, coastal lagoons, and deltas are affected by waves capable of reworking former deposits. Cuspate forelands and spits have been linked to seiches (tidal nodes or wind-generated edge waves). Regressive spits and beach-ridge plains reflect wave effects within estuaries.

Bibliography

Alvarado Ortega, M., 2010. *Barranquilla: ciudad con río y mar. En Barranquilla: Como vamos*. Barranquilla: Universidad del Norte, 24 pp.

Anthony, E. J., 1995. Beach-ridge development and sediment supply: examples from West Africa. *Marine Geology*, **129**, 175–186.

Augustinus, P. G. E. F., 1980. Actual development of the chenier coast of Suriname (South America). *Sedimentary Geology*, **26**, 91–113.

Augustinus, P. G. E. F., Hazelhoff, L., and Kroon, A., 1989. The chenier coast of Suriname. Modern and geological development. *Marine Geology*, **90**, 269–281.

Belperio, A. P., Harvey, N., and Bouman, R. P., 2002. Spatial and temporal variability in the Holocene sea-level record of the South Australia coastline. *Sedimentary Geology*, **150**, 153–169.

Bértola, G., 1994. *Geomorfología y sedimentología de los ambientes mareales de la bahía de Samborombón (provincia de Buenos Aires)*. Thesis Doctoral, La Plata, Facultad de Ciencias Naturales y Museo, UNLP, pp. 56–57.

Bird, E. C. F., 2002. Physical setting and geomorphology of coastal lagoons. In Kjerve, B. (ed.), *Coastal Lagoon Processes*. Elsevier Oceanographic Series, Vol. 60, pp. 9–39.

Borrego, J., Morales, J. A., and Pendón, J. G., 1993. Holocene filling of an estuarine lagoon along the mesotidal coast of Huelva: the Piedras River mouth, Southwestern Spain. *Journal of Coastal Research*, **9**(1), 242–254.

Buynevich, I., Asp, N., FitzGerald, D., Cleary, W., Klein, A., Siegle, E., and Angulo, R., 2005. Mud in the surf: nature at work in a Brazilian Bay. *Eos*, **86**(33), 301–308.

Byrne, J. V., Le Roy, D. O., and Riley, C. M., 1959. The chenier plain and its stratigraphy, Southwestern Louisiana. *Transactions, Gulf Coast Society of Geological Sciences*, IX, pp. 237–259.

Cavalotto, J. L., Violante, R. A., and Colombo, F., 2005. Evolución y cambios ambientales de la llanura costera de la cabecera del Río de la Plata. *Revista de la Asociación Geológica Argentina*, **60**(2), 353–367.

Chappell, J., and Grindrod, J., 1984. Chenier plain formation in Northern Australia. In Thom, B. G. (ed.), *Coastal Geomorphology in Australia*. Australia: Academic Press, Chap. 10, pp. 197–231.

Cook, P. J., and Polach, H. A., 1973. A chenier sequence at Broad Sound, Queensland, and evidence against a Holocene high sea level. *Marine Geology*, **14**, 253–268.

Correa, I. D., Alcántara-Carrió, J., and González, D. A., 2005. Historical and recent shore erosion along the Colombian Caribbean coast. *Journal of Coastal Research*, **SI49**, 52–57.

Curray, J. R., Emmel, F. J., and Crampton, P. J. S., 1969. Holocene history of a strand plain, lagoonal coast, Nayarit, Mexico. In Ayala-Castañares, A., and Phleger, F. B. (eds.), *Lagunas Costeras, un Simposio*. Simposio Internacional de lagunas costeras, UNAM, UNESCO, pp. 63–100.

Davis, R. A., and Clifton, H. E., 1987. Sea-level change and the preservation potential of wave-dominated and tide-dominated coastal sequences. In Nummedal, D., Pilkey, O. H., and Howard, J. D. (eds.), *Sea-Level Fluctuation and Coastal Evolution*. SEPM Special Publication, 41, pp. 167–178.

Domínguez, J. M. L., 2009. The coastal zone of Brazil. In Dillenburg, S. R., and Hesp, P. A. (eds.), *Geology and Geomorphology of Holocene Coastal Barriers of Brazil*. Dordrecht: Springer. Lecture Notes in Earth Sciences, Chap. 2, pp. 17–51.

Dominguez, J. M. L., Martin, L., and Bittencourt, A. C. S. P., 1987. Sea-level history and Quaternary evolution of river-mouth associated beach-ridge plains along the east-southeast Brazilian coast: a summary. In Nummedal, D., Pilkey, O. H., and Howard, J. D. (eds.), *Sea Level Fluctuation and Coastal Evolution*. Tulsa: SEPM (Society of Economic Paleontologists and Mineralogists). Society of Economic Paleontologists and Mineralogists. SEPM Special Publication 41, pp. 115–127.

Dougherty, A. J., and Dickson, M. E., 2009. Sea level and storm control on the evolution of a chenier plain, Firth of Thames, New Zealand. *Marine Geology*, 307–310, 58–72.

Fitzgerald, D. M., Baldwin, C. T., Ibrahim, N. A., and Humphries, S. M., 1992. Sedimentologic and morphologic evolution of a - beach-ridge barrier along an indented coast: Buzzards Bay, Massachusetts. *SEPM Special Publication*, **48**, 65–75.

Fornari, M., 2010. *Evolucao sedimentary holocénica da retrobarreira na regiao de Jaguaruna-Laguna, Santa Catarina, Brasil*. Unpublished thesis, Sao Paulo, Universidade de Sao Paulo, 262 pp.

Hesp, P. A., and Short, A. D., 1999. Barrier morphodynamics. In Short, A. D. (ed.), *Handbook of Beach and Shoreface Morphodynamics*. New York: Wiley, Chap. 14, pp. 307–333.

Isla, F. I., 1989. Holocene sea-level fluctuation in the southern hemisphere. *Quaternary Science Reviews*, **8**, 359–368.

Isla, F. I., 1995. Coastal lagoons. In Perillo, G. M. E. (ed.), *Morphology and Sedimentology of Estuaries*. Oxford: Elsevier, Chap. 9, pp. 241–272.

Isla, F. I., 1998. Uniformitarismo transgresivo o regresivo? *Boletin Paranaense de Geociencias*, **46**, 165–174.

Isla, F. I., Vilas, F. E., Bujalesky, G., Ferrero, M., Gonzalez Bonorino, G. Y., and Arche Miralles, A., 1991. Gravel drift and wind effects over the macrotidal San Sebastián Bay, Tierra del Fuego. *Marine Geology*, **97**, 211–224.

Kumar, N., and Sanders, J. E., 1974. Inlet sequence: a vertical succession of sedimentary structures and textures created by the lateral migration of tidal inlets. *Sedimentology*, **21**, 491–532.

Lucke, J. B., 1934. A theory of evolution of coastal lagoon deposits on shore lines of emergence. *Journal of Geology*, **42**, 561–584.

Martin, L., and Domínguez, J. M. L., 1994. Geological history of coastal lagoons. In Kjerve, B. (ed.), *Coastal Lagoon Processes*. Oxford: Elsevier. Elsevier Oceanographic Series, Vol. 60, pp. 41–68.

Martin, L., and Suguio, K., 1992. Variation of coastal dynamics during the last 7000 years recorded in beach-ridge plains associated with river mouths: example from the central Brazilian coast. *Palaeogeography, Palaeoclimatology, Palaeoecology*, **99**, 119–140.

Martin, L., and Suguio, K., 1997. *Geologia do Quaternario costeiro do litoral norte do Rio de Janeiro e do Espírito Santo*. Belo Horizonte: CPRM/FAPESP. 94 pp.

Martínez, J. I., Yokoyama, Y., Gómez, A., Delgado, A., Matsuzaki, H., and Rendón, E., 2010. Late Holocene marine terraces of the Cartagena region, southern Caribbean: the product of neotectonism or a former high stand in sea-level? *Journal of South American Earth Sciences*, **29**, 214–224.

Martin, L., Suguio, K., Dominguez, JML., Flexor, JM., 1997. *Geologia do Quaternaio costeiro do litoral norte do Río de Janeiro e do Espírito Santo*. FAPESP, CPRM, Belo Horizonte, Brazil, 112 pp.

Martínez, J. O., González, J. L., Pilkey, O. H. and Neal, W. J., 1995. *Tropical Barrier Islands of Colombia's Pacific Coast*. Journal of Coastal Research, 11 (2):432–453.

Meldahl, K. H., 1995. Pleistocene shoreline ridges from tide-dominated and wave-dominated coasts: northern California and western Baja California, Mexico. *Marine Geology*, **123**, 61–72.

Nascimento, D. R., Jr., 2010. *Evolucao sedimentary holocenica do delta do Rio Tubarao, Estado de Santa Catarina*. Unpublished thesis, Sao Paulo, Universidade de Sao Paulo, 231 pp.

Nieuwenhuyse, A., and Kroonenberg, S. B., 1994. Volcanic origin of Holocene beach ridges along the Caribbean coast of Costa Rica. *Marine Geology*, **120**, 13–26.

Otvos, E. G., 2000. Beach ridges, definition and significance. *Geomorphology*, **32**, 83–108.

Otvos, E. G., and Price, A., 1979. Problems of chenier genesis and terminology – an overview. *Marine Geology*, **31**, 251–263.

Penland, S., and Suter, J. R., 1989. The geomorphology of the Mississippi River chenier plain. *Marine Geology*, **90**, 231–258.

Price, W. A., 1954. *Environment and Formation of the Chenier Plain*. College Station: A&M College of Texas, Department of Oceanography. 7 pp.

Price, W. A., 1955. Environment and formation of the chenier plain. *Quaternaria*, **2**, 75–86.

Reineck, H. E., and Singh, I. B., 1980. *Depositional Sedimentary Environments*. Berlin: Springer.

Rhodes, E. G., 1982. Depositional model for a chenier plain, Gulf of Carpentaria, Australia. *Sedimentology*, **29**, 201–221.

Rodríguez-Ramírez, A., and Yáñez-Camacho, C. M., 2008. Formation of chenier plain of the Doñana marshland (SW Spain): observations and geomorphic model. *Marine Geology*, **254**, 187–196.

Roy, P. S., Cowell, P. J., Ferland, M. A., and Thom, B. G., 1995. Wave-dominated coasts. In Carter, R. W. G., and Woodroffe, C. D. (eds.), *Coastal Evolution: Late Quaternary Shoreline Morphodynamics*. Cambridge: Camdridge Press, Chap. 4, pp. 121–186.

Schnack, E. J., Fasano, J. L., and Isla, F. I., 1982. The evolution of Mar Chiquita lagoon, Province of Buenos Aires, Argentina. In Colquhoun, D. J. (ed.), *Holocene Sea-Level Fluctuations: Magnitudes and Causes*. Columbia, SC: Igcp 61, University of South Carolina, pp. 143–155.

Schofield, J. C., 1960. Sea level fluctuations during the last 4,000 years as recorded by a chenier plain, Firth of Thames, New Zealand. *New Zealand Journal of Geology and Geophysics*, **3**, 467–485.

Short, A. D., 1988. The South Austalian coast and Holocene sea-level transgression. *Geographical Review*, **78**(2), 119–136.

Tanaka, A. P. B., 2010. *Sedimentología, cronología e dinámica progradacional das planicies costeiras de campos Verdes e Ji (Laguna, SC)*. Unpublished thesis, Sao Paulo, Instituto de Geociencias, USP, 113 pp.

Tanner, W. F., 1995. Origin of beach ridges and swales. *Marine Geology*, **129**, 149–161.

Toldo, E. E., 1991. *Morfodinámica da Laguna dos Patos, Rio Grande do Sul*. Porto Alegre: Pesquisas, UFRGS, Vol. 18(1), pp. 58–63.

Vilas, F., Arche, A., Ferrero, M., and Isla, F., 1999. Subantarctic macrotidal flats, cheniers and beaches in San Sebastián Bay, Tierra del Fuego, Argentina. *Marine Geology*, **160**, 301–326.

Warne, A. G., Meade, R. H., White, W. A., Guevara, E. H., Gibeaut, J., Smyth, R. C., Aslan, A., and Tremblay, T., 2002. Regional controls on geomorphology, hydrology, and ecosystem integrity in the Orinoco Delta, Venezuela. *Geomorphology*, **44**, 273–307.

Wells, J. T., and Coleman, J. M., 1977. Nearshore suspended sediment variations, central Surinam coast. *Marine Geology*, **24**, M47–M54.

Wells, J. T., and Coleman, J. M., 1981. Periodic mudflat progradation: Northeastern coast of South America: a hypothesis. *Journal of Sedimentary Petrology*, **51**(4), 1069–1075.

Wells, J. T., and Kemp, G. P., 1981. Atchafalaya mud stream and recent mudflat procreation: Louisiana chenier plain. *Transactions, Gulf Coast Society of Geological Sciences*, XXXI, pp. 409–416.

Wong, T. E., de Kramer, R., de Boer, P. L., Langereis, C., and Sew-A-Tjon, J., 2009. The influence of sea-level changes on tropical coastal lowlands; the Pleistocene Coropina Formation, Suriname. *Sedimentary Geology*, **216**, 125–137.

Woodroffe, C. D., 2002. *Coasts: Form, Process and Evolution*. Cambridge: Cambridge University Press.

Woodroffe, C. D., Curtis, R. J., and McLean, R. F., 1983. Development of a chenier plain, Firth of Thames, New Zealand. *Marine Geology*, **53**, 1–22.

Xiqing, C., 1996. An integrated study of sediment discharge from the Changjiang River, China, and the delta development since the Mid-Holocene. *Journal of Coastal Research*, **12**(1), 26–37.

Xitao, Z., 1989. Cheniers in China: an overview. *Marine Geology*, **90**, 311–320.

Yan, Q., Xu, S., and Shao, X., 1989. Holocene cheniers in the Yangtze delta, China. *Marine Geology*, **90**, 311–320.

Zenkovitch, V. P., 1959. On the genesis of cuspate spits along lagoon shores. *Journal of Geology*. **67**, 269–277.

Cross-references

Coastal Barriers
Delta Plain
Sand Ridge
Spit

CLEAN WATER ACT

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Synonyms

Federal Water Pollution Control Act Amendments

Definition

The US Congress enacted the Clean Water Act in 1972 (P.L. 92-500, the 1972 Amendments to the Federal Water Pollution Control Act), which is the principal law dealing with polluting activity in streams, rivers, lakes, and estuaries of the USA (Copeland, 2006). The Clean Water Act established a water quality standards approach for regulating water quality, with the US Environmental Protection Agency responsible for developing national water quality criteria. A waterbody found to be in violation of a water quality standard was to be listed as "impaired" with consideration of the establishment of a total maximum daily load (TMDL) of the pollutant in violation of the standard (Lee et al., 2005).

Description

The Federal Water Pollution Control Act was originally enacted in 1948 and later amended in 1972 as the Clean Water Act (P.L. 92-500). Subsequent amendments were made in 1977 (P.L. 95-217), 1981 (P.L. 97-117), and 1987 (P.L. 100-4). As noted by Copeland (2006), the Clean Water Act consists of two main parts: "regulatory provisions that impose progressively more stringent requirements on industries and cities in order to meet the statutory goal of zero discharge of pollutants, and provisions that authorize federal financial assistance for municipal wastewater treatment construction."

Bibliography

Copeland, C., 2006. *Water Quality: Implementing the Clean Water Act*. Washington, DC: Congressional Research Service. CRS Report for Congress, Order Code RL33466.

Lee, G. F., and Jones-Lee, A., 2005. Clean Water Act, water quality criteria/standards, TMDLs, and weight-of-evidence approach for regulating water quality. In *Water Encyclopedia: Water Law and Economics*. Hoboken, NJ: Wiley, pp. 598–604.

Cross-references

Anoxia, Hypoxia, and Dead Zones
Eutrophication
Halogenated Hydrocarbons
Nonpoint Source Pollution
Oil Pollution
Polycyclic Aromatic Hydrocarbons
Trace Metals in Estuaries
Water Quality

CLIMATE CHANGE

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Synonyms

Climate variability

Definitions

Climate change (including climate variability) refers to regional or global changes in mean climate state or in patterns of climate variability over decades to millions of years often identified using statistical methods and sometimes referred to as changes in long-term weather conditions (IPCC, 2012). Climate is influenced by changes in continent-ocean configurations due to plate tectonic processes, variations in Earth's orbit, axial tilt and precession, atmospheric greenhouse gas (GHG) concentrations, solar variability, volcanism, internal variability resulting from interactions between the atmosphere, oceans and ice (glaciers, small ice caps, ice sheets, and sea ice), and anthropogenic activities such as greenhouse gas emissions and land use and their effects on carbon cycling.

Introduction

Earth's climate has varied over all timescales due to changes in global energy balance and radiative forcing caused by changes in solar radiation reaching Earth's atmosphere, volcanism, Earth's orbital configuration (precession, tilt, eccentricity), atmospheric greenhouse gas concentrations, and ocean basin-continent distributions. Regional and global climate changes can be amplified or dampened by complex feedback mechanisms involving sea-ice albedo, methane release from permafrost and marine sediments, land surface vegetation cover, ice sheet dynamics, and atmosphere-ocean-land exchange of carbon dioxide. In addition to natural climate changes, there is substantial evidence from instrumental records, climate modeling, and paleoclimate reconstructions that humans have influenced global and regional climate.

Estuaries, inlets, bays, fjords, tidal marshes, and other coastal systems are directly or indirectly affected by climate change. Instrumental records from stream gauges, water quality measurements, and, more recently, satellites provide trends in salinity, temperature, turbidity, dissolved oxygen, and many other parameters that can be linked to regional climate change and variability. Climate, hydrological, and ecosystem modeling studies are another approach to understanding climate impacts. For example, there are growing efforts to project estuarine response to elevated greenhouse gas concentrations (Najjar et al., 2010) some using downscaling methods that link global climate and regional models (Hostetler et al., 2011). Because instrumental records are usually limited to the last few decades, a third approach employs paleoclimatic and paleoecological reconstructions, obtained from geochemical, physical, or biological proxies recovered from sediment cores (Cronin and Walker, 2006; Gooday et al., 2009). Paleo-reconstructions provide direct evidence for past climate impacts and prehistorical baseline conditions for ecosystem restoration, impact assessment, and planning.

Identifying climate impacts on modern estuaries is complicated by multiple environmental stresses from a wide range of local and regional anthropogenic activity such as land use changes associated with urbanization, agriculture, and other activities (Willard and Cronin, 2007; Canuel et al., 2010). These factors can make the attribution of observed changes in estuarine environments to specific causes very challenging.

This chapter summarizes climate impacts on estuaries during the mid-Holocene to late Holocene interglacial period (the last ~7,000 years), which is the period since postglacial sea-level rise stabilized and modern coastal systems took their modern form. Thus, this chapter applies to both prehistorical natural climate variability and climate change since the onset of the Anthropocene, sometimes defined as the period since the industrial revolution beginning ~1750–1800 CE (Common Era) (Gale and Hoare, 2012). Climate impacts can be grouped into five broad, interconnected categories: regional precipitation, sediment processes, temperature (global and regional), biogeochemical processes, and sea-level rise.

Regional precipitation

Climate change has direct impacts on estuaries through its effects on regional rainfall patterns. Seasonal and/or mean annual precipitation in watersheds is often highly correlated with river discharge into estuaries, which in turn affects salinity patterns and circulation. For example, in a partially mixed, microtidal estuary like Chesapeake Bay, river discharge, along with wind and tidal forcing, affects buoyancy-driven circulation, stratification, the development of a pycnocline, and oxygen exchange between upper and deeper layers (Schubel and Pritchard, 1986).

As a consequence, river discharge affects nutrient influx, phytoplankton blooms, dissolved oxygen, water quality, and ecosystem functioning such that excess nutrient loading coupled with greater river discharge has led to estuarine eutrophication on a global scale (Diaz and Rosenberg, 2008; Kemp et al., 2009; Howarth et al., 2011).

Internal modes of climate variability

The term "internal modes of climate variability" is often used to refer to climate changes that are not forced by radiative forcing from GHGs and solar and volcanic activities but rather interactions between the atmosphere, oceans, and ice sheets. The most widely recognized climate patterns are called the El Niño-Southern Oscillation (ENSO), North Atlantic Oscillation (NAO), Pacific Decadal Oscillation (PDO), and Atlantic Multidecadal Oscillation (AMO). Many studies have demonstrated a strong connection between internal modes of climate variability over interannual to multidecadal timescales and estuarine circulation, salinity, and dissolved oxygen (DO). Using a global dataset, Gilbert et al. (2010) could identify a secular pattern of decreasing DO between 1976 and 2000 that was more evident in coastal regions than in the open ocean. However, they also stressed that when interpreting the twentieth century patterns of oxygen concentrations, decadal climate variability can impose large-amplitude oscillations larger than the overall linear trend (see Garcia et al., 2005). Some examples of climate variability impacting regional rainfall, river discharge, estuarine salinity, and, in some cases, nutrient flux include studies of the PDO (Xu et al., 2012), the NAO (Cronin et al., 2005; Prasad et al., 2010), ENSO (Swart et al., 1996; Schmidt et al., 2001; Cronin et al., 2002), and the AMO (Enfield et al., 2001).

There are also well-established links between climate variability and marine biological systems (Mantua et al., 1997, Drinkwater et al., 2003; Pershing et al., 2005; Greene and Pershing, 2007). Cloern et al. (2010) showed that biological communities in San Francisco Bay are sensitive to ocean currents, temperatures, and coastal upwelling connected to PDO variability and North Pacific gyre circulation. Paerl et al. (2013) showed that climate-driven changes in river discharge to North Carolina estuaries altered the composition and biomass of phytoplankton communities. ENSO- and NAO-connected climate variability also influences outbreaks of infectious diseases on a global scale (Lafferty, 2009; Morand et al., 2013) and, in particular, viruses, bacteria, and infectious disease outbreaks in coastal waters (Lipp et al., 2001; Rose et al., 2001).

Two specific aspects of climate that deserve attention are extended droughts or wet periods and extreme events such as tropical cyclones. Evidence from tree-rings, corals, sediments, molluscan isotopes, and speleothems shows that droughts are an inherent part of Holocene climate. Quantitative reconstructions of precipitation show that North America (Cook et al. 2014) and Europe (Büntgen et al., 2010) have experienced decadal, continent-scale droughts over the past millennium. Multiple paleo-reconstructions based on several proxies show that droughts frequently affected mid-Atlantic climate and Chesapeake Bay watershed (Stahle et al., 1998; Cronin et al., 2005, Saenger et al., 2006; Harding et al., 2010). Precipitation changes over centennial timescales also affected coastal systems, such as changes in runoff and productivity in Chilean fjords during the latter part of the Little Ice Age from ~1600 to the 1800s (Rebolledo et al., 2008).

Although specific weather events cannot be directly linked to climate change, there is nonetheless concern that changing climate might increase the frequency and intensity of tropical storms (Nicholls et al., 2007), which can severely impact estuaries. In one of the first intensive studies of hurricane impacts, the June 1972 storm *Agnes* in the eastern United States, there were widespread, long-lasting effects on Chesapeake Bay circulation, salinity, water quality, and ecosystems (Bailey et al., 1975; Davis et al., 1977). Similarly, three hurricanes that hit coastal North Carolina in 1999 caused 50- to 500-year floods, lowered salinity, and enhanced nitrogen loading to Pamlico Sound, which together had multiyear effects on coastal ecosystems (Paerl et al., 2001). Large storms also affect coastal wetlands notably through the impacts of storm surge, wind, and freshwater flushing on wetland soil dynamics and elevation (Cahoon, 2006).

Modeling precipitation changes and impacts

One challenge in estuarine research is predicting future precipitation/streamflow changes due to higher CO_2 concentrations. In the mid-Atlantic region of the eastern United States (Chesapeake Bay, the Delaware Bay, and Hudson River Estuary), impacts on streamflow ranged from a decrease of 40 % to an increase of 30 %, although results varied by season (Najjar et al., 2009). In a study of San Francisco Bay, Knowles and Cayan (2002) found that changes in winter snowpack and reduction in spring runoff would lead to elevated salinity (see Cloern et al., 2011). Future progress in this emerging field will come from linking downscaled climate models with watershed and estuarine hydrodynamic models, especially as improvements are made in predicted precipitation response to future climate change.

Global and regional temperature

Compared to precipitation-driven changes, the impacts of changing temperature may be less obvious in the short term, but nonetheless aquatic temperatures are important in estuarine functioning and ecosystems. Global mean annual and regional ocean temperatures are expected to rise over future decades to centuries due to elevated atmospheric CO_2 concentrations (Najjar et al., 2009), and in theory, this warming might lead to poleward range shifts in temperature-sensitive species (Helmuth et al., 2002; Przeslawski et al., 2012). Moreover, there is indisputable evidence that the world's oceans have been warming for at least the last 50 years (Levitus et al., 2012), and paleoclimate records show that marine species experienced large climate-driven biogeographic range shifts over 10^4–10^7 year timescales. These shifts are best documented in marine sediment records of major microfossil groups (diatoms, dinoflagellates, foraminifera, radiolarian, ostracodes) during glacial-interglacial cycles of the 500,000 years when Earth's mean annual temperature fell ~5 °C during glacial periods (Kucera et al., 2005). In addition to open-ocean sea faunal and floral biogeographic shifts, paleo-records from estuaries and coasts also show Holocene temperature-induced biogeochemical and productivity changes such as the sedimentary record of LIA cooling in Kagoshima Bay, Japan (Kuwae et al., 2007).

In addition to large-scale range shifts, several indirect impacts of rising temperatures deserve mention: reduced sea ice, especially in marginal subarctic seas; coal bleaching; expanded geographic ranges of harmful algal bloom species; and mangrove species expansion among others (Nicholls et al., 2007). Case studies include the Bering-Chukchi Seas (Grebmeier, 2012), the Changjiang River Estuary (Ma et al., 2009), Mediterranean coastal systems (Bensoussan et al., 2010), Narragansett Bay, Rhode Island (Nixon et al., 2009), and the Gulf of Mexico (Bianchi et al., 2013).

Sediment processes

Coastal sedimentary processes influenced by climate include erosion (in the watershed and estuary), transport (in suspension and along river and estuarine bottoms), and deposition in an estuary, bay, or fjord. However, deciphering climate impacts on sedimentation is difficult due to large-scale anthropogenic activities. On the global scale, Syvitski et al. (2005) estimate that humans account for 2.3 ± 0.6 billion metric tons per year but that sediment retention in reservoirs, totaling 100 billion metric tons (bmt) in recent decades, reduces the sediment reaching the world's coasts by 1.4 ± 0.3 bmt per year (see Milliman and Farnsworth, 2011). On a regional scale, Saenger et al. (2008) found that postcolonial agricultural land clearance in the Chesapeake Bay watershed increased sediment accumulation rates by several times, but there were complex leads and lags related to climatic factors.

Nonetheless, preindustrial climate changes are known to affect sediment flux to coastal systems. For example, in subpolar fjords in Svalbard, Szczucinski et al. (2009) found that post-Little Ice Age temperature increase and glacier retreat had large impacts on sediment accumulation.

Within an estuary or bay, sediment affects a variety of factors including turbidity, light penetration, and the distribution of submerged aquatic vegetation (including sea grasses). This applies both to clastic sediment, often referred to as mineral matter, and particulate organic material, much of which is produced by algal productivity fueled by high nutrient concentrations. Sediment also plays an important role in the development of estuarine turbidity maximum zones (ETM, also called turbidity maximum zones, TMZ), a characteristic feature of many estuaries. It has long been known that trapping of suspended material in ETMs can be enhanced by increased vertical stratification due to large freshwater influx (Geyer, 1993). The physics of circulation near these salinity gradients are such that they trap clastic sediment and phytoplankton-derived organic material that has been transported to or resuspended within the ETM, resulting in high nutrient concentrations (Uncles et al., 2006;

Doxaran et al., 2009). As zones of complex salinity variability, nutrient dynamics, planktonic productivity, and fish spawning and growth, ETMs are important estuarine features forced by climate, river discharge, salinity, and sediment transport.

Despite the complexity of processes controlling sediment, land-to-estuary sediment flux, estuaries will continue to be vulnerable to future changes in climate, including the incidence and intensity of extreme storm events.

Biogeochemical processes

In addition to biogeochemical changes related to nutrient and oxygen dynamics discussed above, changes in ocean carbonate chemistry due to the uptake of anthropogenic CO_2 by the world's ocean, often referred to as "ocean acidification" (OA), pose complex, taxon-specific, and still poorly understood impacts on marine life (Hendriks et al., 2010; Wittmann and Pörtner, 2013; Kroeker et al., 2013). It is estimated that mean global ocean pH has been lowered by 0.1 pH units since ~1750 and may decrease by 0.3–0.4 pH units by 2100 (Pelejero et al., 2010). For comparison, glacial-interglacial cycles of the last 400 ka may have experienced changes of between 0.15 and 0.3 pH units. Although anthropogenic driven pH changes cannot be directly compared to natural events due to differing rates and boundary conditions, paleoclimate studies show that over multimillion year timescales, past natural acidification events had large effects on marine organisms (Kump et al., 2009; Pelejero et al., 2010; Hönisch et al., 2012).

Currently, the study of OA impacts on coastal marine organisms is a growing field for corals (Hoegh-Guldberg et al., 2007), molluscs (Talmage and Gobler, 2009; Waldbusser et al., 2011, 2014; Gobler and Talmage, 2013), and other taxonomic groups (Ries et al., 2009; Kroeker et al., 2010). Some case studies suggest that pH has fallen in recent decades in some coastal systems. For example, pH fell from ~8.2 to 7.9 in the last 30 years in Chesapeake Bay (Waldbusser et al., 2011); Feely et al. (2010) estimate that 24–49 % of observed pH lowering in parts of Puget Sound, a deep estuary in the Pacific NW, was due to influx of seasonal upwelled ocean water, that is, global OA, as distinct from in situ remineralization via respiration. Complicating the issue of causality of observed changes in coastal pH, Pelejero et al. (2005) found that pH variation in a southwest Pacific Ocean coral was related to multidecadal climate variability in the Interdecadal Pacific Oscillation. In addition, other factors, such as reduced freshwater influx and higher salinity, may affect estuarine pH.

Sea-level rise

Sea-level rise (SLR) is one of the most challenging yet misunderstood concerns for estuaries and other coastal systems. No fewer than five global and four regional processes influence relative sea level along any particular coast (Cronin, 2012). Global factors include thermosteric ocean expansion (increase in ocean volume, Willis et al., 2010), melting land-based ice from glaciers (increases ocean mass and mean global sea level), melting parts of the Greenland and Antarctic Ice Sheets (increases ocean mass and sea level, Hanna et al., 2013), reservoir storage (decreases mean sea level), and terrestrial water depletion (increases mean sea level, Konikow, 2011). Regional processes (excluding rapid tectonic movement) include glacio-isostatic adjustment (GIA, Peltier and Fairbanks, 2006) due to viscoelastic response of Earth's mantle to melting large ice sheets since the last glacial period ~20 ka (local GIA can also occur due to glacier melting), elastic deformation of Earth's crust due to changes in gravity and rotation (Tamisiea and Mitrovica, 2011), local groundwater withdrawal, and long-term thermal subsidence of the crust (typically minimal).

The contribution of each factor will vary regionally, but nonetheless, from the standpoint of estuaries and other coastal systems, several points deserve emphasis. Global mean sea level has been rising at rate of 3.1 mm year^{-1} over the past few decades (perhaps an acceleration over rates averaged for the last century), mostly due to thermosteric expansion and land ice melting. Some studies suggest that SLR is already affecting large estuaries such as Chesapeake Bay (Hilton et al., 2008; Murphy et al., 2011) and coastal wetlands (Cahoon et al., 2006). In addition, although no consensus exists on future SLR, rates are expected to increase and glacier and ice sheet mass balance loss is likely to dominate SLR the rest of the twenty-first century. Consequently, the modeling study by Hong and Shen (2012) on the impacts of future SLR on Chesapeake Bay is illustrative, finding that primary effects on salinity, stratification, circulation, nutrient retention, and dissolved oxygen varied spatially, seasonally, and interannually. In addition, if as expected, tidal ranges and wave heights increase, severe storms would become an even larger concern in some estuaries (Najjar et al., 2010). Finally, geological records show that in the past, SLR rates reached and at times exceeded ~10–15 mm year^{-1} in the absence of abrupt increase in greenhouse gas forcing. The implication is that, although the many factors that govern coastal ecosystem functioning cannot be oversimplified, the ability of some sensitive systems, notably mangroves, salt marshes, and coral reefs, to "keep up" with SL, that is, to accrete at the same rate of SL rise, remains a major concern.

Summary

Climate changes throughout geological history have influenced estuaries and coastal systems in a variety of ways and over all timescales. Similarly, future climate change will influence estuaries, perhaps at an accelerated rate, notably through effects on salinity and temperature, dissolved oxygen concentrations, nutrient and sediment flux, biogeochemical processes, and coastal ecosystem functioning. Sea-level rise, altered rainfall patterns leading to extreme droughts and wet periods, and biogeochemical changes associated with ocean acidification are among the most important research topics associated with climate that will likely see great progress in the next few years.

Bibliography

Bailey, J. F., Patterson, J. L., and Paulhus, J. L. H., 1975. *Hurricane Agnes Rainfall and Floods, June–July 1972*. Washington, DC: United States Government Printing Office. Geological Survey Professional Paper 924.

Bensoussan, N., Romano, J.-C., Harmelin, J.-G., and Garrabou, J., 2010. High resolution characterization of northwest Mediterranean coastal waters thermal regimes: to better understand responses of benthic communities to climate change. *Estuarine, Coastal and Shelf Science*, **87**, 431–441.

Bianchi, T. S., Allison, M. A., Zhao, J., Li, X., Comeaux, R. S., Feagin, R. A., and Kulawardhana, R. W., 2013. Historical reconstruction of mangrove expansion in the Gulf of Mexico: linking climate change with carbon sequestration in coastal wetlands. *Estuarine, Coastal and Shelf Science*, **119**, 716.

Büntgen, U., Trouet, V., Frank, D., Leuschner, H. H., Friedrichs, D., Luterbacher, J., and Esper, J., 2010. Tree-ring indicators of German summer drought over the last millennium. *Quaternary Science Reviews*, **29**, 1005–1016.

Cahoon, D. R., 2006. A review of major storm impacts on coastal wetland elevations. *Estuaries and Coasts*, **29**, 889–898.

Cahoon, D. R., Hensel, P. F., Spencer, T., Reed, D. J., McKee, K. L., and Saintilan, N., 2006. Coastal wetland vulnerability to relative sea-level rise: wetland elevation trends and process controls. In Verheven, J. T. A., Beltman, B., Bobbink, R., and Whigham, D. F. (eds.), *Wetlands and Natural Resource Management Ecological Studies 190*. New York: Springer, pp. 271–292.

Canuel, E. A., Brush, G., Cronin, T. M., Lockwood, R., and Zimmerman, A. R., 2010. Paleoecology studies in Chesapeake Bay: a model system for understanding interactions between climate, anthropogenic activities and the environment. In Gibson, J., Saunders, K., Gell, P., and Tibby, J. (eds.), *Application of Paleoenvironmental Techniques in Estuarine Studies*. New York: Springer.

Cloern, J. E., Hieb, K. A., Jacobson, T., Sansó, B., Di Lorenzo, E., Stacey, M. T., Largier, J. L., Meiring, W., Peterson, W. T., Powell, T. M., Winder, M., and Jassby, A. D., 2010. Biological communities in San Francisco Bay track large-scale climate forcing over the North Pacific. *Geophysical Research Letters*, **37**, L21602.

Cloern, J. E., Knowles, N., Brown, L. R., Cayan, D., Dettinger, M. D., Morgan, T. L., Schoellhamer, D. H., Stacey, M. T., van der Wegen, M., Wagner, R. W., and Jassby, A. D., 2011. Projected evolution of California's San Francisco Bay-Delta-River system in a century of climate change. *PLoS ONE*, **6**, e24465.

Cook, B. I., Smerdon, J. E., Seager, R., and Cook, E. R., 2014. Pan-continental droughts in North America over the last millennium. *Journal of Climate* , **27**, 383–397, doi: http://dx.doi.org/10.1175/JCLI-D-13-00100.1.

Cronin, T. M., 2012. Rapid sea-level rise. *Quaternary Science Reviews*, **56**, 11–30.

Cronin, T. M., and Walker, H., 2006. Restoring coastal ecosystems and abrupt climate change. *Climatic Change*, **74**, 369–374.

Cronin, T. M., Dwyer, G. S., Schwede, S. B., and Dowsett, H., 2002. Climate variability from Florida Bay Sedimentary Record: Possible teleconnections to ENSO, PNA, and CNP. *Climate Research*, **19**, 233–245.

Cronin, T. M., Thunell, R., Dwyer, G. S., Saenger, C., Mann, M. E., Vann, C., and Seal, R. R., II, 2005. Multiproxy evidence of Holocene climate variability from estuarine sediments, eastern North America. *Paleoceanography*, **20**, PA4006.

Davis, J., Laird, B., Ruzecki, E. P., Schubel, J. R., Huggett, R. J., Anderson, A. M., Wass, M. L., Marasco, R. J., and Lynch, M. P. (eds.), 1977. *The Effects of Hurricane Agnes on the Chesapeake Bay Estuarine System*. Baltimore: The Johns Hopkins University Press. Chesapeake Research Consortium Publication No. 54.

Diaz, R. J., and Rosenberg, R., 2008. Spreading dead zones and consequences for marine ecosystems. *Science*, **321**, 926–929.

Doxaran, D., Froidefond, J.-M., Castaing, P., and Babin, M., 2009. Dynamics of the turbidity maximum zone in a macrotidal estuary (the Gironde, France): observations from field and MODIS satellite data. *Estuarine, Coastal and Shelf Science*, **81**, 321–332.

Drinkwater, K. F., Belgrano, A., Borja, A., Conversi, A., Edwards, M., Greene, C. H., Ottersen, G., Pershing, A. J., and Walker, H., 2003. The response of marine ecosystems to climate variability associated with the North Atlantic Oscillation. In Hurrell, J. W., Kushnir, Y., Ottersen, G., and Visbeck, M. (eds.), *The North Atlantic Oscillation: Climatic Significance and Environmental Impact*. Washington, DC: American Geophysical Union. Geophysical Monograph, Vol. 134, pp. 211–234.

Enfield, D. B., Mestas-Nunez, A. M., and Trimble, P. B., 2001. The Atlantic multidecadal oscillation and its relation to rainfall and river flows in the continental U.S. *Geophysical Research Letters*, **28**, 2077–2080.

Feely, R. A., Alina, S. R., Newton, J., Sabine, C. L., Warner, M., Devol, A., Krembs, C., and Maloy, C., 2010. The combined effects of ocean acidification, mixing, and respiration on pH and carbonate saturation in an urbanized estuary. *Estuarine, Coastal and Shelf Science*, **88**, 442–449.

Gale, S. J., and Hoare, P. G., 2012. The stratigraphic status of the Anthropocene. *The Holocene*, **22**, 1491–1494.

Garcia, H. E., Boyer, T. P., Levitus, S., Locarnini, R. A., and Antonov, J., 2005. On the variability of dissolved oxygen and apparent oxygen utilization content for the upper world ocean: 1955 to 1998. *Geophysical Research Letters*, **32**, L09604.

Geyer, W. R., 1993. The importance of suppression of turbulence by stratification on the Estuarine Turbidity Maximum. *Estuaries*, **16**, 113–125.

Gilbert, D., Rabalais, N. N., Díaz, R. J., and Zhang, J., 2010. Evidence for greater oxygen decline rates in the coastal ocean than in the open ocean. *Biogeosciences*, **7**, 2283–2296.

Gobler, C. J., and Talmage, S. C., 2013. Short and long term consequences of larval stage exposure to constantly and ephemerally elevated carbon dioxide for marine bivalve populations. *Biogeosciences*, **10**, 2241–2253.

Gooday, A. J., Jorissen, F., Levin, L. A., Middelburg, J. J., Naqvi, S. W. A., Rabalais, N. N., Scranton, M., and Zhang, J., 2009. Historical records of coastal eutrophication-induced hypoxia. *Biogeosciences*, **6**, 1707–1745.

Grebmeier, J. M. 2012. Shifting patterns of life in the Pacific Arctic and sub-arctic seas. *Annual Review Marine Science*, **4**, 63–78.

Greene, C. H., and Pershing, A. J., 2007. Climate drives sea change. *Science*, **315**, 384–385.

Hanna, E., Navarro, F. J., Pattyn, F., Domingues, C. M., Fettweis, X., Ivins, E. R., Nicholls, R. J., Ritz, C., Smith, B., Tulaczyk, S., Whitehouse, P. L., and Zwally, H. J., 2013. Ice-sheet mass balance and climate change. *Nature*, **498**, 51–59.

Harding, J. M., Spero, H. J., Mann, R., Herbert, G. S., and Sliko, J. L., 2010. Reconstructing early 17th century estuarine drought conditions from Jamestown oysters. *Proceedings of the National Academy of Sciences*, **107**, 10549–10554.

Helmuth, B., Harley, C. D. G., Halpin, P. M., Oandapos, D. M., Hofmann, G. E., and Blanchette, C. A., 2002. Climate change and latitudinal patterns of intertidal thermal stress. *Science*, **298**, 1015–1017.

Hendriks, I. E., Duarte, C. M., and Álvarez, M., 2010. Vulnerability of marine biodiversity to ocean acidification: a meta-analysis. *Estuarine, Coastal and Shelf Science*, **86**, 157–164.

Hilton, T. W., Najjar, R. G., Zhong, L., and Li, M., 2008. Is there a signal of sea-level rise in Chesapeake Bay salinity? *Journal of Geophysical Research*, **113**, C09002.

Hoegh-Guldberg, Q. O., Mumby, P. J., Hooten, A. J., Steneck, R. S., Greenfield, P., Gomez, E., Harvell, C. D., Sale, P. F., Edwards, A. J., Caldeira, K., Knowlton, N., Eakin, C. M., Iglesias-Prieto, R., Muthiga, N., Bradbury, R. H., Dubi, A., and Hatziolos, M. E., 2007. Coral Reefs under rapid climate change and ocean acidification. *Science*, **318**, 1737–1742.

Hong, B., and Shen, J., 2012. Responses of estuarine salinity and transport processes to potential future sea-level rise in the Chesapeake Bay. *Estuarine, Coastal and Shelf Science*, **104–105**, 33–45.

Hönisch, B., Ridgewell, A., Schmidt, D. N., Thomas, E., Gibbs, S. J., Slujif, A., Zeebe, R., Kump, L., Martindale, R. C., Greene, S. E., Kiessling, W., Ries, J., Zachos, J. C., Royer, D. L., Barker, S., Marchitto, T. M., Jr., Moyer, R., Pelejero, C., Ziveri, P., Foster, G. L., and Williams, B., 2012. The geological record of ocean acidification. *Science*, **335**, 1058–1063.

Hostetler, S. W., Alder, J. R., and Allan, A. M., 2011. *Dynamically downscaled climate simulations over North America: methods, evaluation, and supporting documentation for users*. U.S. Geological Survey Open-File Report 2011-1238.

Howarth, R., Chan, F., Conley, D. J., Garnier, J., Doney, S. C., Marino, R., and Billen, G., 2011. Coupled biogeochemical cycles: eutrophication and hypoxia in temperate estuaries and coastal marine ecosystems. *Frontiers in Ecology and the Environment*, **9**, 18–26.

IPCC, 2012. Managing the risks of extreme events and disasters to advance climate change adaptation. In Field, C. B., Barros, V., Stocker, T. F., Qin, D., Dokken, D. J., Ebi, K. L., Mastrandrea, M. D., Mach, K. J., Plattner, G.-K., Allen, S. K., Tignor, M., and Midgley, P. M. (eds.), *A Special Report of Working Groups I and II of the Intergovernmental Panel on Climate Change*. Cambridge: Cambridge University Press.

Kemp, W. M., Testa, J. M., Conley, D. J., Gilbert, D., and Hagy, J. D., 2009. Temporal responses of coastal hypoxia to nutrient loading and physical controls. *Biogeosciences*, **6**, 2985–3008.

Knowles, N., and Cayan, D. R., 2002. Potential effects of global warming on the Sacramento/San Joaquin watershed and the San Francisco estuary. *Geophysical Research Letters*, **29**, 38-1–38-4.

Konikow, L. J., 2011. Contribution of global groundwater depletion since 1900 to sea-level rise. *Geophysical Research Letters*, **38**, L17401.

Kroeker, K. J., Kordas, R. L., Crim, R. N., and Singh, G. G., 2010. Meta-analysis reveals negative yet variable effects of ocean acidification on marine organisms. *Ecology Letters*, **13**, 1419–1434.

Kroeker, K. J., Kordas, R. L., Crim, R., Hendriks, I. E., Ramajo, L., Singh, G. S., Duarte, C. M., and Gattuso, J.-P., 2013. Impacts of ocean acidification on marine organisms: quantifying sensitivities and interaction with warming. *Global Change Biology*, **19**, 1884–1896.

Kucera, M., Rosell-Melé, A., Schneider, R., Waelbroeck, C., and Weinelt, M., 2005. Multiproxy approach for the reconstruction of the glacial ocean surface (MARGO). *Quaternary Science Reviews*, **24**, 813–819.

Kump, L. R., Bralower, T. J., and Ridgwell, A., 2009. Ocean acidification in deep time. *Oceanography*, **22**, 94–107.

Kuwae, M., Okuda, N., Miyasaka, H., Omori, K., Takeoka, H., and Sugimoto, T., 2007. Decadal- to centennial-scale variability of sedimentary biogeochemical parameters in Kagoshima Bay, Japan, associated with climate and watershed changes. *Estuarine, Coastal and Shelf Science*, **73**, 279–289.

Lafferty, K. D., 2009. The ecology of climate change and infectious diseases. *Ecology*, **90**, 888–900.

Levitus, S., Antonov, J. I., Boyer, T. P., Baranova, O. K., Garcia, H. E., Locarnini, R. A., Mishonov, A. V., Reagan, J. R., Seidov, D., Yarosh, E. S., and Zweng, M. M., 2012. World ocean heat content and thermosteric sea level change (0–2000 m) 1955–2010. *Geophysical Research Letters*, **39**, L10603.

Lipp, E. K., Schmidt, N., Luther, M. E., and Rose, J. B., 2001. Determining the effects of El Niño-Southern Oscillation Events on coastal water quality. *Estuaries*, **24**, 491–497.

Ma, Z., Xu, Z., and Zhou, J., 2009. Effect of global warming on the distribution of *Lucifer intermedius* and *L. hanseni* (Decapoda) in the Changjiang Estuary. *Progress in Natural Science*, **19**, 1389–1395.

Mantua, N. J., Hare, S. R., Zhang, Y., Wallace, J. M., and Francis, R. C., 1997. A Pacific Interdecadal climate oscillation with impacts on salmon. *Bulletin of the American Meterological Society*, **78**, 1069–1079.

Milliman, J. D., and Farnsworth, K. L., 2011. *River Discharge to the Coastal Ocean – A Global Synthesis*. Cambridge: Cambridge University Press.

Morand, S., Owers, K. A., Waret-Szkuta, A., McIntyre, K. M., and Baylis, M., 2013. Climate variability and outbreaks of infectious diseases in Europe. *Nature Scientific Reports*, **3**, doi: 10.1038/srep017714.

Murphy, R. R., Kemp, W. M., and Ball, W. P., 2011. Long-term trends in Chesapeake Bay seasonal hypoxia, stratification, and nutrient loading. *Estuaries and Coasts*, **34**, 1293–1309.

Najjar, R., Patterson, L., and Graham, S., 2009. Climate simulations of major estuarine watersheds in the Mid-Atlantic region of the US. *Climatic Change*, **95**, 139–168.

Najjar, R. G., Pyke, C. R., Adams, M. B., Breitburg, D., Hershner, C., Kemp, M., Howarth, R., Mulholland, M. R., Paolisso, M., Secor, D., Sellner, K., Wardrop, D., and Wood, R., 2010. Potential climate-change impacts on the Chesapeake Bay. *Estuarine, Coastal and Shelf Science*, **86**, 1–20.

Nicholls, R. J., Wong, P. P., Burkett, V. R., Codignotto, J., Hay, J., McLean, R., Ragoonaden, S., and Woodroffe, C. D., 2007. Coastal systems and low-lying areas. Climate Change 2007: impacts, adaptation and vulnerability. In Parry, M. L., Canziani, O. F., Palutikof, J. P., van der Linden, P. J., and Hanson, C. E. (eds.), *Contribution of Working Group II to the Fourth Assessment Report of the Intergovernmental Panel on Climate Change*. Cambridge: Cambridge University Press, pp. 315–356.

Nixon, S. W., Fulweiler, R. W., Buckley, B. A., Granger, S. L., Nowicki, B. L., and Henry, K. M., 2009. The impact of changing climate on phenology, productivity, and benthic–pelagic coupling in Narragansett Bay. *Estuarine, Coastal and Shelf Science*, **82**, 1–18.

Paerl, H. W., Bales, J. D., Ausley, L. W., Buzzelli, C. P., Crowder, L. B., Eby, L. A., Fear, J. M., Go, M., Peierls, B. L., Richardson, T. L., and Ramus, J. S., 2001. Ecosystem impacts of three sequential hurricanes (Dennis, Floyd, and Irene) on the United States' largest lagoonal estuary, Pamlico Sound, NC. *Proceedings of the National Academy of Science*, **98**, 5655–5660.

Paerl, H. W., Hall, N. S., Peierls, B. L., Rossignol, K. L., and Joyner, A. R., 2013. Hydrologic variability and its control of phytoplankton community structure and function in two shallow,

coastal, lagoonal ecosystems: the Neuse and New River estuaries, North Carolina, USA. *Estuaries and Coasts*, doi:10.1007/s12237-013-9686-0.
Pelejero, C., Calvo, E., McCulloch, M. T., Marshall, J. F., Gagan, M. K., Lough, J. M., and Opdyke, B. N., 2005. Preindustrial to modern interdecadal variability in coral reef pH. *Science*, **309**, 2204–2207.
Pelejero, C., Calvo, E., and Hoegh-Guldberg, O., 2010. Paleo-perspectives on ocean acidification. *Trends in Ecology and Evolution*, **25**, 332–344.
Peltier, W. R., and Fairbanks, R. G., 2006. Global glacial ice volume and last glacial maximum duration from an extended Barbados sea level record. *Quaternary Science Reviews*, **25**, 3322–3337.
Pershing, A. J., Greene, C. H., Jossi, J. W., O'Brien, L., Brodziak, J. K. T., and Bailey, B. A., 2005. Interdecadal variability in the Gulf of Maine zooplankton community, with potential impacts on fish recruitment. *ICES Journal of Marine Science*, **62**, 1511–1523.
Prasad, M., Sapiano, M., Anderson, C., Long, W., and Murtugudde, R., 2010. Long-term variability of nutrients and chlorophyll in the Chesapeake Bay: a retrospective analysis, 1985–2008. *Estuaries and Coasts*, **33**, 1128–1143.
Przeslawski, R., Falkner, I., Ashcroft, M. B., and Hutchings, P., 2012. Using rigorous selection criteria to investigate marine range shifts. *Estuarine, Coastal and Shelf Science*, **113**, 205–212.
Rebolledo, L., Sepúlveda, J., Lange, C. B., Pantoja, S., Bertrand, S., Hughen, K., and Figueroa, D., 2008. Late Holocene marine productivity changes in northern Patagonia-Chile inferred from a - multi-proxy analysis of Jacaf channel sediments. *Estuarine, Coastal and Shelf Science*, **80**, 314–322.
Ries, J., Cohen, A. L., and McCorkle, D. C., 2009. Marine calcifiers exhibit mixed responses to CO_2-induced ocean acidification. *Geology*, **37**, 1131–1134.
Rose, J. B., Epstein, P. R., Lipp, E. K., Sherman, B. H., Bernard, S. M., and Patz, J. A., 2001. Climate variability and change in the United States: potential impacts on water and foodborne diseases caused by microbiologic agents. *Environmental Health Perspectives*, **109**, 211–221.
Saenger, C., Cronin, T. M., Thunell, R., and Vann, C., 2006. Modeling river discharge and precipitation from estuarine salinity in the northern Chesapeake Bay: application to Holocene paleoclimate. *The Holocene*, **16**, 1–11.
Saenger, C., Cronin, T. M., Willard, D., Halka, J., and Kerhin, R., 2008. Increased terrestrial to ocean sediment fluxes in the northern Chesapeake Bay with twentieth century land alteration. *Estuaries and Coasts*, **31**, 492–500.
Schmidt, N., Lipp, E. K., Rose, J. B., and Luther, M., 2001. Analysis of ENSO related trends in Florida precipitation and streamflow. *Journal of Climate*, **14**, 615–628.
Schubel, J. R., and Pritchard, D. W., 1986. Responses of upper Chesapeake Bay to variations in discharge of the Susquehanna River. *Estuaries*, **9**, 236–249.
Stahle, D. W., Cleaveland, M. K., Blanton, D. B., Therrell, M. D., and Gay, D. A., 1998. The lost colony and Jamestown drought. *Science*, **280**, 564–567.
Swart, P. K., Dodge, R. E., and Hudson, H. J., 1996. A 240-year stable oxygen and carbon isotopic record in a coral from South Florida: implications for the prediction of precipitation in southern Florida. *Palaios*, **11**, 362–375.
Syvitski, J. P. M., Vörösmarty, C. J., Kettner, A. J., and Green, P., 2005. Impact of humans on the flux of terrestrial sediment to the global coastal ocean. *Science*, **308**, 376–380.
Szczucinski, W., Zajaczkowski, M., and Scholten, J., 2009. Sediment accumulation rates in subpolar fjords – impact of post-Little Ice Age glaciers retreat, Billefjorden, Svalbard. *Estuarine, Coastal and Shelf Science*, **85**, 345–356.
Talmage, S. C., and Gobler, C. J., 2009. The effects of elevated carbon dioxide concentrations on the metamorphosis, size, and survival of larval hard clams (*Mercenaria mercenaria*), bay scallops (*Argopecten irradians*), and Eastern oysters (*Crassostrea virginica*). *Limnology and Oceanography*, **54**, 2072–2080.
Tamisiea, M. E., and Mitrovica, J. X., 2011. The moving boundaries of sea level change: understanding the origins of geographic variability. *Oceanography*, **24**, 24–39.
Uncles, R. J., Stephens, J. A., and Law, D. J., 2006. Turbidity maximum in the macrotidal, highly turbid Humber Estuary, UK: Flocs, fluid mud, stationary suspensions and tidal bores. *Estuarine, Coastal and Shelf Science*, **67**, 30–52.
Waldbusser, G. G., and Salisbury, J. E., 2014. Ocean acidification in the coastal zone from an organism's perspective: multiple system parameters, frequency domains, and habitats. *Annual Reviews in Marine Science*, **6**, 11.1–11.27.
Waldbusser, G. G., Voigt, E. P., Bergschneider, H., Green, M. A., and Newell, R. I. E., 2011. Biocalcification in the Eastern Oyster (*Crassostrea virginica*) in relation to long-term trends in Chesapeake Bay pH. *Estuaries and Coasts*, **34**, 221–231.
Willard, D. W., and Cronin, T. M., 2007. Paleoecology and ecosystem restoration: case studies from Chesapeake Bay and the Florida Everglades. *Frontiers in Ecology and the Environment*, **5**, 491–498.
Willis, J. K., Chambers, D. P., Kuo, C.-Y., and Chum, C. K., 2010. Global sea level rise. *Oceanography*, **23**, 26–35.
Wittmann, A. C., and H.-O. Pörtner, 2013. Sensitivities of extant animal taxa to ocean acidification. *Nature Climate Change*, **3**, 995–1001.
Xu, J., Long, W., Wiggert, J. D., Lanerolle, L. W. J., Brown, C. W., Murtugudde, R., and Hood, R. R., 2012. Climate forcing and salinity variability in Chesapeake Bay, USA. *Estuaries and Coasts*, **35**, 237–261.

Cross-references

Barrier Island
Estuarine Circulation
Estuarine Geomorphology
Eutrophication
River-Dominated Estuary
Saltmarshes
Sediment Erosion
Sediment Transport
Shoreline Changes
Storm Surges

COASTAL BARRIERS

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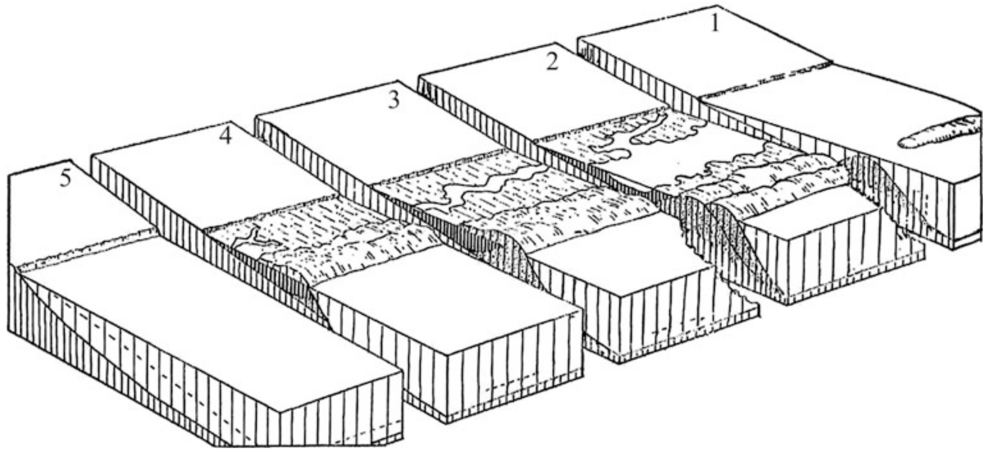
Definition

A "coastal barrier" is a barrier that lies between a sea/lake/lagoon and some landform or feature that is non-coastal or at least more landward than the immediate modern or Holocene coastal landform or group of landforms. It may



Coastal Barriers, Figure 1 Stages in the development and retrogression of a barrier. (After Davis (1912) from Johnson (1919)). The models illustrate some of the barrier types from attached barrier (*5*) to barrier islands (*1*).

be that the next landward feature is another coastal barrier or ancient rocks. In this use of the term, it is merely a barrier between the water and other land or landforms, and it is likely consistent with the original use of the term (Johnson, 1919).

Description

In the American literature, it has been common to refer to a barrier as a barrier island, due to the predominance of these types of barriers on the East Coast of the USA. However, a barrier island is only one type of a suite of types of coastal barriers. Dillenburg and Hesp (2009) state that "a coastal barrier is a shore parallel structure, formed by an accumulation of sand, gravel, shells, and small amounts of organic material due to the action of waves, tides and winds" (p. 1). In some cases, a barrier may not be shore parallel, particularly where spits are forming (Zenkovich, 1967).

Hesp and Short (1999) define a coastal barrier as "a shore-parallel, sub-aerial and sub-aqueous accumulation of detrital sediment formed by waves, tides and aeolian processes. It constitutes a definable coastal landform or sequence of landforms which is clearly separate in age, lithology, and/or form from adjacent, underlying or landward landforms. The barrier may block off or impound drainage from the hinterland, but this is not a prerequisite for definition as a barrier" (p. 308). Coastal barriers may be progradational (building seawards), retrogradational (eroding landwards), or stable (Morton, 1994), and they may be transgressive where sea level is rising and regressive where sea level is falling (Hesp and Short, 1999).

There are a variety of coastal barrier types and many varying terms for these types. On an evolutionary continuum, coastal barriers range from barrier islands (i.e., a barrier separated from the mainland by a lagoon or sea with no connections to the mainland at either end, thus a true island) to attached barriers (i.e., barriers that are attached to the mainland and may merely be a beach or have dune fields which transgress or climb the mainland terrain) (Figure 1). Where the barrier is predominantly a beach and attached to the mainland, it has also been termed a bayhead beach (Johnson, 1919) and mainland beach (Roy et al., 1994). In between these types are various types such as barrier spits (barrier connected at one end to the mainland) and bay barriers (barrier connected at both ends and extending across a bay) (Shepard, 1960; Dillenburg and Hesp, 2009).

The subaerial morphology of coastal barriers may range from a beach and backshore, overwash terrace, overwash fans and nebkha (discrete small dunes), beach and foredune, beach and beach ridge, multiple foredunes (relict foredune plains), beach ridge plains, foredune and blowouts, parabolic dunefields and transgressive dunefields, or combinations of these. All of the dune types may be found on progradational, retrogradational, or stable barriers.

Bibliography

Davis, W. M., 1912. *Die Erklarende Beschreibung der Landformen*. Berlin: Teubner.

Dillenburg, S. R., and Hesp, P. A., 2009. Coastal barriers: an introduction. In Dillenburg, S. R., and Hesp, P. A. (eds.), *Geology and Geomorphology of Holocene Coastal Barriers of Brazil*. Berlin: Springer. Lecture Notes in Earth Sciences, Vol. 107, pp. 1–15.

Hesp, P. A., and Short, A. D., 1999. Barrier morphodynamics. In Short, A. D. (ed.), *Handbook of Beach and Shoreface Morphodynamics*. New York: Wiley, pp. 307–333.

Johnson, D. W., 1919. *Shore Processes and Shoreline Development*. New York: Wiley.

Morton, R. A., 1994. Texas barriers. In Davis, R. A., Jr. (ed.), *Geology of Holocene Barrier Island Systems*. New York: Springer, pp. 75–114.

Roy, P. S., Cowell, P. J., Ferland, M. A., and Thom, B. G., 1994. Wave dominated coasts. In Carter, R. W. G., and Woodroffe, C. D. (eds.), *Coastal Evolution, Late Quaternary Shoreline*

Morphodynamics. Cambridge: Cambridge University Press, pp. 121–186.
Shepard, F. P., 1960. Gulf coast barriers. In Shepard, F. P., Phleger, F. B., and van Andel, T. H. (eds.), *Recent Sediments, Northwest Gulf of Mexico*. Tulsa: American Association of Petroleum Geologists.
Zenkovich, V. P., 1967. *Processes of Coastal Development*. Edinburgh: Oliver and Boyd.

Cross-references

Backbarrier
Barrier Island

COASTAL BAYS

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Synonyms

Coastal lagoon; Estuary

Definition

Coastal bays are bodies of water of variable size, shape, and morphology formed by the indentation or concavity of the coastline of an ocean or sea. Some investigators include coastal lagoons or enclosed embayments in this definition (Anderson et al., 2010; Glibert et al., 2010). Coastal bays are substantially smaller than a bight (e.g., New York Bight), gulf (e.g., Gulf of Mexico), sea (Sea of Japan), or sound (Long Island Sound).

Description

A range of coastal bays exists worldwide from water bodies largely exposed to the sea to those totally enclosed or nearly totally enclosed by the sea. As stated by Oertel (2005), "headland shores are often seaward of an inundated shore forming an irregular coastline." These headlands usually consist of harder rocks or consolidated sediments more resistant to erosion than those underlying the inundated shore. As a result they form promontories, with the softer substrate, which erodes more rapidly, forming the concavity of the shoreline. This process can lead to the formation of coastal bays.

Bibliography

Anderson, I. C., Stanhope, J. W., Hardison, A. K., and McGlathery, K. J., 2010. Sources and fates of nitrogen in Virginia coastal bays. In Kennish, M. J., and Paerl, H. W. (eds.), *Coastal Lagoons: Habitats of Environmental Change*. Boca Raton: CRC Press, Taylor and Francis, pp. 43–72.
Glibert, P. M., Boyer, J. N., Heil, C. A., Madden, C. J., Sturgis, B., and Wazniak, C. S., 2010. Blooms in lagoons: different from those of river-dominated estuaries. In Kennish, M. J., and Paerl, H. W. (eds.), *Coastal Lagoons: Habitats of Environmental Change*. Boca Raton: CRC Press, Taylor and Francis, pp. 91–114.
Oertel, G. F., 2005. Coastal lakes and lagoons. In Schwartz, M. L. (ed.), *Encyclopedia of Coastal Science*. Dordrecht: Springer, pp. 263–266.

Cross-references

Coastal Lagoons
River-Dominated Estuary
Tectonic Eustasy

COASTAL CLIFFS

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Synonyms

Fossil cliffs; Notches

Definition

Coastal cliffs are steep erosional slopes bordering the sea or estuary, in rocky as well as in sedimentary environments. They are common features all over the world (Trenshaile, 1987; Sunamura, 1992; Griggs and Trenshaile, 1994).

Description

In coastal erosive environments, the shore often takes the form of a steep slope or cliff (smaller forms ~1 m in size are termed notches). When formed in rocks, the cliff may terminate in the sea or estuary. In most cases, there is a small active shore just beneath the cliff. In uplifted areas, fossil cliffs (or dead cliffs) may occur. They reflect previous sea-level positions. Along subsiding coasts, one may find submarine drawn cliffs.

Cliff erosion usually feeds lateral accumulation of beach material by long-shore drift. The position with respect to mean sea level of the breaking point between an actively wave-washed shore and a steeply rising cliff foot depends on a number of different dynamic factors (which may change over time).

Bibliography

Griggs, G. B., and Trenshaile, A. S., 1994. Coastal cliffs and platforms. In Carter, R. W. G., and Woodroffe, C. D. (eds.), *Coastal Evolution*. Cambridge: Cambridge University Press, pp. 425–450.
Sunamura, T., 1992. *Geomorphology of Rocky Coasts*. Chichester: Wiley.
Trenshaile, A. S., 1987. *The Geomorphology of Rocky Coasts*. Oxford: Oxford University Press.

Cross-references

Coastal Landforms
Uplifted Coasts

COASTAL EROSION CONTROL

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Definitions

Coastal erosion is a natural or anthropogenic process in which sediment is worn away from the shoreline and seafloor due to natural and anthropogenic factors, such as storms, boat wakes, tidal currents, and rising sea levels.

Erosion control refers to erosion mitigation techniques based on soft and hard structural shoreline stabilization methods and nonstructural measures.

Hard structural stabilization refers to shoreline erosion control approaches based on the construction of man-made structures, such as seawalls, breakwaters, and groins.

Soft structural stabilization refers to shoreline erosion mitigation and control measures based on soft methods, namely, sand, pebble, or gravel fill, such as beach nourishment.

Nonstructural measures refer to any coastal erosion control strategy that does not involve man-made construction or other physical measures, but is based on good practices, policies, and education aimed at reducing anthropogenic and natural impacts, such as land use restriction and zoning.

Introduction

Estuaries are transitional zones where marine and riverine environments meet, i.e., where freshwater from a river mixes with saltwater from the sea. Here, many habitats, species, and ecological communities exist, and their ecosystem and naturalistic values are widely recognized. Along coasts, river banks, and nearshore profiles are governed by sediment transport equilibrium (i.e., erosion and deposition phenomena). Alteration of this equilibrium may result in shoreward recession, leading to land loss. Since estuarine areas are often used for tourism and recreational purposes, this leads to financial loss as well as ecological community and biodiversity impairment. State and local authorities thus aim to protect estuarine areas and to mitigate erosion. A review of the causes of coastal erosion and an evaluation of the erosion processes are provided below, together with an analysis of erosion control measures.

Causes of coastal erosion

There are many causes of coastal erosion processes attributable to natural and anthropogenic factors. A classification can be based on the temporal scale of such factors, distinguishing between short- and long-term events. Natural processes consist of short-term events that are generally the result of storms and river floods (i.e., high-energy content events), while long-term events relate to sea-level rise (Pranzini and Rossi, 1995; Khalil, 1997), tidal cycles, tectonic events, coastal subsidence (Khalil, 1997), climate change, river regimes, and discharge flux (Medina and Lopez, 1997).

In regard to anthropogenic factors, sediment loss is generally due to medium- and long-term events, such as decreasing sediment supply to coastal physiographic units (Simeoni et al., 1997; Eronat, 1999; Loizidou and Iacovou, 1999), deforestation in coastal and riverine watersheds (Eronat, 1999), non-sustainable man-made coastal structures and urban development (Fathallah and Gueddari, 2001; Rakha and Abul-Azm, 2001), flow regime and engineering structure changes, and riverbed sand and gravel extraction (PAP/RAC, 2000).

Erosion processes

Suspension and bed load transport can be distinguished (Fredsøe and Deigard, 1994), the latter mainly causing the loss of grain size material on the seabed and affecting long-term and short-term shoreline evolution. Such phenomena are primarily modeled by physical and mathematical and generally consider hydrodynamic and morphological factors. Hydrodynamic models are based on the classical equations of motion, with vertical averaged velocities (two-dimension models), and wave propagation, refraction, diffraction, and breaking. They evaluate velocity flow fields under generic forcing (Figure 1).

Morphological models consider seabed characteristics and evaluate bed load and shoreline evolution, for both long-term and short-term conditions (Figure 2).

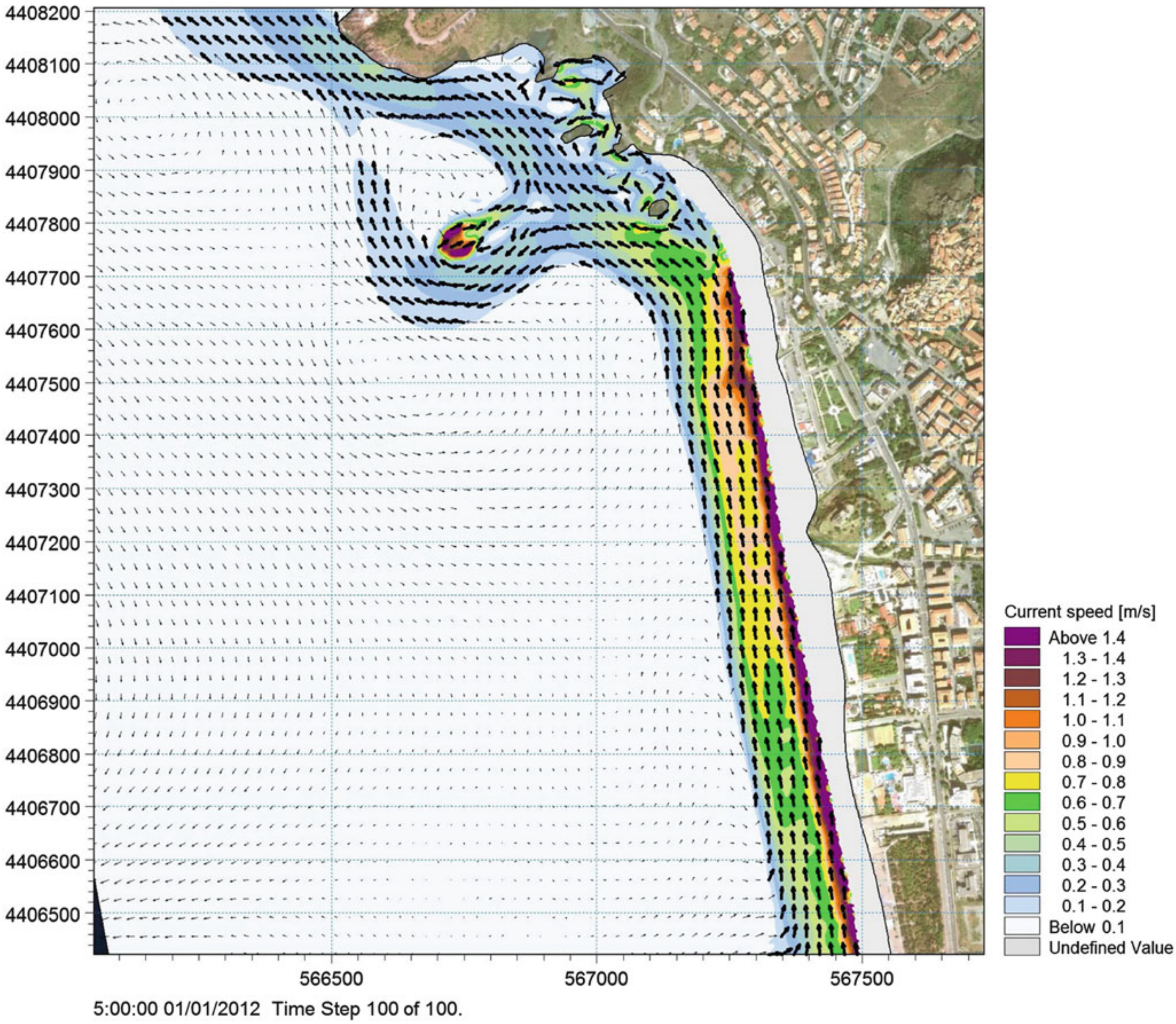
Physical models are based on two-dimensional and three-dimensional laboratory scale similarity models that use channels or large basins (Figure 3) to study both bed load processes and evaluate the effectiveness of the structural design.

Finally, it should be pointed out that both physical and mathematical models require a detailed knowledge of seabed morphology, waves and currents (i.e., river or tidal currents or wave-generated currents), bathymetric surveys, and sediment characterization (pebble or sand beaches, sediment grain size, and erodibilty changes).

Coastal erosion risk mitigation

A coastal erosion control project should be developed with reference to coastal cells (Eurosion, 2004), which are lengths of coastlines in which a complete sediment balance can be identified. Spatial and temporal erosion phenomena scales should also be identified. Acute and structural erosion can be distinguished. Acute erosion is connected to waves, wind, and tidal action and typically covers temporal and spatial scales up to 1 month and from about 1 m to 100 km, respectively. On the contrary, structural erosion involves larger temporal and spatial scales, from about 1 month to 100 years and from 1 to 1,000 km, respectively (Safecoast, 2008).

Depending on spatial scale, coastal erosion control measures can target (1) specific river bank sections and



Coastal Erosion Control, Figure 1 Numerical results of hydrodynamic coastal field at Scalea beach, Italy.

(2) larger areas or entire estuarine areas. Moreover, erosion control measures should consider the environment energy level, distinguishing between low-energy areas and ocean-facing beaches, the latter being higher-energy systems. This distinction is useful for focusing on the best erosion control project, which is strongly site specific. Coastal erosion control methods can be mainly classified as structural and nonstructural measures. Structural measures can then be further classified as “hard” and “soft” alternatives. Structural measures involve permanent concrete, rock, or wooden engineering structures, since nonstructural measures generally refer to best management practices or to actions not involving man-made structures.

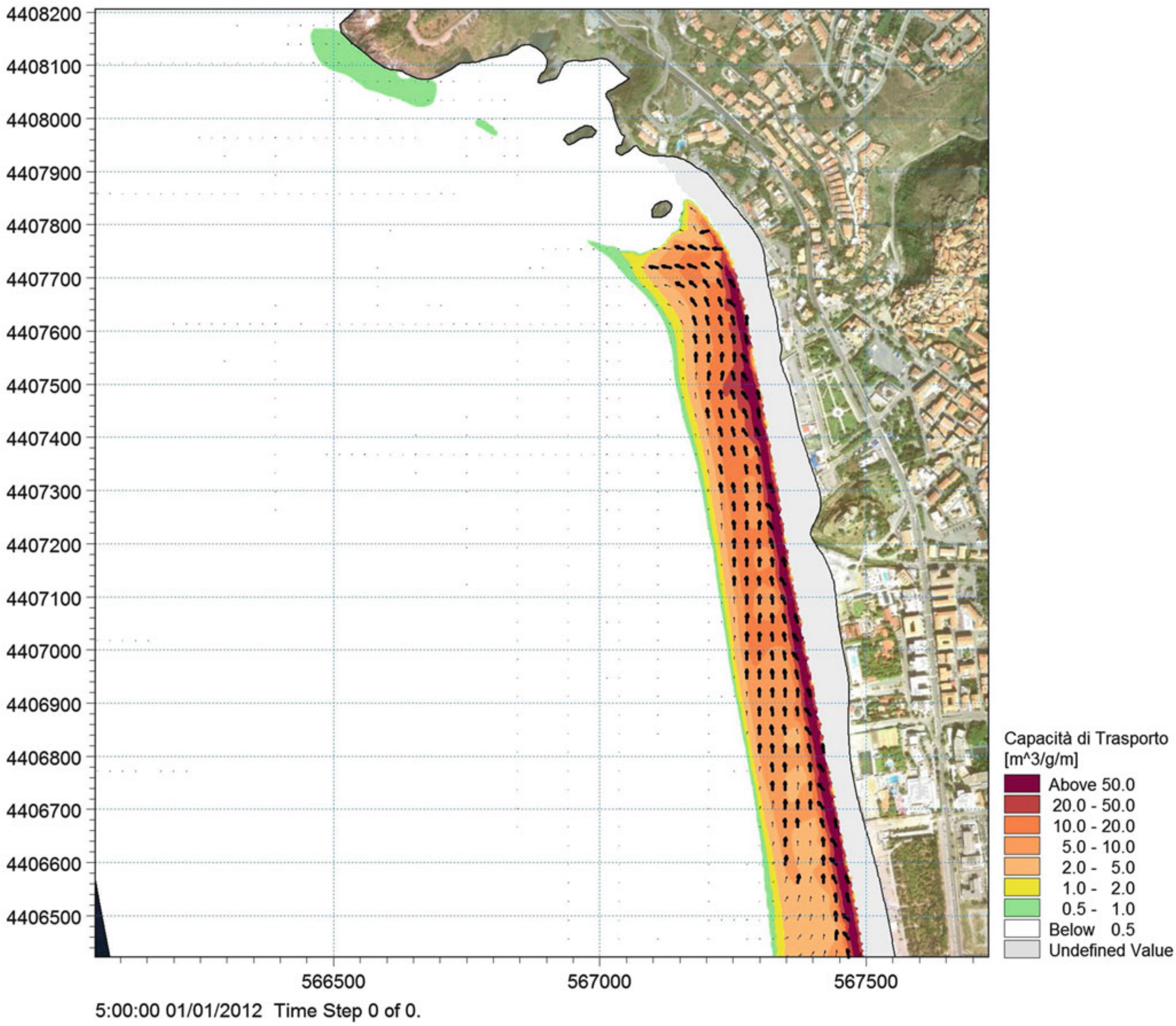
The main hard erosion control measures are:

- Breakwaters
- Revetments, seawalls, and bulkheads
- Groins

These structures are designed to protect the areas behind them by “fixing” the shoreline or by modifying the flow-field circulation by “trapping” sediment.

The main soft erosion control measures are:

- Nourishment
- Rip-rap, gabions, and paved-lining revetments
- Marsh sills
- Planting vegetation

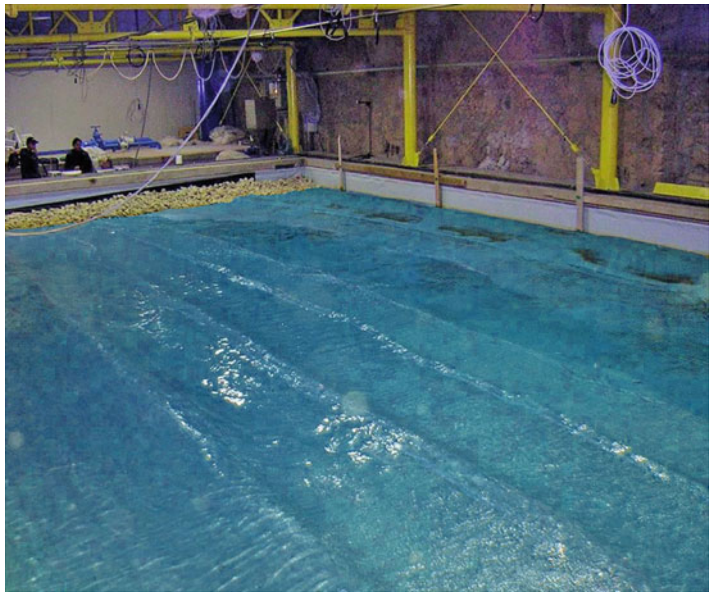


Coastal Erosion Control, Figure 2 Numerical results of bed load transport at Scalea beach, Italy.

Rip-rap, gabions, paved-lining revetments, groins, marsh sills, and planting vegetation are generally effective for river bank protection, while in wave-exposed areas, coastal erosion control measures should be chosen depending on whether cross-shore or long-shore transport is dominant, with a preference for shorel-parallel structures (such as revetments, attached and detached breakwaters) or perpendicular structures (such as groins), respectively. Furthermore, good coastal planning and management practices that promote land use rehabilitation targets and eco-sustainable tourism are also suitable coastal erosion control measures. Lastly, the choice of erosion control method is strongly dependant on the environmental, social, and cultural characteristics of the areas concerned and on legislative, policy, and economic aspects. Thus, the combination of different methods is often an effective strategy.

Structural measures: hard stabilization

Hard stabilization methods are based on protective structures designed to stabilize the shore and to prevent waves and tides from reaching the area or to trap sediment. Such structures typically act by reducing incident wave energy and changing flow-field circulation. Stabilization effectiveness is strongly dependant on wave exposure, the latter being influenced by structure inclination and orientation (e.g., perpendicular or parallel to the coastline),

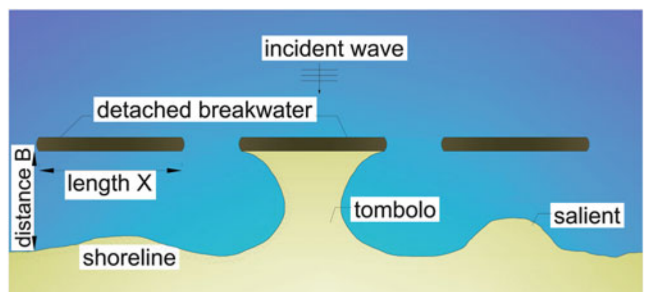


Coastal Erosion Control, Figure 3 Large basin simulation: example of an experimental investigation.

permeability/impermeability, overtopping, etc. The main hard structures are described below.

Breakwaters

Breakwaters (Figure 4) are structures designed to protect shorelines from erosion by shielding waves (i.e., reducing incident wave energy) and changing littoral transport conditions. These structures can be directly connected to the shoreline or constructed shore-parallel, respectively, attached or detached. In the first case, breakwaters act as a revetment, protecting adjacent upland areas against scour induced by waves and currents, while detached breakwaters act by allowing sand accumulation from the original shoreline to the landward breakwater. Such sand accumulation is called a "tombolo" or "salient," depending on whether or not the breakwater is reached by sand. Accretionary beach features are characterized by a dimensionless ratio X/B (Herbich, 1991), X representing the breakwater length and B the breakwater distance from the original shoreline. For $X/B > 1$, sediment deposition and accumulation behind breakwater forms until the shoreline is connected to the structure (permanent tombolo), while for $X/B < 1$, sediment forms from the shoreline in the lee of the structure, without reaching the breakwater (salient). Since sediment transport phenomena are in equilibrium, sand accumulation leads to the formation of an erosion zone. With reference to design and construction aspects, breakwaters can be classified as follows: (1) rock or concrete units with trapezoidal cross sections, (2) prefabricated triangular-shaped concrete units, and (3) sand-filled containers (caissons) with geotextile units.

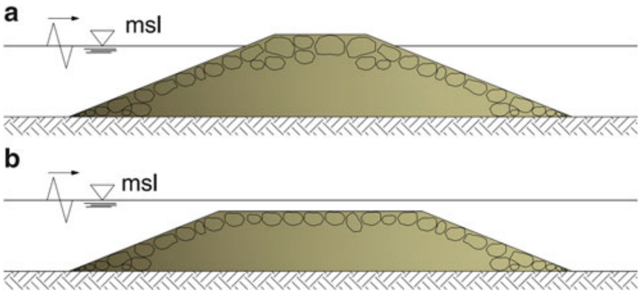


Coastal Erosion Control, Figure 4 Detached breakwaters: definition sketch.

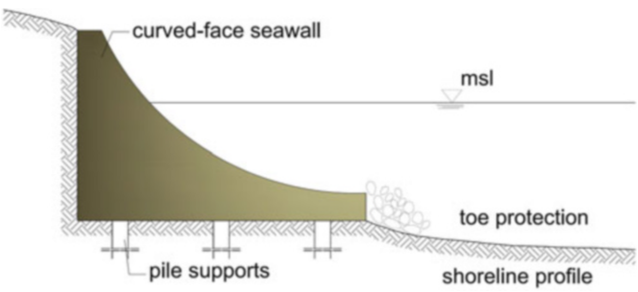
Breakwaters can be emergent and submerged, depending on crest position with respect to water level, namely, if the crest is positioned entirely below or above mean sea level (Figure 5). Both emergent and submerged breakwaters have a strong impact on sediment fluxes and on morphodynamic evolution, the latter depending on structure-induced waves, circulation fields, and wave overtopping. In order to avoid negative morphological effects, such as local scours, these aspects should be preliminarily and carefully analyzed. Lastly, it should be pointed out that during storm conditions, breakwater cannot stop or dissipate most of the waves, which result in low effectiveness, and thus other methods such as supplementary nourishment may be required.

Revetments, seawalls and bulkheads

Revetments, seawalls, and bulkheads are shore-parallel structures built adjacent to the shoreline. The main



Coastal Erosion Control, Figure 5 (**a**) Emergent; (**b**) submerged breakwaters.



Coastal Erosion Control, Figure 6 Seawall.

difference lies in their functional aspects (U.S. Army Corps of Engineers, 1984). In fact, revetments are stone or concrete structures built adjacent to the shoreline and are designed to protect the underlying soil from erosion. A larger stone layer is generally placed on the frontwater, with a smaller layer filter placed below it. The latter prevents underlying soil washing, while the main erosion protection is guaranteed by the upper stone layer.

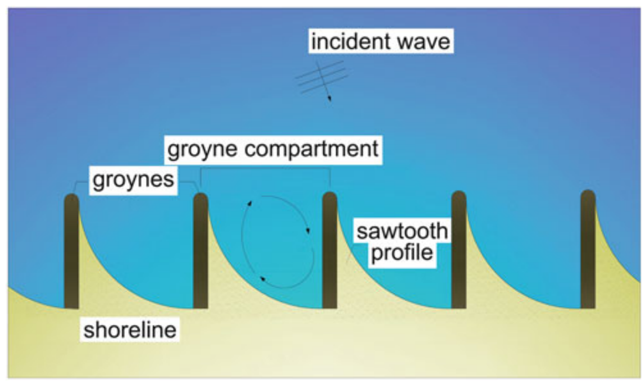
Seawalls are primarily designed to protect the shore against wave action; bulkheads are retaining walls designed to provide protection in low-to-moderate wave energy environments. More specifically, seawalls are massive concrete or stone, vertical or sloped structures (Figure 6), with rubble, curved, or stepped face. A combination stepped-curved face may also be constructed. Revetment and seawall slopes should be no steeper than 1:3, and the length of the structure should be carefully selected to avoid erosion of the adjacent coastline.

Bulkheads (Figure 7) are designed to protect adjacent upland areas and to retain the soil behind them. A sufficiently large embedded wall is required, and tie rods may be used to increase the stability of the structure. Toe protection is required for all these structures so as to prevent local scour.

Groins

Groins are long, narrow coastal structures used both on open beaches and in estuaries, altering nearshore tidal flow patterns and deflecting currents. They are placed

Coastal Erosion Control, Figure 7 Wooden bulkheads (Rogers and Skrabal, 2001).



Coastal Erosion Control, Figure 8 Groins: definition sketch.

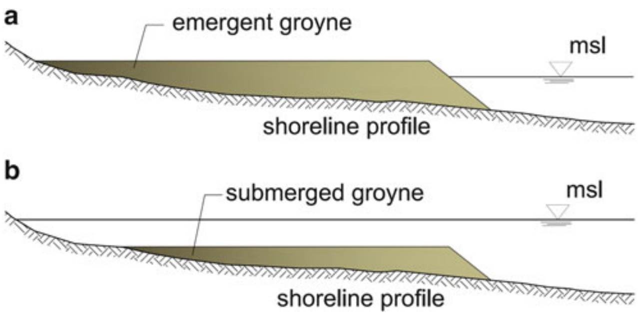
perpendicular or slightly perpendicular to the shoreline (Figure 8). Such structures are generally constructed in groups, so that compartments between adjacent groins can be identified, trapping sediments in each of them and extending coast longevity.

Groins are most effective when long-shore currents prevail in one direction.

Generally, two kinds of groins (Figure 9) can be distinguished (van Rijn, 2010; van Rijn, 2011):

- Impermeable, high-crested structures, usually made of sheet piling or concrete structures. Crest levels are 1 m above MSL (mean sea level). A full long-shore current block is expected, so that the sand within each compartment is retained. A typical sawtooth shoreline profile should be created, thus increasing groin spacing.
- Permeable, low-crested structures, with crest level between MLW (mean low water) and MHW (mean high water). These structures act as flow resistance, reducing the littoral drift in the inner surf zone. A regular shoreline should be created.

Groins should only be constructed along coasts with recession rates in excess of 2 m/year. Their length should moreover be extended over the inner surf zone



Coastal Erosion Control, Figure 9 (**a**) Emergent; (**b**) submerged groins.

(Basco and Pope, 2004; Kana et al., 2004). High-crested, impermeable groin length (*L*) and spacing (*S*) typically varies between 50 and 100 m and between 1.5 and 3 times the length of the groin, respectively (van Rijn, 2011). Finally, groins induce local scour at the toe of the structures and thus require regular maintenance.

Structural measures: soft stabilization

Nourishment

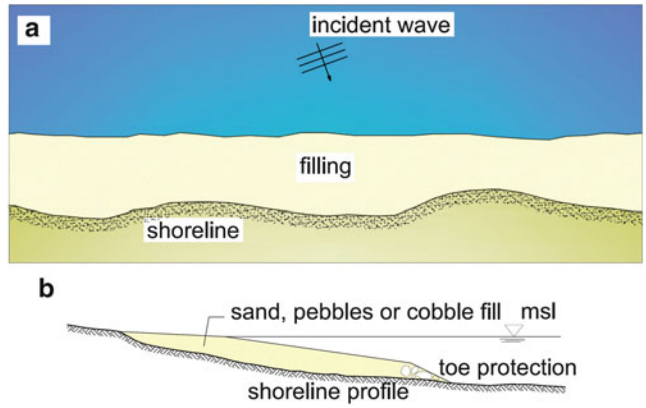
Nourishment is a shoreline stabilization method using sand, pebbles, or gravel beach fill (Figure 10). A key parameter of successful nourishment design is the choice of fill material. In fact, sediment should have the following main characteristics (Department of Boating and Waterways and State Coastal Conservancy, 2002):

- No contamination
- Fine grain size fraction
- Grain size comparable to or larger than in situ material

With reference to uncontaminated sediments, the introduction of contaminated material to coastal systems not only compromises habitats and ecosystems but also creates financial loss in terms of tourism, these areas being often used for recreational purposes.

With regard to the choice of sediment grain size, sediment with comparable or larger than in situ material size characteristics is preferable. In fact, comparably, grain size material tends to have the same behavior as in situ material, while larger sediment generally results in a more stable solution. Vice versa, fine grain size usually results in less stable solutions and accelerated erosion.

Sediment grain size and contamination level are strongly dependant on the source of the nourishment material. The latter generally includes dredged sediment from harbor construction and maintenance, lagoon restoration, and offshore and inland (e.g., damming rivers dredging) sources. Fill material may be placed (1) on the dry beach (dune nourishment); (2) on the beach cross section, on the dry portion and near the waterline, and across the entire beach cross section (i.e., above and below water); and (3) offshore as a sand bar (National Research Council, 1995).



Coastal Erosion Control, Figure 10 Nourishment: (**a**) plane view; (**b**) cross section.

In dune nourishment configuration 1, fill material is placed high above the waterline. This configuration provides effective protection against storm waves, but no expansion in dry beach width and no increase in recreational coastal areas. In configuration 2, an immediate increase in beach width (i.e., recreational areas) is observed. Furthermore, once placed, fill material is redistributed offshore and alongshore below wave and current action until a stable configuration is achieved. If fill material is placed both above and below the waterline, an already stable configuration is attempted, and there is little offshore sand redistribution, which leads to minimal changes in dry beach width. Finally, in configuration 3, fill material is placed in the surf zone, and the sand gradually moves onshore below wave and current action, thus increasing the beach width.

In regard to environmental aspects, nourishment can have a strong impact on aquatic habitats on the seabed, near the shoreline. Thus, species ecosystem response tolerance and the burial adaptability thereof should also be considered. In all cases described above, because of the nourishment design characteristics, waves and currents gradually remove some of the sediment, and periodic maintenance is required.

Rip-rap, gabions, and paved-lining revetments

Shoreline revetments may be constructed using rip-rap revetments (Figure 11), gabions, and paved linings (Figure 12) that are wire cages filled with stones and placed as revetment along the shoreline and river banks, in vertical stacked or sloped configurations.

Marsh sills

Marsh sills are shore-parallel structures designed to protect planted wetland vegetation. An offshore wood or rock mound (sill) and an intertidal area are created between the sill and upland (Figure 13). Protection is achieved thanks to existing or planted vegetation in the intertidal zone, which dissipates wave energy, preventing it from reaching the upland (Rogers and Skrabal, 2001). An added value of this strategy is its ability to promote the creation of natural habitats.

Planting vegetation

Vegetation plays an important role in promoting shoreline stabilization, reducing wave and current energy, and trapping incoherent sediment in radical apparatuses. Thus, planting vegetation is considered a shoreline stabilization method, although its effectiveness is strongly site specific. In fact, planting vegetation generally allows good erosion control to be achieved in low-energy environments, such as in estuarine tidal zones, while on the contrary, in high-energy environment, vegetation seems ineffective. Finally, there should be a preference for the planting of native species.

Coastal Erosion Control, Figure 11 Rip-rap revetments (Trowell, 2012).

Nonstructural measures

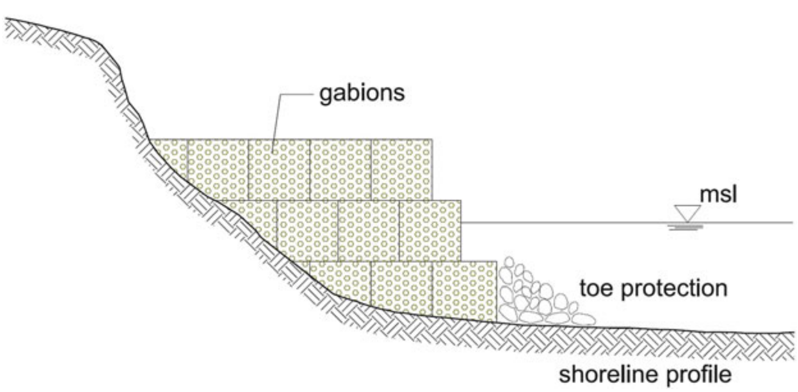
Policy and planning techniques

Policy and planning techniques for erosion control relate to strategies for coastal area use and anthropogenic pressure control, based on the introduction of sustainable coastal area management development logic. In fact, man's use of coastal and estuarine areas for promoting economic activities generally requires intensive development and accelerated estuarine area modification processes, often leading to natural equilibrium alteration and increased vulnerability. Policy planning techniques involve a large number of factors, the majority of which are described herein.

Firstly, estuarine area management project measures are based on specific environmental policies, legislations (e.g., European Bird, 1979 and Habitats, 1992 Directives, Natura 2000 ecological networks of protected areas to name but a few), and project management development. The latter is generally based on "prevention" and "protection" measures, attending to primary coastal erosion risk management and consisting of both structural and nonstructural measures (Safecoast, 2008). Examples of "prevention" strategies are relocation, zoning, space allocation and reservation, coastal erosion risk education, and communication and raising awareness. Examples of "protection" strategies are based on building and maintaining structures for erosion control (such as breakwaters, nourishment and groins).

Good project management should be based on a precise site evaluation and decision logic systems, with focus on the following aspects (Safecoast, 2008):

- Physical and environmental characteristics
- Economic and ecological values, assessed by stakeholders and others, such as engineers, scientists, politicians, land use planners, and the public affected
- Historical and cultural background
- Policy measures and a general set of rules capable of identifying different scenarios (e.g., land use restriction)



Coastal Erosion Control, Figure 12 Gabions: definition sketch.

Coastal Erosion Control, Figure 13 Marsh sills (Modified from Trowell, 2012).

- Coastal dynamics, effectiveness of erosion control measures, erosion phenomena, and spatial and temporal scales

On this basis, the resulting project will be the "optimum solution," chosen in continuity with existing management policies, protection measures, and related operational procedures, such as institutional arrangement, operational responsibilities, and financing.

Conclusions

Natural and anthropogenic factors cause alterations in coastal sediment transport equilibrium, thus resulting in coastal erosion, with ecological and financial loss. Furthermore, cultural identity, resources, and recreational coastal land use are strongly compromised or lost.

Hard and soft control measures are effective tools for mitigating such phenomena. Hard stabilization methods in particular mainly consist of concrete and rock structures such as groins, revetments, breakwaters, etc., while soft methods consist of shoreline protection measures based on beach nourishment and the planting of vegetation, thus allowing recreational tourist areas to be developed. Good practices in the planning and management of coastal areas are also effective coastal erosion control measures and important targets for many states. The choice of stabilization methods should be the "optimal" method among possible solutions and should be selected starting from a hydrodynamic, environmental, social, and cultural site characterization. Finally, the combined use of different stabilization methods enables considerable coastal erosion control objectives to be achieved.

Bibliography

Basco, D. R., and Pope, J., 2004. Groin functional design guidance from the coastal engineering manual. *Journal of Coastal Research*, (Special Issue 33) 121–130.

Department of Boating and Waterways and State Coastal Conservancy, 2002. California beach restoration study, ch. 4, pp. 3–6.

Eronat, A. H., 1999. Altinova Madra Creek region coastal erosion study. In *Proceedings of the MEDCOAST 99 – EMECS 99 Joint Conference: Land – Ocean Interactions – Managing Coastal Ecosystems*, Vol. 3, pp. 1525–1540.

EUROSION, 2004. Living with coastal erosion in Europe: sediment and space for sustainability. A guide to coastal erosion management practices in Europe. Report.

Fathallah, S., and Gueddari, M., 2001. Sedimentary processes and shoreline changes along the Sousse Coast, Eastern Tunisia. In *Proceedings of the Fifth International Conference on the Mediterranean Coastal Environment*, Vol. 3, pp. 1457–1466.

Fredsøe, J., and Deigaard, R., 1994. *Mechanical of Coastal Sediment Transport*. Singapore: World Scientific Publishing.

Herbich, J. B., 1991. *Coastal and Ocean Engineering*. Houston: Gulf Publishing Company.

Kana, T. W., White, T. E., and McKee, P. A., 2004. Management and engineering guidelines from groin rehabilitation. *Journal of Coastal Research* (Special Issue 33), 57–82.

Khalil, S., 1997. Critical problems of the Egyptian Mediterranean coastal zones. In *Proceedings of the Third International Conference on the Mediterranean Coastal Environment, MEDCOAST 97*, Vol. 1, pp. 513–521.

Loizidou X. I., and Iacovou, N. G., 1999. Anthropogenic coastal erosion and shoreline management in Cyprus. In *Proceedings of the MEDCOAST' 99-EMECS 99 Joint Conference: Land – Ocean Interactions -Managing Coastal Ecosystems*, Vol. 3, pp. 1501–1509.

Medina, J. M., and Lopez, J. S., 1997. Strong erosion scenario due to disequilibrium of solid transport rate: the case of Torrox Beach(Malaga). In *Proceedings of the Second International*

Conference on the Mediterranean Coastal Environment, MEDCOAST 97, Vol. 2, pp. 1234–1239.
National Research Council, 1995. *Beach Nourishment and Protection*. Washington, DC: National Academy Press.
PAP/RAC, 2000. Report on the Albanian coastal erosion, PAP/RAC Mission to Albania to Assess the Problem of Coastal Erosion-Consultants' Reports, UNEP/MAP Priority Actions Program Regional Activity Centre, 2000.
Pranzini, E., and Rossi, L., 1995. Physical model test in the framework of beach nourishment performance evaluation. In *Proceedings of the Second International Conference on the Mediterranean Coastal Environment'95*, Vol. 2, pp. 1145–1159.
Rakha, K. A., and Abul-Azm, A. G., 2001. Modelling of shoreline changes along the north western Egyptian coast. In *Proceedings of the Fifth International Conference on the Mediterranean Coastal Environment'01*, Vol. 3, pp. 1467–1478.
Rogers, S. M., and Skrabal, T. E., 2001. *Managing Erosion on Estuarine Shorelines*. Raleigh: North Carolina State University. The Soundfront Series, pp. 2–37.
Safecoast, 2008. Chapter 5: Strategy development to manage coastal risk. Synthesis Report Safecoast. The Hague: Safecoast project team, pp. 81–109.
Simeoni U., Pano N., and Ciavola P., 1997. The coastline of Albania: morphology, evolution and coastal management issues. In Evolution des côtes méditerranéennes, Bulletin de l'Institut Oceanographique, de Monaco, Commission Internationale pour l'Exploration Scientifique de la mer Méditerrannées, Monaco, 18, pp. 151–168.
Trowell, S., 2012. Estuarine shoreline stabilization design and technique. In *Estuarine Shorelines: Value, Regulations, and Stabilization*. Beaufort: N.C. Division of Coastal Management.
U.S. Army Corps of Engineers, 1984. *Shore Protection Manual*, 4th edn. Washington, DC: U.S. Army Engineer Waterways Experiment Station, Coastal Engineering Research Center/U.S. Government Printing Office, Vol. 2.
van Rijn, L. C., 2010. Coastal erosion control based on the concept of sediment cells. European Commission, Concepts and Science for Coastal Erosion Management, Conscience.
van Rijn, L. C., 2011. Coastal erosion and control. *Ocean & Coastal Management*, **54**(12), 867–887.

Cross-references

Bulkheads
Revetments
Sediment Grain Size
Shore Protection

COASTAL INDICATORS

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Synonyms

Coastal state indicators

Definition

Quantitative/qualitative statements or measured/observed parameters that can be used to describe existing situations and measure changes or trends over time (UNESCO, 2006, p. 11), concerning the state of an estuary or other coastal feature.

Aims and concepts

Indicators aim to convey a complex message in a simple manner. Their three main functions are to simplify the information, quantify the target system, and provide a facilitator tool in the communication process between different stakeholders (UNESCO, 2006). For this reason, coastal indicators constitute an extremely useful tool in coastal management since they can translate observations, models, and scientific interpretation – which are too complex and difficult to be used directly in the managing processes – in a simplified form to coastal managers. Therefore, indicators facilitate the integration of scientific know-how in coastal zone planning and management (e.g., UNESCO, 2003; UNESCO, 2006; NOAA, 2010) and reduce the risk of failure in the communication process between scientists and coastal managers (e.g., van Koningsveld, 2003; van Koningsveld et al., 2005; Jiménez, 2010). Thus, in the determination of adequate coastal indicators to describe a particular system, it is fundamental to receive the input of coastal managers and scientists in a joint effort to define the adequate indicators (van Koningsveld, 2003). The former are able to assess what information will be of most value to the management, while the latter can determine what might be possible to measure based on existing or potential technology and scientific understanding (van Rijn, 2010). As an example, shoreline position is one of the most commonly used indicators to determine coastal morphodynamic state.

Bibliography

Jiménez, J., 2010. *Coastal State Indicators at the Conscience Case Study Sites. Concepts and Science for Coastal Erosion Management (Conscience) Project*. Deliverable D17, Deltares, the Netherlands.
NOAA, 2010. *Coastal Zone Management Act - Performance Measurement System: Contextual Indicators Manual*. Charleston, SC: NOAA National Ocean Service, Office of Ocean and Coastal Resource Management.
UNESCO, 2003. *A Reference Guide on the Use of Indicators for Integrated Coastal Management*. IOC Manuals and Guides, 45; ICAM Dossier 1, Paris.
UNESCO, 2006. *A Handbook for Measuring the Progress and Outcomes of Integrated Coastal and Ocean Management*. IOC Manuals and Guides, 46; ICAM Dossier 2, Paris.
van Koningsveld, M., 2003. *Matching Specialist Knowledge with End User Needs*. PhD Thesis, Enschede, The Netherlands, University of Twente.
van Koningsveld, M., Davidson, M., and Huntley, D., 2005. Matching science with coastal management needs: the search for appropriate coastal state indicators. *Journal of Coastal Research*, **21**, 399–411.
van Rijn, L., 2010. *Description of Coastal State Indicators. Concepts and Science for Coastal Erosion Management (Conscience) Project*. Deliverable D9, Deltares, the Netherlands.

COASTAL LAGOONS

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Synonyms

Coastal bays; Coastal lakes; Coastal ponds

Definition

Coastal lagoons are shallow brackish or marine water bodies separated from the ocean by a barrier island, spit, reef, or sand bank (Colombo, 1977; Barnes, 1980; Kjerfve, 1994; Kennish and Paerl, 2010a). Depending on the extent of the barriers, they may be partially or totally enclosed, although most are connected at least intermittently to the open ocean by one or more restricted tidal inlets. Oertel (2005) called the smaller, totally enclosed systems coastal lakes or coastal ponds. Those with outlets to the sea are termed coastal lagoons and coastal bays, depending on their shapes.

Introduction

Coastal lagoons form on low-lying coasts such as along the Atlantic and Gulf coasts of the USA, where they are particularly extensive, covering ~2,800 km of shoreline (Nichols and Boon, 1994). They are much less common on most other coasts, occupying only ~12 % of the coastal shorelines worldwide. The Antarctic is the only continent devoid of coastal lagoons, while they are most prominent along the coasts of Africa (17.9 % of the coastline) and North America (17.6 %) and less conspicuous along the coasts of Asia (13.8 %), South America (12.2 %), Australia (11.4 %), and Europe (5.3 %) (Barnes, 1980; Kennish and Paerl, 2010a).

The size and shape of coastal lagoons vary considerably, although they are usually oriented with their long axis parallel to the shoreline, as exemplified by the Barnegat Bay-Little Egg Harbor system in New Jersey (USA) (Figure 1) (Kennish, 2001). However, some lagoonal water bodies have a triangular or delta shape with v-shaped landward margins, as demonstrated by the Rehoboth Bay and Assawoman Bay in Delaware (USA) (Oertel, 2005). They range in size from a few square kilometers up to 10,000 km^2 as in the case of the expansive Lagoa dos Patos in Brazil (Bird, 1994).

Formation

The genesis of coastal lagoons is closely linked to the formation of coastal barriers which separate flooded basins landward from the coastal ocean. According to de Beaumont (1845), the barriers form by the upbuilding of bars and shoals. Gilbert (1885) attributed barrier formation to the progradation of spits which creates shallow embayments behind them. McGee (1890) advanced an inundation model of coastal lagoon formation whereby a rising sea floods lowland areas. Oertel (2005) supported the models of Gilbert (1885) and McGee (1890) as the two main modes by which coastal lagoons form.

Physical-chemical characteristics

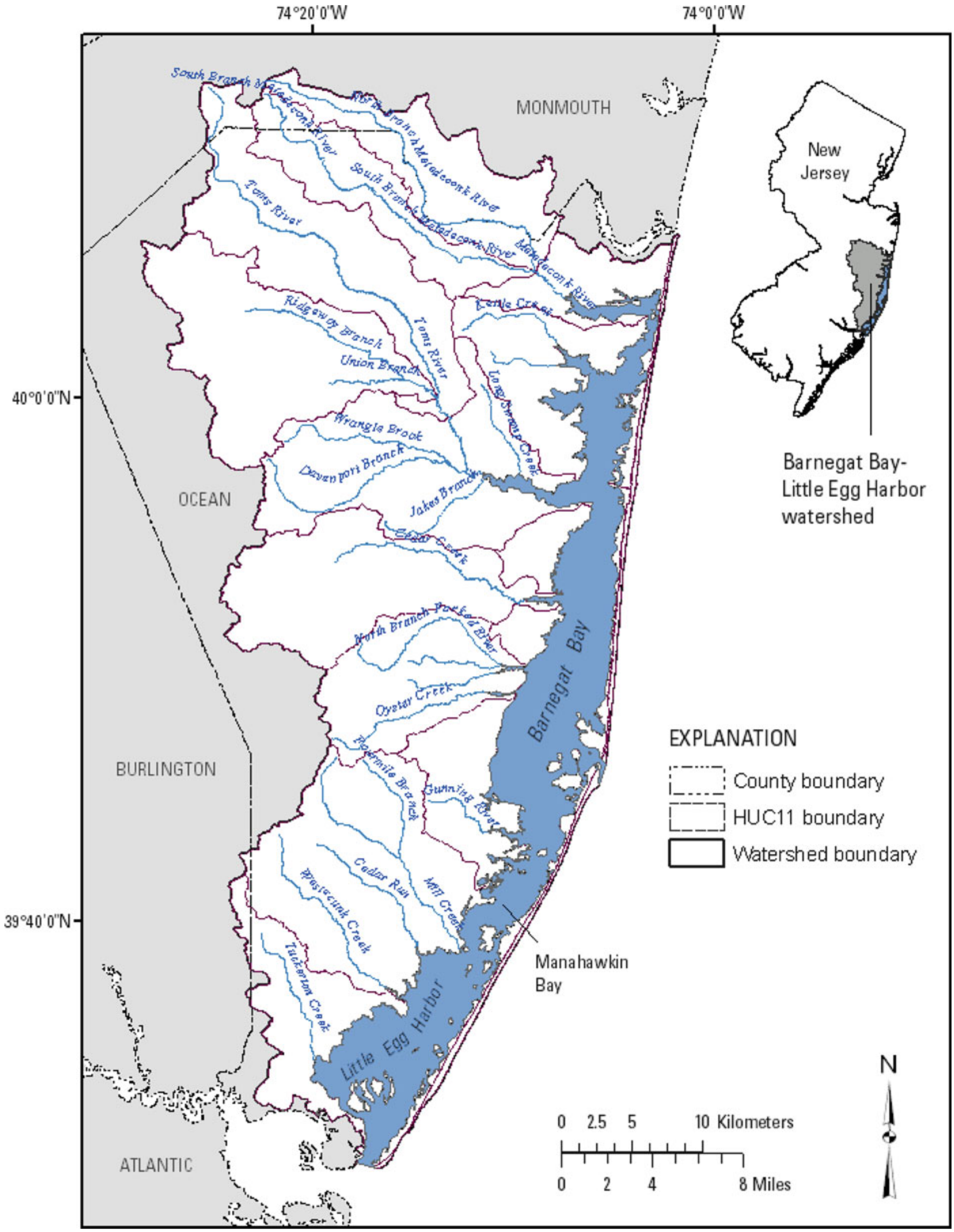
The basin morphometry and circulation of coastal lagoons differ considerably from those of larger, river-dominated estuaries. Coastal lagoons are shallow, generally averaging less than 2–3 m in depth, but depths of up to 30 m have been recorded in some tidal channels of these systems (Oertel, 2005; Kennish and Paerl, 2010b). They are generally well mixed by wave and current action. Because coastal lagoons receive relatively small volumes of freshwater input, tidal exchanges through narrow inlets play a significant role as a driver of lagoonal circulation. Most coastal lagoons are microtidal systems.

The physical-chemical processes taking place in coastal lagoons depend greatly on multiple factors, notably the size and configuration of the tidal inlets, expanse and development of bordering watersheds, amount of freshwater input, tidal prism, wind velocity and direction, and water depth (Alongi, 1998; Kennish and Paerl, 2010a). As stated by Kennish and Paerl (2010a), "Variations in precipitation and evaporation, surface runoff, and groundwater seepage, together with fluxes in wind forcing, account for large differences in advective transport in lagoonal estuaries. Storm and wind surges, overwash events, inlet configurations, land reclamation, construction of dams, dikes and artificial bars, as well as channel dredging events, are important drivers of hydrological change in these systems."

Because of the extreme enclosure of most coastal lagoons by barriers and the limited tidal exchange with ocean waters, these shallow systems tend to have protracted water residence times. As a result, coastal lagoons are susceptible to accumulation of pollutants from coastal watersheds and airsheds. They are also easily impacted by overwash events driven by extreme climate events such as hurricanes that can transport large amounts of beach and coastal ocean sediments into these backbays. This was the case in New Jersey when superstorm Sandy made landfall on October 29, 2012, creating a storm surge exceeding 4 m in some areas and dumping more than 1.5 million cubic meters of beach sand into Barnegat Bay-Little Egg Harbor. Similar events have been recorded for other coastal lagoons impacted by hurricanes and extratropical storms.

Sediments

Coastal lagoons receive terrigenous sediment from streams and rivers draining coastal watersheds. These sediments often consist of fine silts and clays, much of which flocculate and are deposited at the mouth of the influent systems. Fine-grained sediments also accumulate near the lagoonal shoreline in proximity to salt marshes which facilitate deposition of silts and clays. However, in some



Coastal Lagoons, Figure 1 Barnegat Bay-Little Egg Harbor, a coastal lagoon located along the central New Jersey coastline (USA). Note the coastal watershed draining to the lagoon and the barrier island system forming the eastern boundary. Figure 1 from US Geological Survey, West Trenton, New Jersey.

coastal lagoons, the influx of sediments from land sources is minimal, and a significant amount of sediment accumulating in various areas of the lagoonal basin is the result of sediment reworking of the lagoonal floor. This is the case in Barnegat Bay-Little Egg Harbor, New Jersey (Psuty, 2004; Psuty and Silveira, 2009), as well as many other temperate coastal lagoons of North America (Oertel, 2005). Coarser sediments generally are found in proximity to the backbarriers and tidal inlets. These sediments, which are typically better sorted than those near the mainland, primarily derive from marine and backbarrier sources via storm surge and overwash events which build washover fans, and tidal currents through inlets which build ebb-tidal deltas and other sandy deposits in the lagoonal basin.

Biotic production

Coastal lagoons are characterized by high levels of biotic production. This is so because the photic zone extends to the lagoonal floor in most areas, and they usually receive considerable amounts of nutrients from the surrounding watersheds which stimulate primary production. Benthic algal and seagrass production can exceed phytoplankton production in coastal lagoons. In addition, there is strong benthic-pelagic coupling; in coastal lagoons the effects of biogeochemical cycling, bioturbation, and other interactions between the bottom sediments and the overlying water column may be far greater than those in deeper estuaries. Nutrients may be recycled many times before exiting inlets to the coastal ocean due to protracted water residence times which account for high rates of productivity per unit nutrient input (Kennish and Paerl, 2010b).

The range of annual primary production in coastal lagoons is large ($\sim$50–$>$500 g C m^{-2} year^{-1}). Based on the classification of Nixon (1995), many coastal lagoons fall within the range of eutrophic conditions (300–500 g C m^{-2} year^{-1}) or even exhibit hypereutrophic conditions ($>$500 g C m^{-2} year^{-1}) (Nixon, 1995). The high primary production in these water bodies, together with the input of organic matter from adjoining wetlands and external systems, supports rich faunal communities, with many species utilizing these environments seasonally. Benthic macrofaunal productivity in coastal lagoons amounts to $\sim$20–200 g ash-free dry weight m^{-2} year^{-1}, with zooplankton productivity being as much as 50 % of this amount. Nekton productivity in turn ranges from $\sim$10 % to 100 % of the zooplankton productivity in these systems (Alvarez-Borrego, 1994). Coastal lagoons also provide ideal nursery and feeding habitats for many marine fauna (Kennish and Paerl, 2010b; Day et al., 2012).

Anthropogenic effects

Coastal lagoons are used for fisheries and aquaculture, energy production, biotechnology, transportation, shipping, and many other human uses (Pauly and Yáñez-Arancibia, 1994; Kennish and Paerl, 2010b). Watersheds surrounding coastal lagoons are often heavily populated and developed because of the great commercial and recreational value of these water bodies, their exceptional ecosystem services, and the access they afford to coastal ocean waters. However, altered land use/land cover of upland areas associated with increasing population growth and development, together with escalating human activities in the coastal lagoons themselves, has impacted their structure and function and compromised their ecological integrity (Kennish and Paerl, 2010b). For example, the removal of natural vegetation, compaction of soils, and construction of impervious surfaces promote nutrient runoff into the lagoons, hastening their nutrient enrichment and eutrophication (Kennish, 1997; Kennish, 2002).

Eutrophication of coastal lagoons and estuaries is on the increase worldwide (Nixon, 1995; Kennish et al., 2008; Kennish, 2009; Kennish and Paerl, 2010a), and it poses the greatest threat to the ecological integrity of these valuable ecosystems (Kennish and de Jonge, 2011). Eutrophication leads to an array of cascading changes in ecosystem structure and function such as decreased dissolved oxygen levels, increased microalgal and macroalgal abundance, occurrence of harmful algal blooms (HABs), loss of seagrass habitat, reduced biodiversity, declining fisheries, imbalanced food webs, altered biogeochemical cycling, and diminished ecosystem services (Nixon, 1995; Kennish, 1997; Kennish et al., 2008; Kennish and Paerl, 2010b).

Because of their extreme enclosure and restricted circulation, coastal lagoons are highly susceptible to accumulation of chemical contaminants such as polycyclic aromatic hydrocarbons, halogenated hydrocarbons, and metals. Bottom sediments serve as a repository and secondary pool of these hazardous substances. Volatile organics and plastics are also a potential threat to organisms inhabiting these environments. Oil spills are particularly detrimental. Pathogens delivered to lagoonal systems in stormwater runoff subsequent to rainfall events frequently compromise their water quality, although such events are usually ephemeral.

The shorelines of many coastal lagoons are altered by housing and bulkhead construction, which interferes with natural processes and directly impacts habitat. The siting of marinas along these shorelines, oil and gasoline leakages from fixed installations, sanitation-tank releases from boats, sewage wastewater discharges, and dredging activities adversely affect lagoonal organisms. Aquaculture operations can markedly degrade water quality in confined areas. In many systems, organic loading contributes to elevated BOD levels and significant oxygen depletion leading to system impairment.

Conclusions

Coastal lagoons are highly productive, enclosed water bodies that are heavily utilized by humans. They are complex physiographic features susceptible to eutrophication and other anthropogenic impacts due to their relatively low freshwater inputs, shallow depths, restricted circulation, poor flushing, limited ocean exchange, and protracted water residence times. As a result, coastal lagoons are beset by similar problems such as depleted dissolved oxygen, habitat loss and alteration, and, in some cases, altered ecosystem structure and function. Indicators of eutrophication are widespread in these shallow water bodies, including elevated chlorophyll *a* levels, HABs, submerged aquatic vegetation loss, and impacted biotic communities and harvestable fisheries. Progressive eutrophication of coastal lagoons can lead to permanent loss of essential habitat, diminished aquatic life support, and a marked decline in human use. Because of their enclosure, coastal lagoons are also susceptible to chemical contaminant inputs, pathogens, and organic carbon loading. The hardening of lagoonal shorelines, constructing of installations, and dredging of sediments physically alter habitats which also impacts biotic communities and their sustainability.

Bibliography

Alongi, D. M., 1998. *Coastal Ecosystem Processes*. Boca Raton: CRC Press.

Alvarez-Borrego, S., 1994. Secondary productivity in coastal lagoons. In Gonenc, I. E., and Wolflin, J. P. (eds.), *Coastal Lagoons: Ecosystem Processes and Modeling for Sustainable Use and Development*. Boca Raton: CRC Press.

Barnes, R. S. K., 1980. *Coastal Lagoons*. Cambridge: Cambridge University Press.

Bird, E. C. F., 1994. Physical setting and geomorphology of coastal lagoons. In Kjerfve, B. (ed.), *Coastal Lagoon Processes*. Amsterdam: Elsevier, pp. 9–39.

Colombo, G., 1977. Lagoons. In Barnes, R. S. K. (ed.), *The Coastline*. London: Wiley, pp. 63–81.

Day, J. W., Kemp, M. W., Yánez-Arancibia, A., and Crump, B. C., 2012. *Estuarine Ecology*, 2nd edn. Hoboken: Wiley-Blackwell.

de Beaumont, E., 1845. *Lecons de Geologie Pratique*. Paris: P. Bertrand.

Gilbert, G. K., 1885. Lake Bonneville. U.S. Geological Survey, Monograph 1, Washington, DC.

Kennish, M. J., 1997. *Practical Handbook of Estuarine and Marine Pollution*. Boca Raton: CRC Press.

Kennish, M. J., 2001. Physical description of the Barnegat Bay-Little Egg Harbor estuarine system. *Journal of Coastal Research*, **SI 32**, 13–27.

Kennish, M. J., 2002. Environmental threats and environmental future of estuaries. *Environmental Conservation*, **29**, 78–107.

Kennish, M. J., 2009. Eutrophication of mid-Atlantic coastal bays. *Bulletin of the New Jersey Academy of Science*, **54**, 5–12.

Kennish, M. J., Livingston, R. J., Raffaelli, D., and Reise, K., 2008. Environmental future of estuaries. In Polunin, N. (ed.), *Aquatic Ecosystems: Trends and Global Prospects*. Cambridge: Cambridge University Press, pp. 188–208.

Kennish, M. J., and Paerl, H. W., 2010a. Coastal lagoons: critical habitats of environmental change. In Kennish, M. J., and Paerl, H. W. (eds.), *Coastal Lagoons: Critical Habitats of Environmental Change*. Boca Raton: CRC Press, pp. 1–15.

Kennish, M. J., and Paerl, H. W. (eds.), 2010b. *Coastal Lagoons: Critical Habitats of Environmental Change*. Boca Raton: CRC Press.

Kennish, M. J., and de Jonge, V. N., 2011. Chemical introductions to the systems: diffuse and nonpoint source pollution from chemicals (nutrients: eutrophication). In Kennish, M. J., and Elliott, M. (eds.), *Treatise on Estuarine and Coastal Science, Vol. 8, Human-Induced Problems (Uses and Abuses)*. Oxford: Elsevier, pp. 113–148.

Kjerfve, B., 1994. Coastal lagoons. In Kjerfve, B. (ed.), *Coastal Lagoon Processes*. Amsterdam: Elsevier, pp. 1–8.

McGee, W. D., 1890. Encroachments of the sea. *Forum*, **9**, 437–449.

Nichols, M. M., and Boon, J. D., 1994. Sediment transport processes in coastal lagoons. In Kjerfve, B. (ed.), *Coastal Lagoon Processes*. Amsterdam: Elsevier, pp. 157–219.

Nixon, S. W., 1995. Coastal eutrophication: a definition, social causes, and future concerns. *Ophelia*, **41**, 199–220.

Oertel, G. F., 2005. Coastal lakes and lagoons. In Schwartz, R. L. (ed.), *Encyclopedia of Coastal Science*. Dordrecht: Springer, pp. 263–266.

Pauly, D., and Yáñez-Arancibia, A., 1994. Fisheries in coastal lagoons. In Kjerfve, B. (ed.), *Coastal Lagoon Processes*. Amsterdam: Elsevier, pp. 377–399.

Psuty, N. P., 2004. Morpho-sedimentological characteristics of the Barnegat Bay-Little Egg Harbor Estuary. In Davis, D. W., and Richardson, M. (eds.), *The Coastal Zone: Papers In Honor of H. Jesse Walker*. Baton Rouge: Louisiana State University. Geoscience and Man, Vol. 38, pp. 81–92.

Psuty, N. P., and Silveira, T. M., 2009. Geomorphological evolution of estuaries: the dynamic basis for morpho-sedimentary units in selected estuaries in the northeastern United States. *Marine Fisheries Review*, **71**(3), 34–45.

Cross-references

Backbarrier
Barrier Island

COASTAL LANDFORMS

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Synonyms

Coastal geomorphic forms

Definition

Coastal landforms are those formed and modified by various geological and oceanographic processes. The present-day coastal land was carved out during the Late Quaternary, particularly during the Holocene. Anthropogenic activities enhance the coastal landform changes.

Introduction

The coasts are dynamic and their morphology is continually changing in response to various processes operating at different rates. Climate change and sea-level variations during the Quaternary period have strongly influenced the geomorphic and sedimentation processes in the coastal regions, and much of the coastal land was carved out during the Late Quaternary period. Sea-level changes during the Holocene have influenced the evolution of coastal environments such as estuaries/lagoons and barrier complexes and controlled the sedimentation in the coastal environments (Narayana and Priju, 2004). The evolution and subsequent changes of coastal landforms were influenced by various factors, viz., the coastal processes, sea-level changes, and tectonics. These landforms are modified by a variety of dynamic processes and the driving forces include framework geology, oceanographic processes, river-mouth processes, sediment supply, and human activity (FitGerald et al., 2008). Coastal landforms are extremely variable and coastal habitats change over a range of spatial and temporal scales, and recognition of these variations is necessary for effective planning and management (Woodroffe, 2007). Barrier islands, wetlands, and other parts of coastal systems might have a threshold, and, when the limits of threshold are exceeded, the landforms become unstable and prone to irreversible changes in form and position (Williams and Gutierrez, 2009).

Coastal systems exhibit two distinct types of coastal landforms: depositional and erosional. Erosional coastal landforms typically exhibit high relief and rugged topography, which include sea cliffs, wave-cut platforms, and stacks. The depositional coastal landforms include barrier islands, beach ridges, cheniers, tidal flats, mudflats, etc. In this chapter, we focus on the depositional coastal landforms and their characteristics.

Barrier Islands

Coastal barriers and spits are often regarded as similar coastal forms in terms of beach deposition projecting across coastal bays. While barriers tend to bridge the bay by joining the mainland at each end, spits are only attached at one end. However, many barriers show cross-barrier breaks or breaches through which the sea may enter on a permanent or intermittent basis, thus forming barrier islands (Figure 1). Coastal barriers are complex constructional morphological features involving deposition by waves, wave-generated currents, tidal currents, and wind activity (Hayes, 1979). A barrier exhibits two morphodynamic units – a seaward beach face and a landward facing back-barrier slope – and these two units develop when the barrier is gravel dominated (Orford et al., 1996). As sand becomes the dominant component, a third environment comprising aeolian dunes can appear at the top of the beach face (barrier crest) and spread onto the backslope. Current flows may have been responsible for the initial submarine platform under the barrier, but with wave action forcing, onshore migration of the barrier takes place in combination with fine sedimentation characteristics of the low-energy back-barrier bay. Tidal currents also become dominant once barrier islands appear.

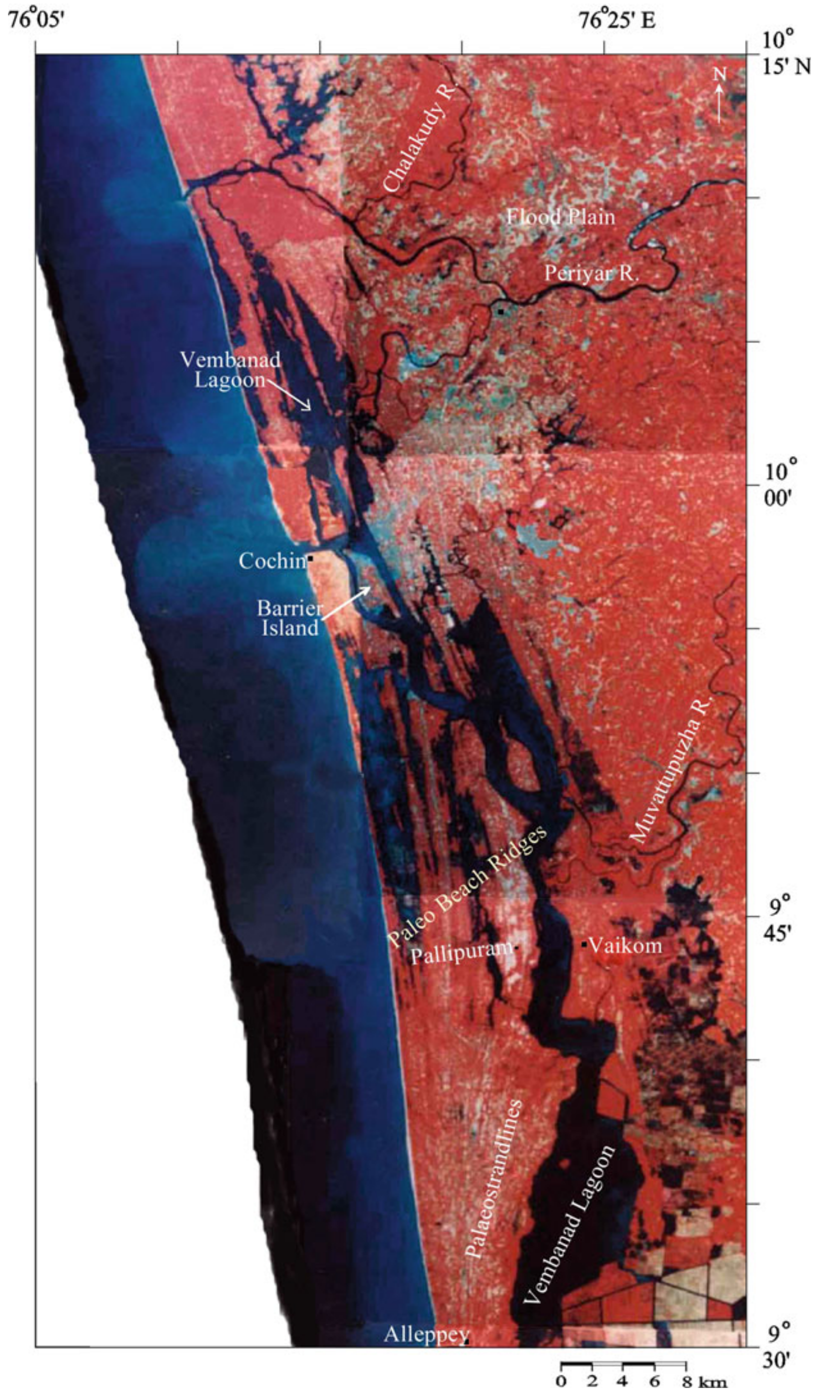
Sediment supply and the type of sediment are major controls on barrier development with which a behavioral distinction can be drawn between sand-dominated barriers and gravel-dominated barriers. This distinction has a spatial basis with gravel-dominated barriers being more prevalent in mid-upper latitudes compared to sand-dominated barriers, which reflects the greater potential of coarse material in high latitudes as a residue of late Quaternary glacigenic processes.

Coastal bars

Coastal bars can be broadly defined as aggradational ridges of sediments whose formation, morphology, and behavior are determined by interactions between waves, currents, tides, local slope, and grain size. Bars occur along beach, river delta, estuary, and continental shelf environments with a wide range of sizes, types, and orientation (Figures 1 and 2). Beach morphology undergoes cycling change, promoting offshore sediment transport and bar formation during winter; while during summer when the oceanographic conditions are calmer, the landward migration of the bar and eventual welding to the beach face take place. However, the existence of such "winter" and "summer" profiles is not universal, as both barred and non-barred profiles occur at all times in some areas, while in others only one type may persist throughout the year. Furthermore, cyclic beach response at timescales much shorter than seasons can result in barred profiles (Short, 1979). Types of bars are often distinguished based on their alongshore planform shape and orientation relative to the shoreline as linear, shore-parallel, sinuous, or crescentic with a trough separating them from the shoreline. Some of the coastal areas consist of alternating transverse bars, welded to the shoreline and are separated by channels occupied by rip currents (Figure 2). Bar type is strongly related to wave energy level with linear bars developing under high-energy conditions, crescentic bars during intermediate energy, and transverse bars during lower wave energy levels. Under very low-energy conditions, a bar may become fully welded to the beach and appear as a flat terrace at low tide. These types of bar configurations are common on microtidal beaches and may grade into each other as energy levels vary. A number of classifications exist describing both bar types and the continuum of bar evolution (e.g., Greenwood and Davidson-Arnott, 1979; Short and Aagaard, 1993; Wijnberg and Kroon, 2002).

Coastal lagoons

The term lagoon describes a stretch of salt water separated from the sea by a low sandbank or coral reef (Figures 1 and 3). Coastal lagoons (Figure 3) are mostly estuarine, usually shallow, and have generally been partly or wholly sealed off from the sea by the deposition of spits or barriers, by localized tectonic subsidence, or by the growth of coral reefs. They are best formed on transgressive coasts, particularly where the continental margin has a low gradient, and sea-level rise is slow. The lagoons are ephemeral features and their depths and areal extent gradually decrease due to sedimentation from inflowing rivers, as well as accumulation of sediment washed in from the sea, wind-blown material, and chemical and organic deposits. Lagoons range in size from less than a kilometer to more than a 1000 km, and they occur on about 12 % of the length of the world's coastline (Bird, 2000). They can be classified on the basis of infilling or increasing in size (Nichols, 1989). The infill of some



Coastal Landforms, Figure 1 Satellite image showing various coastal landforms such as barrier island, lagoon, paleo-strandlines, and flood plains along southwest coast of India.

lagoons, particularly those that are parallel to the shore, may involve the development of cuspate forms that divide the lagoon into a series of segments. These divisions have been attributed to winds blowing along the length of the lagoon producing waves which build spits that isolate the lagoon into separate basins.

Mudflats

Mudflats (Figure 4) occur along low-energy shorelines that are well supplied with silt and clay-sized sediments, particularly on many estuarine margins, delta shorelines, and areas of open coast subject to low wave energy. Such settings are usually dominated by tidal processes, and the characteristic

Coastal Landforms, Figure 2 Well-developed shore-attached transverse bars and adjacent deeper rip channels at Lighthouse Beach, New South Wales (Short, 1979) (Source: http://www.ozcoasts.gov.au/conceptual_mods/beaches/wdb.jsp).

Coastal Landforms, Figure 3 Hypersaline coastal lagoon at Mar Menor, Iberian Peninsula, Spain (Source: http://www.latorreholiday.co.uk/3.html).

landforms of muddy coasts – salt marshes, mangrove swamps, and tidal flats – are often well developed under macrotidal conditions (Hayes, 1975). Enormous quantities of muddy sediment are supplied by some of the world's major rivers, and their estuaries and deltas often feature extensive shore-attached mud banks. Consequent to the development of wide estuarine mouths, a lot of seawater enters through them during high tide and submerges low-lying flats adjacent to the river mouth forming tidal flats. Both estuarine and open coast mud banks are highly dynamic landforms, which exhibit seasonal variability in response to variations in river flow and wave energy.

Fine sediments in suspension can be transported over long distances by coastal currents. These fine particles undergo flocculation and, once they are flocculated, settle from suspension rapidly giving rise to muddy deposits/mudflats near estuarine/river mouths. Flocculation is influenced by a variety of factors, notably salinity, fluid shear, and suspended sediment concentration (Lick and Huang, 1993). The effect of these processes may vary over quite short spatial and temporal scales, especially in estuaries, where mixing of freshwater and saltwater occurs and where marked variation in flow intensity occurs at tidal timescales. The cohesive nature of muddy sediments makes

Coastal Landforms, Figure 4 Present-day mudflats of Sado River Estuary, Portugal (Source: http://geologicalintroduction.baffl.co.uk/?p=323).

their behavior far more complex than that of non-cohesive sands. Flocculated sediments typically settle from suspension far more rapidly than their constituent mineral particles, and the stability of natural muddy deposits is governed not only by physical processes but also by the activity of a rich and diverse biota (Paterson, 1997).

Beaches

Beach is a wave-dominated accumulation of sediment located between wave base and the upper swash limit. A beach system is a product of the interaction of waves, tide, and sediment, and hence, beaches exist in a wide spectrum of wave, tide, and sediment combinations and geological settings. Beach systems occur in all tide ranges, in all latitudes, and in all climates. Beaches composed of fine sand through boulders may range from low-energy to high-energy systems exposed to persistent 2–3-m-high swell which breaks across wide surf zones. All beaches contain three dynamic zones – wave shoaling, wave breaking, and swash–backwash. The wave shoaling zone extends from the modal wave base where average waves can entrain and move sediment shoreward, to the outer breakpoint. The wave shoaling zone is dominated by asymmetrical wave orbital motions which produce a concave upward profile. It extends out to depths of 30 m or more which may lie 2–3 km offshore on high-energy coasts, while on low-energy coasts, it may only extend to low tide – a few meters from the shore. The surf zone, located between the breakpoint and shoreline, has the greatest potential for complex dynamic processes and resulting topography and bedforms. The width of the surf zone depends on the surf zone gradient, a function of sand size and wave height. The width may vary from a few meters on a steep reflective beach, typically 50–100-m-wide on a single bar intermediate beach, and up to several hundred meters on a high-energy dissipative beach. Surf zone topographic features include shore-parallel bars and troughs with waves breaking over the bars and reforming in the troughs. Surf zone bedforms reflect the changing velocity and direction of currents and depth of water and range from flat bed over the shallow bars to wave orbital and shore perpendicular current ripples in the troughs, to shore-parallel seaward migrating ripples in the rip channels (Short, 1979).

Beaches are of three types, which refer to the morphodynamic character of a beach system: wave-dominated, tide-modified, and tide-dominated beaches. Wave-dominated beaches occur where waves are high relative to the tide range. This can be defined quantitatively by the relative tide range (Masselink and Short, 1993)

$$\mathrm{RTR} = \mathrm{TR}/\mathrm{Hb} \tag{1}$$

where TR is the spring tide range and Hb the average breaker wave height. When $\mathrm{RTR} < 3$, beaches are tide dominated; when $3 < \mathrm{RTR} < 15$, they are tide modified; and, when the $\mathrm{RTR} > 15$, they become tide dominated.

Shingle beach

The term “shingle” has been used to describe sediments composed of mainly rounded pebbles, larger in diameter than sand (>2 mm) but smaller than boulders (<200 mm) (Figure 5). In many locations, shingle is mixed with sand, silt, clay, or organic debris, resulting in a “mixed” sediment beach (Kirk 1980), but all shingle and boulder beaches can be regarded as different types of “coarse clastic” beach (Carter and Orford, 1991). Shingle coasts form in wave-dominated locations where suitably sized material is available, and they occur in high latitudes and temperate shores, which were affected by Quaternary glaciation.

Coastal Landforms, Figure 5 Sand and shingle beach, Blakeney Point, Sheringham (Source: http://www.geograph.org.uk/photo/2019176).

Coastal Landforms, Figure 6 Low amplitude beach cusps on Panambur beach, near Mangalore, west coast of India. Wide surf zone is seen on the background (Picture by K.S. Jayappa).

In general, shingle coasts have received less scientific attention than sandy and muddy shorelines, as they are much less common worldwide. However, recently there has been an increasing awareness of the geomorphologic, ecological, and engineering significance of shingle coasts in the contexts of sea-level change, flood defense, and habitat conservation (Packham et al., 2001).

Beach cusps

Beach cusps are crescentic, concave-seaward, and regularly spaced features occurring along the shorelines (Figure 6). The term "beach cusp" has been used for features with spacing ranging from 10 cm to many hundreds of meters; the term "swash cusp" has been used for rhythmic beach features with a spacing less than tens of meters (Hughes and Turner, 1999). Beach cusps are most commonly associated with medium to coarse sands, shingle, or mixed sand–shingle sediments on steep beaches, demonstrating significant wave reflection (Nolan et al., 1999). Multiple sets of cusps may be present at different levels on beaches of high tidal range. Beach cusps consist of embayments or swales separated by triangular horns which are normally comprised of coarser sediments. Under low-energy conditions, oscillatory flows, horn

Coastal Landforms, Figure 7 *Beach ridges* indicating the past sea level and abundant sediment supply along southwest coast of India.

divergent flows, and horn convergent flows will develop, and under high-energy conditions, sweeping flows and swash-jet flows can occur (Masselink et al., 1998).

Beach ridges
Beach ridges are azonal accumulation forms developed on seashores. They are usually subparallel ridges of sand, gravel, or pebble, as well as detritus of shell, situated in the foreshore zone, which is the boundary of low and high water range (Figure 7). Older complexes of beach ridges appear in the backshore zone, which lies above the high water range. Beach ridges forming at the present day are roughly parallel to the coast. Two types of beach ridges may develop on a progradational sea coast (Carter, 1986). The first type is a result of gradual accretion and coalescing of swash bars during transport of a deposit by wave action, and the second type is connected with longshore bar emergence during low wave energy conditions and simultaneous fall of sea level. They are constructed mainly by landward dipping laminae, and the morphology of these ridges is more complicated.

Beach ridges are also partly developed by the processes of aeolian deflation and accumulation. There is often an accumulated cover of aeolian deposits on earlier formed ridges, stabilized by vegetation. As a result, on the beach ridges, irregular hummock dunes or parallel foredune ridges can be situated (Carter and Wilson, 1990). Beach ridges are good paleogeography indicators of past wave regimes, sediment supply, sediment source, climatic conditions, sea-level change, and also isostatic emergence or submergence of land. Hence, beach ridges can be used to reconstruct past relative sea-level changes and the history of deposits.

Chenier ridges/plains
The name "chenier" derives from the French word chene, meaning oak, which grows on the coast of Louisiana, USA. Chenier ridges (cheniers) are elongated beach ridges with sand or shell composition and are separated laterally from other cheniers on a chenier plain, by fine grained sediments (Figure 8). Chenier ridges frequently bend landward at the downdrift end and branch in a fan-like fashion. Cheniers are found on generally low wave energy, low gradient, muddy shorelines, and in areas where there is an abundant sediment supply (Augustinus, 1989). Cheniers can be up to 6 m high, tens of kilometers in length, and hundreds of meters wide.

Raised beaches
A raised beach is a relict depositional landform comprising mostly wave-transported sedimentary material and preserved above and landward of the active shoreline (Figure 9). Raised beaches were first described by Jamieson in 1908, when he stated that raised beaches form along coasts or lake shorelines and are well recognized as indicators of a fall in relative sea level. In certain situations, multiple raised beaches may form adjacent to one another, producing a beach ridge plain or strandplain (Otvos, 2000). The elevated position of a raised beach relative to active shoreline processes may be the product of one or more of the following mechanisms: (1) tectonic uplift (Garrick, 1979), (2) isostatic rebound related to ice-unloading of a land mass (Smith et al., 2000), (3) depositional regression involving delivery of sediment to a shoreline at a rate sufficient to allow formation and stranding of successive beaches (Thom, 1984), and (4) forced regression whereby eustatic sea-level fall leads to abandonment of a shoreline (Murray-Wallace and Belperio, 1991).

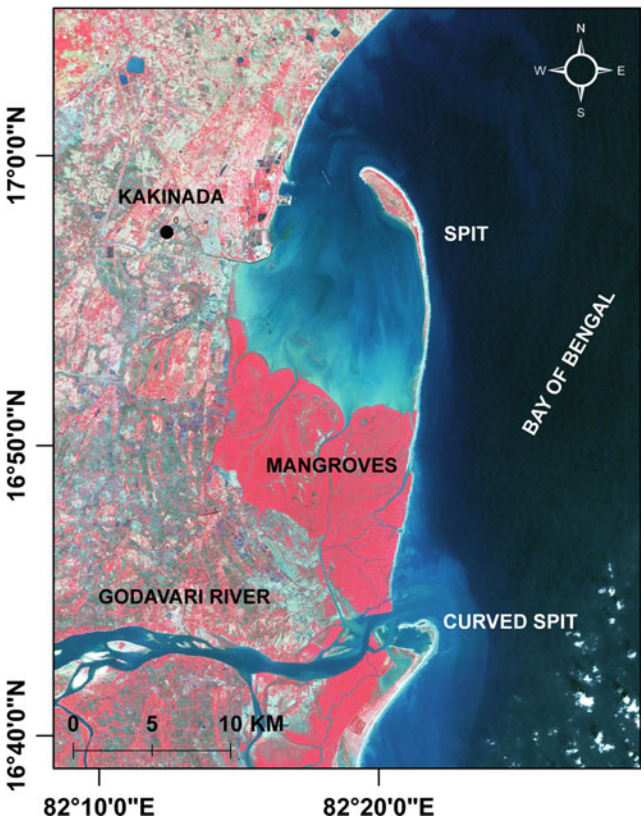
Spits
Spits are essentially narrow depositional embankment-type features that show a dominance of longshore sediment deposition over cross-shore sediment movement (Figure 10). A spit's elongation relative to width is an

Coastal Landforms, Figure 8 Chenier plains separated laterally by other cheniers, indicating low wave energy and low gradient coasts.

Coastal Landforms, Figure 9 Beach rock at Okha, Gujarat coast, India, indicating the raised beach (Picture by P. Hanamgond).

indication of both the coastal sediment availability and net longshore-directed, wave-generated transport potential. Coastal configuration also plays an important role in the formation of spits (Carter, 1988). Spits are found on an irregular coastline where sediment availability and wave power allow a constructional smoothing of the coastline. Sand-dominated spits are the most common, whereas gravel-dominated spits occur in mid-upper latitudes, where gravel is a major component of coastal sediments. As spits are essentially a product of breaking wave activity, mud-dominated spits are unlikely to be observed. The presence of a spit generates a back-spit energy lee with low-energy currents and fine sediment stores (tidal banks and marshes).

Spits develop where wave refraction cannot accommodate the sudden changes of coastal trend and when rapid reduction in breaker approach angle reduces the longshore drift rate to zero at this point. This allows beach deposition to overshoot the directional shift in coastline. The spit builds from this depositional nucleus, and its orientation is a function of wave refraction accommodating to the changing nearshore bathymetry induced by the presence of the spit (Carter, 1988). Spits show a sequence of planform changes that are related to variation in both sediment supply and longshore transport potential and are best developed when nearshore wave approach is angled along the spit. A spit is connected to the coast and its proximal sediment source by the neck, while a spit extension occurs at the spit's distal end or terminus. A spit is usually the subaerial expression of a larger submarine feature, the distal position of which is the spit ramp. Ramp deposition controls spit growth and usually has a high fine sediment



Coastal Landforms, Figure 10 Spit on the Godavari River mouth, near Kakinada, India, as viewed by lands at satellite.

proportion related to wave-generated currents. As most of the sediment for the spit platform is supplied by longshore transport, it mimics sediment availability to the superstructure, though tending toward finer sediment (Ollerhead and Davidson-Arnott, 1995).

The spit platform requires an increasing sediment volume as the spit progressively builds into deeper water and as the volume of the superstructure generally remains the same. Spit elongation rates will decline over time if the longshore sediment supply rate does not increase. Thus, sediment supply rate is a major control on spit development. Rapid wave shoaling and landward curvature of the breaking wave crest at the spit terminus with steep bathymetric gradients lead to curvature of the distal structure against the general trend of the spit (Figure 10). High volume, but episodic, sediment supply can lead to drift-aligned spits, where the spit plan outline is essentially rectilinear despite overlapping recurves (Carter and Orford, 1991). This scenario is often associated with the initial formation of spits in a disjointed coastline, where sediment supply is formed from isolated finite sediment sources (Orford et al., 1996). Spits evolve generally in shore-parallel direction; however, inlet shoreline curvature may produce shore-normal orientations. Shore-normal spits generally have landward trends. The migration of spits into inlets influences the efficiency of an inlet to transfer water between the sea and back-barrier lagoon and generally requires a morphodynamic response by the inlet (Ortel, 1985). Spits generally retreat under rising relative sea level through overwashing and hence rollover.

Tombolo

A tombolo is a sandbar or a barrier that joins an island with a mainland or another island, resulting from longshore drift or the migration of an offshore bar toward the coast (Figure 11). Tombolos are constructive features, occurring along shorelines of submergence that are protected from large waves and where islands are common. Sediment supply is predominantly derived from the islands, yet some may also come from erosion of the shoreline, fluvial materials, underwater reefs, and offshore glacial deposits. Several types of tombolos – single, double, multiple, forked, parallel, and complex tombolos – are reflective of the coastal system (e.g., wave mechanisms) from which they are derived. Tombolos can restrict flow between the sea and intertidal zone, forming a lagoon and altering the local ecology.

Strandflats

The strandflat was first described by Reusch (1894), and its possible origin was first explained in detail by Nansen (1922). The word "strandflat" is used for the shallow sea along the western Norwegian coast and also along coasts in Arctic and Antarctic areas that have been covered by ice sheets during the Quaternary ice age. Apart from long stretches of the west coast of Norway where the strandflat is an almost continuous feature, the strandflat has also been recognized in areas as far apart as the South Shetland Isles, Alaska, and western Scotland. The low areas of strandflat often appear as broad glacially molded coastal rock platforms and backed by high cliffs (Figure 12). These shore platforms generally exhibit considerable local relief.

The processes of strandflat formation include marine abrasion, subaerial weathering, glacial erosion, frost shattering, and cold climate shore erosion. The strandflat is primarily the result of sea-ice erosion and frost shattering during the Quaternary, and the surfaces are later modified by marine and glacial erosional processes (Larsen and Holtedahl, 1985). The strandflat surfaces produced by cold climate shore processes must have been repeatedly overwhelmed by ice sheets and subject to marine processes during numerous intervals of cold climate throughout the Quaternary.

Tidal delta

Tidal deltas are large sand bodies formed within, or in the vicinity of, tidal inlets. Flood-tidal deltas form landward of the inlet mouth, under the influence of flood-tidal currents. The major morphological features of flood-tidal deltas typically include a seaward-dipping flood ramp.

Coastal Landforms, Figure 11 Tombolo on Om Beach, Gokarn, west coast of India (Picture by P. Hanamgond).

Coastal Landforms, Figure 12 Flat coastal plains, crisscrossed by glacial rivers, are known as strandflats. This view shows a 3-km-wide strandflat along the coast of Oscar II Land to the south of Engelskbukta (Source: http://www.swisseduc.ch/glaciers/arctic-islands/arctic-07-en.html?id=8).

However, landward sand movement occurs through the migration of sand waves under the action of flood currents; subtidal flood channels, which extend into the inlet and which dissect the partly intertidal landward portion of the delta; marginal ebb-aligned spits; and spillover lobes formed by the action of ebb currents over the lower parts of the ebb shield (Hayes, 1980).

Ebb-tidal deltas occur seaward of the inlet, predominantly under the influence of ebb-tidal currents and wave action. These deltas are usually comprised of an ebb channel, maintained by strong tidal currents; linear bars, formed by wave–current interactions along the margins of the ebb channel; a terminal lobe formed at the distal (seaward) end of the ebb channel, where the tidal current diminishes; and sandsheets or swash platforms formed by wave action adjacent to the ebb channel characterized by migrating swash bars.

The morphology of the tidal deltas is characterized by tidal prism, configuration of the inlet and adjacent shoreline, wave climate, and the rate of littoral sediment

Coastal Landforms, Figure 13 Tidal creeks at the southern end of Great Bay (Source: http://www.nhdfl.org/about-forests-and-lands/bureaus/natural-heritage-bureau/photo-index/Deletions/tidal-creek-bottom.aspx).

transport. In microtidal areas, flood deltas are often better developed than their ebb counterparts, owing to the dominance of landward, wave-driven, sediment transport. Ebb delta morphology is generally more variable than that of flood deltas, owing to the importance of regional and local contrasts in wave climate (Boothroyd, 1985) and due to the close coupling of delta processes with wider coastal morphodynamics. Ebb delta volume increases with the tidal prism and decreases with inlet width/depth ratio and wave energy. Under conditions of low wave energy, ebb deltas are typically more elongated and extend farther seaward.

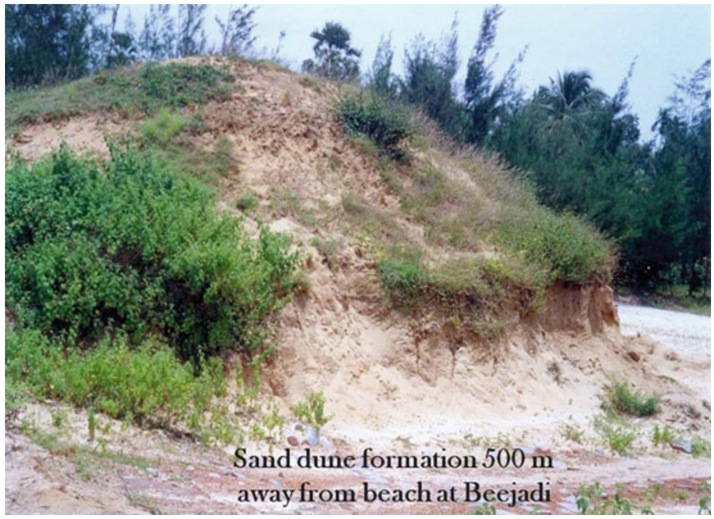
Another important landform in tidal delta/tidal flats is the tidal creek (Figure 13). Creeks occur extensively on mudflats and muddy coasts, mangrove swamps, and salt marsh surfaces (Eisma, 1998). Tidal creeks often have a high drainage density because of the large volumes of water that they drain (Pethick, 1984). The morphology of the creeks is also often distinctive. Although some may bear a superficial resemblance to dendritic river channel networks, flow along them is bidirectional (French and Stoddart, 1992). They have a tendency to taper upstream and flare downstream (Fagherazzi and Furbish, 2001), and their discharge is determined by the tidal prism. In areas with a large tidal range or rapid seaward progradation, creek systems may be markedly linear in form. In areas with cohesive sediments, creeks have steep edges, whereas in sandier areas, they tend to be shallower and wider.

Coastal dunes

Sand dunes are ubiquitous landforms along many of the world's coastlines and are indicative of periods when wind, sediment supply, vegetation, and local climate all coexisted at suitable levels to result in dune deposition (Figure 14). Once established within an accommodation space, coastal dune fields represent responsive geomorphological landscapes that react closely with changes to forcing parameters (Jackson and Cooper, 2011). Morphological behavior of dune fields is characterized by changes in climate which drives local precipitation, temperature, and wind stress over dune landforms. Sea-level rise, associated with rapid climate change scenarios, is normally tied with instability at the front edge of a sand dune coast (Carter, 1991; Saye and Pye, 2007). Under such scenario, there will be a predicted increase in vertical growth and eventual mobility of foredunes, leading to a transgressive response across the rest of the dune field (Figure 14).

Dune fields are classified in to four types – foredunes, blowout, parabolic, and transgressive dune fields. Foredunes are shore-parallel dune ridges formed on the top of the backshore by aeolian sand deposition within vegetation. They may range from scattered hummocks or nebkha, relatively flat terraces, to markedly convex ridges. Active foredunes occupy a foremost seaward position, but not all foremost dunes are foredunes. Other dune types may occupy a foremost position on eroding coasts or coasts where foredunes are unable to form. Foredunes generally fall into two main types – incipient and established foredunes.

A blowout is a saucer-, cup-, bowl-, or trough-shaped depression formed by wind erosion on a preexisting sand deposit. The adjoining accumulation of sand, the depositional lobe, derived from the depression and possibly other sources, is normally considered part of the blowout (Nordstrom et al., 1990). Blowout morphology may be highly variable, ranging from cigar-shaped, V-shaped, scooped hollow, and cauldron and corridor types, from pits to elongated notches, troughs or broad basins, and saucer and trough blowouts (Cooper, 1967). Saucer blowouts are semicircular or saucer shaped and often appear as shallow dishes. Deeper cup- or bowl-shaped blowouts



Coastal Landforms, Figure 14 Sand dunes showing the vertical growth covered with vegetation, west coast of India (Picture by K.S. Jayappa).

may evolve from them. Trough blowouts are generally more elongate, with deeper deflation floors and basins and with steeper, longer erosional lateral walls or slopes.

Parabolic dunes are typically U- and V-shaped dunes characterized by short to elongate, trailing ridges which terminate downwind in U- or V-shaped depositional lobes. The depositional lobes may be simple, relatively featureless sandsheets, or textured with a variety of dune forms (e.g., transverse dunes, barchanoidal dunes, etc.).

Transgressive dune fields are well developed on high wind and wave energy coasts with significant sediment supply and in all climatic regions. Transgressive dune fields and sheets are aeolian sand deposits formed by the downwind or alongshore movement of sand over vegetated to semi-vegetated terrain. Such sheets and dune fields may range from quite small (hundreds of meters in alongshore and landward extent) to megadune size fields. They may be completely unvegetated, partially vegetated, or fully vegetated (Nordstrom et al., 1990). Dune fields are covered with a variety of superimposed dune forms. They have also been termed mobile dunes, migratory dunes, mendano, and machair.

Coral reefs

Corals are organisms that secrete a calcareous exoskeleton and are major contributors to a coral reef. Coral reefs are natural structures of calcium carbonate made largely from the skeletons of hard corals and coralline algae. Some modern reefs have been forming for millions of years and can stretch for hundreds of kilometers off tropical coasts.

Reefs can be broadly classified as spatially heterogeneous, three-dimensional structures which have morphological form that is different from that of the underlying substrata. The term reef has been used to classify a whole host of organic and inorganic structures including stone reefs, oyster reefs, coral reefs, atolls, and algal reefs.

Reefs are found in temperate to tropical marine ecosystems, with the most prominent reef types, corals and atolls, being found in tropical and subtropical zones. Algal reefs and bioherms are commonly found in more moderate climatic zones, such as the Mediterranean. In temperate regions, reefs are often more like bioherms or biostromes in structure.

Coral reefs are found mainly between 25°N and 25°S latitudes. The reef-building (herm atypic) corals prefer sea-surface temperatures between 25 and 29 °C. Hermatypic corals mostly occur in the "photic" zone, where sufficient sunlight can penetrate for their symbiotic algae for photosynthesize. The distribution of fossil coral reefs suggests that sea-surface temperatures have constrained their spread since their appearance in the early Triassic (Birkeland, 1997). Coral reefs were alternately exposed and drowned as temperatures and sea levels oscillated during the Quaternary. During glacial periods, when sea levels were low, the distribution of coral reefs was much less and in marginal areas of the modern coral seas (like the Hawaiian Islands), the reefs died out entirely (Grigg, 1988). Owing to cooler temperatures, coral reefs grew at slower rates, and many were comparatively ephemeral. As temperatures increased and sea levels rose at the end of the glacial periods, reefs gradually became reestablished across wider areas of the seas.

Depending on oceanographic factors, upward-growing coral reefs were either able to "keep up" with rising postglacial sea level or form a drowned reef (Neumann and MacIntyre, 1985). Drowned reefs occur in many parts of the Pacific and Indian Oceans. Most of the coral reefs have failed to keep up with rising sea level associated with climate change and sea-level history, paleolatitude, seawater temperature, and light (Flood, 2001). In many parts of

Coastal Landforms, Figure 15 Photograph showing the 8-km-wide Atafu Atoll located in the southern Pacific Ocean (Source: http://earthobservatory.nasa.gov/IOTD/view.php?id=37753).

the world, coral reefs are found raised above their modern counterparts and, as such, often provide important insights into reef structure and history.

The morphology and genesis of coral reefs vary significantly. On the basis of their form, reefs may be divided into atoll reefs, barrier reefs, and fringing reefs (Nunn, 1994). Barrier and atoll reefs are older, often being composed of reefs of many different ages; reef upgrowth during postglacial periods has been followed by subaerial exposure, and erosion during the glacial periods followed by renewed upgrowth.

Fringing reefs are juvenile and the youngest and most ephemeral of the three forms, and they grow outwards from a coast. They are located close to the land and indeed cannot exist very far away from the land. Unlike atoll reefs and barrier reefs, most fringing reefs are formed as discrete units during the recent period of postglacial sea-level rise. Most of them began growing from shallow depths on the flanks of a tropical coastline when ocean-water temperatures at the end of the glacial period became suitable for reef growth (Neumann and MacIntyre, 1985). Fringing reefs are mostly affected by humans.

Atolls

Coral atolls, dispersed widely throughout the warm waters of the tropical Pacific and Indian Oceans, are among the world's most impressive biogenic landforms (Nunn, 2010). The classic exposition of atoll origin was first explained by Charles Darwin in 1836, where he observed a barrier reef surrounding Mo'orea Island during his visit to the Keeling Islands. Darwin set out his theory of atoll development which involved the upward growth of a coral reef in response to the subsidence of its foundations (Darwin, 1842). Darwin's elegant theory was founded on the premise of a subsiding volcanic island and the corresponding upward growth of friging and barrier coral reefs keeping pace with the rising relative sea levels (Terry and Goff, 2013). Darwin suggested that it was the tendency of ancient volcanic islands in the oceans to subside, but their coral fringe could stay alive only if it was able to grow upward at the same rate. Thus, modern atoll reefs are only veneers of living coral growing atop a coral framework composed largely of the skeletal remains of dead hermatypic (reef-building) corals.

Atolls are generally subcircular rings of coral reef (Figure 15) surrounding a lagoon with no dry land other than occasional islands (called *motu*) made from sand and gravel-sized detritus thrown up on the reef during storms (Nunn, 1994). The word "atoll" should be applied only to the reef, but sometimes the term is used more loosely to refer to *motu*. In the Pacific, where some of the world's oldest atolls exist, many have reef foundations dating from at least the Oligocene. It is a surprise to know how such organic structures remain intact despite the continuous buffering of storm waves, earthquakes, and even nuclear weapons tests. Johnston Atoll in the central Pacific, where the US chemical weapon stocks are being destroyed, lost its southern flank in a series of huge landslides predating its discovery by humans. On the other hand, part of Moruroa Atoll in French Polynesia, where 98 subterranean tests of nuclear bombs were carried out between 1981 and 1991, has subsided as a direct consequence of nuclear tests (Keating, 1998). Many atolls exhibit major arcuate "bight-like" structures in their plan form. These departures from circular or elliptical forms are indicative of geomorphological processes that cannot be ignored (Terry and Goff, 2013).

Summary

The coastal landforms, carved out during the Late Quaternary period, are influenced and modified primarily by sea-level changes, ocean and river-mouth processes, sediment nature and its supply, and human activity. The important depositional coastal landforms are barrier islands, coastal bars, lagoonal systems, and mudflats that are characterized mostly by tides, waves, currents, and grain size, as well as beaches and spits that occur in all tide and wave conditions. In the formation of spits, wave refraction and longshore drift, apart from coastal configuration, play an important role. Beach ridges and chenier plains are indicators of past sea-level changes, climatic conditions, wave regime, and sediment supply. Coastal dunes represent geomorphological landscapes that are reworked by climate drivers such as local precipitation, temperature, and wind. Transgressive dune fields indicate high wind and wave energy coasts with large sediment supply.

Coral reefs mainly confine to 25°N – 25°S latitudes, and their occurrence suggests that sea-surface temperatures play an important role in their distribution. Based on the morphology, reefs are classified into atoll, barrier, and fringe reefs. Darwin (1842) was the first who proposed a theory on evolution of coral reefs, and he hypothesized that subsiding volcanic islands correspond with upward growth of fringing and barrier coral reefs keeping pace with relative sea-level rise.

The impact of human activities is the primary concern for the fragile nature of coastal landforms and coastal ecosystems.

Bibliography

Augustinus, P. G. E. F., 1989. Cheniers and chenier plains: a general introduction. *Marine Geology*, **90**, 219–229.

Bird, E., 2000. *Coastal Geomorphology: An Introduction*. New York: John Wiley and Sons.

Birkeland, C. E., 1997. *Life and Death of Coral Reefs*. New York: Chapman and Hall.

Boothroyd, J. C., 1985. Tidal inlets and tidal deltas. In Davis, R. A. (ed.), *Coastal Sedimentary Environments*, 2nd edn. New York: Springer, pp. 445–532.

Carter, R. W. G., 1986. The morphodynamics of beach-ridge formation: Magilligan: Northern Ireland. *Marine Geology*, **73**, 191–214.

Carter, R. W. G., 1988. *Coastal Environments*. New York: Academic.

Carter, R. W. G., 1991. Near-future sea level impacts on coastal dune landscapes. *Landscape Ecology*, **6**, 29–39.

Carter, R. W. G., and Orford, J. 1991. The sedimentary organization and behavior of drift-aligned barriers, Coastal Sediments '91. American Society Civil Engineers, New York, pp. 934–948.

Carter, R. W. G., and Wilson, P., 1990. The geomorphological, ecological and pedological development of coastal foredunes at Magilligan Point, Northern Ireland. In Nordstrom, K. E., Psuty, N., and Carter, R. W. G. (eds.), *Coastal Dunes: Form and Process*. Chichester: Wiley, pp. 129–157.

Cooper, W. S., 1967. *Coastal Sand Dunes of California*. Boulder: Geological Society of America Memoir.

Darwin, C., 1842. *The Structure and Distribution of Coral Reefs*. London: Smith Elder & Company.

Eisma, D., 1998. *Intertidal Deposits: River Mouths, Tidal Flats, and Coastal Lagoons*. Boca Raton: CRC Press.

Fagherazzi, S., and Furbish, D. J., 2001. On the shape and widening of salt marsh creeks. *Journal of Geophysical Research*, **106**, 991–1003, doi:10.1029/1999JC000115. issn: 0148-0227.

FitGerald, D. M., Fenster, S. M., Argow, B., and Buynevich, I. V., 2008. Coastal impacts due to sea level rise. *Annual Reviews of Earth and Planetary Sciences*, **36**, 601–647.

Flood, P. G., 2001. The 'Darwin Point' of Pacific Ocean atolls and guyots: a reappraisal. *Palaeogeography, Palaeoclimatology, Palaeoecology*, **175**, 147–152.

French, J. R., and Stoddart, D. R., 1992. Hydrodynamics of salt marsh creek systems: implications for marsh morphological development and material exchange. *Earth Surface Processes and Landforms*, **17**, 235–252.

Garrick, R. A., 1979. Late Holocene uplift at Te Araroa, Ease Cape, North Island, New Zealand. *New Zealand Journal of Geology and Geophysics*, **22**, 131–139.

Greenwood, B., and Davidson-Arnott, R. G. D., 1979. Sedimentation and equilibrium in wave-formed bars: a review and case study. *Canadian Journal of Earth Sciences*, **16**, 312–322.

Grigg, R. W., 1988. Paleoceanography of coral reefs in the Hawaiian Emperor chain. *Science*, **240**, 1737–1743.

Hayes, M. O., 1975. Morphology of sand accumulations in estuaries. In Cronin, L. E. (ed.), *Estuarine Research*. New York: Academic, Vol. 2, pp. 3–22.

Hayes, M. O., 1979. Barrier island morphology as a function of tidal and wave regime. In Leatherman, S. P. (ed.), *Barrier Islands*. New York: Academic.

Hayes, M. O., 1980. General morphology and sediment patterns in tidal inlets. *Sedimentary Geology*, **26**, 139–156.

Hughes, M., and Turner, I., 1999. The beach face. In Short, A. D. (ed.), *Handbook of Beach and Shoreface Morphodynamics*. Chichester: John Wiley and Sons, pp. 119–144.

Jackson, D. W. T., and Cooper, J. A. G., 2011. Coastal dune fields in Ireland: rapid regional response to climatic change. *Journal of Coastal Research*, **SI 64**, 293–297.

Keating, B. H., 1998. Nuclear testing in the Pacific from a Geological Perspective. In Terry, J. P. (ed.), *Climate and Environmental Change in the Pacific*. School of Social and Economic Development, The University of the South Pacific, pp. 113–144.

Kirk, R. M., 1980. Mixed sand and gravel beaches: morphology, processes and sediments. *Progress in Physical Geography*, **4**, 189–210.

Larsen, E., and Holtedahl, H., 1985. The Norwegian strandflat: a reconsideration of its age and origin. *Norsk Geologiske Tidsskrift*, **65**, 247–254.

Lick, W., and Huang, H., 1993. Flocculation and the physical properties of flocs. In Mehta, A. J. (ed.), *Nearshore and Estuarine Cohesive Sediment Transport*. Washington, DC: American Geophysical Union, pp. 21–39.

Masselink, G., and Short, A. D., 1993. The effect of tide range on beach morphodynamics and morphology: a conceptual beach model. *Journal of Coastal Research*, **9**, 785–800.

Masselink, G., Hegge, B. J., and Pattiaratchi, C. B., 1998. Beach cusp morphodynamics. *Earth Surface Process: Land*, **22**, 1139–1155.

Murray-Wallace, C. V., and Belperio, A. P., 1991. The last interglacial shoreline in Australia – a review. *Quaternary Science Reviews*, **10**, 441–461.

Nansen, F., 1922. The strandflat and isostasy, Videnskapelkapets Skrifter 1. Math.-Naturw. Kl. (Kristiana), 11.
Narayana, A. C., and Priju, C. P., 2004. Evolution of coastal landforms and sedimentary environments of the late quaternary period along central Kerala, southwest coast of India. *Journal of Coastal Research*, **SI 39**, 1898–1902.
Neumann, A. C., and MacIntyre, I., 1985. Reef response to sea-level rise: keep-up, catch-up or give-up. *Proceedings of the 5th International Coral Reef Congress*, **3**, 105–110.
Nichols, M. M., 1989. Sediment accumulation rates and sea-level rise in lagoons. *Marine Geology*, **88**, 201–219.
Nolan, T. J., Kirk, R. M., and Shulmeister, J., 1999. Beach cusp morphology on sand and mixed sand and gravel beaches, South Island, New Zealand. *Marine Geology*, **157**, 185–198.
Nordstrom, K. F., Psuty, N. P., and Carter, R. W. G., 1990. *Coastal Dunes: Form and Process*. London: Wiley.
Nunn, P. D., 1994. *Oceanic Islands*. Oxford: Blackwell.
Nunn, P. D., 2010. Pacific atolls: a world apart. In Migon, P. (ed.), *Geomorphological Landscapes of the World*. New York: Springer, pp. 349–356.
Oertel, G. F., Fowler, J. E., and Pope, J., 1985. History of erosion and erosion control efforts at Tybee Island, Georgia. Miscellaneous Paper CERC-85-1, Vicksburg, Army Engineer Waterways Experiment Station U.S.
Ollerhead, J., and Davidson-Arnott, R. G. D., 1995. The Evolution Of Buctouche Spit, New Brunswick, Canada. *Marine Geology*, **124**, 215–236.
Orford, J. D., Carter, R. W. G., and Jennings, S. C., 1996. Control domains and morphological phases in gravel- dominated coastal barriers. *Journal of Coastal Research*, **12**, 589–605.
Otvos, E. G., 2000. Beach ridges: definitions and significance. *Geomorphology*, **32**, 83–108.
Packham, J. R., Randall, R. E., Barnes, R. S. K., and Neal, A. (eds.), 2001. *Ecology and Geomorphology of Coastal Shingle*. New York: Westbury Publishing.
Paterson, D. M., 1997. Biological mediation of sediment erodibility. In Parker, R., and Watts, J. (eds.), *Cohesive Sediments*. New York: John Wiley and Sons, pp. 215–229.
Pethick, J., 1984. *An Introduction to Coastal Geomorphology*. New York: Edward Arnold.
Reusch, H., 1894. Strandflaten, et nyt træk i Norges geografi. *Norges Geologiske Undersokelse*, **14**, 1–14.
Saye, S. E., and Pye, K., 2007. Implications of sea level rise for coastal dune habitat conservation in Wales, UK. *Journal of Coastal Conservation*, **11**, 31–52.
Short, A. D., 1979. Three dimensional beach-stage model. *Journal of Geology*, **87**, 553–571.
Short, A. D., and Aagaard, T., 1993. Single and multi- bar beach change models. *Journal of Coastal Research*, **SI 15**, 141–157.
Smith, D. E., Cullingford, R. A., and Firth, C. R., 2000. Patterns of isostatic land uplift during the Holocene: evidence from mainland Scotland. *The Holocene*, **10**, 489–501.
Terry, J. P., and Goff, J., 2013. One hundred and thirty years since Darwin: reshaping the theory of atoll formation. *The Holocene*, **23**, 615–619.
Thom, B. G., 1984. *Coastal Geomorphology in Australia*. Sydney: Academic.
Wijnberg, K. M., and Kroon, A., 2002. Barred beaches. *Geomorphology*, **48**, 103–120.
Williams, S. J., and Gutierrez, B. T., 2009. Sea-level rise and coastal change: causes and implications for the future coasts and low-lying regions. *Shore & Beach*, **77**, 13–21.
Woodroffe, C. D., 2007. The natural resilience of coastal systems: primary concepts. In McFadden, L., Penning-Rowsell, E., and Nicholls, R. J. (eds.), *Managing Coastal Vulnerability*. Amsterdam: Elsevier, pp. 45–60.

Cross-references

Back Dune
Bar
Climate Change
Coastal Barriers
Coastal Bays
Coastal Lagoons
Deltas
Estuarine Beaches
Foredune
Mangroves
Saltmarshes
Secondary Dune
Spit
Tidal Flat
Tides

COASTAL RISKS: FLOODS

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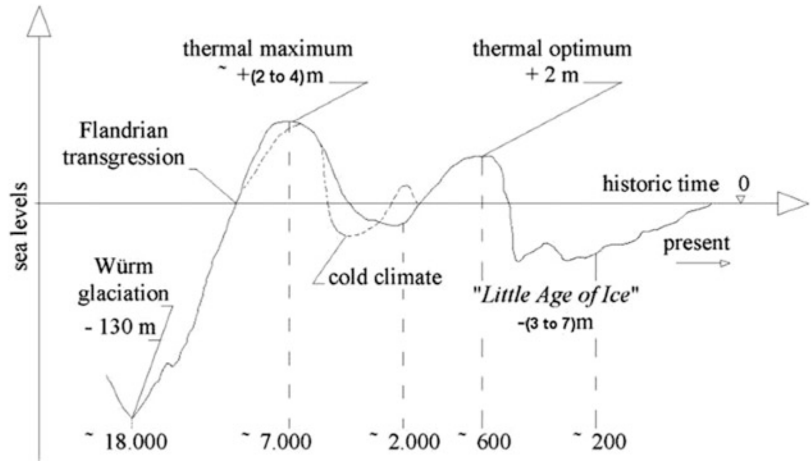
Definition

Flooding of coastal lands is primarily due to inundation by the sea during storms and other natural events (e.g., tsunamis) that increase coastal population risk. Flooding also occurs from inland waters when storm water levels hinder evacuation.

Description

Coastal risks have very different origins and etiologies, but current vulnerability assessments focus mainly on erosion and climate-induced floods. Generalized fluid-dynamic erosion is the reason for most shore and coastal protection measures. Tectonic plate movements may generate tsunami waves, and the resulting coastal risks can reach great levels for vulnerable settlements. Hurricanes and regional monsoons can also cause vast erosion and damage to coastal zones.

Eustatic sea level is increasing due to climate change (Figure 1), but crustal isostasy, tectonics, and coastal plain subsidence cause variability of sea level at the local level. Accordingly, this trend of increasing global sea level can be accentuated in some places or attenuated and even reversed in others. The intensity and frequency of coastal flooding depend not only on eustasy but also on other climate-related factors such as low-pressure systems and strong winds that can raise the average sea level above the current tide and generate temporary increases in basin water levels that cause inland coastal floods (Diez et al., 2011, 2012). A temporary rise in sea level can act as a dam at the mouth of a river causing blockage of river drainage and a rise in water levels on the river, on its



Coastal Risks: Floods, Figure 1 Global Climate Evolution since last Glaciation expressed on mean sea level and based on geo-historic approaches. Scales vary decreasing to present time and level.

floodplain, and on surrounding areas (Audiencia Territorial de Valencia, 1991).

Unlike marine floods that are caused by seawater in limited areas, coastal floods may be due to a variety of different causes, and they may affect much larger coastal zones. When such floods are directly sourced to pluvial or fluvial waters, persistent high sea level is the main cause of the flood duration.

Historically, structural coasts were selected for settlements mainly because of security, health, safety, and economic reasons. However, sedimentary coasts offered greater productivity, and these areas soon attracted large populations to their plain littoral/deltaic hillocks. Mediterranean seaside ("maritime") cities were always located beside a castled rock-hill (e.g., Athens, Haifa, Genoa, Malaga, and Monaco), whereas coastal plain cities were established on the landward side of relatively dry and high grounds (e.g., Rome on its seven hills, Valencia or Alexandria on delta hillocks, Venice on a relict barrier island of the Po Delta, etc.).

Coastal cities and towns often spread into lowland areas. Enormous conurbations are at a huge risk of flooding today as a result. The case of New Orleans is paradigmatic: its older settlements barely suffered from floods caused by Hurricane Katrina, meanwhile most of its later developments and lower quarters were catastrophically flooded. The great European delta formed at the outlet of the Elm, Rhine, and Meuse rivers has required drastic transformations to protect large cities from flood risk.

Flood risk can never be totally eradicated. Therefore, each vulnerable coastal development now requires a risk management plan to deal with the hazard. Flood risk has become a datum for analysis and resiliency management.

Bibliography

Audiencia Territorial de Valencia, 1991. Criminal Sentence 56/1982, about Tous dam failure. October 23, 1990. Valencia.

Diez, J. J., Paz, R., Esteban, M. D., López-Gutiérrez, J. S., and Negro, V., 2011. Urban coastal flooding and climate change. *Journal of Coastal Research*, **SI 64**, 205–209. Szczecin, Poland.

Diez, J. J., Esteban, M. D., López-Gutiérrez, J. S., and Negro, V., 2012. Meteocean influence on inland and coastal floods in the east of Spain. *Journal of Coastal Research* (Online), ISSN 1551–5036, Palm Beach, FL.

Cross-references

Climate Change
Coastal Erosion Control
Shoreline Changes
Submerged Coasts
Submergent Shoreline
Uplifted Coasts

COASTAL SQUEEZE

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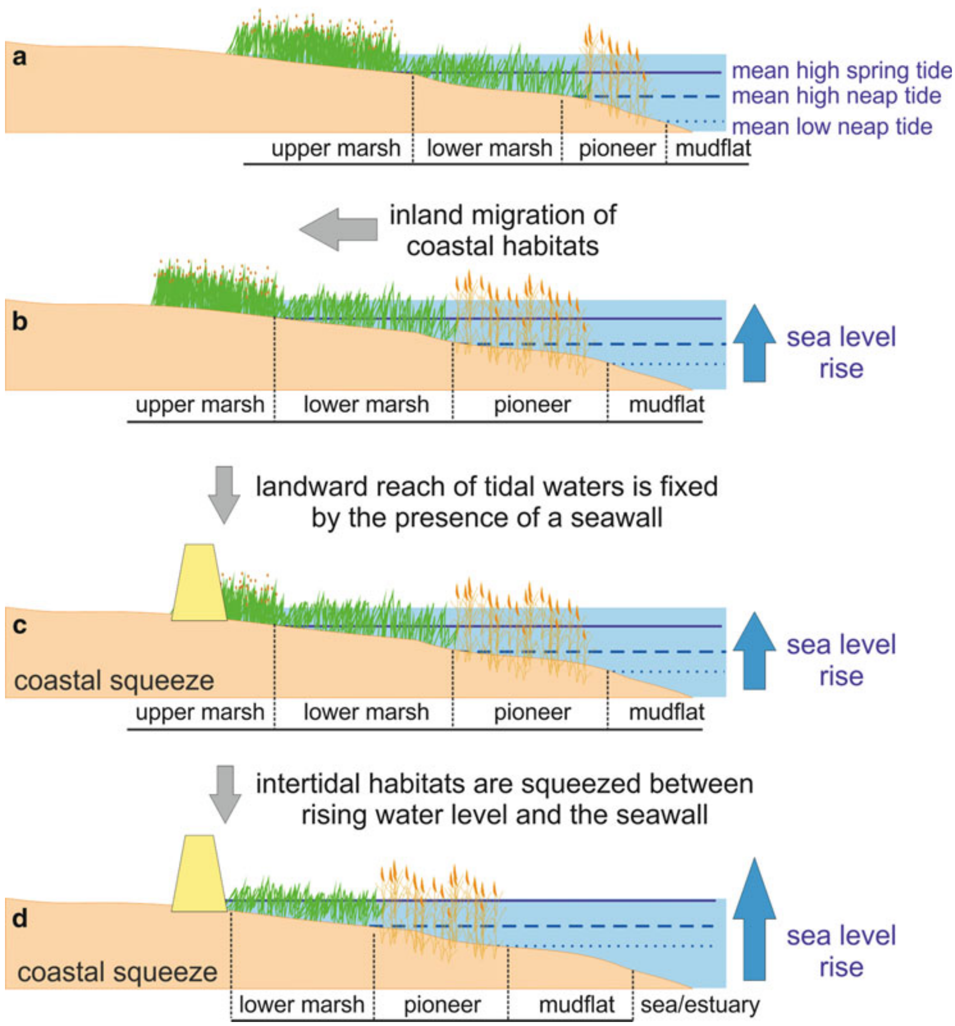
Definition

Coastal squeeze refers to the loss of intertidal habitats due to rising sea levels along coastlines fixed by hard engineering structures. The term coastal squeeze should not be used to refer to losses due to natural processes (Pontee, 2013).

Natural coasts can dynamically adjust to changing meteorological and climatic conditions. In natural systems, rising sea levels usually result in a landward movement of habitats (Figure 1a, b). Salt marshes, for example, depending on a number of interacting physical and biotic variables, can migrate inland and accrete vertically, naturally adjusting to sea-level rise. The natural



Coastal Squeeze, Figure 1 The elevation in relation to the tidal range is one of the key factors determining the type of intertidal habitat that may develop in a particular location (**a**). Natural habitats tend to migrate inland as a response to rising sea levels (**b**). As a result of this migration the intertidal area may expand or reduce depending, for example, on the coastal topography. Hard engineering structures will invariably fix the landward limit of intertidal areas (**c**), which will be reduced in extent as sea levels rise and more land becomes permanently inundated (**d**). The loss of coastal habitats due to rising sea levels in front of artificially fixed shorelines is known as coastal squeeze.

landward migration of habitats is prevented in coastlines "fixed" by hard coastal engineering, leading to coastal squeeze (French, 1997).

The type of intertidal wetland that may be established at any particular location is influenced (among other variables) by their position within the tidal range (Figure 1a). The vertical zonation of marshes reflects the tolerance of species to inundation (Pennings and Calloway, 1992), i.e., more tolerant species are found at lower elevations. Coastal defences fix the upper boundary of intertidal habitats (Figure 1c, d); therefore, a rise in sea level will gradually increase the frequency and duration of inundation and ultimately result in loss of intertidal area (as lower areas become permanently submerged). Depending on the range of elevations in relation to the water levels, increased exposure to inundation may lead to a shift in the types of marsh communities and/or the loss of habitats. Mudflats may occupy areas formerly dominated by pioneer marshes (Figure 1d); these might shift to higher ground or will disappear if suitable conditions are not available. The same process applies to other types of marshes.

Coastal squeeze and land reclamation are often cited as the main causes for the loss of intertidal habitats (e.g., Doody, 2012). Coastal squeeze is not the only cause for the loss of intertidal habitats. Hughes and Paramor (2004) argue that coastal squeeze would lead first to the loss of upper marshes, while the loss of pioneer marshes is most commonly observed. The authors suggest that increases in the abundance of the polychaete *Nereis* might be the cause of widespread loss of pioneer marshes in southeast England. The impact of storms along the coast of the Gulf of Mexico has been identified as one of the main reasons for the increased rate of wetland loss in the United States in the period 2004-2009 when compared with the previous five years (Dahl and Stedman, 2013). The loss of salt marshes is particularly concerning as they provide natural coastal protection and other valuable ecosystems services.

Bibliography

Dahl, T. E., and Stedman, S. M., 2013. *Status and Trends of Wetlands in the Coastal Watersheds of the Conterminous United States 2004 to 2009*. U.S. Department of the Interior, Fish and Wildlife Service and National Oceanic and Atmospheric Administration, National Marine Fisheries Service, 46 p.

Doody, J. P., 2012. Coastal squeeze and managed realignment in southeast England, does it tell us anything about the future? *Ocean & Coastal Management*, doi:10.1016/j.ocecoaman.2012.05.008.

French, F. W., 1997. *Coastal and Estuarine Management*. London: Routledge.

Hughes, R. G., and Paramor, O. A. L., 2004. On the loss of saltmarshes in south-east England and methods for their restoration. *Journal of Applied Ecology*, **41**, 440–448.

Pennings, S. C., and Calloway, R. M., 1992. Salt marsh plant zonation: the relative importance of competition and physical factors. *Ecology*, **73**, 681–690.

Pontee, N., 2013. Defining coastal squeeze: a discussion. *Ocean & Coastal Management*, **84**, 204–207.

COASTAL WETLANDS

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Definition

Coastal wetlands are habitats in close proximity to oceanic or estuarine waters that are directly influenced by tides and are covered permanently or periodically with salt or brackish water. In the broad sense, they include submerged seagrass communities, tidal flats, and emerged salt marshes and mangrove forests.

Coastal wetlands

Tidal flats are ecotonal areas between land and sea and can extend from the subtidal through the intertidal and into the supratidal zones. Although they often appear barren, they can be highly productive and support large animal populations. They occur throughout the world in areas with significant fine-grained sediment deposition. Seagrasses are submerged flowering plants occurring in protected shallow estuaries generally with soft sediments. They can form extensive beds that are important habitat for a large number of animal species and play an important ecological role in the nearshore estuarine environment. See Tidal Flat for more details. Below we consider emergent salt marshes and mangroves.

We distinguish between coastal wetlands (situated at or near the coast with direct influence of seawater salinity) and tidal wetlands which can include freshwater areas a considerable distance from the ocean but whose hydrology is still influenced by tidal phenomena that can propagate substantial distances upland (Rey et al., 2012a). More or less distinct vegetation zones dictated principally by tidal inundation are common in coastal wetlands, but vegetation mosaics and mixes are also widespread, and sharp transitions often occur in response to slight environmental gradients (Marani et al., 2013).

The structure and function of coastal wetlands are determined by many interconnected processes and feedback loops that operate at varying temporal and spatial scales (Berger et al., 2008). Top-down constraints such as landform and climate interact with bottom-up effects such as local competition and individual plant photosynthesis to affect the structure and function of a given wetland. Twilley and Rivera-Monroy (2005) divided these processes into three types: *regulators* (non-resource factors such as salinity, climate, etc.), *resources* (factors used by organisms for growth such as nutrients, sunlight, and space), and *hydroperiod* (the duration, frequency, and depth of flooding).

The hydrological pattern of coastal wetlands is the dominant factor affecting their structure and function (Mitsch and Gosselink, 2007). Hydrology affects many biotic and abiotic processes which in turn may modify hydrology. Examples of these include primary and secondary productivity, soil and water chemistry including anaerobiosis, nutrient cycling, salinity, biological diversity, carbon cycling, sedimentation dynamics, and microbial metabolism. In addition to climate and basin morphology, tidal flooding and flow through tidal creeks and channels (Perillo, 2009) are often the most important components of coastal wetland hydrological dynamics. Other important factors include subsurface composition, precipitation, surface flows, ground water flows, and evapotranspiration. Sedimentation dynamics, which includes production, transport, and sediment storage and is heavily influenced by hydrology, also plays a critical role in wetland function and maintenance (D'Alpaos et al., 2012).

The modification and transport of chemicals through coastal wetland ecosystems (biogeochemical cycling) result from a complex matrix of chemical, physical, and biological processes, again, with numerous feedback mechanisms, and give rise to many of the well-known wetland functions such as carbon sequestration/export, nutrient exports, and many others. Biogeochemical cycling interacts with marsh hydrology and geomorphology to determine physical and biological conditions within a given wetland. Major chemical cycles in coastal wetlands include those of nitrogen, sulfur, iron, manganese, carbon, and phosphorous.

The structure and function of wetland biological communities are closely tied to and have a heavy influence on other wetland processes (Marani et al., 2007). Primary production by marsh vegetation directly or indirectly influences a variety of food webs including wetlands, estuarine, oceanic, and terrestrial-based ones. Vegetation, together with wetland physiographic features such as tidal creeks and elevation discontinuities, provides critical habitat for a wide variety of terrestrial and aquatic organisms. Ecological interactions such as competition and predation are also crucial in modifying community structure. Belowground bacterial activity has important effects upon biogeochemical cycles. Wetland chemistry and hydrology directly influence plant and animal communities, but biological activity, for example, the burrowing of fiddler crabs (McCraith et al., 2003) or emergent plant metabolism (Gribsholt et al., 2003; Gribsholt and Kristensen, 2003), can have important consequences for wetland hydrology and chemistry.

Importance of coastal wetlands

Coastal wetlands have great ecological importance because of their biodiversity (Gopal and Junk, 2013) and productivity. They are often critical habitats for protected and endangered species and for species of commercial or recreational fishery value. Wetlands are integral components of coastal hydrological processes and function in flood control and in retention, transport, and storage of carbon, sediments, nutrients, and pollutants. Wetlands often filter contaminants originating higher up in the watershed and act as sinks for excess nutrients, thus contributing to the maintenance of estuarine and nearshore oceanic water quality. Wetlands often provide erosion protection and sediment stabilization. Wetlands also have high recreational and aesthetic values that make these areas desirable for human habitation. Additionally, the coastal zone in general is highly valuable economically, with many important facilities such as ports and airports and the industrial/commercial development that they attract often situated there. As a result, more than 40 % of the US population lives in coastal counties (NOAA, 2013), and over 44 % of the world's population lives within 150 km of the coast (UN, 2013).

General distribution of coastal wetlands

Below is a very general outline of coastal wetland distribution throughout the world. It only offers broad descriptions, and individual localized sites may depart significantly from regional norms.

Polar coastal wetlands

High-latitude coastal wetlands consist of salt and brackish water marshes and laida (wetlands inundated by both salt water during storms and freshwater during snowmelt). They occur along most coasts in the Northern Hemisphere, with some of the more extensive ones occurring along Hudson and James Bays and along the coastal plains of Alaska and the Yukon (Martini et al., 2013). Large wetlands also occur along the Russian coast and in major river deltas of the region. Similar wetlands do not occur in the Southern Hemisphere because of a dearth of ice-free substrate (Martini et al., 2013).

North America

Farther south, between New Brunswick and Nova Scotia in the Gulf of Maine, the Bay of Fundy wetlands occur at the approximate subarctic-temperate transition zone. This area has one of the largest tides in the world, with spring tidal ranges of close to 15 m. Extensive low marshes are populated almost exclusively by *S. alterniflora*, whereas in the high marsh *S. patens* is most widespread and *Phragmites australis and Iva frutescens* occur along the upper edge. Within New England, regional differences associated with climate and human impacts exist, but in general, marshes standing on marine peat have similar vegetative composition as above. *Juncus gerardii* is common as an upland fringe, and in areas with substantial freshwater inputs, a brackish community consisting of *Scirpus americanus*, several *Typha* species, *Zizania aquatica*, and *Phragmites australis* occurs (Nixon and Oviatt, 1973; Pratolongo et al., 2013).

From New England south to northern Florida, marshes develop behind protective barrier island complexes. Throughout this area, *S. alterniflora* dominates the low marsh, with a tall form occurring in areas with longer flooding periods and a short form where daily tidal inundation lasts only for a short time. In the high marsh, *S. patens* or *Juncus roemerianus* may form monospecific stands or the two species may codominate, often forming complex spatial patchworks. As in New England, transitional brackish areas exist which in this case are dominated by *Spartina cynosuroides*.

In Florida, south of 30°N latitude, mangroves gradually replace salt marshes, but narrow bands of salt marsh can be found throughout the state. Three mangrove species occur in Florida: the red mangrove (*Rhizophora mangle*), the black mangrove (*Avicennia germinans*), and the white mangrove (*Laguncularia racemosa*). A variety of herbaceous halophytes often occupy the mangrove understory; examples include *S. alterniflora*, *Batis maritima*, and *Salicornia virginica*.

East of the Mississippi deltaic wetlands, along the northern Gulf Coast of North America, grass/rush marshes can be found usually directly in front of the open ocean due to the low tidal energy in the region. Clearly delimited low marsh and high marsh plant zones are often evident, but convoluted plant community mixtures can be just as common (Montague and Wiegert, 1990). Generally, *S. alterniflora* forms relatively narrow bands along the shoreline and is then replaced by black needle rush (*Juncus roemerianus*), but both species can also form extensive monocultures (Kurz and Wagner, 1957; Rey et al., 2012a).

The Mississippi deltaic plain region supports extensive wetland complexes (approximately 7,250 km^2) in six major drainage basins that represent a time series of shifts in the major channel of the river. The youngest basins support mostly freshwater marshes because of their shallow depth and the large volume of freshwater inflow. The older basins support a variety of wetland types including extensive salt marshes and forested wetlands. West of the Mississippi, coastal wetlands tend to occur along protected shores behind barrier islands and along protected bays, in once flooded ancient river valleys.

Along the Pacific coast of North America, coastal wetlands are sparse because of the rugged terrain. Small isolated wetlands along river valleys are dominated by *Spartina foliosa* (a species that also exhibits short and tall forms) accompanied by various succulents. Farther north, the San Francisco Bay area supports wetlands dominated by *S. foliosa* in the low marsh and *S. virginica*, *Jaumea carnosa*, *Triglochin maritima*, and *D. spicata* in the high marsh. In the Pacific Northwest, low marshes are dominated by halophytic succulents such as *Salicornia virginica*, *Jaumea carnosa*, and *Triglochin maritima* as well as several grasses and sedges such as *Distichlis spicata* and *Carex lyngbyei*. Tufted hair grass (*Deschampsia cespitosa*) commonly dominates in high marshes, accompanied by mixes of many other species.

Central America and Caribbean

In the Caribbean islands, mangroves predominate as fringe vegetation along the coast, in protected bays and lagoons, and as overwash islands that are often completely flooded by each tidal cycle. In Central America, more extensive mangals develop, and herbaceous marshes can be found as small isolated pockets or as narrow fringes in front of mangrove formations.

South America

As in North America, only small isolated wetlands can be found along the Pacific coast because of the rugged terrain associated with the Andean chain. Extensive wetlands occur on the Atlantic coast. Mangroves can be found throughout the northern part of the continent, with 90 % of the coverage by approximately 10 South American species found in Brazil, Colombia, Venezuela, Ecuador, and Suriname (FAO, 2007). Typical salt marshes occur farther south and are dominated by *S. alterniflora* in the low marsh and *S. densiflora* in the high marsh along with associates such as *Limonium brasiliense*, *Juncus acutus*, and *Distichlis spicata*. As one approaches Tierra del Fuego, the region becomes arid and cold and vegetation everywhere becomes scant.

Europe

Northern European marshes are characterized by a pioneer zone thinly vegetated with *Spartina anglica* and *Salicornia* spp., an intermediate zone populated with a variety of halophytes such as *Aster tripolium*, and an upper marsh zone where *Festuca rubra*, *Juncus gerardii*, and *Elymus athericus* predominate. Recently, the latter species has been invading the mid and low marsh areas (Pratolongo et al., 2013). Large salt marshes exist along the Atlantic coast and along the North, Baltic, and Walden seas. Coastal wetlands in the Mediterranean region are commonly associated with river mouths; important deltaic wetlands include the Camargue (Rhone) in France and the Ebro Delta (Ebro) in Spain. In this area uncharacteristically, shrubby species such as *Sarcocornia fruticosa* predominate in the low marsh and *Limonium* spp. in the mid-marsh, and various *Juncus* species usually populate the high marsh.

Asia

The variety of climatic and edaphic conditions, the extent of the coastline, and the high frequency of embayments, islands, flats, estuaries, and river deltas, particularly in S.E. Asia, result in the highest biodiversity and the greatest areal coverage of mangroves in the world. Over 55 species of true mangroves occur in Asia (FAO, 2007). In East Asia, extensive coastal marshes occur along deltas formed by major rivers such as the Yangtze and Huang He. *S. alterniflora* was introduced to several areas in China in the late 1970s and has displaced many of the native plants, particularly in the lower marsh areas (Lu and Zhang, 2013).

Oceania

This region, which includes Australia, Papua New Guinea, New Zealand, and the S. Pacific islands, has a very high mangrove biodiversity (close to 50 species) but low areal extent of mangroves. They are found in protected bays, estuaries, lagoons, and coral atolls in the region. Close to 75 % of the mangrove coverage is concentrated in Australia (FAO, 2007). In Australia, *Spartina anglica* can occur seaward of mangrove areas. In more temperate areas, more conventional salt marsh zonation occurs.

Africa

Mangroves occur throughout the continent. On the east coast, mangroves often occur as narrow fringes except where large deltas (e.g., in Mozambique and Tanzania) allow the development of more extensive forests. Much broader forests develop in the west coast, culminating in the vast mangal associated with the Niger River delta that extends up to 40 km inland and supports very large trees that can reach 40 m in height. Salt marshes occur on the Mediterranean, Red Sea, and Indian Ocean coasts often behind mangroves. In Southern Africa, mangroves occur as far south as Angola, and extensive salt marshes south of there (Hughes et al., 1992).

Impacts to coastal wetlands

Human pressures currently cause the greatest impacts to wetlands. These include outright habitat loss due to residential, industrial, and agricultural development and associated infrastructure; or habitat degradation as a result of pollution, hydrological changes, and other impacts from surrounding human activities. Pest control activities mostly for mosquitoes can impact coastal wetlands, but habitat management for mosquito control has also been used as a marsh restoration tool (Rey et al., 2012b). In the United States, there were wetland losses of 146,200 ha in coastal watersheds of the eastern seacoast between 1998 and 2004 in spite of overall gains in wetland coverage during the same period (Stedman and Dahl, 2008). Between 2004 and 2009, salt marsh and estuarine emergent areas declined by 45,140 ha (Dahl, 2011). Worldwide mangrove losses between 1980 and 2005 have been estimated at more than 3.5 million ha (FAO, 2007) but the actual losses may be significantly greater (Giri et al., 2011).

As habitats that bridge marine and terrestrial ecosystems, coastal wetlands are particularly vulnerable to sea level changes and increased frequency of storms and other extreme events produced by climate change (Hopkinson et al., 2008). Depending upon circumstances, coastal wetlands may keep up with the relative rise, be lost, be degraded, or migrate landward in response to sea level rise (Gilman et al., 2007).

Summary

Coastal wetlands include seagrass communities, tidal flats, coastal salt marshes, and mangrove forests. They are important and complex ecosystems whose structure and function are determined by a large number of biotic and abiotic processes including non-resource factors such as salinity and climate, resources used by organisms for growth, and hydroperiod, with the latter being the dominant factor in salt marshes. These areas have great ecological importance due to the value and diversity of ecosystem functions that they provide. However, because of the desirability of the coastal zone for human habitation and associated infrastructure, coastal habitat degradation and loss is a serious problem worldwide. Also, because of their location in the interphase between the sea and the land, these habitats are particularly vulnerable to sea level changes and increased frequency of storms and other extreme events produced by climate change.

Bibliography

Berger, U., Rivera-Monroy, V. H., Doyle, T. W., Dahdouh-Guebas, F., Duke, N. C., Fontalvo-Herazo, M. L., Hildenbrandt, H., Koedam, N., Mehlig, U., Piou, C., and Twilley, R. R., 2008. Advances and limitations of individual-based models to analyze and predict dynamics of mangrove forests: a review. *Aquatic Botany*, **89**, 260–274.

D'Alpaos, A., Da Lio, C., and Marani, M., 2012. Biogeomorphology of tidal landforms: physical and biological processes shaping the tidal landscape. *Ecohydrology*, **5**, 550–562.

Dahl, T. E., 2011. *Status and Trends of Wetlands In the Conterminous United States 2004–2009*. Washington, D.C.: National Oceanic and Atmospheric Administration, National Marine Fisheries Service, and U.S. Fish and Wildlife Service.

FAO, 2007. *The World's Mangroves 1980–2005*. Rome: FAO Forestry Papers 153. FAO – Food and Agriculture Organization of the United Nations.

Gilman, E., Ellison, J., and Coleman, R., 2007. Assessment of mangrove response to projected relative sea-level rise and recent historical reconstruction of shoreline position. *Environmental Monitoring and Assessment*, **124**, 105–130.

Giri, C., Ochieng, E., Tieszen, L. L., Zhu, Z., Singh, A., Loveland, T., Masek, J., and Duke, N., 2011. Status and distribution of mangrove forests of the world using earth observation satellite data. *Global Ecology and Biogeography*, **20**, 154–159.

Gopal, B., and Junk, W. J., 2013. Biodiversity in wetlands: an introduction. In Gopal, B., Junk, W. J., and Davis, J. A. (eds.), *Biodiversity in Wetlands: Assessment, Function, and Conservation*. Leiden: Backhuys Publishers, pp. 1–10.

Gribsholt, B., and Kristensen, E., 2003. Benthic metabolism and sulfur cycling along an inundation gradient in a tidal *Spartina anglica* salt marsh. *Limnology and Oceanography*, **48**, 2151–2162.

Gribsholt, B., Kostka, J. E., and Kristensen, E., 2003. Impact of fiddler crabs and plant roots on sediment biogeochemistry in a Georgia saltmarsh. *Marine Ecology Progress Series*, **259**, 237–251.

Hopkinson, C. S., Lugo, A. E., Alber, M., Covich, A. P., and Van Bloem, S. J., 2008. Forecasting effects of sea-level rise and windstorms on coastal and inland ecosystems. *Frontiers in Ecology and the Environment*, **6**, 255–263.

Hughes, R. H., Hughes, J. S., and Bernacsek, G. M., 1992. *A Directory of African Wetlands*. Gland: International Union for the Conservation of Nature (IUCN).

Kurz, H., and Wagner, K., 1957. *Tidal marshes of the Gulf and Atlantic coasts of Northern Florida and Charleston, South Carolina*. Tallahassee: Florida State University Studies.

Lu, J. B., and Zhang, Y., 2013. Spatial distribution of an invasive plant *Spartina alterniflora* and its potential as biofuels in China. *Ecological Engineering*, **52**, 175–181.

Marani, M., D'Alpaos, A., Lanzoni, S., Carniello, L., and Rinaldo, A., 2007. Biologically-controlled multiple equilibria of tidal landforms and the fate of the venice lagoon. *Geophysical Research Letters*, **34**, L11402, doi:10.1029/2007GL030178.

Marani, M., Da Lio, C., and D'Alpaos, A., 2013. Vegetation engineers marsh morphology through multiple competing stable states. *Proceedings of the National Academy of Sciences of the United States of America*, **110**, 3259–3263.

Martini, I. P., Jefferies, R. L., Morrison, R. I. G., and Abraham, K. F., 2013. Polar coastal wetlands: development, structure, and land use. In Perillo, G. M. E., Wolanski, E., Cahoon, D. R., and Brinson, M. M. (eds.), *Coastal Wetlands. An Integrated Ecosystem Approach*. Amsterdam: Elsevier, pp. 119–156.

McCraith, B. J., Gardner, L. R., Wethey, D. S., and Moore, W. S., 2003. The effect of fiddler crab burrowing on sediment mixing and radionuclide profiles along a topographic gradient in a southeastern salt marsh. *Journal of Marine Research*, **61**, 359–390.

Mitsch, W. J., and Gosselink, J. G., 2007. *Wetlands*, 4th edn. Hoboken: John Wiley & Sons.

Montague, C. L., and Wiegert, R. G., 1990. Salt Marshes. In Myers, R. L., and Ewel, J. J. (eds.), *Ecosystems of Florida*. Orlando: University of Central Florida Press, pp. 481–516.

NOAA. 2013. *National Coastal Population Report*. National Ocean Service, Washington, D.C., 22 pp.

Nixon, S. W., and Oviatt, C. A., 1973. Ecology of a New England salt-marsh. *Ecological Monographs*, **43**, 463–498.

Perillo, G. M. E., 2009. Tidal courses: classification, origin and functionality. In Perillo, G. M. E., Wolanski, E., Cahoon, D. R., and Brinson, M. M. (eds.), *Coastal Wetlands. An Integrated Ecosystem Approach*. Amsterdam: Elsevier, pp. 185–209.

Pratolongo, P. D., Kirby, J. R., Plater, A., and Brinson, M. M., 2013. Temperate coastal wetlands: morphology, sediment processes, and plant communities. In Perillo, G. M. E., Wolanski, E., Cahoon, D. R., and Brinson, M. M. (eds.), *Coastal Wetlands. An Integrated Ecosystem Approach*. Amsterdam: Elsevier, pp. 89–118.

Rey, J. R., Carlson, D. B., and Brockmeyer, R., 2012a. Coastal wetland management in Florida: environmental concerns and human health. *Wetlands Ecology and Management*, **20**, 197–211.
Rey, J. R., Walton, W. E., Wolfe, R. J., Connelly, C. R., O'Connell, S. M., Berg, J., and Laderman, A. D., 2012b. North American wetlands and mosquito control. *International Journal of Environmental Research and Public Health*, **2012**, 4537–4605.
Stedman, S.-M., and Dahl, T. E., 2008. *Status and Trends of Wetlands in The Coastal Watersheds of the Eastern United States 1998 to 2004*. Washington, D.C.: National Oceanic and Atmospheric Administration, National Marine Fisheries Service, and U.S. Fish and Wildlife Service.
Twilley, R. R., and Rivera-Monroy, V. H., 2005. Developing performance measures of mangrove wetlands using simulation models of hydrology, nutrient biogeochemistry, and community dynamics. *Journal of Coastal Research*, **40**, 79–93.
UN (United Nations), 2013. UN Atlas of the Oceans. http://www.oceansatlas.org/servlet/CDSServlet?status=ND0xODc3JjY9ZW4mMzM9KiYzNz1rb3M~. Accessed April, 2013.

Cross-references

Barrier Island
Cordgrass
Deltas
Estuarine Deltaic Wetlands
Fiddler Crabs
Food Web/Trophic Dynamics
Mangroves
Nutrient Dynamics
Saltmarshes
Sandflat
Sea-Level Change and Coastal Wetlands
Tidal Flat
Wetlands

CORDGRASS

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Synonyms

Salt-marsh cordgrass; *Spartina* spp.

Definition

Cordgrass refers to species within the *Spartina* genus of grasses (Poaceae). Cordgrass is the dominant plant in salt marshes in many regions.

Description

The genus *Spartina* includes 15 species (ITIS, 2013). Cordgrass species have long slender leaves and tall inflorescences with many spikelets, each holding many seeds. They are perennial, wind pollinated, and, often, dispersal limited.

Cordgrass species are halophytic, with special adaptations like salt-excreting glands and aerenchymous rhizomes for saline and anoxic soil conditions. They are often the dominant plant in edaphically stressful intertidal salt-marsh ecosystems. In western Atlantic and Gulf of Mexico salt marshes, *Spartina alterniflora* is the dominant plant species, confined to the low marsh by competition with its congener, *S. patens*, which dominates the high marsh (Bertness, 1991). *S. densiflora* is the dominant plant in South American salt marshes.

Cordgrass species are highly invasive and often able to hybridize with each other. *S. townsendii* and *S. anglica* species originated by hybridization within the last 150 years (Daehler and Strong, 1996). *S. densiflora* is invasive in Spain (Nieva et al., 2005) and in California, where it has hybridized with the native *S. foliosa* (Ayres et al., 2008). *S. anglica*, *S. alterniflora*, and *S. patens* have been introduced in China (An et al., 2007) and in the western United States (Daehler and Strong, 1996).

Cordgrass has been introduced intentionally for shoreline stabilization (Ranwell, 1967). Cordgrass is a marsh builder and a true ecosystem engineer (Gedan et al., 2011). Cordgrass is also valuable as livestock fodder, central to salt-marsh food webs, and efficient at sequestering carbon and nitrogen.

Bibliography

An, S. Q., Gu, B. H., Zhou, C. F., Wang, Z. S., Deng, Z. F., Zhi, Y. B., Li, H. L., Chen, L., Yu, D. H., and Liu, Y. H., 2007. *Spartina* invasion in China: implications for invasive species management and future research. *Weed Research*, **47**, 183–191.
Ayres, D. R., Grotkopp, E., Zaremba, K., Sloop, C. M., Blum, M. J., Bailey, J. P., Anttila, C. K., and Strong, D. R., 2008. Hybridization between invasive *Spartina densiflora* (Poaceae) and native *S. foliosa* in San Francisco Bay, California, USA. *American Journal of Botany*, **95**, 713–719.
Bertness, M. D., 1991. Zonation of *Spartina patens* and *Spartina alterniflora* in New England salt marsh. *Ecology*, **72**, 138–148.
Daehler, C. C., and Strong, D. R., 1996. Status, prediction and prevention of introduced cordgrass *Spartina* spp. invasions in Pacific estuaries, USA. *Biological Conservation*, **78**, 51–58.
Gedan, K. B., Kirwan, M. L., Wolanski, E., Barbier, E. B., and Silliman, B. R., 2011. The present and future role of coastal wetland vegetation in protecting shorelines: answering recent challenges to the paradigm. *Climatic Change*, **106**, 7–29.
ITIS, 2013. *Spartina* [WWW Document]. Integrated Taxonomic Information System. http://www.itis.gov (accessed Jun 26, 2013).
Nieva, F. J. J., Castellanos, E. M., Castillo, J. M., and Figueroa, M. E., 2005. Clonal growth and tiller demography of the invader cordgrass *Spartina densiflora* Brongn. at two contrasting habitats in SW European salt marshes. *Wetlands*, **25**, 122–129.
Ranwell, D. S., 1967. World resources of *Spartina townsendii* (sensu lato) and economic use of *Spartina* marshland. *Journal of Applied Ecology*, **4**, 239–256.

Cross-references

Coastal Wetlands
Estuarine Deltaic Wetlands
Marsh Islands
Salt Marsh Accretion
Saltmarshes
Wetlands

CULTURAL SEAFOOD MANAGEMENT

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Definition

Cultural seafood management is seafood management based on customary (traditional) fishing rights and the integration of traditional knowledge in estuarine and coastal management.

Introduction

In most countries artisanal fisheries have existed for centuries with fishers using traditional and local knowledge in community or citizen-based management systems. While these systems are usually small scale, there is increasing interest in expanding or integrating these systems into larger and more complex systems (Reis and D'Incao, 2000). This need is driven by the knowledge that these habitats have changed and that seafood resources are limited. Increasingly there is recognition that traditional ecological knowledge (TEK) held by indigenous communities can play a valuable role in the management of natural resources including fisheries (Mathew, 2011). Practical skills, wisdom, and knowledge accumulated over successive generations can contribute to knowledge about species, their distributions, life histories, and behavior (Butler et al., 2012).

Throughout the world, estuaries are used for obtaining seafood and, when traditional knowledge is used with Western science and management knowledge (SMK), there are opportunities for generating a diversity of information for problem solving. Inshore marine fisheries such as those found in estuaries are often small; however, they may involve multiple species that are taken for subsistence. For example, in the islands of Torres Strait between Queensland and Papua New Guinea, people have high consumption of seafood based on up to 350 species, one of the highest numbers per capita in the world.

There are various levels of governance for seafood resources, and increasingly these are shared with government agencies. Management systems are generally region specific and may be driven by cultural keystone species. Cultural values, co-management, and power sharing of estuarine resources within a legislative framework occur in many countries, the first nation groups in North America, Melanesia, northern Australia, and New Zealand. The focus here is to use selective examples to illustrate the use of integrated management systems and their effectiveness in protecting seafood resources for future generations.

Traditional Ecological Knowledge (TEK) and Western Science and Management Knowledge (SMK)

Indigenous knowledge is used as a general term when describing knowledge systems of indigenous peoples, whereas traditional ecological knowledge is defined by Berkes (2003) as "a cumulative body of knowledge, practice and belief, evolving by adaptive processes and handed down through generations by cultural transmission." In a recent review, Thornton and Scheer (2012) use the term local and traditional ecological knowledge (LTK) which is described as a body of knowledge and not just a collection of facts. This knowledge is highly valued and has been used politically and strategically for territorial claims in many places including Oceania, North America, Central America, South Asia, and Southeast Asia. TEK/LTK is a knowledge-practice-belief complex, which is operated in many small fishing communities and where the fishery depends mainly on local knowledge of species rather than management systems, institutions, or world views. Knowledge can be grouped into nine cultural domains, as used by the Torres Strait islanders who occupy more than 150 islands between Queensland, Australia, and Western Province, PNG (Smyth et al., 2006, cited in Butler et al., 2012). Cultural domains include food source and skills associated with gathering, how it is used, including trade, environmental knowledge associated with collecting the food and "totem," the connection to various groups and mythological aspects, ceremony, beliefs, and art.

In contrast, the Western science management knowledge base (SMK) is usually underpinned by federal, state laws, and regulations implemented by agencies and departments. In most countries the responsibility for the marine and estuarine environment is split between multiple government agencies (with responsibilities defined by law) and operated by defense forces, coast guards, wildlife agencies, environmental protection agencies, health departments, fisheries, local authorities, and others. The areas under their control are large, and the managements driven by economic gains.

There are obvious differences in the form of the knowledge base resulting from SMK, which consists of published reports, qualitative analyses, and sharing of information, compared with TEK, where the information is transmitted orally and often only within families. Jokiel et al. (2011) compared major aspects of the traditional and Western knowledge systems for inshore reefs in Hawaii and listed other major differences including authority, enforcement, and resource monitoring.

Cultural keystone species

Ecologists recognize that some species have a key role in the structure and functioning of ecosystems and are essential in maintaining ecological balance. Similarly, in many societies, plants and animals can shape the cultural identity of a people. Several species together play a social role by interacting with each other forming a cultural grouping similar to a keystone guild. Cultural keystone species include green turtles and dugong for fishers from developing regions of Melanesia (Garibaldi and Turner, 2004). Their importance is reflected in the fundamental roles these species play in diet, materials, medicine, and/or spiritual practices. In Torres Strait, these keystone species have been important in creating links and understanding between TEK and SMK. The success of these is because of their cross-cultural values for the islanders and government conservation directives for the species. Another spin-off from the arrangement includes the establishment of a ranger program allowing community-based management of invertebrate species such as trochus and beche de la mer. Some estuarine cultural keystone species are large and iconic marine species, which have extensive ranges into coastal waters and provide ecosystem services and benefit from national as well as international beneficiaries. This has resulted in cross-scale partnerships between multiple indigenous communities and state and national government agencies. In broadening the approach from small-scale management to larger areas, several studies have reported successes in conservation and restoration by recognizing and focusing on cultural keystone species.

Traditional fisheries management

In many parts of the world, estuarine systems are threatened; anthropogenic changes have caused changes to estuarine habitats and their ecology. Furthermore, demographic changes, social political pressures, urbanization, education, commercialization, and technological advances have led to changed perceptions about the value of marine resources. Traditional community-based fisheries are usually small scale. Fishing is an integral part of many estuarine communities with immense cultural significance including beliefs about their origins and traditions. Ruddle (1993) provided numerous examples of traditional management from the Asia-Pacific region, including examples which he regarded as unsuccessful and those that were successful. He suggested several ways forward, favoring the use of legislation to reinforce but specify the power of traditional rights. In South Asia, one community-based traditional management system is Padu, which is based on caste and gender, a managed prawn or shrimp fishery in local lagoons where fishermen catch shrimp as they migrate from the estuary back to the sea. In recognizing the pressures exerted on management systems, previously unlicensed fishers from Kerala, Southern India, challenged the decision to prevent them from fishing in traditional areas and in the municipal courts gained official access to the fishery. The groups were well organized and charged with facilitating equitable access, providing social responsibility and providing mechanisms for rule making and resolving conflicts (Lobe and Berkes, 2004). More recently, Coulthard (2011) analyzed the Padu system in the Pulicat lagoon, India, where there are more than 30,000 artisan fishers. Here she describes the system as being in a fragile state due to poverty and reduced income because of widely fluctuating catches. Still the fishers remain loyal to the Padu system which provides political power, social standing, and prestige. This system is quite different to the systems used by larger commercial fisheries which run alongside the traditional fisheries usually with science-based stock assessments for individual species. They use fishery-specific tools (not discussed here) such as licensing, temporal closures, gear restrictions, and size limits. In the future it seems likely that such intervention measures will need to be included in traditional fisheries so that they can be of benefit to future generations.

Natural resource management

For centuries the Polynesians, who inhabited Hawaii, used a management system "ahupua'a," an integrated watershed management system between freshwater and the nearby marine coastal environment, based on ecosystem linkages between the mountains and the sea (Jokiel et al., 2011). This concept gained prominence in New Zealand, where centuries of observation and the continued practice of gathering mahinga kai and kaimoana, Ngāi Tahu whāui have built a unique body of experience and knowledge which is important for understanding the environment and maintaining its health and well-being. Mauri, which is an important part of Māori culture, is the energy or life force which is sacred and a spiritual link to the past the present and the future. Pauling (2003) developed cultural tools for mountains to the sea natural resource monitoring, including evaluating mahinga kai (food) and resources. The state of the Takiwā assessments, Te Āhuatanga o te Ihutai, was undertaken to establish the cultural health of the Avon-Heathcote Ihutai Estuary, a small estuary near Christchurch, in the South Island of New Zealand, where modifications of the estuary and the previous release of human waste made the estuary unsafe and culturally unacceptable for gathering kai moana. Surveys were undertaken to provide an assessment of the current health and provide ideas about how future management might improve the cultural health of the catchment. Cultural health scores were based on the status of the site, suitability to harvest mahinga kai, physical and legal access, site pressure, degree of modification, and the identification of valued and pest species. These were used with other assessment tools, stream health monitoring, bacterial water quality, and electric fishing. In 2007, when the first assessments were made, the cultural values of the catchment were poor and these did not improve in the 2012 assessment (Lang et al., 2012). Improvement had been expected because of the removal of treated waste effluent from the estuary. Unfortunately, a series of large earthquakes damaged the infrastructure. Raw sewage was released into the rivers and estuary; in addition, sediments were disturbed. These events most likely explain the lack of improvement in cultural health of the habitat.

Cooperative management or co-management

In an early review, Sen and Neilsen (1996) described fisheries co-management as an arrangement where responsibility for resource management is shared between the government and user groups. It is an element of community fisheries management. Governments and scientists have concerns about the impacts of harvesting on populations, and the conflicting perceptions about the status of the shellfish stocks have amplified the need for co-management. Using examples from small-scale fisheries from Africa, Asia, the Caribbean, Europe, North America, and the Pacific, he concluded that most of the examples were at an early stage of implementation and the reason for introducing measures was because of overexploitation of the fishery stock or to resolve conflicts between users. Co-management is seen as one way to increase the resilience in the system to environmental and other changes. This management system depends on the integration of knowledge, the processes used, and the degree of power sharing (Wilson et al., 2006; Berkes, 2009). Three stages are described in co-management (Plummer, 2006; Butler et al., 2012). These are (1) "independence," where there is limited interaction between government and local people; (2) "association," the start of an exchange of information and resource evaluation and shared vision; and (3) "integration," where there is a sharing of the consequences of actions and resolving conflicts. Traditional fisheries often contain unlicensed fishing for general or cultural use, whereas some co-fisheries arrangements are controlled under treaties. In Torres Strait, these include dugong, green turtle (separate management plans for each), and the reef fisheries, which include both fish and invertebrates (Kwan et al., 2006). The objectives of the plans are to achieve sustainability, revive TEK, and allow islanders control in decision making. Although this is an example of co-management, Weiss et al. (2012) suggests that this is dominated by top-down government management. One aspect of co-management is that it depends on resources; for example, in Torres Strait, the fisheries management is well resourced compared with neighboring Melanesian nations (Butler et al., 2012). Successful co-management is a knowledge partnership and can be difficult where it involves indigenous peoples whose knowledge is based on different worldviews (Berkes, 2009). Using science together with traditional knowledge is not simply a synthesis of the two kinds of knowledge, but an ability to develop mutual respect and trust which may not always succeed (Spak, 2005).

Ecosystem-based management

The ecosystem-based management tool was originally developed in response to the impacts of fishing on fish stocks and habitat degeneration due to natural and anthropogenic factors. Fishers were targeted because their communities have biological, oceanographic, economic, social, and cultural aspects which can contribute to the fisheries management (Aswani, 2011). The challenge is to validate the information and create policies and legal ways to integrate information into fishery-management systems (Mathew, 2011). Some suggested ways of doing this included recognizing fishers as holders of knowledge and accepting customary law. It was also suggested that governments and managers need to guarantee the approaches to conservation will be fair and developed under a system of co-management. It was felt that scientists need to overcome their reluctance to use nonspecialist knowledge instead of data. They also need to support the requirements of the communities. Finally, mechanisms need to be established to guarantee that knowledge sharing will benefit and not harm the fishers. Knowledge should be used to conserve the fish stocks and protect the habitat for long-term food security. There should be conflict resolutions in place and the propriety rights of traditional knowledge will remain with the providers.

Integrated coastal management

This is defined as a continuous and dynamic process allowing decisions to be made for the sustainable use of development and protection of coastal and marine areas and resources. The Haida Nation in northern British Columbia has been resource owners and managers for thousands of years on Haida Gwaii, where they apply traditional knowledge and experience to fisheries management. Their approach is based on Haida ethics, principles or values of respect, balance, interconnectedness, seeking wise council, and responsibility. They demonstrate a commitment to responsible and respectful management of marine resources and ecosystems. Plans for the Pacific North Coast Integrated Management Area (PNVIMA) are being led by the Council of the Haida Nation, under the Haida constitution in an example of local community-based and co-governance. The lead agency is Fisheries and Oceans Canada and their approach is for collaborative management with Aboriginal peoples, a change from the previous regime where government policies tended to displace first nations from marine resources (Jones et al., 2010). This initiative was launched in March 2009 and is ongoing, strengthened by court decisions and policies relating to Aboriginal rights and title. This approach has resulted in the establishment of a network of marine-protected areas with levels of protection influenced by first nations and marine resource management that includes principles of social equity and ecosystem justice.

Conclusions

There is overwhelming support for including traditional knowledge in resource management of estuarine and coastal marine areas and an increasing number of examples where this has been shown to benefit all users. Worldwide, the extent of local or traditional knowledge is patchy; some has been lost and in some countries local fisheries knowledge may be largely exclusive to women (Ruddle and Hickey, 2008). There is therefore an urgent need to systematically collect as well as archive cultural, historic, and contemporary information. These different sources of information can then be combined using geo-spatial information systems as suggested by De Freitas et al. (2009). Their multilayer GIS database for artisanal fisheries in Brazil integrates both traditional and scientific data allowing analyses of catch data for target species and highlighting estuarine areas that are likely to be under pressure from overfishing. The GIS tools integrate and translate complex data into an accessible format using maps, and there are widespread applications from such a geodatabase, both culturally and commercially. Also, when updated regularly, they can be used to rapidly respond to changing conditions or emergencies that may require the use of management tools such as temporary closures.

By incorporating traditional knowledge, integrated coastal management is seen as the way forward for the protection and use of marine and estuarine resources (Aswani et al., 2012). The planning and implementing of this management must be backed by legislation that clearly defines the rights of indigenous and government stakeholders. This is the case in New Zealand, where the Treaty of Waitangi is a legal partnership between the British Crown and Māori and the Conservation Act directs the Department of Conservation to establish co-management arrangements with Māori. Mātaitai reserves are authorized by the Minister of fisheries to manage and control seafood harvesting in keeping with local sustainable management practices. Tangata tiaki/kaitiaki recommends bylaws and issues customary food authorizations, and, while commercial fishing may not be allowed, both Māori and non-Māori are allowed to fish in reserve areas. A lack of communication and mistrust can impede negotiations between indigenous and government representatives, and so a framework is required to integrate information in a respectful way (Gratani et al., 2011) and guarantee that the proprietary rights to traditional knowledge remain with the providers. It is hoped that new management systems for seafood will reduce conflicts between users, decrease overexploitation, and revitalize conservation traditions that may have been lost.

Bibliography

Aswani, S., 2011. Socioecological approaches for combining ecosystem-based and customary management in Oceania. *Journal of Marine Biology*, 2011, ID 845385, 13 pp.

Aswani, S., Christie, P., Muthiga, N. A., Mahon, R., Primavera, J. H., Cramer, L. A., Barbier, E. B., Granek, E. F., Kennedy, C. J., Wolanski, E., and Hacker, S., 2012. The way forward with ecosystem-based management in tropical contexts: reconciling with existing management systems. *Marine Policy*, **36**, 1–10.

Berkes, F., 2003. Rethinking community-based conservation. *Conservation Biology*, **18**, 621–630.

Berkes, F., 2009. Evolution of co-management: role of knowledge generation, bridging organisations and social learning. *Journal of Environmental Management*, **90**, 1692–1702.

Butler, J. R. A., Tawake, A., Skewes, T., Tawake, L., and McGrath, V., 2012. Integrating traditional ecological knowledge and fisheries management in the Torres Strait, Australia: the catalytic role of turtles and dugong as cultural keystone species. *Ecology and Society*, **17**(4), 34.

Coulthard, S., 2011. More than just access to fish. The pros and cons of fisher participation in a customary marine tenure (Padu) system under pressure. *Marine Policy*, **35**, 405–4011.

De Freitas, D. M., and Tagliani, P. R. A., 2009. The use of GIS for the integration of traditional and scientific knowledge in supporting artisanal fisheries management in southern Brazil. *Journal of Environmental Management*, **90**, 2071–2080.

Garibaldi, A., and Turner, N., 2004. Cultural keystone species: implications for ecological conservation and restoration. *Ecology and Society*, **9**(3), 1. http://dx.doi.org.

Gratani, M., Butler, J. R. A., Royee, F., Valentine, P., Burrows, D., Canendo, W. I., and Anderson, A. S., 2011. Is validation of indigenous ecological knowledge a disrespectful process? A case study of traditional fishing poisons and invasive fish management from the wet tropics, Australia. *Ecology and Society*, **16**, 25. http://dx.doi.org.

Jokiel, P. L., Rogers, K. S., Walsh, W. J., Polhemus, D. A., and Wilhelm, T.A., 2011. Marine resource management in the Hawaiian Archipelago: the traditional system in relation to the western approach. *Journal of Marine Biology*, Article ID151682, 16 pp.

Jones, R., Rigg, C., and Lee, L., 2010. Haida marine planning: first nations as a partner in marine conservation. *Ecology and Society*, **15**, 12. http://www.ecologyandsociety.org.

Kwan, D., Marsh, H., and Delan, S., 2006. Factors influencing the sustainability of customary dugong hunting by a remote indigenous community. *Environmental Conservation*, **33**, 164–171.

Lang, M., Orchard, S., Falwasser, T., Rupene, M., Williams, C., Tirikatene-Nash, N., and Couch, R., 2012. State of the Takiwā 2012, *Te Āhuatanga o Te Ihutai. Cultural health assessments of the Avon-Heathcote Estuary and its catchment*. Mahaanui Kurataiao Ltd.

Lobe, K., and Berkes, F., 2004. The Padu system of community-based fisheries management: change and local institutional innovation in south India. *Marine Policy*, **28**, 271–281.

Mathew, S., 2011. Fishery-dependent information and the ecosystem approach: what role can fishers and their knowledge play in developing countries? *ICES Journal of Marine Science*, **68**, 1805–1808.

Pauling, C., 2003. *Ki Uta Ki Tai–Mountains to the Sea Natural Resource Management: A Scoping Document for Developing Mountains to the Sea Natural Resource Management Tools for Ngai Tahu*. Christchurch, NZ: Te Rununga o Ngāi Tahu.

Plummer, R., 2006. Sharing the management of a river corridor: a case study of the co-management process. *Society and Natural Resources*, **19**, 709–721.

Reis, E. G., and D'Incao, F., 2000. The present status of artisanal fisheries of extreme southern Brazil: an effort towards community-based management. *Ocean and Coastal Management*, **43**, 585–595.

Ruddle, K., 1993. External forces and change in traditional community-based fishery management systems in the Asia-Pacific Region. *Maritime Anthropological Studies*, **6**, 1–37.

Ruddle, K., and Hickey, F. R., 2008. Accounting for the mismanagement of tropical nearshore fisheries. *Environmental Development and Sustainability*, **10**, 657–675.

Sen, S., and Nielsen, J.-R., 1996. Fisheries co-management: a comparative analysis. *Marine Policy*, **20**, 405–418.

Spak, S., 2005. The position of indigenous knowledge in Canadian co-management organizations. *Anthropologica*, **47**, 233–246.

Thornton, T. F., and Scheer, A. M., 2012. Collective engagement of local and traditional knowledge and science in marine environments. *Ecology and Society*, **17**(3), 8. http://dx.doi.org.

Weiss, K., Hamann, M., Kinney, M., and Marsh, H., 2012. Knowledge exchange and policy influence in a marine resource governance network. *Global Environmental Change*, **22**, 178–188.

Wilson, D. C., Ahmed, M., Siar, S. V., and Kanagaratnam, U., 2006. Cross-scale linkages and adaptive management: fisheries co-management in Asia. *Marine Policy*, **30**, 523–533.

Cross-references

Adaptive Management
Ecosystem-Based Management
Estuarine Habitat Restoration
Fish Assemblages

CYANOBACTERIA

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Synonyms

Blue-green algae

Definition

Cyanobacteria or cyanophyceae (initially known as blue-green microalgae) are nonmotile and planktonic photosynthetic prokaryotes, belonging to the kingdom Eubacteria, division of Cyanophyta. They are common in some extreme environments and occasionally form dense blooms harmful to estuarine environments.

Introduction

Cyanobacteria (photosynthetic prokaryotes) are the Earth's oldest known oxygenic photoautotrophs (Pearl and Otten, 2013). The fossil records of Schopf et al. (2000) suggest that cyanobacteria have been present on earth for at least 3.5 billion years, being distributed worldwide from polar to equatorial latitudes (Vincent, 2000; Wynn-Williams, 2000). Their proliferation during the Precambrian era (~3.5 bya) dramatically altered the previously anoxic biosphere which led to the evolution of higher terrestrial plant and animal life (Schopf et al., 2000). Thus, the cyanobacteria group constitutes a large and morphologically diverse phylum with more than 4,000 isolates and 19 of the most important taxa (species).

Many genera have the ability to fix atmospheric nitrogen (N_2) (through an anaerobic process), while they can store phosphorus (P) and sequester iron (Fe) and a range of essential trace metals (Whitton, 2012). These traits enable them to exploit both nutrient-scarce and nutrient-enriched, diverse terrestrial and aquatic environments worldwide. The cyanobacteria present a range of attributes that give them, in certain environmental conditions, a clear competitive growth advantage over planktonic algae, and therefore they thrive in all kinds of environments (Gomes et al., 2012). Besides planktonic forms, benthic cyanobacteria constitute the principal colonizers at the interface between sediments and water, where they affect fluid flow dynamics and structure formation (Whitton and Potts, 2000). Once classified as microalgae, the cyanobacteria produce photosynthetic pigments (chlorophyll *a* and/or other accessory pigments such as phycocyanin, allophycocyanin, and phycoerythrin) (Briand et al., 2003).

Here, the status of cyanobacteria in estuarine environments is reviewed, their biological and ecological features, and roles in primary sedimentary structures. In addition, the effects of anthropogenic and climate change on cyanobacteria blooms and toxicities are examined.

Ecobiology of cyanobacteria

Cellular morphological features of cyanobacteria are very diverse, including spherical, ovoid, and cylindrical unicellular species, as well as multicellular colonial and filamentous forms (Couté et al., 2001). Some species are able to differentiate specialized cells: (1) heterocysts which are able to fix nitrogen in water under N-limited conditions; and (2) akinetes which tolerate stressful conditions such as periods of high temperature or drought. Cyanobacterial species are sometimes difficult to identify due to their high phenotypic plasticity (Briand et al., 2003).

These organisms comprise a unique phylogenetic group of bacteria that perform oxygenic photosynthesis (Hackenberg et al., 2011). In addition, cyanobacteria occupy diverse ecological niches and exhibit enormous diversity in terms of their habitats, physiology, morphology, and metabolic capabilities (Beck et al., 2012). In fact, cyanobacteria are able to establish competitive growth in almost any environment where there is, at least temporarily, water and sunlight (Badger et al., 2006; Esteves-Ferreira et al., 2013).

The most recent taxonomic classification of cyanobacteria is based on the so-called polyphasic approach (Johansen and Casamatta, 2005). In this approach, molecular phylogenetic analyses are the basic criteria for classification of genera and species, with the cytological and morphological markers (synapomorphic and autapomorphic characters) and the ecology (habitat preference, life strategy, and ecophysiology) considered an integral part of the taxonomic definition, with additional important biochemical and molecular markers (Komarek and Mares, 2012).

Two morphological types are distinguished within the cyanobacteria group: (1) the filamentous species forming elongated cell chains (trichomes) often bundled together (multi-trichomous species); and (2) the coccoid species forming spheroidal cells often arranged in cell clusters (Staley et al., 1989; Whitton and Potts, 2000). In estuarine ecosystems, cyanobacteria are primary producers that use light energy to synthesize organic matter from mineral nutrients and CO_2 (photosynthesis). Their specific physiologic capabilities enable them to compete very efficiently with other photosynthetic microorganisms and to regulate their buoyancy (by means of gas vacuoles). Thus, they can colonize different depths in the water column depending on the location of nutrients and availability of light (Klemer et al., 1982; Walsby et al., 2001).

Their possession of accessory pigments, such as phycoerythrin, enable several cyanobacteria species to carry out photosynthesis at depths that receive only green light and where, in addition, nutrients are more abundant than at the surface (as in the case of surface waters rapidly depleted following spring algal proliferations). Cyanobacterial pigments, as well as mycosporin-like amino acids, are involved in their capacity to resist ultraviolet radiation in surface waters, giving them another advantage over some phytoplankton (Castenholz et al., 2000).

The cyanobacteria are poorly grazed by zooplankton due to their production of mucilage layers (Mur et al., 1999; Goleski et al., 2010). Recent data reveal that cyanobacteria have adopted a mode of defense depending on grazer pressure (i.e., they are able to modify their defense reaction according to the actual risk of grazing) (Gomes et al., 2012). The synthesis of different toxins by many cyanobacterial blooms gives them a selective advantage, since some zooplanktonic predators are susceptible to these toxins and thus avoid eating cyanobacteria (Jacquet et al., 2004; Oliver et al., 2010).

The biology and ecology of cyanobacteria have been extensively studied throughout the world during the two last decades due to their expansion and proliferation in most aquatic environments (Pearl and Otten, 2013). Several inner ecophysiological strategies allow the cyanobacteria to exploit anthropogenic modifications of aquatic environments (specifically nutrient over-enrichment and hydrologic alterations). Thus when conditions of light and water column stability are favorable, cyanobacteria may proliferate creating a competitive advantage over other species of phytoplankton (Smith and Bennet, 1999; Peter et al., 2002). Because benthic cyanobacteria are the principal colonizers of the interface between sediments and water, they can affect fluid flow dynamics and structure formation in biofilms and microbial mats. Therefore, they greatly influence the sedimentary dynamics of peritidal depositional systems as noted by Vincent et al. (2000) and Noffke et al. (2003).

Cyanobacteria status in estuarine ecosystems

Due to their salinity gradients, estuaries provide a large variety of aquatic habitats for native populations of marine, brackish, and freshwater planktonic species (Telesh, 2004). The spatial zoning and functional characteristics of estuaries result in biologically active zones with high concentrations of bacteria and microalgae (Golubkov et al., 2001). Further, the microbial communities of estuarine ecosystems are susceptible to rapid changes in response to the flux of environmental conditions. Thus, the flux of dissolved and suspended organic and inorganic material, in addition to hydrological variations, significantly affects microbial abundance, diversity, and activity in the estuarine ecosystems (Bouvy et al., 2010).

Given sufficient nitrogen inputs, estuarine and coastal marine environments can be driven by phosphorus limitation which contributes to greater far field nitrogen enrichment and eutrophication at greater distances (Howarth et al., 2011). Nutrient loading from coastal watersheds and upstream systems typically deliver higher quantities of nutrients than those entering from coastal ocean waters (Galloway et al., 2004; Fennel et al., 2006).

In estuarine waters with salinities greater than 8–10 ppt, planktonic cyanobacteria capable of N-fixation are largely absent (Howarth and Marino, 2006; Marino et al., 2006; Howarth and Paerl, 2008). A decrease in planktonic N-fixation in estuaries has been attributed in part to high levels of sulfates in seawater, making the assimilation of molybdenum (an element required for N-fixation) difficult. This leads to slow potential growth rates of N-fixing cyanobacteria (heterocystus cyanobacteria, where N-fixation occurs only in heterocyst cells) exposed to grazing by zooplankton and benthic animals (Chan et al., 2006).

Cyanotoxicity and cyanobacterial blooms

A notable increase in occurrence and intensity of cyanobacteria toxic blooms has been observed worldwide over the last several decades (Eiler and Bertilsson, 2004; Pearson and Neilan, 2008; Rinta-Kanto et al., 2009). For major cyanobacterial genera involved in harmful blooms, the optimal growth rates and bloom potentials have increased with higher water temperatures; thus global warming may be playing a key role in the expansion and persistence of bloom-forming cyanobacterial taxa (Pearl and Fulton, 2006).

A recent study by Pearl et al. (2013) showed how cyanobacterial surface blooms may locally increase surface water temperatures due to light energy absorption via an array of photosynthetic and photoprotective pigments (chlorophylls, carotenoids, and phycobilins). This represents a positive feedback mechanism by which cyanobacterial bloom species can optimize their growth rates leading to competitive dominance over eukaryotic phytoplankton. Global warming, therefore, may enhance cyanobacterial dominance in the plankton as reported by Bonilla (2012).

Cyanobacterial blooms are complex microbial assemblages, consisting of many representatives from characterized phyla (Pope and Patel, 2008; Li et al., 2011; Wilhelm et al., 2011). The morphological features of organisms within a bloom appear as associative microbial assemblages analogous to biofilms (Zehr et al., 1995; Reid et al., 2000; Omoregie et al., 2004; Burke et al., 2011).

The initiation, maintenance, and subsequent decline of cyanobacteria blooms depend to a large extent on the availability of nitrogen (N) and phosphorus (P) (Levich, 1996). It also depends on the ratios of N and P, selecting for organisms capable of fixing atmospheric nitrogen over those lacking this physiology (Klausmeier et al., 2004).

Among the harmful cyanobacteria species cited elsewhere, the most common toxin producing cyanobacteria N_2-fixing genera are *Anabaena*, *Aphanizomenon*, *Cylindrospermopsis*, *Lyngbya*, *Nodularia*, *Oscillatoria*, and *Trichodesmium*, while the non-N_2 fixers are *Microcystis* and *Planktothrix* which thrive in fresh and estuarine environments as well as in marine systems (Pearl et al., 2013). The major harmful toxins produced by toxic or harmful cyanobacteria are large classes of natural polyketides compounds, nonribosomal peptides, or a mixture of both (Moreira et al., 2013). Their biosynthesis is performed by a family of multi-enzymatic complexes called nonribosomal peptide synthetases (NRPS) and polyketide synthases (PKS) organized into repeated functional units known as modules (Carmichael, 1992; Cane et al., 1999).

The "harmful" environmental aspect of cyanobacterial blooms described by Pearl et al. (2013) begins with a loss of water clarity, suppression of aquatic macrophytes, and negative effects on invertebrate and fish habitats. Consequently, the bacterial decomposition of dying blooms may lead to oxygen depletion (hypoxia and anoxia) and subsequent fish kills.

Smith et al. (2008) also indicates that cyanobacterial odorous and bioactive metabolites have a negative impact on aquaculture organisms. The toxins cause mortality of aquaculture organisms or harm consumers consuming the seafood products via accumulation of hepatoxins, cytotoxins, neurotoxins, dermatoxins, and brine shrimp/molluscan toxins. Some metabolites degrade the nutritional state of aquaculture species (inhibitors of proteases and grazer deterrents). Aquaculture species or aquaculture workers can be seriously impacted by dermatoxins, irritant toxins, hepatoxins, and cytotoxins.

The cyanobacterium *Microcystis aeruginosa* is the most common bloom-forming and hepatotoxin-producing species of cyanobacteria. It is known to produce the hepatotoxic heptapeptide microcystin in a variety of forms (Kaebernick et al., 2000). Microcystin binds to the multispecific bile acid transport system, subsequently causing toxic effects on hepatocytes. The effect is the inhibition of eukaryotic protein phosphatases PP2A and PP1 (Ppp1, Ppp2, Ppp4, Ppp5, and Ppp6) that are involved in tumor promotion and genotoxicity (Moreira et al., 2013).

The most recent data on microcystin is that it occurs worldwide. Anthropogenic nutrient loading, rising temperatures, enhanced vertical stratification, and an increase in residence time favor cyanobacterial dominance and CyanoHAB proliferation in a wide range of aquatic ecosystems (Pearl and Otten, 2013).

Summary

Cyanobacteria can adapt to dramatic changes in hydrobiological conditions. They have numerous physiological adaptations and mechanisms that enable them to take advantage of environmental changes and extremes that influence the biosynthesis of cyanotoxins for several cyanobacterial species.

The occurrence of harmful cyanobacteria has been linked to an increase in nutrient pollution in aquatic ecosystems. Future climate change is predicted to cause shifts in species composition of cyanobacterial blooms favoring invasive species since modern global distributions of cyanobacterial species result from differences in evolutionary adaptations and phenotypic traits.

Bibliography

Badger, M. R., Price, G. D., Long, B. M., and Woodger, F. J., 2006. The environmental plasticity and ecological genomics of the cyanobacterial CO2 concentrating mechanism. *Journal of Experimental Botany*, **57**, 249–265.

Beck, C., Knoop, H., Axmann, I., and Steuer, R., 2012. The diversity of cyanobacterial metabolism: genome analysis of multiple phototrophic microorganisms. *BMC Genomics*, **13**(1), 56.

Bonilla, S., Aubriot, L., Soares, M. C. S., Gonzalez-Piana, M., Fabre, A., Huszar, V. L. M., Lurling, M., Antoniades, D., Padisak, J., and Carla, K., 2012. What drives the distribution of the bloom-forming cyanobacteria *Planktothrix agardhii* and *Cylindrospermopsis raciborskii*? *FEMS Microbiology Ecology*, **79**, 594–607.

Bouvy, M., Arfi, R., Bernard, C., Carré, C., Got, P., Pagano, M., and Troussellier, M., 2010. Estuarine microbial community characterization as indicators of human induced changes (Senegal River, West Africa). *Estuarine, Coastal and Shelf Science*, **87**, 573–582.

Briand, J. F., Jacquet, S., Bernard, C., and Humbert, F., 2003. Health hazards for terrestrial vertebrates from toxic cyanobacteria in surface water ecosystems. *Veterinary Research*, **34**, 361–377.

Burke, C., Thomas, T., Lewis, M., Steinberg, P., and Kjelleberg, S., 2011. Composition, uniqueness and variability of the epiphytic bacterial community of the green alga *Ulva austrealis. ISME Journal*, **5**, 590–600.

Cane, D. E., and Walsh, C. T., 1999. The parallel and convergent universes of polyketide synthases and non ribosomal peptide synthetases. *Chemical Biology*, **6**, R319–R325.

Carmichael, W., 1992. Cyanobacteria secondary metabolites: the cyanotoxins. *Journal of Applied Bacteriology*, **72**, 445–459.

Castenholz, R. W., and Garcia-Pichel, F., 2000. Cyano- bacterial responses to UV-radiation. In: Whitton, B. A., and Potts M. (eds.), *The Ecology of cyanobacteria, their diversity in time and space*, Dordrecht: Kluwer Academic Publishers, pp. 591–611.

Chan, F., Marino, R., Howarth, R. W., and Pace, M. L., 2006. Experimental tests of ecological constraints on planktonic nitrogen fixation in saline estuaries: II. Grazing controls on cyanobacterial population dynamics. *Marine Ecological Progress Series*, **309**, 41–53.

Couté, A., and Bernard, C., 2001. Les cyanobactéries toxiques. In Frémy, J. M., and Lassus, P. (eds.), Toxines d'algues dans l'alimentation, Ed. Ifremer, pp. 89–108.

Eiler, A., and Bertilsson, S., 2004. Composition of freshwater bacterial com- munities associated with cyanobacterial blooms in four Swedish lakes. *Environmental Microbiology*, **6**(1), 228–1243.

Esteves-Ferreira, A. A., Corrêa, D. M., Carneiro, A. P. S., Rosa, R. M., Loterio, R., and Araújo, W. L., 2013. Comparative evaluation of different preservation methods for cyanobacterial strains. *Journal of Applied Phycology*, **25**, 919–929.

Fennel, K., Wilkin, J., and Levin, J., 2006. Nitrogen cycling in the Middle Atlantic Bight: results from a three-dimensional model and implications for the North Atlantic nitrogen budget. *Global Biogeochemical Cycle*, **20**, GB3007, doi:10.1029 / 2005GB002456.

Galloway, J. N., et al., 2004. Nitrogen cycles: past, present, and future. *Biogeochemistry*, **70**, 153–226.

Goleski, J. A., Koch, F., Marcoval, M. A., Wall, C. C., Jochem, F. J., and Peterson, B. J., 2010. The role of zooplankton grazing and nutrient loading in the occurrence of harmful cyanobacterial blooms in Florida Bay. *Estuaries and Coasts*, **33**, 1202–1215.

Golubkov, M., Balushkina, E. V., Anokhina, L. E., Nikulina, V. N., Orlova, M. I., Panov, V. E., and Umnova, L. P., 2001. The role of biological active zones in the organic pollution and purification of the Neva Estuary. *Proccedings of the Zoological Institute of the Russian Academy of Sciences*, **289**, 95–100.

Gomes, L. N. L., Corrêa Oliveira, S. M. A., Giani, A., and Gomez, E. S., 2012. Association between biotic and abiotic parameters and the occurrence of cyanobacteria in a Brazilian. *Environment Monitoring Assessment*, **184**, 4635–4645.

Hackenberg, C., Kern, R., Hüge, J., Stal, L. J., Tsuji, Y., Kopka, J., Shiraiwa, Y., Bauwe, H., and Hagemann, M., 2011. Cyanobacterial lactate oxidases serve as essential partners in N2 fixation and evolved into photo-respiratory glycolate oxidases in plants. *Plant Cell*, **23**, 2978–2990.

Howarth, R. W., and Marino, R., 2006. Nitrogen as the limiting nutrient for eutrophication in coastal marine ecosystems: evolution views over three decades Limnol. *Oceanography*, **51**(1, part 2), 364–376.

Howarth, R. W., and Paerl, H., 2008. Coastal marine eutrophication: control of both nitrogen and phosphorus is necessary. *Proceedings of the National Academy of Sciences of the United States of America*, **105**, E103.

Howarth, R. W., Chan, F., Conley, D. J., Garnier, J., Doney, S. C., Marino, R., and Billen, G., 2011. Coupled biogeochemical cycles: eutrophication and hypoxia in temperate estuaries and coastal marine ecosystems. *Front Ecology and Environment*, **9**(1), 18–26, doi:10.1890/100008.

Jacquet, S., Briand, J. F., Leboulanger, C., Avois-Jacquet, C., Druart, L., and Anneville, O., 2004. The proliferation of the toxic cyanobacterium *Planktothrix rubescens* following restoration of the largest natural French lake (Lac du Bourget). *Harmful Algae*, **4**, 651–672.

Johansen, J. R., and Casamatta, D. A., 2005. Recognizing of cyanobacterial diversity through adoption of a new species paradigm. *Algalogical Studies*, **117**, 71–93.

Kaebernick, M., Neilan, B. A., and Börner, T., 2000. Light and the transcriptional response of the microcystin biosynthesis gene cluster. *Applied and Environmental Microbiology*, **66**(8), 3387–3392.

Klausmeier, C. A., Litchman, E., Daufresne, T., and Levin, S. A., 2004. Optimal nitrogen-to-phosphorus stoichiometry of phytoplankton. *Nature*, **429**, 171–174.

Klemer, A. R., Feuillade, J., and Feuillade, M., 1982. Cyanobacterial blooms: carbon and nitrogen have opposite effects on the buoyancy of *Oscillatoria. Science*, **215**, 1629–1631.

Komarek, J., and Mares, J., 2012. Update to modern taxonomy of fresh water planktic heterocytous cyanobacteria. *Hydrobiologia*, **698**, 327–351.

Levich, A., 1996. The role of nitrogen-phosphorus ratio in selecting for dominance of phytoplankton by cyanobacteria or green algae and its application to reservoir management. *Journal of Aquatic Ecosystems Stress Recoveries*, **5**, 55–61.

Li, N., Zhang, L., Li, F., Wang, Y., Zhu, Y., Kang, H., Wang, S., and Qin, S., 2011. Metagenome of microorganisms associated with the toxic Cyanobacteria *Microcystis aeruginosa* analyzed using the 454 sequencing platform. *Chin J Oceanol Limnol*, **29**, 505–513.

Marino, R., Howarth, W., Pace, M.L., and Likens, L.G., 2006. A mesocosm test of ecological controls on planktonic cyanobacterial growth and N2 fixation in saline estuaries: I. Nutrients and grazing. *Marine Ecology Progress Series*, **309**, 25–39.

Moreira, C., Vasconcelos, V., and Antunes, A., 2013. Phylogeny of microcystins: evidence of a biogeographical trend? *Current Microbiology*, **66**, 214–221.

Mur, L., Skulberg, O., and Utkilen, H., 1999. Cyanobacteria in the environment. In Chorus, I., and Bartram, J. (eds.), *Toxic Cyanobacteria in Water*. London/NewYork: E & FN Spon, pp. 15–40.

Noffke, N., Gerdesb, G., and Klenkec, T., 2003. Benthic cyanobacteria and their influence on the sedimentary dynamics of peritidal depositional systems (siliciclastic, evaporitic salty, and evaporitic carbonatic). *Earth-Science Reviews*, **62**, 163–176.

Oliver, R. L., Mitrovic, S. M., and Rees, C., 2010. Influence of salinity on light conditions and phytoplankton growth in a turbid river. *River Research and Applications*, **26**, 894–903.

Omoregie, E. O., Crumbliss, L. L., Bebout, B. M., and Zehr, J. P., 2004. Determination of nitrogen-fixing phylotypes in *Lyngbya sp*. and *Microcoleus* chthonoplastes cyanobacterial mats from Guerrero Negro, Baja California, Mexico. *Applied and Environmental Microbiology*, **70**, 2119–2128.

Pearl, H. W., and Fulton, R. S. I. I. I., 2006. Ecology of harmful cyanobacteria. In Graneli, E., and Turner, J. (eds.), *Ecology of Harmful Marine Algae*. Berlin: Springer, pp. 95–107.

Pearl, H. W., and Otten, T. G., 2013. Harmful cyanobacterial blooms: causes, consequences, and controls. *Microbial Ecology*, **65**, 995–1010.

Pearson, L. A., and Neilan, B. A., 2008. The molecular genetics of cyanobacterial toxicity as a basis for monitoring water quality and public health risk. *Current Opinion Biotechnology*, **19**, 281–288.

Peter, M. V., Cassman, K. E., Cleveland, C., Crews, T., Field, C. B., Grimm, N. B., Howarth, R. B., Marino, R., Martinelli, R., Rastetti, R. B., and Sprent, J. I., 2002. Towards an ecological understanding of biological nitrogen fixation. *Biogeochemistry*, **57**(58), 1–45.

Pope, P. B., and Patel, B. K. C., 2008. Metagenomic analysis of a freshwater toxic cyanobacteria bloom. *FEMS Microbiology Ecology*, **64**, 9–27.

Reid, R. P., Visscher, P. T., Decho, A. W., Stolz, J. F., Bebout, B. M., Dupraz, C., Macintyre, I. G., Paerl, H. W., Pinckney, J. L., Prufert-Bebout, L., Steppe, T. F., and Desmarais, D. J., 2000. The role of microbes in accretion, lamination and early lithification of modern marine stromatolites. *Nature*, **406**, 989–992.

Rinta-Kanto, J. M., Konopko, E. A., and DeBruyn, J. M., 2009. Lake Erie Microcystis: relationship between microcystin production, dynamics of genotypes and environmental parameters in a large lake. *Harmful Algae*, **8**, 665–673.

Schopf, J. W., 2000. The fossil record: tracing the roots of the cyanobacterial lineage. In Whitton, B. A., and Potts, M. (eds.), *The ecology of cyanobacteria*. Dordrecht: Kluwer Academic, pp. 13–35

Smith, V. H., and Bennet, S. J., 1999. Nitrogen: phosphorus supply ratios and phytoplankton community structure in lakes. *Archives fur Hydrobiologie*, **146**, 37–53.

Smith, J. L., Boyer, G. L., and Zimba, P. V., 2008. A review of cyanobacterial odorous and bioactive metabolites: impacts and management alternatives in aquaculture. *Aquaculture*, **280**, 5–20.

Staley, J. T., Bryant, M. P., Pfennig, N., and Holt, J. G., 1989. *Bergey's Manual of Systematic Biology*. Baltimore: Williams & Wilkins, 2648 pp.

Telesh, I. V., 2004. Plankton of the baltic estuarine ecosystems with emphasis on new estuary: a review of present knowledge and research perspectives. *Marine Pollution Bulletin*, **49**, 206–219.

Vincent, W. F., 2000. Cyanobacterial dominance in the polar regions. In Whitton, B. A., and Potts, M. (eds.), *The Ecology of Cyanobacteria, Their Diversity in Time and Space*. Dordrecht: Kluwer, pp. 321–340.

Walsby, A. E., Dubinsky, Z., Kromkamp, J. C., Lehmann, C., and Schanz, F., 2001. The effects of diel changes in photosynthetic coefficients and depth of *Planktothrix rubescens* on the daily integral of photosynthesis in Lake Zurich. *Aquatic Sciences*, **63**, 326–349.

Whitton, B. A., 2012. *The Ecology of Cyanobacteria II: Their Diversity in Time and Space*. Dordrecht: Springer.

Whitton, B. A., and Potts, M., 2000. *The Ecology of Cyanobacteria: Their Diversity in Time and Space*. Berlin: Springer, 0792347358.

Wilhelm, S. W., Farnsley, S. E., Lecleir, G. R., Layton, A. C., Satchwell, M. F., Debruyn, J. M., Boyer, G. L., Zhu, G., and Paerl, H. W., 2011. The relationships between nutrients, cyanobacterial toxins and the microbial community in Taihu (Lake Tai), China. *Harmful Algae*, **10**, 207–215.

Wynn-Williams, D. D., 2000. Cyanobacteria in deserts – life at the limit? In Whitton, B. A., and Potts, M. (eds.), *The Ecology of Cyanobacteria, Their Diversity in Time and Space*. Dordrecht: Kluwer, pp. 341–366.

Zehr, J. P., Mellon, M., Braun, S., Litaker, W., Steppe, T., and Paerl, H. W., 1995. Diversity of heterotrophic nitrogen fixation genes in a marine cyanobacterial mat. *Appl Environ Microbiol*, **61**, 2527–2532.

Cross-references

Algal Blooms
Phytoplankton
Phytoplankton Blooms
Toxic Blooms