B

BARRIER ISLANDS

Definition and occurrence

Barrier islands are elongate, shore-parallel accumulations of unconsolidated sediment (primarily sand), some parts of which are supratidal, that are separated from the mainland by bays, lagoons, or wetland complexes. They are most abundant along the coastlines of the trailing edges of continental plates and of epicontinental seas and lakes (e.g., Caspian and Black Seas). They do not occur on coasts with tidal ranges greater than around 4 m, because their primary mechanism wave action is not focused long enough at a single level during the tidal cycle to form the island and the strong tidal currents associated with such large tides transport the available sand to offshore regions. Barrier islands do occur, primarily as spit forms, on leading edge and glaciated coasts, but they are a minority coastline type in those areas.

Barrier islands are the dominant coastline type along the Atlantic and Gulf coasts of the United States, most of them having been formed within the last 4,000–5,000 years during a near stillstand of sea level.

Barrier islands on depositional coasts

Depositional, coastal plain shorelines typically have barrier islands located between major river deltas and estuaries. Two types of barrier islands may be present, those that consistently migrate landward (retrograding) and those that build seaward (prograding). The island type depends upon the ratio of relative sea-level change to sediment supply. Diminished or low sediment supply and/or rapid sea-level rise promotes the development of retrograding barrier islands and *vice versa* for prograding barrier islands.

Retrograding barrier islands are composed of coalescing washover fans and terraces that are overtopped at high tides, usually several times a year (Figure B1). Stratigraphically, a relatively thin wedge of sand and shell of the washover terrace overlies backbarrier sediments, which are typically composed of muddy sediments deposited in the lagoons or wetlands behind the islands. These islands are impractical sites for human development because of their constant landward migration.

Prograding barrier islands are composed of multiple beach ridges. Many have a drumstick configuration because of selective sand accumulation on the updrift end of the island (the direction sand comes from). The most notable changes on prograding islands occur when the adjacent tidal inlets migrate or when the inlets expand dramatically during hurricane storm surges.

Stratigraphically, prograding barrier islands are composed of sand 8–10 m thick (thickness depends upon wave size), which has prograded over offshore muds (Figure B1). When human development occurs on these islands, buildings are usually secure from all but the most extreme hurricanes, if they have been set back an adequate distance from the front-line dunes and tidal inlets. But that security will vanish if a major rise in sea level occurs (Hayes, 1996).

The morphology of prograding barrier islands is controlled by a combination of wave and tidal forces (Hayes, 1979; Davis, 1994a). Under *wave-dominated* conditions, which most commonly occur in microtidal areas (tidal range ≤2 m; Davies, 1964), the barriers are long, typically tens of kilometers, with widely spaced inlets that have large flood-tidal deltas and small ebb-tidal deltas. Washover fans are common and the islands are flanked on the landward side by bays and/or lagoons (Figure B2). Barrier islands along *mixed-energy* coasts (Hayes, 1979), which typically occur in mesotidal areas (tidal range = 2–4 m), are stunted and short (usually <10 km) with abundant tidal inlets that contain large ebb-tidal deltas and small to nonexistent flood-tidal deltas. These islands are flanked on the landward side by complex tidal channels, tidal flats, and wetlands (Figure B3). Barrier islands do not occur on *tide-dominated* coasts.

Origin of barrier islands

Barrier islands are thought to most commonly originate in one of the three possible ways: (1) by spit elongation (Fisher, 1967); (2) retreating transgressive barrier islands (Swift, 1975); and (3) a process termed *transgressive–regressive interfluve hypothesis* by Hayes (1994). In many parts of the world, it is clear that the source of sand for the existing barrier islands originated from an updrift headland, and as a spit extended away from the headland it was cut into segments during storms, creating tidal inlets that eventually attained permanence. Swift (1975) stated that barrier islands originated at a lower stand of sea level and migrated over the drowning coastal plain as sea level rose during the early Holocene. The "primary barrier" of Pierce and Colquhoun (1970), the nucleus for many prograding barrier islands, no doubt originated in this way. According to the transgressive–regressive interfluve hypothesis, as sea level rose, the transgressive barrier eventually perched on the topographic high of an interfluve located between major alluvial valleys that were carved during the last Pleistocene lowstand (Figure B4). Once sea level stabilized around 4500 years BP, prograding barrier islands developed in areas with adequate sediment supply. As the island grew, beach ridges prograded away from the interfluve, with major tidal inlets forming at both ends of the island over the antecedent lowstand valleys (Moslow, 1980). This mechanism explains the origin of many of the major prograding barrier islands along the coast of the southeastern United States.

Barrier islands on leading edge and glaciated coasts

The somewhat rare barrier islands on the leading edge, west coast of the United States are, for the most part, relatively short spits that have built

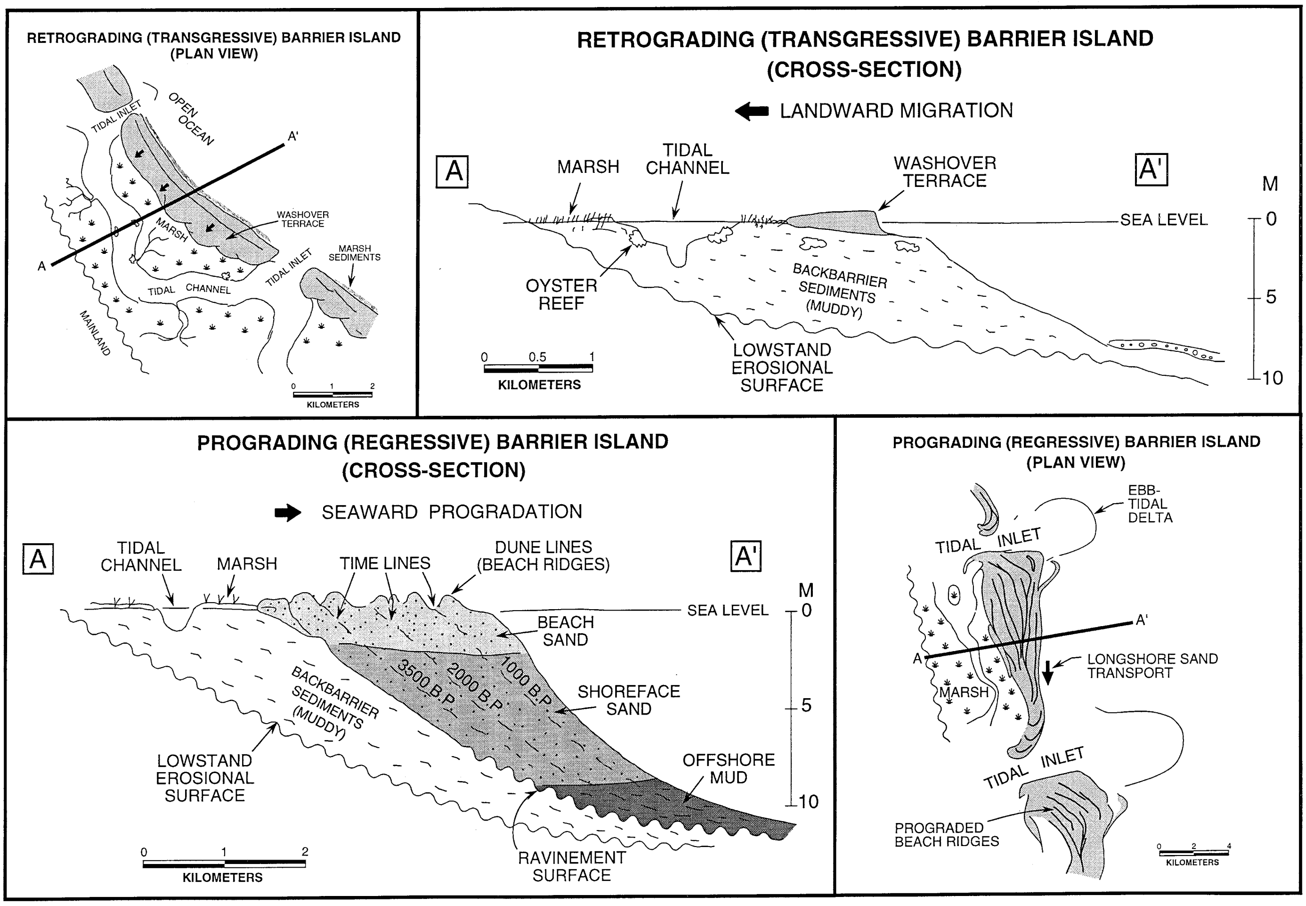

Figure B1 Morphology and stratigraphy of prograding and retrograding barrier islands.

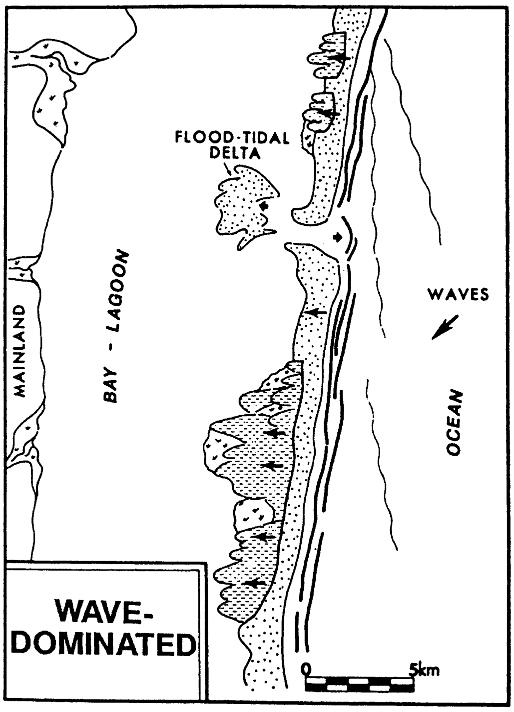

Figure B2 Typical morphology of a prograding, wave-dominated barrier-island shoreline.

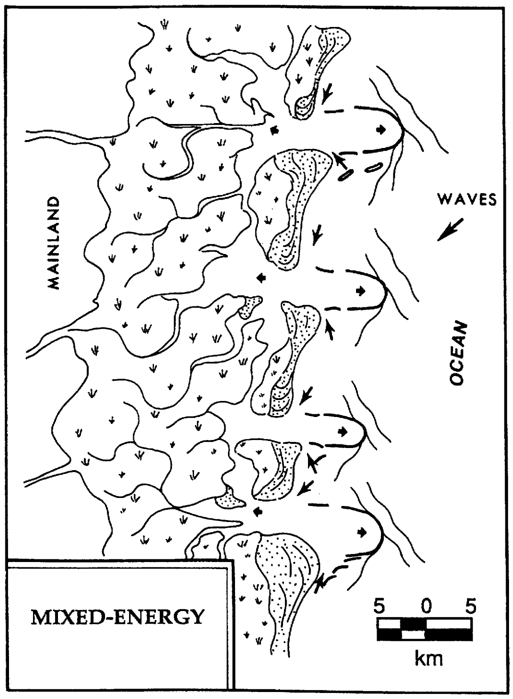

Figure B3 Typical morphology of a prograding, mixed-energy barrier-island shoreline.

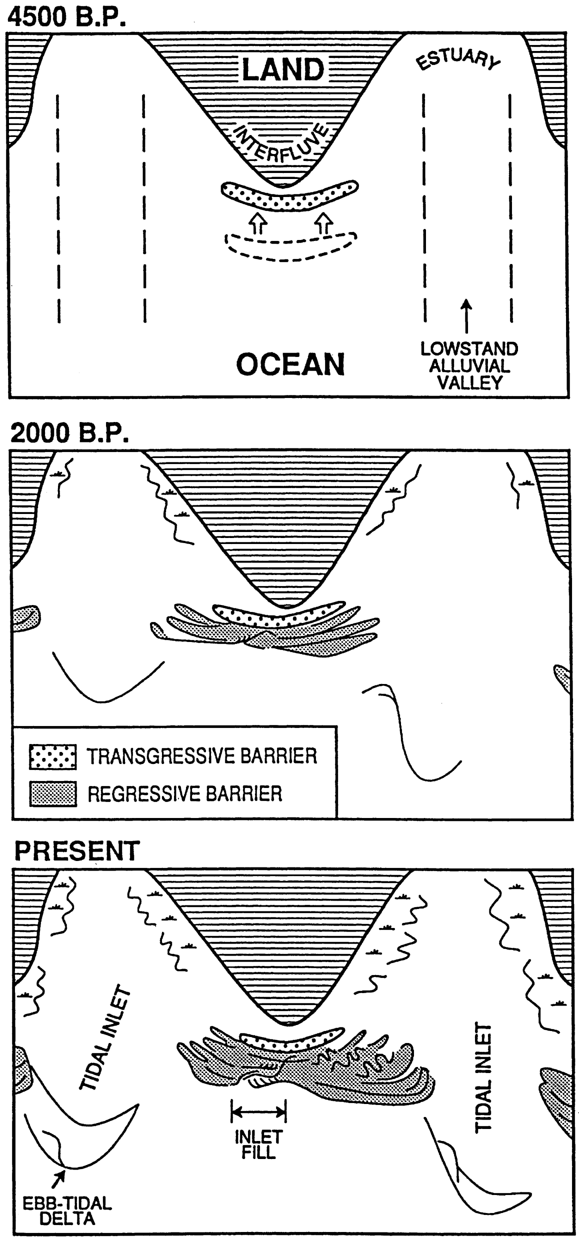

Figure B4 Model for the origin of a prograding, mixed-energy barrier-island along the southeastern US coastline–transgressive–regressive interfluve hypothesis. Based on Pierce and Colquhoun (1970) and Moslow (1980).

away from rocky headlands or river mouths. River discharge controls the shape of the spit during high discharge, and waves control it during low discharge (Dingler and Clifton, 1994; Smith, *et al.*, 1999).

Although occurring in a wide variety of types, which were classified into six major categories by Fitzgerald and Van Heteren (1999), barrier islands make up <25% of the glaciated coastline of New England. Most of these islands originate as spits, which are transformed into a variety of forms by tidal and wave action. Antecedent topography and geology also play important roles in shaping the morphology of the barrier islands along this complex coastline.

Further reading

For further information on the subject of barrier islands the reader is referred to: Schwartz (1973), Siringan and Anderson (1993), Davis (1994b), Moslow and Heron (1994), and Sexton and Hayes (1996).

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Bibliography

Davies, J.L., 1964. A morphogenic approach to world shorelines. *Zeitschrift für Geomorphologie*, **8**: 27–42.
Davis, R.A., Jr., 1994a. Barrier Island systems—a geologic overview. In Davis, R.A., Jr. (ed.), *Geology of Holocene Barrier Island Systems*. New York: Springer-Verlag, pp. 1–46.
Davis, R.A., Jr. (ed.), 1994b. *Geology of Holocene Barrier Island Systems*. New York: Spring-Verlag.
Dingler, J.R., and Clifton, H.E., 1994. Barrier systems of California, Oregon, and Washington. In Davis, R.A., Jr. (ed.), *Geology of Holocene Barrier Island Systems*. New York: Springer-Verlag, pp. 115–205.
Fitzgerald, D.M., and Van Heteren, S., 1999. Classification of Paraglacial Barrier Systems: coastal New England, USA. *Sedimentology*, **46**: 1083–1108.
Fisher, J.J., 1967. Origin of Barrier Island Chain Shoreline. *Geological Society of America Special Paper 115*, pp. 66–67.
Hayes, M.O., 1979. Barrier Island morphology as a function of tidal and wave regime. In Leatherman, S.P. (ed.), *Barrier Islands from the Gulf of St. Lawrence to the Gulf of Mexico*, New York: Academic Press, pp. 1–27.
Hayes, M.O., 1994. The Georgia Bight Barrier System. In Davis, R.A., Jr. (ed.), *Geology of Holocene Barrier Island Systems*. New York: Springer-Verlag, pp. 233–304.
Hayes, M.O., 1996. Barrier Island Basics. *Geotimes*, **41**: 17.
Leatherman, S.P. (ed.), 1979. *Barrier Islands from the Gulf of St. Lawrence to the Gulf of Mexico*. New York: Academic Press.
Moslow, T.F., 1980. *Stratigraphy of Mesotidal Barrier Islands*. PhD Thesis, Depart Geology, University of South Carolina.
Moslow, T.F., and Heron, S.D., Jr., 1994. The outer banks of North Carolina. In Davis, R.A., Jr. (ed.), *Geology of Holocene Barrier Island Systems*. New York: Springer-Verlag, pp. 47–74.
Pierce, J.W., and Colquhoun, D.J., 1970. Holocene evolution of a portion of the North Carolina Coast. *Geological Society of America. Bulletin*, **81**: 3697–3714.
Schwartz, M.L. (ed.), 1973. *Barrier Islands*. Stroudsburg: Dowden, Hutchinson & Ross.
Sexton, W.J., and Hayes, M.O., 1996. Holocene deposits of reservoir-quality sand along the central south Carolina coastline. *American Association of Petroleum Geologists Bulletin*, **80**: 831–855.
Siringan, F.P., and Anderson, J.B., 1993. Seismic facies, architecture, and evolution of the Bolivar roads tidal/delta complex, East Texas Gulf Coast. *Journal of Sedimentary Petrology*, **63**: 794–808.
Smith, D.G., Meyers, R.A., and Jol, H.M., 1999. Sedimentology of an Upper-Mesotidal (3.7 m) Holocene Barrier, Willapa Bay, SW Washington, U.S.A.. *Journal of Sedimentary Research*, **69**: 1290–1296.
Swift, D.J.P., 1975. Barrier-island genesis; evidence from the Central Atlantic Shelf, Eastern USA. *Sedimentary Geology*, **14**: 1–43.

Cross-references

Barrier
Changing Sea Levels
Spits
Tidal Inlets
Tide-Dominated Coasts
Wave-Dominated Coasts

BARRIER

A barrier (coastal barrier) is an elongated coastal ridge of deposited sediment built-up by wave action above high tide level offshore or across the mouths of inlets or embayments. It is usually backed by a lagoon or swamp, which separates it from the mainland or from earlier barriers. A barrier, thus defined, is distinct from a bar, which is submerged at least at high tide (Shepard, 1952), and from reefs of biogenic origin (see Coral Reefs). Most barriers consist of sand, but some contain gravel as well as sand, and others consist entirely of gravel (shingle): see Gravel Barriers. Chesil Beach, on the south coast of England, is a well-known shingle barrier, and similar features are seen on the southeast coast of Iceland, and on the east and south coasts of South Island, New Zealand. Commonly the gravel has been derived from glacial or periglacial deposits, as on the north coast of Alaska and the southern shores of the Baltic Sea.

Barriers are said to occupy about 13% of the world's coastline (Leontyev and Nikiforov, 1965). They are most extensive on the Gulf and Atlantic coasts of North America and the ocean coasts of Australia, South Africa, and eastern South America, but they also occur on a smaller scale elsewhere, notably in Sri Lanka and New Zealand. Some barriers are transgressive, migrating landward across lagoon and swamp deposits; others remain in position, or are widened seaward by progradation, usually indicated by successively formed beach or dune ridges. Transgressive barriers occur where sediment is washed or blown over into backing lagoons or swamps, particularly during storms. Low sectors of a barrier through which storm waves or surges flow are called swashways, and sediment swept over a barrier through these is deposited as a washover fan on the inner shore.

On some coasts there are multiple barriers, the inner and older barriers being of Pleistocene age bordered (and overlain) by outer and younger barriers of Holocene age. Thus the Gippsland Lakes in southeastern Australia are enclosed by an Inner Barrier and an Outer Barrier (the Ninety Mile Beach), separated by lagoons and swamps, and with relics of an earlier, Prior Barrier that predates the enclosure of the existing Lakes (Bird, 1973). In this case, barriers have developed seaward of earlier marine coastlines, but evidence of preceding exposure to the open sea is not always present, particularly in lagoons where the enclosing barriers have been transgressing landward during a phase of rising sea level, as on the Siberian coast (Zenkovich, 1967).

The term barrier beach indicates a single, narrow elongated ridge (usually less than 200 m wide) built parallel to the coast, without surmounting dunes. A barrier island is bordered by transverse gaps (tidal inlets, lagoon entrances, river outlets), which may be migratory and subject to closure; it usually bears beach ridges, dunes and associated swamps, and minor lagoons, and may incorporate recurves (Schwartz, 1973). Examples include Scolt Head Island on the east coast of England, and several along the Atlantic coast of the United States. A barrier with many interruptions becomes a barrier island chain. A barrier attached to the mainland at one end can be termed a barrier spit, as on the coast north of the Columbia River in Washington State (where Long Beach is a barrier spit partly enclosing Willapa Bay); one built across the mouth of an embayment a bay (baymouth) barrier. There are also mid-bay barriers and bay-head barriers, defined by their position.

In general, barriers are found on coasts where the tide range is small (as on the southern Baltic coast), and become chains of barrier islands where currents produced by larger tides maintain transverse gaps (as on the Danish, German, and Dutch North Sea coasts).

Barriers can form in various ways (Schwartz, 1971), multiple causality being related to the nature and supply of sediment, the transverse profile of the coast, tide range, wave conditions, and relative sea-level change. Some barriers may have formed by the emergence of nearshore swash bars as sea-level fell (e.g., Knotten, on the Danish island of Laesø), but many developed during and after the Late Quaternary marine transgression by submergence of pre-existing sand ridges and the shoreward drifting of sea floor sediment. Of these, some are still transgressive (as on parts of the Atlantic coast of the United States) while others have become anchored, and widened by progradation (as on parts of the southeast Australian coast). Growth and landward transgression of barriers has been demonstrated in recent years on parts of the coast of the Caspian Sea, which has risen about 2 m since 1977 (Kaplin and Selivanov, 1995).

Barrier spits may show features indicative of longshore growth, such as former recurved terminations on the landward side (as on Orfordness in England or the Langue de Barbarie in West Africa), but others have been built and widened by wave-deposited sediment from the sea floor (as on Clatsop Spit in Oregon), and many result from combinations of onshore and longshore sediment drifting.

The shaping of barriers can be traced with reference to patterns of beach and dune ridges indicating stages in their growth, and from their stratigraphy, which may indicate phases of upward growth, landward movement, and seaward progradation, as in Van Straaten's (1965) classic study of barriers on the Netherlands coast and Thom's (1984) study of sand barriers in eastern Australia. Barriers of unconsolidated sand are readily re-shaped by wave and wind action, but where barrier sediments have become lithified (e.g., the Pleistocene dune calcarenites in Australia and elsewhere) they are more durable, and may show cliffing (as in the inner barriers of the Coorong in South Australia). Some barriers incorporate segments of pre-existing terrain, such as the glacial moraines on Long Island in New York, Walney Island in northwest England and Sylt on the German North Sea coast.

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Bibliography

Bird, E.C.F., 1973. Australian coastal barriers. In Schwartz, M.L. (ed.), *Barrier Islands*. Stroudsburg: Dowden, Hutchinson & Ross, pp. 410–426.

Kaplin, P., and Selivanov, O.A., 1995. Recent coastal evolution of the Caspian Sea as a model for coastal responses to the possible acceleration of global sea level rise. *Marine Geology*, **124**: 161–175.

Leontyev, O.K., and Nikiforov, L.G., 1965. Reasons for the world-wide occurrence of barrier beaches. *Oceanology*, **5**: 61–67.

Schwartz, M.L., 1971. The multiple causality of barrier islands. *Journal of Geology*, **79**: 91–94.

Schwartz, M.L. (ed.), 1973. *Barrier Islands*. Stroudsburg: Dowden, Hutchinson & Ross.

Shepard, F.P., 1952. Revised nomenclature for coastal depositional features. *American Association of Petroleum Geologists*, **36**: 1902–1912.

Thom, B.G., 1984. Sand barriers of eastern Australia: Gippsland—a case study. In Thom, B.G. (ed.), *Coastal Geomorphology in Australia*. Sydney: Academic Press, pp. 233–261.

Van Straaten, L.M.J.U., 1965. Coastal barrier deposits in south and north Holland, *Meded. Geol. Stichting*, **17**: 41–87.

Zenkovich, V.P., 1967. *Processes of Coastal Development* (trans. O.G. Fry, ed. J.A. Steers). Edinburgh: Oliver and Boyd.

Cross-references

Barrier Islands
Bars
Coral Reefs
Drift and Swash Alignments
Gravel Barriers
Spits

BARRIERS, GRAVEL—See GRAVEL BARRIERS

BARS

Sedimentary ridges, both symmetric and asymmetric, and generally larger than bedforms that characterize the upper shoreface of coastal zones dominated by waves are called *wave-formed bars*. They were recognized as early as 1845 on the marine coasts of Europe (Elie de Beaumont), by 1851 in the Great Lakes of North America (Desor), and subsequently on marine and lacustrine coasts worldwide (see Schwartz, 1982, pp. 135–139). However, confusion still surrounds this term because of its use for ridges with a wide range of size, morphology, location, and orientation relative to the shoreline. Also, the term *bar* has been used in a variety of environments, from subaerial to those dominated by tidal currents or river currents. Furthermore, the present understanding of the origin(s) and dynamics of *wave-formed bars* is still incomplete.

Shepard (1950) called shore-parallel ridges and troughs *longshore bars* and *troughs*, equating them with the terms *ball* and *low* of Evans (1940), and associated them with plunging breakers. He emphasized the *seasonality* of such bars on the west coast of the United States, and subsequently terms such as *winter* and *summer, storm* and *normal*, and *storm* and *swell* have been applied to denote the presence or absence of bars. Although a correlation between profile form and storm waves or season may exist in some localities (e.g., Inman *et al.*, 1993), it is not universal. Both *barred* and *non-barred* profiles occur at times in some areas, while in others only one profile type may persist throughout the year. There is usually a distinct *relaxation time* between the forcing conditions and bar adjustment; thus in the short-term, bars are generally in a transient state. In the longer term (years to decades), wave-formed bars represent the *equilibrium morphology* for many coastal environments.

Bar morphology

Wave-formed bars are most clearly identified as near-symmetrical or asymmetrical undulations in the upper shoreface profile (Figure B5). They occur intertidally and subtidally, and may range in number from one to more than thirty, this number often varying through time. Short and Aagaard (1993) introduced a bar parameter, $B^* = x_s/gT^2 \tan \beta$, to

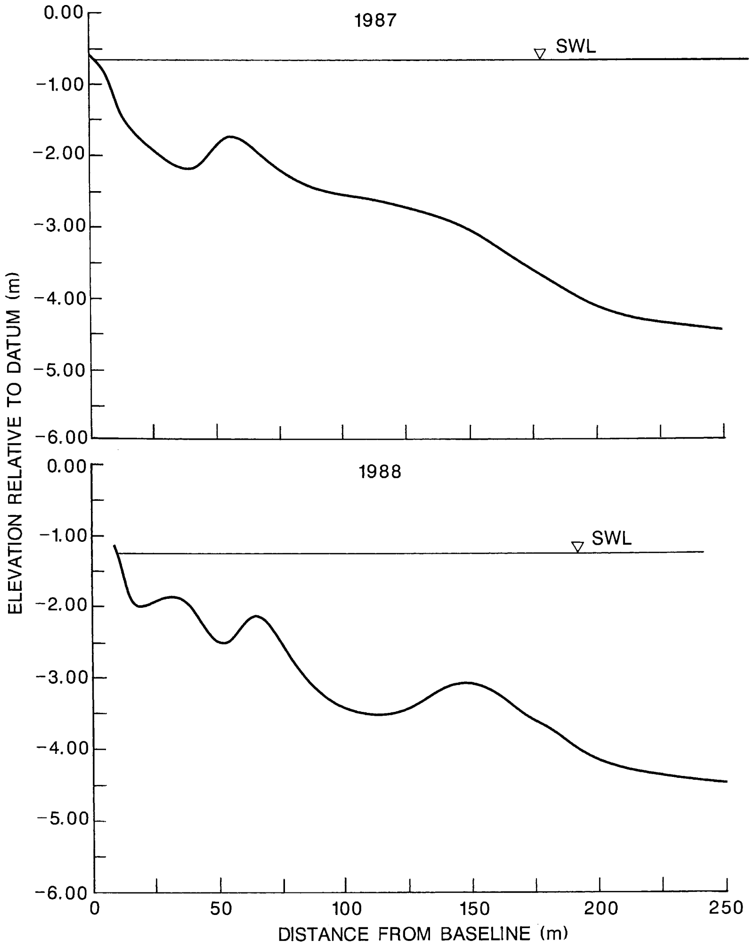

Figure B5 Typical barred profiles from a sandy nearshore environment in the Canadian Great Lakes. The profiles were surveyed in successive years along the same transect. Note the differing number and position of the wave-formed bars at the same location, even though the mean beach slope is the same.

identify the number of bars on a linear sloping shoreface (tan β) terminating at a constant depth at a distance offshore, x_s. When $B^* < 20$, the profile is non-barred, for $B^* = 20$–50, 1 bar occurs; for $B^* = 50$–100, 2 bars, for $B^* = 100$–400, 3 bars; and for $B^* > 400$ there are 4 bars. Crest heights above the adjacent trough can range from less than a decimeter (Carter, 1978) to more than 4.75 m (Greenwood and Mittler, 1979). In plan view, they form continuous or compartmentalized, linear, sinuous, or crescentic patterns, and range from shore-parallel to shore-normal in orientation, often producing periodic or *rhythmic* topography both alongshore and cross-shore. The morphometry of bars has been studied by Greenwood and Davidson-Arnott (1975), Hands (1976), and Reussink *et al.* (2000) in order to define the equilibrium form and dynamics induced by a specific set of environmental constraints.

Bar classification

A *universal* classification of *wave-formed bars* does not yet exist, and indeed it may never be possible to define perfectly mutually exclusive classes. A simple descriptive classification based on morphology and the associated environmental constraints is illustrated in Table B1 (Greenwood and Davidson-Arnott, 1979). The group names are those in common use, and the definitive paper describing each type is cited. Other classifications are based on the concept that bars are part of a temporal sequence of beach profile evolution and that they are scaled to that of the controlling wave process. The morphological sequence is controlled by incident wave energy (high and low frequency) and was identified either through aerial photographs or more recently through video-imagery (e.g., Short, 1979; Lippmann and Holman, 1990; see Figure B6). Many coastal environments do not experience such sequential behavior.

Ridge and runnel topography (*Type I*) is found on low-angle, macro- to meso-tidal foreshore slopes dominated by surf action and foreshore drainage during the tidal cycle. Although low in amplitude these bars are usually stable in form and position or migrate only slowly. In contrast, the *cusp-* or *bar-type sand waves* (*Type II*) are extremely dynamic, often destroyed during storms and regenerated as the storm wanes and smaller amplitude, longer period waves propagate shoreward. These bars result from surf bores and swash action near the toe of the swash slope (an alternative name is *swash* bar). Furthermore, they may develop from *Type VI* bars as they migrate relatively rapidly both alongshore and onshore, and in the latter case may *weld* to the foreshore (Davis *et al.*, 1972; Aagaard *et al.*, 1998). Note that there is confusion with respect to the term *ridge and runnel* as used in northwest Europe and North America (Orford and Wright, 1978; Orme and Orme, 1988). Here, the term *ridge and runnel* is restricted to its initial definition by King and Williams (1949); the forms described by Hayes and subsequent workers (Hayes and Boothroyd, 1969) are classified here as *Type II* bars.

Type III multiple parallel bars (e.g., Nilsson, 1973; Exon, 1975) and *Type IV transverse bars* (e.g., Niedoroda and Tanner, 1970; Carter, 1978; Dolan and Dean, 1985) tend to be limited to low-angle shorefaces and small to moderate wave heights, coupled with limited water level shifts. However, they have been identified on more energetic shorelines (Konicki and Holman, 2000).

The number of bars increases with decreasing beach slope (Davidson-Arnott, 1988). The height and spacing of the multiple bars increases in the offshore direction, and bar form is near symmetrical in contrast to the *Type II* group. Transverse bars run normal or obliquely to the shoreline and can range in length from 3 m up to 4 km, with heights from less than 0.05 m up to 2 m and alongshore spacing of the

Table B1 Bar morphologies and environmental constraints

			Morphology					Environmental constraints			
Class	Name	Definitive description	Size (m)	Planform	Profile	Number seaward	Location	Wave energy	Breaking processes	Tidal range	Slope
Type I	Ridge and runnel	King and Williams (1949)	$h \sim 0.2$–1.5 $l \sim 10^3$	Straight, shore parallel	Asymmetric landward	1–4	Intertidal	L–M	Surf-swash, beach drainage	Ma–Me	0.007 0.024
Type II	Cusp- or bar-type sand wave	Sonu (1973)	$h \sim 0.2$–1.5 $l \sim 10^2$	Straight to spit shaped, shore parallel	Asymmetric landward	1–2	Intertidal and low tide terrace	M	Breakers, surf-swash	Me–Mi	>0.01
Type III	Multiple parallel	Zenkovitch (1967)	$h \sim 0.2$–0.75 $l \sim 10^3$	Straight to sinuous, shore parellel	Near symmetric	4–30 or more	Nearshore & intertidal	L–M	Spilling	Mi–Me	<0.01
Type IV	Transverse	Niederoda and Tanner (1970)	$h \sim 0.2$–0.75 $l \sim 10^2$	Straight, shore normal	Symmetric and asymmetric landward	1	Nearshore and intertidal	L	Surf-swash,	Mi spilling	<0.0045
Type V	Nearshore I	Shepard (1950)	$h \sim 0.15$–1.0 $l \sim 10^2$ (?)	Straight, shore parallel	Asymmetric landward	1–2	Nearshore	M–H	Plunging	Mi–Me	<0.1
Type VI	Nearshore II	Evans (1940)	$h \sim 0.25$–3.0 $l \sim 10^3$	Straight, sinuous to crescentic, shore parallel	Asymmetric landward	1–4	Nearshore	M	Spilling	Mi	<0.01

Note: h = bar height; l = bar length; Ma = macro; Me = meso; Mi = micro; L = low; M = moderate; H = high.
Source: Modified from Greenwood and Davidson-Arnott, 1979.

WRIGHT & SHORT (1984)		**LIPPMANN & HOLMAN (1990)**
Beach State 6 *Non-barred, dissipative beach*	==========	**Type H** *Non-barred dissipative; Infragravity wave scaled surf zone*
		Type G *2-D, infragravity wave scaled bar*
Beach State 5 *Longshore Bar & Trough*		**Type F** *3-D, non-rhythmic, infragravity wave scaled bar*
Beach State 4 *Rhythmic Bar & Beach*	==========	**Type E** *Offshore rhythmic, infragravity wave scaled bar*
		Type D *Attached rhythmic, infragravity wave scaled bar*
Beach State 3 *Transverse Bar & Rip*		**Type C** *Attached non-rhythmic, infragravity wave scaled bar*
Beach State 2 *Swash Bar-Shore Terrace*	==========	**Type B** *Incident wave scaled, 2-D bar*
Beach State 1 *Non-barred reflective beach*	==========	**Type A** *Non-barred, reflective, incident wave scale*

Figure B6 Classification and scaling of sequential upper shoreface morphologies. The equivalence between the contrasting sequences of Wright and Short and Lippmann and Holman is indicated (modified after Lippmann and Holman, 1990).

order of 10^0–10^2 m (Carter, 1978; Gelfenbaum and Brooks, 1997). The larger forms may migrate alongshore at rates up to 8 m a^{-1}. Usually, transverse bars are anchored to the shoreline (indeed they appear as an extension of a shoreline protuberance), but Konicki and Holman (2000) recorded the unusual case of transverse bars running offshore from a *Type VI* bar.

The division of *nearshore bars* into two groups is based upon size, stability, and the controlling waveform. *Type V bars* are associated with large plunging breakers, which produce narrow, low amplitude ridges on relatively steep slopes: they lack a well-defined asymmetry and are essentially unstable modifications of non-barred nearshore profiles. *Type VI bars*, in contrast, are relatively large configurations formed seaward of the low water level. Where there is more than one bar, the distance offshore, depth-of-water over the crest, and bar height all usually increase offshore in a regular manner, although in some cases the height decreases after some offshore distance (Lippmann *et al.*, 1993; Ruessink and Kroon, 1994). The volume of sediment in each bar form usually increases consistently offshore. *Type VI* bars may be three-dimensional, sinuous-to-crescentic, and the alongshore length scales may range from 10^2 to 10^3 m (Greenwood and Davidson-Arnott, 1975; Bowman and Goldsmith, 1983). Where more than one bar is rhythmic, the alongshore wavelength decreases shoreward.

Bar genesis

The boundary conditions necessary for bar formation depend upon the longer term evolution of the coast, which dictates the nature of the bed materials (grain mineralogy, size, sorting, etc.), the bathymetric setting (slope, exposure, etc.), and the geographic location (wave climate, tidal regime, etc.). Local forcing conditions for bars have been studied both theoretically and empirically, and by experiments in the laboratory and the field (see van Rijn 1998). In general, barred profiles are associated with large values of both wave steepness and wave height-to-grain size ratios, and are associated with the final stages of shoaling and dissipation of wave energy through breaking, and the complex hydrodynamics, which accompany these processes (Wright *et al.*, 1979). Furthermore, the size of wave-formed bars induces a very strong feedback to the shoaling and breaking process. Although cause and effect are far from clear, it is evident that equilibrium bar profiles can exist only where the time-averaged sediment transport (suspended and bedload) is zero everywhere on that profile.

A large number of specific hypotheses have been proposed for bar formation over the last 50 years and all involve mechanisms for convergence of sediment transport; these hypotheses were primarily related to *Type V and VI* bars and fall into three major groups:

(1) *break point hypotheses* relate bars directly to wave breaking and result from: (i) a seaward transport of sediment entrained by roller or helical vortices under plunging or spilling breakers, respectively (Miller, 1976; Zhang, 1994; Figure B7); (ii) convergence of sediment at the breakpoint through onshore transport associated with increasing asymmetry and skewness of the high-frequency incident waves and offshore transport through set up induced undertow (Dally and Dean, 1984; Dally, 1987; Thornton *et al.*, 1996). However, Sallenger and Howd (1989) concluded that bars are not necessarily coupled to the breakpoint, but can grow and migrate, while within the inner surf zone, landward to the point of initial breaking.

(2) *infragravity wave hypotheses* propose that low frequency waves generated within the surf zone (surf beat) or offshore and reflected produce a convergent pattern of drift velocities, which interact with the large incident short wave oscillatory velocities to induce a range of bar forms from two- to three-dimensional crescentic forms (e.g., Bowen and Inman, 1971; Short 1975; Bowen 1980; Holman and Bowen, 1982; Bowen and Huntley, 1984). These waves can be standing or progressive and can be produced in a number of different ways as a result of energy dissipation during breaking and are frequently related to amplitude modulation of the incident wave field (*groupiness*; Roelvink and Broker, 1993; Reussink, 1998). Alternating scour and deposition by mass

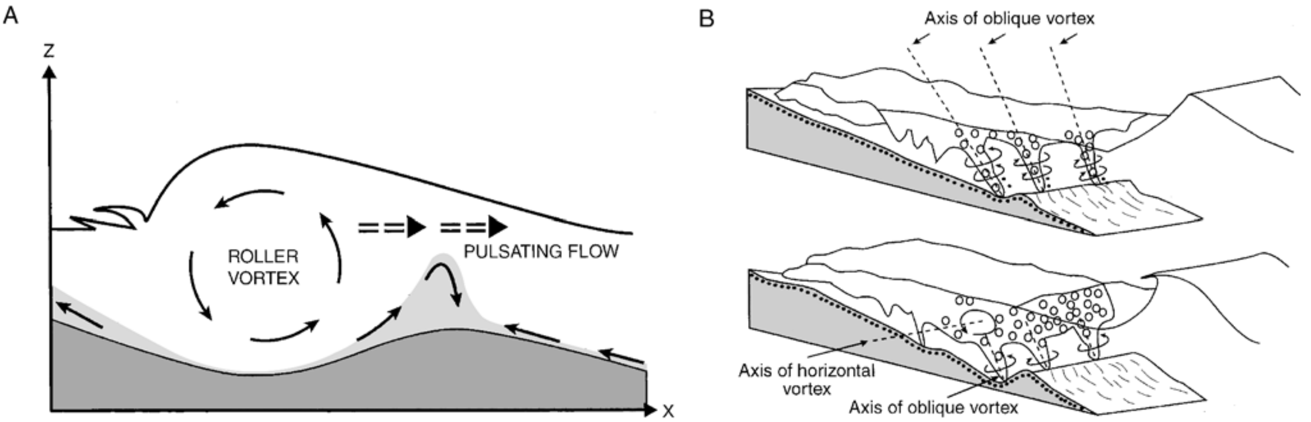

Figure B7 Bar formation by breaking waves: (A) trough scouring by a roller vortex under plunging breakers and offshore sediment transport converging with sediment driven onshore by shoaling waves (modified after Miller, 1976); (B) trough scouring by oblique vortices generated under spilling breakers (modified after Zhang, 1994).

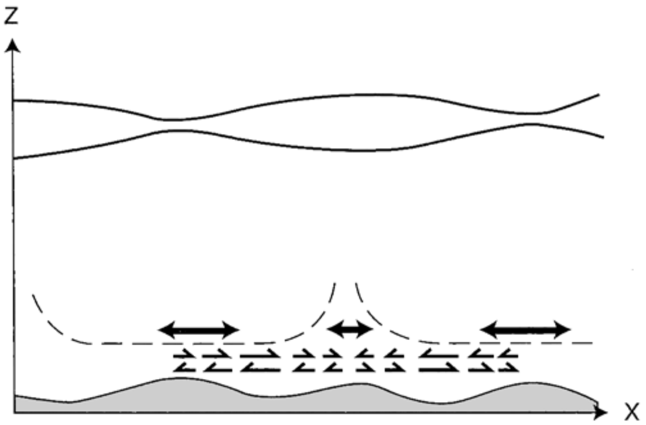

Figure B8 Bar formation as a result of mass transport in the boundary layer of a strongly reflected incident wave. The surface wave envelope is shown as well as the circulation within the bottom boundary layer. Bed load will converge at nodes of the surface elevation and suspended load at antinodes. Note: the boundary layer flow is indicated by single-headed arrows; the mean flow is indicated by double-headed arrows.

transport velocities in the bottom boundary layer generated by standing waves, resulting from the interaction of reflected and incident waves, was shown to occur in the laboratory by Carter *et al.* (1973; Figure B8). The boundary layer was actually segregated. At the bed, drift velocities converge at nodes, while at some distance above the drift velocities converge at antinodes. Under large waves when bars are most active, suspension transport is dominant and therefore sediment will converge and bars will form at the antinodal position of standing waves (e.g., Bowen, 1980). Reflection of waves in the infragravity range was clearly demonstrated by Suhayda (1974) and shown to relate to bar forms by Short (1975) and Katoh (1984). Sediment moves to null positions in the drift velocity field of low-frequency standing (Figure B9(A)) or progressive edge waves (Figure B9(B)), which are periodic both alongshore and offshore. Recent field measurements have clearly shown the importance of group-bound long waves to suspended sediment transport in barred surf zones (e.g., Osborne and Greenwood, 1992), but isolating the drift velocities associated with these secondary waves is difficult. This second-order drift velocity hypothesis requires one dominant wave frequency, which is not common (see Bauer and Greenwood, 1990 for an exception). However, there are a number of suggestions to overcome this inadequacy of the edge wave hypothesis. Aagaard (1990) has argued for the excitation of cutoff mode edge waves (limited by the beach slope) and selection of the dominant mode as that mode which is closest to the wave group period. A phase coupling between the primary orbital motion of a partially standing long wave and groupy short waves was also proposed by O'Hare (1994) to avoid this requirement of narrow bandedness in the infragravity spectrum. Other mechanisms producing a limited number of edge wave frequencies and modes are topographic control (Kirby *et al.*, 1981; Bryan and Bowen, 1996) and interaction of edge waves and the longshore current (Howd *et al.*, 1992). O'Hare and Huntley (1994) propose a leaky wave origin for an inner surf zone bar, which is relatively insensitive to the group period, incident wave height, and the width of the infragravity spectrum.

(3) *self-organization hypotheses* propose that processes associated with the complex, nonlinear feedback between the sand bed and the hydrodynamics give rise to a range of topographic forms. For example, alongshore and offshore sediment movement was proposed under meandering or cellular nearshore circulations produced by (i) instability of longshore flows (Figure B10; Barcilon and Lau, 1973; Hino, 1974; Falques, 1991; Damgaard Christiansen *et al.*, 1994); (ii) coupling between morphodynamic instability and mean flows (Deigaard *et al.*, 1999; Vittorio *et al.*, 1999; Falques *et al.*, 2000); and (iii) Bragg scattering from periodic topography (Heathershaw and Davies, 1985; O'Hare and Davies, 1993; Rey *et al.*, 1995; Yu and Mei, 2000). These mechanisms cannot produce bars directly, but require some initial perturbation of the profile. However, it has been shown that some bar characteristics are not well predicted by these models (e.g., the cross-shore/ alongshore spacing—see Konicki and Holman, 2000). The nonlinear action between shoaling waves and the bed (Boczar-Karakiewicz and Davidson-Arnott, 1987) was also proposed as a mechanism for generating periodic patterns of sediment transport which matched the spacing and general shape of multi-barred shorelines.

The horizontal roller vortex mechanism is most applicable to single *Type V* bars of the US west coast, and justifies the early correlation of bar formation with wave steepness. Multibarred profiles reflect either multiple breakpoints (Dally, 1980; Davidson-Arnott, 1981) or bar formation by distinct differences in wave energy; for example, an outer bar may be produced under storm waves and an inner bar by less energetic conditions (King and Williams, 1949). Water level shifts and coincident shifts in breaker location could also produce a multiple barred system. Oblique, helical vortices were produced under spilling rather than plunging breakers in the laboratory and could account for both single and multiple barred profiles (Zhang, 1994). However, the mass transport velocities under reflected standing waves would perhaps best explain the formation of *Type III* multiple parallel bars; simple reflection of the incident waves could not be the cause, since the length scales of the bars would require much longer periods. The theoretical convergence patterns of drift velocities under standing edge waves provide strong support for their role in forming crescentic *Type VI* bars. Progressive edge waves may be responsible for linear bars of the same group (Huntley, 1980). Further, the edge wave periods necessary to produce the length scales found in nature is of the same order as the well-known *surf beat*. However, the generation of these trapped modes of oscillation still remains ill-defined, even though field observations of low-frequency peaks in the nearshore energy spectrum have been made on barred coasts and related to the presence of edge waves (Huntley, 1980; Bauer and Greenwood, 1990).

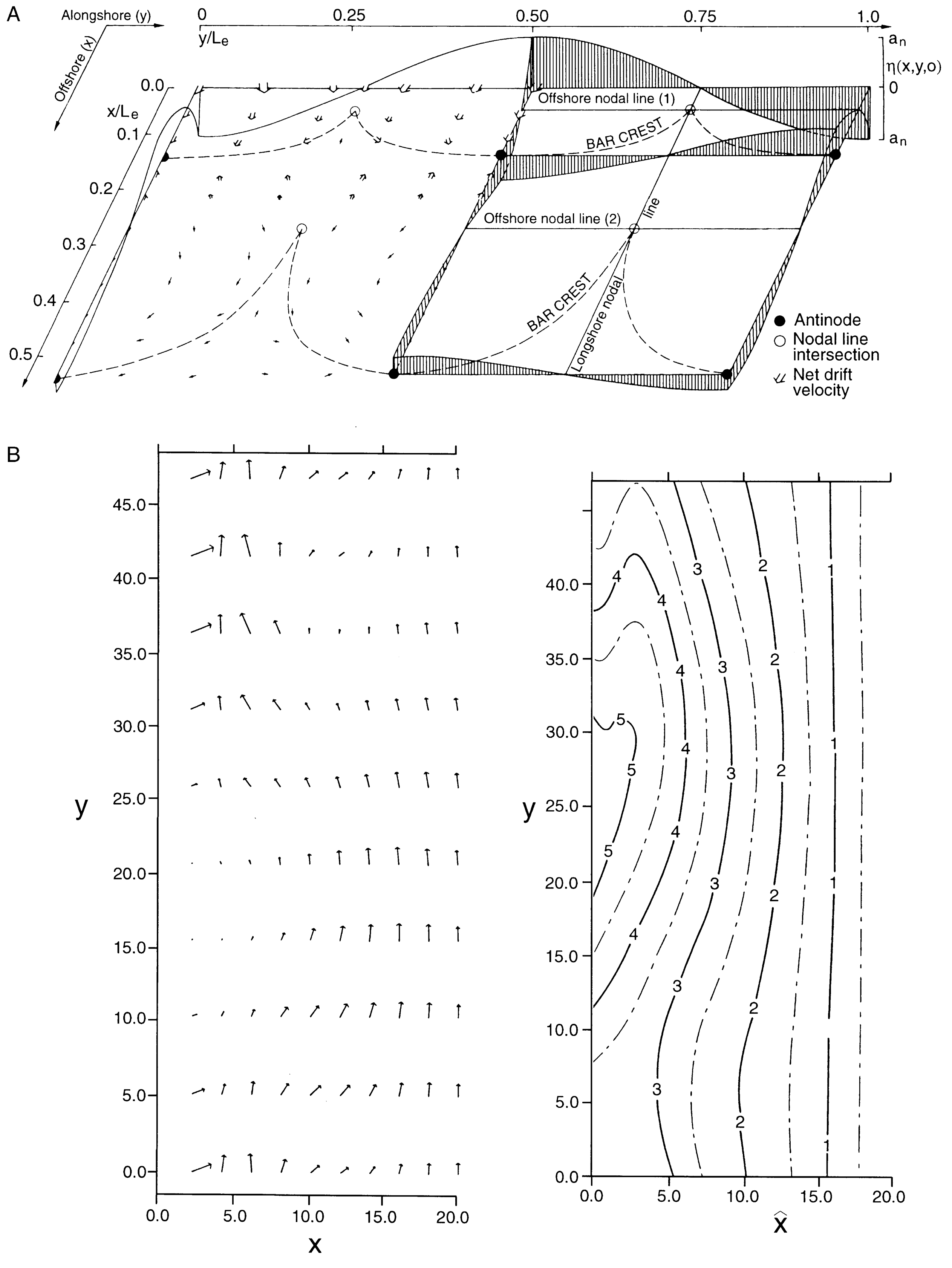

Figure B9 Bar formation by infragravity waves: (A) net drift velocities associated with standing edge waves and the creation of crescentic bars (modified after van Beek, 1974). (B) dimensionless drift velocities and equilibrium nearshore bathymetry associated with the propagation of two edge wave modes (1 and 2) of the same frequency in the same direction (modified after Holman and Bowen, 1982). Note: *y* represents the alongshore direction and *x* the across-shore direction.

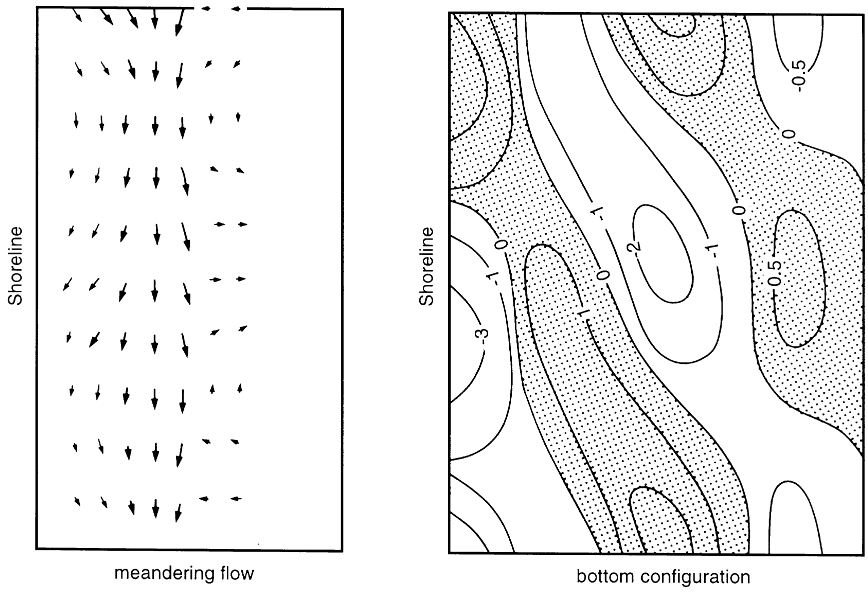

Figure B10 Bar formation due to hydrodynamic instability between longshore currents and the sand bed (modified after Hino, 1974). Note the meandering nature of the longshore flow and the sinuous bar topography that is produced.

Barred topography has long been associated with the occurrence of cellular nearshore circulations (Shepard *et al.*, 1941), and Hino (1974) proposed that an instability of the fluid sediment interface would generate variations in sediment transport resulting in sinuous or crescentic undulations of the surf-zone bed (Figure B9). Certainly the role of rip-cell circulation in bar dynamics has been well documented for *bar-* and *cusp-type sand waves* (Bowen and Inman, 1969; Davis and Fox, 1972; Sonu, 1973; Greenwood and Davidson-Arnott, 1975; Wright and Short, 1984), for *transverse bars* (Niedoroda, 1973), and for *Type VI* bars, both crescentic and straight (Greenwood and Davidson-Arnott, 1979). However, it is also possible that the regularity in nearshore circulations is in fact controlled by the presence of edge waves (Holman and Bowen, 1982). Whichever mechanism initiates bars, there will be feedback between the topography and the hydrodynamics, perhaps giving rise to some "hybrid" model of formation (Holman and Sallenger, 1993).

Bar morphodynamics

In general, the smaller the *wave-formed bar* the more dynamic it is, as there is less sediment involved in morphological changes (Sunamura and Takeda, 1984). However, there is considerable variability in morphodynamic behavior, depending upon bar type, the general environmental constraints, and indeed the antecedent state of the bar (i.e., whether or not it is close to its equilibrium position). Bars also tend to migrate at lower rates as the tidal range increases, since at some stage the bars are being exposed subaerially and remain static at this time. Bar dynamics have generally been related to behavior under specific storm events. However, the magnitude, frequency, and sequencing (*chronology*) of such events may be important in the nearshore, which as a nonlinear dynamical system, is extremely sensitive to feedback processes (see Moller and Southgate, 1997; Southgate and Moller, 2000; see Elgar, 2001 for an alternative view). There now exist at least two long time series of morphological change: **(**1**)** thirty years of annual profiling along 100 km of the Dutch coast (Ruessink and Kroon, 1994); **(**2**)** sixteen years of bathy metry recorded at Duck, NC (Plant *et al.*, 1999). Extensive measurements of the cross-shore location, and alongshore bar shape, are now being made successfully on a near continual basis at a number of locations worldwide using video-imagery (e.g., Lippmann and Holman, 1990; van Enckvort and Ruessink, 2001).

Type I bars are relatively stable in general, although landward migration rates of ~10 m per month have been recorded. Under low energy conditions the ridges have been observed to be: (1) destroyed by storms and regenerated in the post-storm period (Mulrennan, 1992); and (2) formed by storms (Hale and McCann, 1982). *Type II* bars have been shown to migrate at relatively rapid rates, both onshore and alongshore, and *Type III* bars migrate also at a relatively rapid rate. *Type II, IV*, and *VI* bars have been shown to occur as part of a temporal sequence of beach evolution by Wright *et al.*, (1979), Wright and Short (1984), Sunamura (1988), and Lippmann and Holman (1990). This sequence ranges from fully dissipative (barred profile) to fully reflective (non-barred profile) wave conditions, and therefore, is related to the surf similarity parameter ($\varepsilon = a_b\omega^2/g \tan^2\beta$; where a_b = breaker amplitude, ω = incident radian wave frequency, g = the gravitational constant, β = beach slope). In the Australian Model, the two-dimensional shore-parallel longshore bar and trough occurs at the fully dissipative beach stage, the rhythmic bar and beach at an intermediate stage, and the non-barred profile occurs at the fully reflective stage. In regions where a more limited range of waves exist, the beach may simply change between one or two stages, and where the environmental constraints are more restrictive still, then the bars may assume only one characteristic morphology. Further refinement of the stage model used the *Dean Parameter* ($\Omega = H_b/\omega_s T$; where H_b = breaker height in meters; ω_s = sediment fall velocity in meters per second; T = wave period in seconds). Barred profiles occurred when $\Omega > 0.85$ and non-barred profiles occurred when $\Omega < 0.85$ (Wright *et al.*, 1985). Sunamura (1988) used the dimensionless parameter $K^* = H_b^2/gT^2d$, where g is the gravitational constant and d is the grain size, to classify sequences dependent upon erosional or accretional beach stages. Erosion is characterized by $K^* \geq 20$ and is associated with offshore bar migration, slope decreases, and a dissipative state; while $5 \leq K^* \leq 20$ indicates onshore migration and beach accretion. Yet, a further parameter was introduced by Kraus and Larson (1988) to separate barred and non-barred profiles, $P = gH_o^2/\omega_s^3 T$, where H_o = offshore wave height. A value of 9,000 separates barred (greater values) from non-barred profiles (Dalrymple, 1992).

Type VI nearshore bars have been found to migrate onshore, offshore, and alongshore, with offshore rates reaching 2.5 m h^{-1} during storms and erosion/accretion rates of 0.05 m h^{-1} (Sallenger *et al.*, 1985; Aagaard and Greenwood, 1995). Onshore migration rates are generally smaller, but may still reach 1 m h^{-1}. When the *Type VI* bars are three-dimensional, they may migrate alongshore at rates up to 10 m per month (Greenwood and Davidson-Arnott, 1975). Ruessink *et al.* (2000) examined the relative rates of across-shore and alongshore migration using complex empirical orthogonal functions applied to profile data. The alongshore migration rate ranged up to 150 m per day and was strongly related to the alongshore component of the offshore wave energy flux. Short-term variability in bar crest position was shown to be due to changes in the quasi-regular topography, and not to alongshore uniform on–offshore migration. While offshore migration under storms has been clearly related to hydrodynamic forcing, especially the setup-driven undertow (Gallagher *et al.*, 1998) or mean currents modulated by infragravity waves (Aagaard and Greenwood, 1995), the onshore migration of Type VI bars is poorly known. Generally the motion is attributed to skewed fluid velocities and accelerations (Elgar *et al.*, 2001).

On the Dutch coast a multiple bar *Type VI* system exhibited characteristics of a feedback-dominated system, producing cyclic changes over either 4 or 15–18 years (Wijnberg and Terwindt, 1995). Plant and Holman (1997) showed that bars on the east coast of the United States exhibited unpredictable behavior in relation to wave height changes and yet still moved through a sequential pattern of form changes. This paradoxical behavior they related to feedback effects. The forcing for these transitions is as controversial as bar genesis, since direct hydrodynamic forcing has been proposed as well as a self-organization mechanism.

Little work has been done specifically upon bar decay, other than the welding process associated with *Type II* bars (e.g., Davis *et al*., 1972; Aagaard *et al*., 1998). However, the one major exception is the study of the multiple bar system along the Dutch coast. Here, the bar system shifts progressively offshore over time and the outermost bar decays. This has been attributed to the action of highly asymmetric, nonbreaking waves (Larson and Kraus, 1992; Reussink and Kroon, 1994; Wijnberg, 1997). Plant *et al*. (2001) suggest that a morphologic feedback mechanism can lead to bar decay. As bars move onshore under nonbreaking conditions they are also reduced in height; thus they move further away from wave breaking, allowing further bar decay. This has been observed at Duck, NC (Lippmann *et al*., 1993).

Predictive models of bar genesis and dynamics

Because of the relatively poor knowledge of long-term bar behavior and the inadequacy of local sediment transport models for the complex nearshore environment, predictive models for the genesis and dynamics of wave-formed bars are still far from complete. In general models proposed are either (1) process-based models (e.g., Bowen, 1980), or (2) behavior-based models (de Vriend *et al*., 1993). The latter range from: (1) highly parameterized models to predict summer–winter (bar–berm) profiles (e.g., Aubrey, 1979) or sequential bar evolution (Wright and Short, 1984; Sunamura, 1988) to (2) statistically based models for predicting bar dynamics (Aubrey *et al*., 1980) to (3) morphological models to simulate large-scale beach changes (e.g., Cowell *et al*., 1995).

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Bibliography

Aagaard, T., 1990. Infragravity waves and nearshore bars in protected, storm-dominated coastal environments. *Marine Geology*, **94**: 181–203.

Aagaard, T., and Greenwood, B., 1995. Suspended sediment transport and morphological response on a dissipative beach. *Continental Shelf Research*, **15**: 1061–1086.

Aagaard, T., Nielsen, J., and Greenwood, B., 1998. Suspended sediment transport and nearshore bar formation on shallow intermediate-state beach. *Marine Geology*, **148**: 203–225.

Aubrey, D.G., 1979. Seasonal patterns of onshore/offshore sediment movement. *Journal of Geophysical Research*, **84**: 6347–6354.

Aubrey, D.G., Inman, D.L., and Winant C.D., 1980. The statistical prediction of beach changes in southern California. *Journal of Geophysical Research*, **85**: 3264–3276.

Barcilon, A.L., and Lau, J.P., 1973. A model for the formation of transverse bars. *Journal of Geophysical Research*, **78**: 2656–2664.

Bauer, B.O., and Greenwood, B., 1990. Modification of a linear bar-trough system by a standing edge wave. *Marine Geology*, **92**: 177–204.

Boczar-Karakiewicz, B., and Davidson-Arnott, R.G.D., 1987. Nearshore bar formation by non-linear wave processes—a comparison of model results and field data. *Marine Geology*, **77**: 287–304.

Bowen, A.J., 1980. Simple models of nearshore sedimentation: beach profiles and longshore bars. In McCann, S.B. (ed.), *The Coastline of Canada: littoral processes and shore morphology*. Geological Survey of Canada, Special Paper: 80–10, pp. 1–11.

Bowen, A.J., and Huntley, D.A., 1984. Waves, long waves and nearshore morphology. *Marine Geology*, **60**: 1–13.

Bowen, A.J. and Inman, D.L., 1969. Rip currents, 2: laboratory and field observations. *Journal of Geophysical Research*, **74**(23): 5479, 5490.

Bowen, A.J., and Inman, D.L., 1971. Edge waves and crescentic bars. *Journal Geophysical Research*, **76**: 8662–8671.

Bowman, D., and Goldsmith, V., 1983. Bar morphology of dissipative beaches: an empirical model. *Marine Geology*, **51**: 15–33.

Bryan, K., and Bowen, A.J., 1996. Edge wave trapping and amplification on barred beaches. *Journal of Geophysical Research*, **101**: 6543–6552.

Carter, R.W.G., 1978. Small-scale transverse bars in Lough Neagh, Northern Ireland. *Journal Earth Science, Royal Society of Dublin*, **1**: 205–209.

Carter, T.G., Liu, P.L. and Mei, C.C., 1973. Mass transport by waves and offshore sand bedforms. *American Journal of Waterways, Harbors and Coastal Engineering Division*, ASCE **99**: 165–184.

Cowell, P.J., Roy, P.S. and Jones, R.A., 1995. Simulation of LSCB using a morphological behaviour model. *Marine Geology*, **126**: 45–61.

Dally, W.R., 1987. Longshore bar formation—surf beat or undertow? Coastal Sediments '87, ASCE, New York, pp. 71–86.

Dally, W.R., 1980. Numerical model for beach profile evolution. M.S. thesis, Department Civil Engineering, Newark: University of Delaware.

Dally, W.R., and Dean, R.G., 1984. Suspended sediment transport and beach profile evolution. J. Waterway, Port, Coastal and Ocean Eng., **110**: 15–33.

Dalrymple, R.A., 1992. Prediction of storm/normal beach profiles. *Journal of Waterway, Port, Coastal and Ocean Engineering*, **118**: 193–200.

Damgaard Christiansen, E., Deigaard, R., and Fredsøe, J., 1994. Sea bed stability on a long straight coast. In *Proceeding 24th International Conference on Coastal Engineering*, American Society Civil Engineers, New York, pp. 1865–1879.

Davidson-Arnott, R.G.D., 1981. Computer simulation of nearshore bar formation. *Earth Surface Processes and Landforms*, **6**: 23–34.

Davidson-Arnott, R.G.D., 1988. Controls on formation of barred nearshore profiles. *Geographical Review*, **78**: 185–193.

Davis, R.A., Jr., and Fox, W.T., 1972. Coastal processes and nearshore sand bars. *Journal of Sedimentary Petrology*, **42**: 401–412.

Davis, R.A., Jr., Fox, W.T., Hayes, M.O., and Boothroyd, J.C., 1972. Comparison of ridge and runnel systems in tidal and non-tidal environments. *Journal of Sedimentary Petrology*, **42**: 413–421.

De Vriend, H.J., Capobianco, M., Chesher, T., De Swart, H.E., Latteux, B., and Stive, M.J.F., 1993. Approaches to long term modelling of coastal morphology: a review. *Coastal Engineering*, **21**: 225–269.

Deigaard, R., Dronen, N., Fredsoe, J., Jensen, J.H., and Jorgensen, M.P., 1999. A morphological stability analysis for a long straight barred coast. *Coastal Engineering*, **36**: 171–195.

Dolan, T.J., and Dean, R.G., 1985. Multiple longshore sand bars in the upper Chesapeake Bay. *Estuarine, Coastal and Shelf Science*, **5**(21): 727–743.

Elgar, S., 2001. Coastal profile evolution at Duck, North Carolina: a precautionary note. *Journal of Geophysical Research*, **106**: 4625–4627.

Elgar, S., Gallagher, E.L., and Guza, R.T., 2001. Nearshore sandbar migration. *Journal of Geophysical Research*, **106**: 11623–11627.

Evans, O.F., 1940. The low and ball of the east shore of Lake Michigan, *Journal of Geology*, **48**: 467–511.

Exon, N.F., 1975. An extensive offshore sand bar field in the western BalticSea. *Marine Geology*, **18**: 197–212.

Falques, A., 1991. A note on the Barcilon and Lau model for transverse bars. *Revista de Geofisica*, **47**: 191–95.

Falques, A., Coco, G., and Huntley, D.A., 2000. A mechanism for the generation of wave-driven rhythmic patterns in the surf zone. *Journal Geophysical Research*, **105**: 24071–24087.

Gallagher, E., Guza, R.T., and Elgar, S., 1998. Observations of sand bar behaviour on a natural beach. *Journal of Geophysical Research*, **103**: 3203–3215.

Gelfenbaum, G., and Brooks, G.R., 1997. Long term observations of migrating shore-normal bars. *Proceedings of Coastal Dynamics* '97, Plymouth, ASCE, New York, pp. 654–663.

Greenwood, B., and Davidson-Arnott, R.G.D., 1975. Marine bars and nearshore sedimentary processes, Kouchibouguac Bay, New Brunswick, Canada. In Hails, J. and Carr, A. (eds.), *Nearshore Sediment Dynamics and Sedimentation*. London: John Wiley & Sons Ltd., pp. 123–150.

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 Science*, **16**: 313–332.

Greenwood, B., and Mittler, P.R., 1979. Structural indices of sediment transport in a straight, wave-formed, nearshore bar. *Marine Geology*, **32**: 191–203.

Hale, P.B., and McCann, S.B., 1982. Rhythmic topography in a mesotidal low-wave-energy environment. *Journal of Sedimentary Petrology*, **52**: 415–429.

Hands, E.H., 1976. Observations of barred coastal profiles under the influence of rising water levels, Eastern Lake Michigan, 1967–71. U.S. Army, Corps of Engineers, *CERC Technical Report*, 76-1, 113 pp.

Hayes, M.O., and Boothroyd, J.C., 1969. Storms as modifying agents in the coastal environment. In Hayes, M.O. (ed.), *Coastal Environments: NE Massachusetts*. Amherst: Department Geology, University Massachusetts.

Heathershaw, A.D., and Davies, A.G., 1985. Resonant wave reflection by transverse bedforms and its relation to beaches and offshore bars. *Marine Geology*, **62**: 321–338.
Hino, M., 1974. Theory on formation of rip current and cuspidal coast. *Coastal Engineering Japan*, **17**: 23–38.
Holman, R.A., and Bowen, A.J., 1982. Bars, bumps and holes: models for the generation of complex beach topography. *Journal of Geophys. Research*, **87**: 457–468.
Holman, R.A., and Sallenger Jr., A.B., 1993. Sand bar generation: a discussion of the Duck Experiment Series. *Journal of Coastal Research*, **15** (Special Issue): 76–92.
Howd, P.A., Bowen, A.J., and Holman, R.A., 1992. Edge waves in the presence of strong longshore currents. *Journal of Geophys. Research*, **97**: 11357–11371.
Huntley, D.A., 1980. Edge waves in a crescentic bar system. In McCann, S.B. (ed.), *Coastline of Canada: Littoral Processes and Shore Morphology*. Geological Survey of Canada, Paper 80-10, Report 2, pp. 111–121.
Inman, D.L., Elwany, M.H.S., and Jenkins, S.A., 1993. Shorerise and bar–berm profiles on ocean beaches. *Journal of Geophysical Research*, **98**: 18181–18199.
Katoh, K., 1984. Multiple longshore bars formed by long period standing waves. *Report of the Port and Harbour Research Institute*, **23**: 3–46.
King, C.A.M., and Williams, W.W., 1949. The formation and movement of sand bars by wave action. *Geographical Journal*, **112**: 70–85.
Kirby, J.T., Dalrymple, R.A., and Liu, P.L.-F., 1981. Modification of edge waves by barred beach topography. *Coastal Engineering*, **5**: 35–49.
Konicki, K.M., and Holman, R.A., 2000. Transverse bars at Duck, North Carolina. Proceedings of the 25th International. Conference of Coastal Engineering, ASCE, New York, pp. 3588–3599.
Kraus, N.C., and Larson, M., 1988. Beach profile changes measured in the tank for large waves, 1956–57 and 1962. U.S. Army Corps of Engineers, Coastal Engineering Research Center, *Technical Report*. CERC-89-9.
Larson, M., and Kraus, N.C., 1992. Analysis of cross-shore movement of natural longshore bars and material to create longshore bars. U.S. Army Corps of Engineers, CERC Technical Report, DRP-92-5, 89 pp.
Lippmann, T.L., and Holman, R.A., 1990. The spatial and temporal variability of sand bar morphology. *Journal of Geophysical Research*, **95**: 11575–11590.
Lippmann, T.L., Holman, R.A., and Hathaway, K.K., 1993. Episodic, non-stationary behaviour of a two sand bar system at Duck, NC, USA. *Journal of Coastal Research*, **15**(Special Issue): 49–75.
Miller, R.L., 1976. Role of vortices in surface zone prediction: sedimentation and wave forces. In Davis, R.A., Jr. and Ethington, R.L. (eds.), *Beach and Nearshore Sedimentation*. Society of Economic Paleontologists and Mineralogists Special Publication 24, pp. 92–114.
Moller, I., and Southgate, H.N., 1997. Fractal properties of beach profile evolution at Duck, North Carolina. *Journal of Geophysical Research*, **105**: 11489–11507.
Mulrennan, M.E., 1992. Ridge and runnel beach morphodynamics: an example from the central east coast of Ireland. *Journal of Coastal Research*, **8**: 906–918.
Niedoroda, A.W., 1973. Sand bars along low energy beaches. Part 2: Transverse bars, In Coates, D.R. (ed.), *Coastal Geomorphology*. Binghamton: State University of New York, pp. 103–113.
Niedoroda, A.W., and Tanner, W.F., 1970. Preliminary study of transverse bars. *Marine Geology*, **9**: 41–62.
Nilsson, H.D., 1973. Sand bars along low energy beaches Part 1: multiple parallel sand bars of southeastern Cape Cod. In Coates, D.R. (ed.), *Coastal Geomorphology*, Publication In Geomorphology, New York; State University of New York, pp. 99–102.
O'Hare, T., 1994. The role of long waves in sand bar formation: a model exploration. In *Proceedings of Coastal Dynamics, '94*. Barcelona, ASCE, New York, pp. 74–78.
O'Hare, T.J., and Davies, A.G., 1993. Sand bar formation beneath partially standing waves: laboratory experiments and model simulations. *Continental Shelf Research*, **13**: 1149–1181.
O'Hare, T.J., and Huntley, D.A., 1994. Bar formation due to wave groups and associated long waves. *Marine Geology*, **116**: 313–325.
Orford, J.D., and Wright, P., 1978. What's in a name? Descriptive or genetic implications of "ridge and runnel" topography. *Marine Geology*, **28**: M1–M8.
Orme, A.R., and Orme, A.J., 1988. Ridge and runnel enigma. *Geographical Review*, **78**: 169–184.
Osborne, P.D., and Greenwood, B., 1992. Frequency dependent cross-shore suspended sediment transport 2: a barred shoreface, Bluewater Beach, Ontario, Canada. *Marine Geology*, **106**: 25–51.
Plant, N.G., and Holman, R.A., 1997. Strange kinematics of sandbars. *Proceedings of Coastal Dynamics '97*, Plymouth, ASCE, New York, pp. 355–364.
Plant, N.G., Holman, R.A., Freilich, M.H., and Birkemeier, W.A., 1999. A simple model for interannual sandbar behaviour. *Journal of Geophysical Research*, **104**: 15755–15776.
Plant, N.G., Freilich, M.H., and Holman, R.A., 2001. Role of morphologic feedback in surf zone sandbar response. *Journal of Geophysical Research*, **106**: 973–989.
Rey, V., Davies, A.G., and Belzons, M., 1995. On the formation of bars by the action of waves on an erodible bed: a laboratory study. *Journal of Coastal Research*, **11**: 1180–1194.
Roelvink, J.A., and Broker, I., 1993. Cross-shore profile models. *Coastal Engineering*, **21**: 163–191.
Ruessink, B.G., 1998. Infragravity waves in a dissipative multiple bar system. Ph.D. thesis, University of Utrecht, 243 pp.
Ruessink, B.G. and Kroon, A., 1994. The behaviour of a multiple bar system in the nearshore zone of Terschelling, the Netherlands, 1965-1993. *Marine Geology*, **121**: 187–197.
Ruessink, B.G., van Enckevort, I.M.J., Kingston, K.S., and Davidson, M.A., 2000. Analysis of observed two- and three-dimensional nearshore bar behaviour. *Marine Geology*, **64**: 237–257.
Sallenger, A.H., Jr., and Howd, P.A., 1989. Nearshore bars and the breakpoint hypothesis. *Coastal Engineering*, **12**: 301–313.
Sallenger, A.H., Jr., Holman, R.A. and Birkemeier, W.A., 1985. Storm-induced response of a nearshore bar system. *Marine Geology*, **64**: 237–258.
Schwartz, M., 1982. *Encyclopaedia of Beaches and Coastal Environments*. Stroudsburg, PA: Hutchinson Ross Publishing Company.
Shepard, F.P., 1950. Longshore bars and longshore troughs. Beach Erosion Board, Technical *Memo* 15, p. 32.
Shepard, F.P., Emery, K.O., and LaFond, E.C., 1941. Rip currents: a process of geological importance. *Journal of Geology*, **49**: 337–369.
Short, A.D., 1979. Three-dimensional beach stage model. *Journal of. Geology*, **87**: 553–571.
Short, A.D., 1975. Multiple offshore bars and standing waves. *Journal of Geophysical Research*, **80**: 3838–3840.
Short, A.D., and Aagaard, T., 1993. Single and multi-bar beach change models. *Journal of Coastal Research*, **15**(Special Issue): 141–157.
Southgate, H.N., and Moller, I., 2000. Fractal properties of coastal profile evolution at Duck, North Carolina. *Journal of Geophysical Research*, **105**: 11489–11507.
Sonu, C.J., 1973. Three-dimensional beach changes. *Journal Geology*, **81**: 42–64.
Suhayda, J.N., 1974. Standing waves on beaches. *Journal of Geophysical Research* **79**: 3065–3071.
Sunamura, T., 1988. Beach morphologies and their change. Chapter 2, In Horikawa, K. (ed), Nearshore dynamics and coastal processes: theory, measurement and predictive models. Tokyo: University of Tokyo Press, pp. 136–166.
Sunamura, T., and Takeda, I., 1984. Landward migration of inner bars. *Marine Geology*, **60**: 63–78.
Thornton, E.B., Humiston, R.T., and Birkemeier, W., 1996. Bar/trough generation on a natural beach. *Journal of Geophysical Research*, **101**: 12097–12110.
Van Beek, J.L., 1974. Rhythmic patterns of beach topography. *Technical Report*, 187, 46 pp., Coastal Studies, Baton Rouge: Louisiana State University.
Van Enckevort, I.M.J., and Ruessink, B.G., 2001. Effect of hydrodynamics and bathymetry on video estimates of nearshore sand bar position. *Journal Geophysical Research*, **106**: 16969–16979.
Van Rijn, L.C., 1998. *Principles of Coastal Morphology*. Amsterdam: Aqua Publications.
Vittorio, G., de Swart, H.E., and Blondeaux, P., 1999. Crescentic bedforms in the nearshore region. *Journal of Fluid Mechanics*, **381**: 271–303.
Wijnberg, K.M., 1997. On the systematic offshore decay of breaker bars. *Proceedings 25th International Conference Coastal Engineering*, New York, pp. 3600–3613.
Wijnberg, K.M., and Terwindt, J.H.J., 1995. Quantification of decadal morphological behaviour of the central Dutch coast. *Marine Geology* **126**: 301–330.
Wright, L.D., and Short, A.D., 1984. Morphodynamic variability of surf zones and beaches: a synthesis. *Marine Geology*, **56**: 93–118.
Wright, L.D., Short, A.D., and Green, M.O., 1985. Short-term changes in the morphodynamic states of beaches and surf zones: an empirical predictive model. Marine Geology, 62: 339–364.
Wright, L.D., Chappell, J., Thom, B.G., Bradshaw, M.P., and Cowell, P., 1979. Morphodynamics of reflective and dissipative beach and inshore systems, southern Australia, *Marine Geology*, **32**: 105–140.

Yu, J., and Mei, C.C., 2000. Formation of sand bars under surface waves. *Journal of Fluid Mechanics*, **416**: 315–348.
Zhang, D.P., 1994. Wave flume experiments on the formation of longshore bars produced by breaking waves. *Science Report, Institute Geoscience, University Tsukuba*, **15**: 47–105.
Zenkovith, V.P., 1967. *Processes of Coastal Development*. London: Oliver and Boyd.

Cross-references

Beach Features
Beach Processes
Net Transport
Profiling, Beach
Rhythmic Patterns
Surf Zone Processes
Wave–Current Interaction

BAY BEACHES

The length of shore in bays, sounds, lagoons, and estuaries (here termed bays) greatly exceeds the length of ocean shore in many countries. Beaches are common in these bays, but they are often so small and isolated that they escape attention, except in populated locations. The definition of bay in relation to the open coast is somewhat subjective, and bays such as Monterey Bay, California may have wave-energy levels that are among the highest in the world. This discussion is confined to low-energy beaches that occur in mostly enclosed bays where the fetch distances for local wave generation are generally less than 50 km. The principal factors affecting the morphodynamics of these beaches are locally generated waves and wave-induced currents, but wind-induced and tidal currents play a role in morphologic change. Fluvial processes may become dominant at estuarine shores in narrow basins or tributaries. At the low end of the wave-energy continuum, other terms, such as stream bank, intertidal marsh margin, or bay bottom may be more appropriate than beach.

Shore processes

Waves generated by local winds in bays have low heights (usually mean heights <0.2 m and storm wave heights <1.0 m) and short periods (2.0–4.5 s) (Nordstrom, 1992). Ocean waves entering bays play a limited role in beach change where shores do not face ocean entrances (Jackson, 1995). Tidal range affects the vertical distribution of wave energy over the profile, determining the width of the beach and the duration that waves break at any elevation. Bay beaches are usually characterized by a steep upper foreshore with a broad, flat fronting terrace. On tidal beaches, spilling waves break in a broad surf zone across the gently sloping terrace at low tide, but the energy in the waves is low. At high tide, waves reach the upper foreshore with little loss of energy and usually break as plunging waves.

Longshore currents are predominantly generated by the breaking of local wind-waves but refracted ocean waves, tidal flows, and wind drift are locally important and may result in flows bayward of the breaking waves that are opposite flows generated by local wind-waves. Tidal currents are important near channels, projecting headlands, and constrictions in the bay, and they may be the dominant agent of sediment transport on the terrace bayward of the foreshore. Ice forms faster and has a greater influence on mid- and high-latitude bay beaches than on ocean beaches because bay waters are colder in winter, shallower, and less saline; ice lasts longer because low wave energies are slow to remove it. Ship and boat wakes are higher on bay beaches than on ocean beaches because vessels can pass close to the shore, but the average energy in the wakes is usually only a small percentage of the average energy of wind waves in all but the smallest bays.

Water level changes can be locally induced by winds blowing across the bay or they can be induced by flow of water through inlets from surges generated on the open coast. Winds can increase water levels on the downwind side of the bay while lowering water levels on the upwind side, but a large opening to the sea on the downwind side of the bay can result in lower water levels downwind.

Beach and shore characteristics

Beaches comprise a large proportion of the shore in many bays (Nordstrom and Roman, 1996). Important examples include Delaware Bay (Jackson, 1995), Chesapeake Bay (Rosen, 1980; Ward *et al*., 1989), Puget Sound (Downing, 1983; Terich, 1987). Beaches in smaller bays, with limited availability of sand and gravel may be small, highly localized, or confined to ocean entrances. Many beaches have been created in urbanized estuaries where none would occur naturally because wave energies are too low. These artificial beaches are often wider than natural beaches in undeveloped areas. Some new beaches are accidental byproducts of landfill operations; some are created intentionally as new beach recreation areas (Nordstrom, 1992).

Bay beaches may be unvegetated or partially vegetated and composed of sand, gravel, or shell. Surface sediments are often coarser on bay beaches than on ocean beaches with a similar source. Lag gravel is common on the beach surface, formed from particles exhumed by swash or by preferential elimination of fines by low-energy waves. Individual pebbles move readily over the sand surface, and swash excursions create bands of gravel on the upper foreshore.

The depth of mobilization of sediments on the upper foreshore is shallow (e.g., <0.2 m under storm conditions), and the active beach may be only a thin veneer of unconsolidated material overlying an immobile layer of coarse sediments, clay, peat, or a shore platform. Mobilization of sediments on the low tide terrace by waves may occur only to depths of 10–30 mm, and biological activity may play a greater role than wave processes in altering the characteristics of the surface and subsurface (Nordstrom, 1992).

Vegetation plays a greater role in influencing morphologic change on bay beaches than on ocean beaches because of greater abundance of vegetation in bays and the reduced ability of the low-energy waves to move it. Vegetation helps bind bottom sediment and attenuate wave energies; vegetation flotsam in the breaker and surf zones alters the wave and current characteristics and the likelihood of entrainment of beach sediment; vegetation litter in the wrack line forms barriers to waves, currents, and swash uprush.

Bay shorelines are often composed of numerous isolated beaches with different orientations. They have high variability in morphology and rate of erosion over small areas resulting from local differences in fetch, wind direction, stratigraphy, inherited topography, resistant outcrops on the foreshore, variations in submergence rates, and amounts of sediment in eroding formations (Phillips, 1986; Rosen, 1980). Beach compartments are isolated into longshore drift cells defined by deep coves or headlands formed by resistant rock, marsh, or human structures.

The net rate of longshore transport on estuarine beaches varies with orientation, fetch distance, and size of each drift cell and ranges from tens of cubic meters to tens of thousands of cubic meters (Wallace, 1988). Although rates of transport are low, the magnitude of erosion can be high because the quantities of sediment in transport represent a sizable fraction of the total unconsolidated sediment in the active beach. Many bay shores are eroding at greater rates than nearby ocean shores.

Beach change

The upper foreshores of most bay beaches are modally reflective. Conspicuous cyclic morphologic change is confined to the immediate vicinity of the foreshore. Sediment removed from the upper foreshore during high-wave-energy events is deposited on the lower foreshore with a change to a concave upward profile shape. Sediments moved farther offshore onto the terrace form only a thin veneer over the surface instead of forming the break point bar that is prominent on many ocean beaches. Landward and bayward displacement of the entire foreshore profile may also occur while the profile slope is maintained. This parallel-slope retreat and advance is common when sediment exchange is due primarily to longshore transport and is most pronounced near the ends of drift compartments (Nordstrom, 1992).

Resource values of bay beaches

The fronting terrace of a low-energy bay beach has a relatively stable substrate that allows macroscopic plants and fauna to thrive. The upper foreshore is more energetic and may have less species diversity and abundance. Infauna and macroalgae provide prey to juveniles of commercially valuable fish, and the intertidal area provides habitat for recreationally important clams and numerous species of epifauna and infauna. The upper foreshore may be an important spawning area for horseshoe crabs. Fish and invertebrates are prey for foraging birds, especially in the regularly exposed intertidal zone. Wrack from plant litter is inhabited by numerous amphipods and insects. The swash zone and dry upper foreshore are also foraging areas for birds, including upland species.

Bay beaches are not as intensively used for recreation as ocean beaches, but they have important complementary values. They provide convenient surfaces for launching and landing boats and boards for wind surfing. They are favored by parents with children because they provide a safer environment than on the ocean. Many bay beaches are underutilized for recreation because of the unclean appearance of the beaches or lack of awareness of their existence or unique attributes, but ease of access causes bay beaches close to urban areas to have relatively high rates of use.

Shore protection and management

Erosion control strategies for bay beaches may differ from strategies for ocean beaches because of differences in the scale of erosional forces and in the value of resources. Protection programs are facilitated because beach segments are small, isolated drift cells, often under jurisdiction of only one management agency and because small-scale, low-cost protection may be utilized. Low wave energies and gentle offshore gradients make construction of fixed offshore engineering works more practical than on high-energy beaches. Shore-parallel walls are often successful because they can withstand direct attack of local waves; they take up minimal space on the beach and adjacent upland; and they limit the loss of biological resources on the fronting terrace or bay bottom. Projects funded by national or state/ provincial governments are often not economically feasible, resulting in a fragmented approach to protection by individual property owners. Simple engineering principles are often ignored in constructing small-scale protection structures, including lack of filter cloth or weep holes in bulkheads, failure to build structures deep enough to prevent toe scour or high enough to prevent overtopping, weak fastenings, and failure to use adequate sized armor stones or perform maintenance. As a result, there is much evidence of structural failure. Beach fill is increasingly used for protection or recreation, but fill can cover benthic habitat and eliminate shallow-water areas for aquatic plants. Bayside nourishment projects can be inexpensive because only small quantities of fill are required. Fill materials brought in from outside the region may retain their exotic appearance because of limited mixing by low-energy waves.

There has been considerable federal and state intervention in decisions on developing bay shores, especially in productive estuaries, but this intervention is rarely conducted to maintain beach resources. Alternative human uses such as transportation, industrial developments, residences, and boating are compatible with a coastal location according to most policies, and actions to enhance these uses may eliminate beaches. The number and value of bay beaches can be enhanced by implementing beach nourishment operations, altering vegetation, constructing appropriate protection structures, acquiring key sites for public use, and enhancing access. The ease of constructing and maintaining bay beaches and the paucity of quality recreation space in many urban areas make creation of new beaches as surrogates for ocean beaches an attractive option.

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Bibliography

Downing, J., 1983. *The Coast of Puget Sound: Its Processes and Development*. Seattle, WA: Puget Sound Books.

Jackson, N.L., 1995. Wind and waves: influence of local and non-local waves on mesoscale beach behavior in estuarine environments. *Annals of the Association of American Geographers*, **85**: 21–37.

Nordstrom, K.F., 1992. *Estuarine Beaches*. London: Elsevier Applied Science.

Nordstrom, K.F. and Roman, C.T. (ed.), 1992. *Estuarine Shores: Evolution, Environments and Human Alterations*. London: John Wiley & Sons.

Phillips, J.D., 1986. Spatial analysis of shoreline erosion: Delaware Bay, NJ. *Annals of the Association of American Geographers*, **76**: 50–62.

Rosen, P.S., 1980. Erosion susceptibility of the Virginia Chesapeake Bay shoreline. *Marine Geology*, **34**: 45–59.

Terich, T.A., 1987. *Living with the Shore of Puget Sound and the Georgia Strait*. Durham, NC: Duke University Press.

Wallace, R.S., 1988. Quantification of net shore-drift rates in Puget Sound and the Straight of Juan de Fuca, Washington. *Journal of Coastal Research*, **4**: 395–403.

Ward, L.G., Rosen, P.S., Neal, W.J., Pilkey, O.H., Jr., Pilkey, Sr., O.H., Anderson, G.L., and Howie, S.J., 1989. *Living with Chesapeake Bay and Virginia's Ocean Shores*. Durham, NC: Duke University Press.

Cross-references

Beach Erosion
Beach Nourishment
Beach Processes
Dissipative Beaches
Estuaries
Human Impacts on Coasts
Reflective Beaches
Sediment Transport
Shore Protection Structures

BEACH AND NEARSHORE INSTRUMENTATION

Instrumentation in studies of the coast generally, and of the beach and nearshore zone in particular is designed to measure attributes of form and changes in the form (bed) over time, including bedforms; fluid processes related to waves, water level and currents in the water and wind on the beach; and sediment concentration and mass transport rate in the water and on the beach. These measurements may be made at a variety of temporal scales ranging from fractions of a second to months and years and spatial scales ranging from a few square millimeters to hundreds of square kilometers. Some attributes are measured individually, but much of the focus today, and over the past three decades, has been on measurements of morphodynamics, in which the objective is to measure fluid and sediment transport processes and the resulting change in morphology at a temporal scale of minutes to days and occasionally months. Much activity is focused on sandy and to a lesser extent muddy coasts and much of the instrumentation described here is devoted to these, but some work also takes place on the erosion of cohesive clay and bedrock coasts. The highly dynamic nature of the nearshore and swash zone in particular poses many problems for the design of instruments for measuring fluid and sediment transport processes. In addition to the need for very rugged instruments and supports for mounting them, there are difficulties posed by the lack of access to much of the nearshore during storm conditions, and by the presence of bubbles and organic matter in the water column. Ultimately, the instrumentation is designed and deployed to measure particular properties and processes of the beach and nearshore zones, and therefore, this review is organized by the measurement objective rather than particular instrument types.

Measurement of form and changes in form (erosion and deposition)

Erosion of cohesive and bedrock coasts

Where the coastline is developed in bedrock, till, and cohesive muds the focus of attention is usually on the measurement of rates of erosion in relation to the strength attributes of the material and the erosional or forcing processes (Sunamura, 1992). On a small scale, erosion by weathering and abrasion of rock coasts generally takes place so slowly that measurements are made at point locations on a timescale of months to years. The micro erosion meter originally used to measure solution of limestone was adapted for use on intertidal shore platforms (Trudgill *et al*., 1981). It consists of a pointer attached to a micrometer gauge on a mount that can be placed on pins drilled into the rock platform. The mount swivels to allow measurement to be taken at several points around the station so that an average value can be obtained. Measurements are commonly taken at intervals of months or years because of the relatively slow rate of downcutting (Kirk, 1977; Viles and Trudgill, 1984). A cruder version of the instrument has been adapted for measurements of erodability of tills and clays underwater (Askin and Davidson-Arnott, 1981; Davidson-Arnott and Langham, 2000). Because the erosion rates are typically up to several centimeters per year, measurements can be made on a weekly to monthly basis.

Measurements of the erodability of fine-grained cohesive muds commonly found in a variety of marine and estuarine environments, have commonly been made with a variety of benthic flumes—essentially inverted channels of various configurations which can be deployed either on the exposed tidal flat or underwater (Amos *et al*., 1992; Maa *et al*., 1993; Houwing, 1999). Water is circulated through the channel at increasing speeds until the shear on the bed induces erosion and the erosion rate is measured either directly, or indirectly by measurement of suspended sediment concentrations. Recent experiments have also been made with the Cohesive Strength Meter, an automated device which employs a carefully regulated vertical jet of water and monitors the rate of erosion with respect to the impact force (Tolhurst *et al.*, 1999).

Erosion and deposition of sediments

While large-scale changes in form can be measured by a variety of techniques described below, these techniques are usually carried out at finite intervals of days, weeks, or months. Measurement of changes in the bed at particular locations on the timescale of dynamic measurements of fluid flow and sediment transport, typically on the order of minutes to hours, has proved to be surprisingly difficult to do in shallow water. One major problem in the nearshore during storms on sandy coasts is the difficulty of distinguishing the bed from the material immediately above it, which is being transported as bed load or suspended load close to the bed. Techniques for measuring changes in elevation at points in the nearshore range from simple erosion rods to optical and acoustical instruments.

Simple measurements of change in bed elevation and the total depth of activation can be made with rods emplaced along a profile or on a grid which are surveyed before and after a storm (Greenwood *et al.*, 1979). The rods can be emplaced by wading and diving. The maximum scour depth can be resolved by placing a washer on the sand surface and then measuring the depth of burial following the storm. Results from a grid of these can be used to measure volume change in the nearshore (Greenwood and Mittler, 1984). Similarly, thin rods can be used on the subaerial beach to measure erosion by wind and thus provide a comparison volume to measurements of sand transport or deposition. Other approaches involving this simple technology in coastal applications include the use of a bedframe device to measure rates of sediment deposition in foredunes (Davidson-Arnott and Law, 1990, 1996) and the use of Surface Elevation Tablet (SET) stations in measuring net change in saltmarshes (Cahoon *et al.*, 2000). Recently, automated devices which act in a similar fashion have been developed. One approach uses a vertical array of photocells spaced at a small increment, usually on the order of 1 cm, with the bed being distinguished by either a change in the voltage output or a circuit which can detect where the break is between exposed and buried cells (e.g., Lawler, 1992). An alternative method uses the difference in conductivity between sediments and seawater to distinguish the bed level (Ridd, 1992). The value of these instruments is that they are relatively low cost and therefore provide the potential for deployment of sufficient sensors to give reasonable spatial coverage across the surf zone.

It should be possible to detect the bed using a small echo sounder mounted on a support above the bed, though this has proved notoriously difficult when there are large amounts of sediment moving over the bed and in suspension. One adaptation of this approach is to mount the transducer on a frame with a sealed stepper motor that permits it to traverse a section of the bed, thus permitting determination of two-dimensional bedform properties and migration rates (Greenwood *et al.*, 1993). Transducers and miniature versions of sidescan sonar have been used more successfully in deeper water where sediment concentrations are much lower. Recent developments in acoustic doppler technology give a much better definition of the bed. Several versions of acoustic doppler velocity profilers (ADCPs) are available which enable the speed of currents to be detected at incremental distances from the sensor. When pointed downward, these are able to distinguish the bed more precisely than simple sonar devices, because the doppler shift is absent from sediments that are not moving (see section below on Sediment concentration, mass transport rate, and deposition for information relating to the ADCP). Small, relatively cheap acoustic sounders are available for use in air and can be used on the beach to measure changes in elevation of the bed or the water surface in wells installed to measure the water table. These devices can also be mounted on tracks to give a profile of changes in elevation during a transport event and the dimensions of any bedforms that develop.

Measurement of form and form change

Measurement of the dune, beach, and nearshore form on a scale of meters to kilometers has traditionally been done using standard survey and hydrographic techniques. Surveys out to the limit of wading have been carried out with levels and theodolites, and the use of a total station incorporating an electronic distance measurement (EDM) unit and electronic data storage is now standard. These permit rapid surveys over a range of elevations and the output is readily incorporated into a wide range of contouring and geographic information system (GIS) software packages which can produce digital elevation models and permit easy extraction of volume change through repetitive surveys (see Figure B11). In shallow water, depth has traditionally been determined using standard echo sounders mounted on a boat (Gorman *et al.*, 1998). Digital recording has now replaced the standard paper trace and positional data can be recorded simultaneously using a global positioning system (GPS). Towed arrays or acoustic multibeam transducers can be used to give simultaneous mapping of a wide swath, including information on large bedforms (Morang *et al.*, 1997). Better definition of the seafloor and three-dimensional bedform features can be attained with sidescan sonar, which utilizes a towed transducer that emits a signal at right angles to the tow direction and records returns from a swath either side of the transducer (Morang *et al.*, 1997).

The use of GPS which integrates signals from three or more satellites to determine location and elevation for a variety of surveying tasks, is now becoming standard in measuring beach form and change as it is in so many other fields. Simple systems can give positional accuracy of a few meters and elevation to about 10 m. Much greater accuracy can be obtained through the use of differential systems, which simultaneously capture the signal from the satellites and from a land-based station whose position and elevation is known precisely (see Figure B11). Moderate priced differential systems make use of Coast Guard beacons, which are set up along the coast for navigational purposes. These can give positional accuracy of ± 2–3 cm and vertical resolution of about double that, though the accuracy decreases with distance from the beacon. More expensive differential systems use a base station set up over a known position and a rover station for the actual survey. The systems can be used to measure the height and position of particular points but they can also be put in a backpack or on a vehicle allowing continuous recording of a traverse. This permits the mapping of linear features such as the waterline, thalweg of tidal creeks, bar crest, and top and bottom of cliffs, thus permitting much better delineation of these features and permitting more accurate delineation of change through repetitive surveys.

A major problem for morphodynamic experiments in the nearshore and surf zone is to obtain measurements of form change during intense storm events. While measurements of sediment transport and nearshore water motion can be obtained throughout an event, most measurements of form change have been obtained through standard surveys carried out during low wave conditions before and after the event. Some data during storms can be obtained from jetties and from specially constructed platforms that span the surf zone. However, some specialized equipment makes data collection during quite high wave conditions possible. These include various sled devices, which can have either a mast with a prism for measurement by a total station or a GPS station to enable position and elevation to be determined. The sleds may be towed by boat beyond the surf zone and winched onshore or a pulley system may be attached to an anchor seaward of the surf zone, enabling the sled to be pulled offshore without recourse to a boat. One highly specialized instrument is the CRAB used extensively at the CERC facility at Duck, North Carolina to carry out a variety of tasks in the water, including surveys in waves up to 3 m (e.g., Plant *et al.*, 1999).

Production of topographic maps from stereo pairs of aerial photographs has been a standard procedure for five decades but photo rectification and automated contouring have required expensive equipment and are rarely used for small-scale beach studies. However, new developments in video technology and digital photogrammetry are making remote measurements of form change much more practical. Video technology has been applied for more than a decade to measure waves and swash run-up (see below) but it has also been applied to measurement of the position of nearshore bars through time exposure of wave breaking (Konicki and Holman, 2000; Ruessink *et al.*, 2000; Alport *et al.*, 2001). The intensity of wave breaking is captured by creating time exposure images over a period on the order of 10 min and the resultant smooth white bands outline the zones of wave breaking on shallow bar crests and at the beach. Video cameras can also be used to monitor changes in dynamic features such as tidal inlets and associated ebb and flood tidal deltas (Morris *et al.*, 2001).

The use of digital images from still and video cameras to produce digital elevation models (DEMs) through a variety of computer software packages is of especial interest in mapping coastline changes and changes in the morphology of the beach and foredune area (Chandler, 1999). The technique makes use of overlapping pairs of photographs produced either in the traditional way through the movement of a camera installed in a plane, helicopter, or a land-based vehicle, or through the use of images taken from two fixed positions. In the case of aerial photography or moving vehicles, the position of each digital image can be linked to real time positional data provided by DGPS. Where fixed cameras are used on the beach, control points whose position and

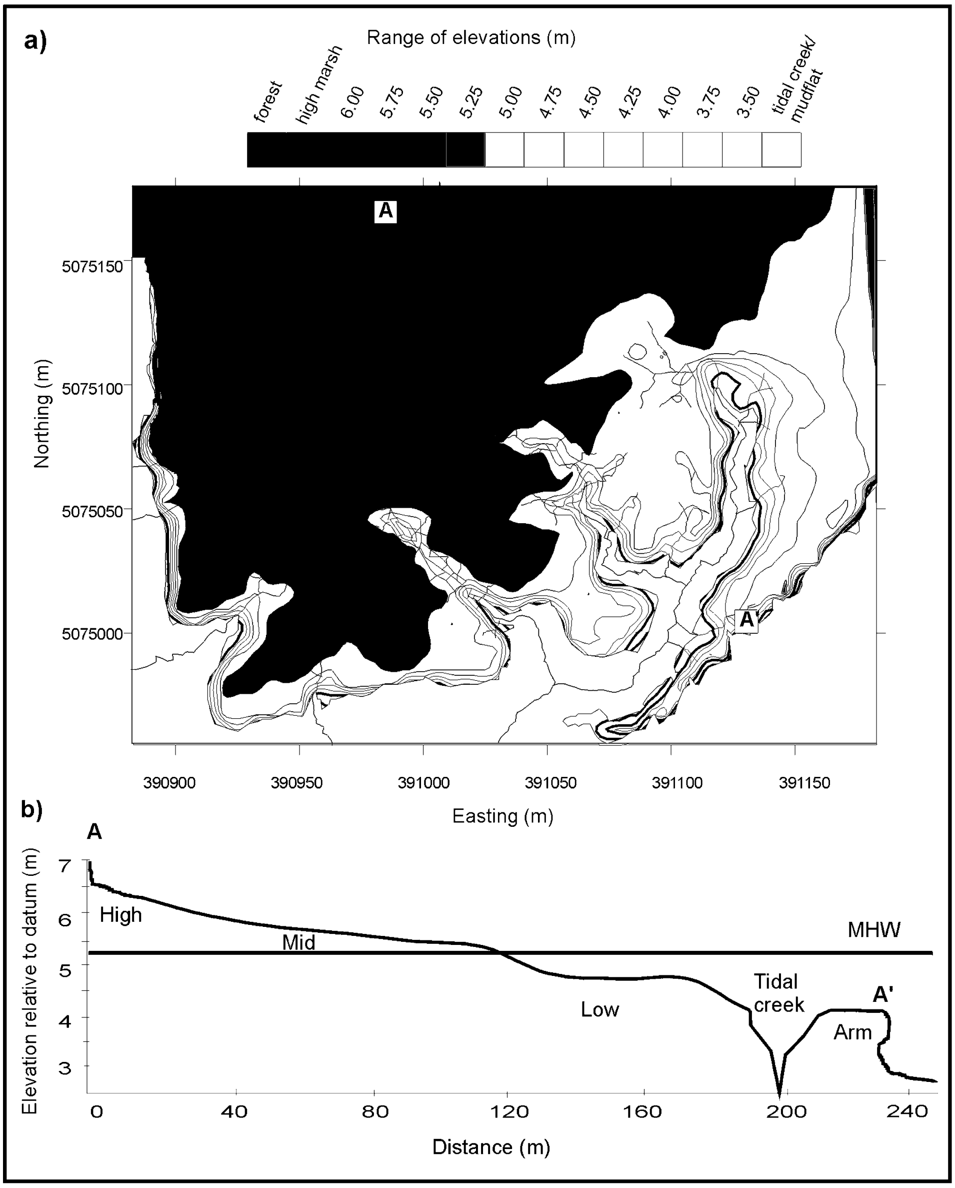

Figure B11 Digital elevation model of a saltmarsh and tidal creeks, Bay of Fundy, Canada, produced from measurements made with a total station and with a DGPS system.

elevation have been surveyed precisely are used to aid in rectification (Hancock and Willgoose, 2001). The advantage of these automated photogrammetric systems is that they can provide a very large number of data points for construction of the DEMs and much of the processing can be automated, thus allowing the evolution of topography over days, weeks, or months to be captured.

On a larger scale, the development of light detection and ranging (LIDAR) technology combined with DGPS permits topographic mapping of both the land surface and the nearshore bed to depths of 10–15 m (Irish and White, 1998; Sallenger *et al.*, 2001). The technology makes use of a laser transmitter/receiver, which transmits laser pulses toward the surface and records the traveling time of the reflected pulse. The pulse is reflected from the land surface and, over water, the return from the bottom can also be detected down to depths that depend on the degree of absorption, scattering, and refraction in the water; these in turn depend on sun angle and intensity and on the degree of turbidity in the water. The system can be deployed in a helicopter or fixed wing aircraft. Apart from the unique ability to map both the land and shallow nearshore, the technique offers a relatively low-cost method for determining topographic changes due to major storms and hurricanes (Sallenger *et al.*, 2001), and for surveying changes in areas such as salt marshes and tidal mud flats which are difficult to access with standard surveying approaches.

Winds, waves, water levels, and currents

Much of the focus in field studies of the beach and nearshore zone is on measuring the morphodynamics of sandy coasts, and to a lesser extent that of muddy coasts. These experiments require measurement of fluid and sediment dynamics over a range of timescales from fractions of a second to hours, days, and months and over spatial scales ranging from a few centimeters to hundreds of meters. Until recently, different instrumentation has been required to measure the fluid dynamics from that measuring sediment dynamics. Over the past three decades, mechanical devices for measuring fluid dynamics have been increasingly replaced by solid-state electronics involving the application of a range of direct and remote technologies. Because of the broad range of forcing variables and the spatial scales involved in determination of sediment transport, a wide range and large number of instruments is typically employed. Instrumentation typically involves measurement of wind speed and direction (both for aeolian transport on the beach and for the dynamics of the water surface), water surface elevation, wave form and direction, and water motion.

Wind speed and direction

Wind speed and direction have traditionally been measured by some form of mechanical cup or propellor-type anemometer and resistance

Figure B12 Array of cup anemometers and wind vanes mounted on towers to measure wind flow over the beach foreshore at Innisfree Beach, Ireland. Two versions of integrating sediment traps are seen on the left—the smaller traps are cylindrical traps after Leatherman (1978) and the other traps are wedge traps (Nickling and McKenna-Neuman, 1997).

wind vane mounted on a mast above the water or land surface. These give good resolution of the horizontal wind velocity at a particular elevation. Typically, in studies of aeolian transport on the beach several vertical arrays of anemometers will be deployed in order to obtain measurements of internal boundary layer development and to estimate the bed shear velocity u_* (Greely *et al.*, 1996; see Figure B12). The vertical flow can be obtained with systems of three propellor type anemometers or with two mounted at 45° angles as in the K-Gill anemometer (Atakturk and Katsaros, 1989). Sonic anemometers now offer the ability to measure fluid motion in all three dimensions, though their size is still large enough to make measurements close the bed (and thus the saltation layer) difficult. Recent modification of the Irwin sensor, a vertical pitot tube that can be mounted flush with the bed (Irwin, 1980), offers the ability to obtain direct measurements of wind stress near the bed with little disturbance to the flow.

Mean water level

Measurements of water surface elevation are collected routinely to measure changes due to tides, storm surge, and wave set-up and set-down across the breaker and surf zones. Traditional mechanical floats have now been replaced by optical and acoustic sensors installed in stilling wells. Most studies of surf zone dynamics have made use of mean values of the surface elevation measured from wave staffs or pressure transducers, though the accuracy of these measurements is on the order of +/− several centimeters. More precise measurements that can be used for investigating detailed mechanisms of nearshore circulation can be obtained through the use of manometer tubes deployed into the surf zone from the shore (Nielsen and Dunn, 1998).

Waves

Field measurements of waves in the inner nearshore and surf zones can be obtained by some form of surface piercing wave staff, which has the advantage of providing a direct measure of the wave form. These systems make use of the conduction of electricity by water, particularly seawater, and record either the change in electrical resistance or capacitance of the system as water rises and falls over a length of uninsulated cable which is part of an electrical circuit (Ribe and Russin, 1974; Timpy and Ludwick, 1985). The change in resistance or capacitance can be conditioned to produce a variation in an output signal which may be a DC current or a frequency. The sensor itself may be fixed to a support jetted into the sand in the nearshore (see Figure B13) or attached to some physical structure such as a jetty or platform. The advantage of the surface staff is that it provides a direct measure of the water surface form and they have been used extensively in many studies, particularly in fetch limited areas where installation can be accomplished during calm conditions (Davidson-Arnott and Randall, 1984; Greenwood and Sherman, 1984).

The disadvantage of wave staffs is that they are subject to high wave forces when deployed in shallow water and they have largely been replaced with some form of pressure transducer housed in a watertight case. These can be deployed some distance below the surface, or on the bed and they are often colocated with other sensors such as electromagnetic current meters and nepelometers (see below). They sense the change in pressure associated with the passage of individual waves. The pressure variation with depth can be predicted from wave theory and thus it is possible to develop a transform function that will relate the recorded variations in pressure to the surface wave form (Lee and Wang, 1984). Since there is usually a spectrum of frequencies present in the pressure transducer record, the transform should be performed for all of the frequencies present. This is not a trivial task, though it can be done routinely in a data analysis program. There is some loss of information on the true form of the surface wave as well as the loss of the higher frequencies with increasing depth of deployment, but this is offset in studies in and close to the breaker and surf zones by the ease of deployment and the reduced exposure to breaking wave forces.

The water surface can also be measured remotely using a video camera to measure the change in surface elevation against a graduated pole or screen. This gives a good measure of the wave form without interference and it enables determination of whether the wave is broken or not. The record can be digitized manually or a computer software algorithm can be used to extract the position of the surface automatically. Video cameras have also been used extensively to extract data on run-up frequencies on the beach (Holman and Sallenger, 1985). Recent developments in LIDAR technology may also permit application to measuring waves (Irish *et al.*, 2001).

Individual wave staffs or pressure transducers provide a picture of the variations in water surface elevation through time—that is, they give

Figure B13 Resistance wave staffs (left) deployed over an intertidal ridge and runnel, Nova Scotia, Canada. A grid of depth of disturbance rods is deployed across the bar (center) and frames supporting OBS and electromagnetic current meters can be seen in the far right.

information on wave height and period but not on the direction of travel. This requires either the deployment of several instruments in an array, which permits determination of the wave direction through a comparison of travel time between various sensors (Bodge and Dean, 1984; Howell, 1992), or the measurement of both the horizontal and vertical components of water motion using a pressure transducer and bidirectional electromagnetic current meter or an acoustic doppler current meter (see following section).

Because of the rapid oscillatory motion associated with wave action in shallow water, mechanical current meters are generally not useful in studies of fluid processes in the inner nearshore and surf zone, though miniature-ducted impellor current meters have proved useful in some locations (Wright *et al*., 1982; Masselink and Hegge, 1995). Measurements of hydrodynamics in the nearshore and surf zone were revolutionized in the 1970s by the development of electromagnetic current meters (EMCMs) and they have been used extensively in almost all field experiments as well as in the laboratory (Huntley and Bowen, 1975; Cushing, 1976). Examples of their use can be found in numerous experiments including the Nearshore Sediment Transport Study (NSTS, Seymour, 1989), the Canadian Coastal Sediment Study ($C^2 S^2$; Willis, 1987) and in the various experiments carried out at the CERC at Duck, North Carolina (Birkemeier *et al*., 1997). The current meters produce a fluctuating magnetic field around the sensor head and measure the voltage generated by fluid flow in the field using Farraday's law. Typically, the instruments have four sensors mounted orthogonally so as to detect flow along two orthogonal axes. The current meter is usually mounted so as to detect horizontal flow, but it is possible to mount it with one axis vertically. The sensors have a fast response time, permitting sampling at frequencies >5 Hz and are able to detect very small mean flows in a highly fluctuating environment (see Figure B14). The majority of EMCMs used in the field have been made by Marsh-McBirney Inc. of Maryland, USA. Large models have a 10.5 cm diameter head and the small ones, which have been used extensively in the surf zone have a 4 cm head. These current meters have been evaluated extensively (Aubrey and Trowbridge, 1985, 1988; Guza, 1988) and their widespread use allows for ease of comparison between different studies. While EMCMs are now being replaced by various forms of acoustical instrument, they are still useful in the breaker and surf zone because of their smaller sensitivity to the presence of air bubbles.

In the past decade, various forms of acoustical doppler instruments have been developed which have been used in laboratory and field experiments. Essentially, they emit an acoustic signal which is reflected by fine material in the water column from a focal point and received by orthogonally mounted transducers. The relative motion in each axis is then determined by the doppler shift of the signal. The acoustic doppler velocimeter (ADV) is the simplest of the instruments and measures velocity at a single point on the order of a few centimeters from the emitting transducer.

Sediment concentration, mass transport rate, and deposition

Obtaining measurements of sediment transport is the third key element in morphodynamic experiments in the coastal zone. There are a large number of instruments available and a variety of approaches have been taken, but much work remains to be done to obtain reliable measurements over a reasonable spatial and temporal scale (White, 1998). Estimates of net longshore transport over periods of weeks, months, or years at a location can be obtained from measurements of the amount trapped at a total barrier, either over a short period at a purpose-built groin (Wang and Kraus, 1999) or at a large jetty. A number of studies have also used fluorescent or radioactive tracers for measurement of longshore sediment transport or of transport pathways within the surf zone. However, the focus here is on instrumentation for instantaneous measurement of the transport rate either directly, or indirectly through the combination of measurements of sediment concentration and net water motion. Direct measurement techniques include various traps and acoustic doppler instruments that measure concentration and velocity simultaneously through some portion of the water column. Indirect techniques for measuring concentration include optical and conductivity devices.

Figure B14 Electromagnetic current meter (left) and three OBS probes mounted on a "goalpost" in the intertidal zone, Skallingen, Denmark. Electronics for the current meter are installed in the waterproof housing secured to a pole jetted into the sand to the left of the goalpost. A profile line of large depth of disturbance rods is visible at the right.

Traps and pumps

A number of devices have been used in attempts to trap sediment suspended in the water column of the net transport in the swash or surf zone, but the oscillatory motion associated with wave action makes this task much more difficult than, for example, under unidirectional flow in a river. Pump samplers have been used with varying degrees of success and various bottles for capturing the suspended sediment load. These all require considerable effort and the logistical difficulties coupled with doubts as to the accuracy of the sampling process have limited their further use. Recently, arrays of streamer traps consisting of long bags of fine mesh fixed to a rigid rectangular inlet have been used to measure transport where there is a net current present. These traps are able to capture large amounts of sediment, but it is still not clear that they can provide reliable estimates of the net transport or that they provide accurate results over a range of conditions.

Optical devices

Optical devices emit light and give a measure of the sediment concentration in the water column at a point either through the degree of attenuation of the light beam or through the amount of light reflected from particles in suspension. They therefore do not provide a direct measure of sediment transport and thus must be collocated with a device such as an EMCM which measures the fluid flow. The light transmitted is generally in a narrow wave band in the infrared range in order to minimize the effects of natural light in the water.

Transmissometers measure the degree of attenuation of the light over a fixed distance separating the emitter from the receiver. They tend to be relatively bulky instruments best suited for work some distance seaward of the breaker zone in depths greater than 10 m where suspended sediment concentrations are relatively low. A Sea Tech transmissometer with a 5 cm path length was developed for use in shallow water (Huntley, 1983) but the much smaller probes associated with instruments measuring reflected light proved more suitable for the inner nearshore and surf zones.

Much of our understanding of the dynamics of suspended sediment transport in the nearshore over the past three decades has come from the use of the optical backscatterance sensor (OBS) originally developed at the University of Washington (Downing *et al.*, 1981) and now produced commercially by D & A instruments. The OBS is a miniature nephelometer which measures the backscatterance of light by sediments suspended in the fluid. It utilizes a narrow infrared beam which has the advantage of minimizing interference by sunlight and confining the sampling volume to a short distance from the probe. The sensor is compact (2.1 cm diameter) with transmitter and receiver mounted next to each other at the end of the probe, thus minimizing flow interference and enabling sensors to be mounted in close proximity to other probes or to electromagnetic current meters with which they are often collocated (see Figures B14, B15(a)). The OBS probe is very rugged, enabling it to be deployed in areas of strong currents and breaking wave impacts, and it is clearly superior to other nephelometers for work in the nearshore marine environment (Greenwood *et al.*, 1990). They are designed to measure suspended sediment concentrations in areas where concentrations may be high and/or may vary rapidly over short time periods (i.e., on the order of 0.25 Hz). They have been used in a wide range of marine environments, including the shoreface and continental shelf (Wright *et al.*, 1991; Kineke and Sternberg, 1992), the breaker and surf zones (Black and Rosenberg, 1994), and estuaries (Kineke *et al.*, 1991).

Provided they are not deployed too close to the bed (interference with the bed itself) or too close to the surface (effects of ambient light) both transmissometers and OBS probes work well. They are linear over a wide range of grain sizes from clay to sand and there has been extensive testing and calibration of both types of instruments in the laboratory and field (Downing and Beach, 1989; Osborne *et al.*, 1993; Greenwood and Jagger, 1995; Bunt *et al.*, 1999; Sutherland *et al.*, 2000). However, they perform best with a narrow range of grain size and calibration, where there is a wide range of grain size they are subject to considerable error (Bunt *et al.*, 1999).

Routine field calibration of the instruments is difficult and laboratory calibration requires quite complex facilities and involves the difficulties of obtaining representative samples of suspended sediment to be returned to the lab for testing. Recent testing of a laser *in situ* scattering and transmissometry (LISST) instrument produced by Sequoia Scientific (Traykovski *et al.*, 1999; Gartner *et al.*, 2001; Mikkelsen and Pejrup, 2001) offers the potential to measure the complete particle size distribution and concentration simultaneously. Initially, this may offer a means of calibrating cheaper, less bulky sensors but further developments may lead to smaller versions which could be deployed close to the bed.

Acoustic Doppler velocity profilers

Acoustical instruments offer the possibility of measuring both particle concentration and velocity simultaneously and over some appreciable portion of the water column, and thus providing a direct measure of the transport rate. Acoustic Doppler velocity profilers (ADVP) use the same basic technology as the ADV described above. However, they measure the return signal in very small increments of time, thus allowing the determination of velocity and concentration in discrete "bins." They can be used from a boat with position fixed by DGPS and can thus give a complete picture of flow over bedforms in estuaries and tidal channels (Best *et al.*, 2001).

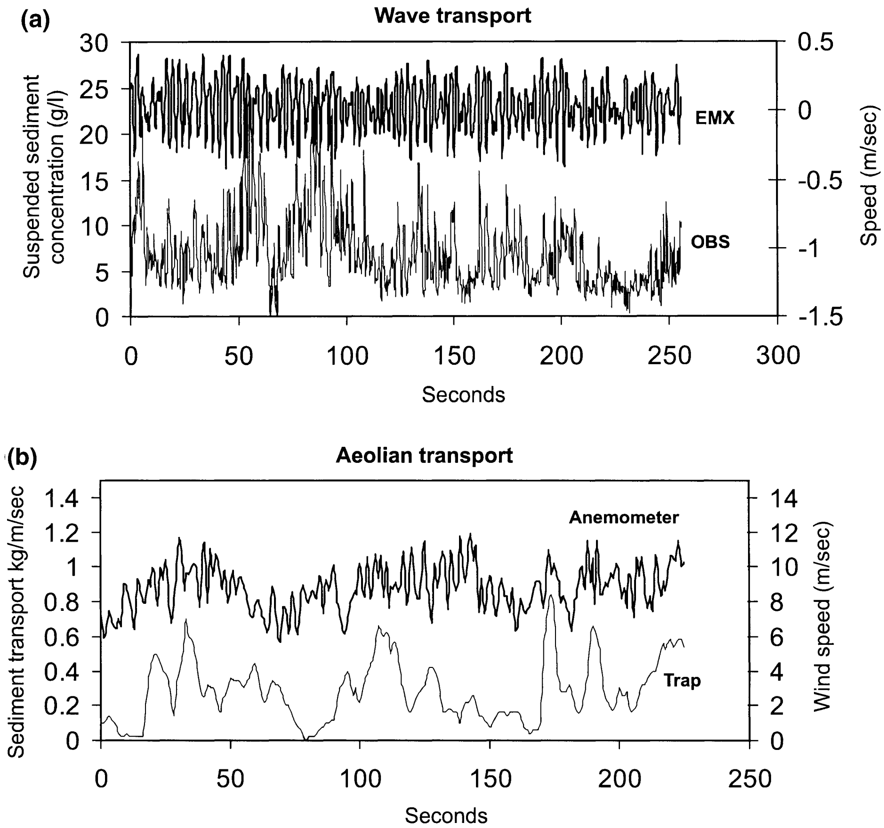

Figure B15 Comparison of flow and sediment transport in the nearshore and on the beach. (a) On-offshore flow speed measured with an electromagnetic current meter and suspended sediment concentration measured with an OBS probe at an elevation of 0.15 m above the bed on the crest of an intertidal sand bar (see Figure B13). The instruments were sampled at 4 Hz. (b) Horizontal wind speed measured with a cup anemometer and sediment transport rate measured with a wedge trap fitted with an electronic balance. The instruments were sampled at 1 Hz and smoothed with a 5 s running mean.

Aeolian sand transport

The design of field instrumentation for measuring sand transport by wind has tended to lag behind that available for measuring transport in the water. Sediment transport from the beach over a period of days or weeks can be measured indirectly by measuring accumulation in vegetated sand dunes by profiling or by the use of a bedframe device (Davidson-Arnott and Law, 1990, 1996).

Most direct measurements of sediment transport have been made with simple vertical traps which are oriented into the wind and which allow the sand captured to collect in the base. The sediment collected over a period of time is weighed to give an average transport rate for the collection period. A major problem is to design the trap so that it is isokinetic, otherwise sand in transport is diverted away from the trap opening by the pressure buildup. Simple vertical traps, which have been used widely may have an efficiency <30%. Wedge-shaped traps have improved aerodynamics and are likely much closer to isokinetic (Nickling and McKenna-Neuman, 1997—see Figure B12). However, these traps are more sensitive to changes in wind direction and will undersample when the wind angle exceeds 5°; thus they can only be used for periods of 15–30 min without attention. A variety of other trap designs are available (Goossen *et al.*, 2000) but all have a number of problems with accuracy.

Horizontal traps offer the opportunity to sample all of the transport load and provide a means for calibrating vertical traps. They require a large pit several meters across and again will integrate total transport over periods of tens of minutes to hours (Greely *et al.*, 1996). Use of a wet horizontal trap can reduce some of the logistics (Wang and Kraus, 1999) because the trap need only be a few centimeters deep.

A number of trap designs are now being used to obtain measurements of the instantaneous mass transport rate, thus permitting comparison of the transport rate with measurements of the wind flow. The trap design of Nickling and McKenna-Neuman has been modified to incorporate a continuous weighing electronic balance (Mckenna-Neuman *et al.*, 2000; see Figure B15(b)). Bauer and Namikas (1998) used the same trap but designed a combination tipping bucket and strain gage to weigh the sand collected over long time periods. The design of Jackson (1996) uses a similar weighing mechanism to that used by Bauer and Namikas, but the trap itself is a circular collection funnel that is mounted flush with the surface. This avoids the problem of isokinetic sampling associated with vertical traps and has the advantage of omni directional collection. However, it measures flux to the surface rather than the total transport rate.

Impact measurement

The drawbacks of trap designs and the need for high speed, continuous sampling of the transport rate have led to the development of several instruments that measure the impact of saltating grains and then attempt to calibrate this to the transport rate. The saltiphone (Arens, 1996) uses a microphone to record impacts and the intensity is then recorded as a voltage signal. While this gives a measure of the relative transport rate, it has proved difficult to calibrate and is sensitive to variations in grain size. The SENSIT (Stockton and Gillette, 1990) responds to the impact of saltating grains on a piezoelectric crystal and counts the number of impacts per second. It has also proved difficult to calibrate to give a measure of the mass transport rate and seems to offer only a relative measure of the grains in saltation. Because of the small sampling area a vertical array of the sensors must be deployed in order to measure the total mass transport rate.

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Bibliography

Alport, M.J., Basson, J., Mocke, G., Naicker, J., and Saltau, C., 2001. Discrimination and analysis of video imaged shorelines and nearshore processes. In *Proceedings Coastal Dynamics '01*. American Society of Civil Engineers, New York, pp. 989–997.

Amos, C.L., Daborn, G.R., Christian, H.A., Atkinson, A., and Robertson, A., 1992. In situ erosion measurements on fine-grained sediments from the Bay of Fundy. *Marine Geology*, **108**: 175–196.

Arens, B., 1996. Rates of aeolian sand transport on a beach in a humid temperate climate. *Geomorphology*, **7**: 3–18

Askin, R.W., and Davidson-Arnott, R.G.D., 1981. Micro erosion meter modified for use underwater. *Marine Geology*, **40**: M45–M48.

Atakturk, S.S., and Katsaros, K.B., 1989. The K-Gill: a twin propellor-vane anemometer for measurements of atmospheric turbulence. *Journal of Atmospheric and Oceanic Technology*, **6**: 509–515.

Aubrey, D.G., and Trowbridge, J.H., 1985. Kinematic and dynamic estimates from electromagnetic current meter data. *Journal of Geophysical Research*, **90**(C5): 9137–9146.

Aubrey, D.G., and Trowbridge, J.H., 1988. Reply (to comments by Guza, 1988). *Journal of Geophysical Research*, **93**(C2): 1344–1346.

Bauer, B.O., and Namikas, S.L., 1998. Design and field test of a continuously weighing tipping-bucket assembly for aeolian sand traps. *Earth Surface Processes and Landforms,* **23**: 1171–1183.

Best, J.L., Kostaschuk, R.A., and Villard, P., 2001. Quantitative visualization of flow fields associated with alluvial sand dunes. *Journal of Flow Visualization*, **4**: 373–381.

Birkemeier, W.A., Long, C.E., and Hathaway, K.K., 1997. DELILAH, DUCK94 and Sandy Duck: three nearshore field experiments. In *Proceedings Coastal Engineering 1996*. American Society of Civil Engineers, New York, pp. 4052–4065.

Black, K.P., and Rosenberg, M.A., 1994. Suspended sand measurements in a turbulent environment: field comparison of optical and pump sampling techniques. *Coastal Engineering*, **24**: 137–150.

Bodge, K.R., and Dean, R.G., 1984. Wave measurement with differential pressure gages. In *Proceedings Coastal Engineering 1984*. American Society of Civil Engineers, pp. 755–769.

Bunt, J.A.C., Larcombe, P., and Jago, C.F., 1999. Quantifying the response of optical backscatter devices and transmissometers to variations in suspended particulate matter. *Continental Shelf Research*, **19**: 1199–1220.

Cahoon, D.R., French, J.R., Spencer, T., Reed, D., and Moller, I., 2000. Vertical accretion versus elevational adjustment in UK saltmarshes: an evaluation of alternative methodologies. In Pye, K., and Allen, J.R.L. (eds.), *Coastal and Estuarine Environments. Journal of the Geological Society*, Special Publication **175**: 223–238.

Chandler, J., 1999. Effective application of automated digital photogrammetry for geomorphological research. *Earth Surface Processes and Landforms*, **24**: 51–63.

Cushing, V., 1976. Electromagnetic water current meter. In *Proceedings of Oceans '76. Proceedings Coastal Engineering 1992*. Institute of Electrical and Electronic Engineers, pp. 298–301.

Davidson-Arnott, R.G.D., and Langham, D.R.J., 2000. The effects of softening on nearshore erosion of a cohesive shoreline. *Marine Geology*, **166**: 145–162.

Davidson-Arnott, R.G.D., and Law, M.N., 1990. Seasonal patterns and controls on sediment supply to coastal foredunes, Long Point, Lake Erie. In Nordstrom, K.F., Psuty, N.P., and Carter, R.W.G. (eds.), *Coastal Dunes: Form and Process*. Chichester: John Wiley & Sons, pp. 177–200.

Davidson-Arnott, R.G.D., and Law, M.N., 1996. Measurement and prediction of long-term sediment supply to coastal foredunes. *Journal of Coastal Research*, **12**: 654–663.

Davidson-Arnott, R.G.D., and Randall, D.C., 1984. Spatial and temporal variations in spectra of storm waves across a barred nearshore. *Marine Geology*, **60**: 15–30.

Downing, J.P., and Beach, R.A., 1989. Laboratory apparatus for calibrating optical suspended solids sensors. *Marine Geology*, **86**: 243–249.

Downing, J.P., Sternberg, R.W., and Lister, C.R.B., 1981. New instrumentation for the investigation of sediment suspension processes in the shallow marine environment. *Marine Geology*, **42**: 19–34.

Gartner, J.W., Cheng, R.T., Wang, P.-F., and Richter, K., 2001. Laboratory and field evaluation of the LISST-100 instrument for suspended particle size determinations. *Marine Geology*, **175**: 199–219.

Goossens, D., Offer, Z., and London, G., 2000. Wind tunnel and field calibration of five aeolian sand traps. *Geomorphology*, **35**: 233–252.

Gorman, L., Morang, A., and Larson, R., 1998. Monitoring the coastal environment; Part IV: mapping, shoreline changes, and bathymetric analysis. *Journal of Coastal Research*, **14**: 61–92.

Greeley, R., Blumberg, D.G., and Williams, S.H. 1996. Field measurements of the flux and speed of wind-blown sand. *Sedimentology*, **43**: 41–52.

Greenwood, B., and Jagger, K., 1995. Sensitivity of optical sensors to grain size variations in the sand mode: implications for transport measurements. In *Proceedings Canadian Coastal Conference*. Canadian Coastal Science and Engineering Association, pp. 383–398.

Greenwood, B., and Mittler, P.R., 1984. Sediment flux and equilibrium slopes in a barred nearshore. *Marine Geology*, **60**: 79–98.

Greenwood, B., and Sherman, D.J., 1984. Waves, currents sediment flux and morphological response in a barred nearshore. *Marine Geology*, **60**: 31–61.

Greenwood, B., Hale, P.B., and Mittler, P.R., 1979. Sediment flux determination in the nearshore zone. In *Proceedings Workshop on Instrumentation for Currents and Sediments in the Nearshore Zone*. National Research Council of Canada, pp. 99–115.

Greenwood, B., Osborne, P.D., Bowen, A.J., Hazen, D.G., and Hay, A.E., 1990. C-COAST: The Canadian Coastal Sediment Transport Programme—suspended sediment transport in the nearshore zone. In *Proceedings Canadian Coastal Conference*. National Research Council of Canada, pp. 319–336.

Greenwood, B., Richards, R.G., and Brander, R.W., 1993. Acoustic imaging of sea-bed geometry: a high resolution remote tracking sonar (HERTSII). *Marine Geology*, **112**: 207–218.

Guza, R.T., 1988. Comments on "Kinematic and dynamic estimates from electromagnetic current meter data" by D.G. Aubrey and J.H. Trowbridge. *Journal of Geophysical Research*, **93**(C2): 1337–1343.

Hancock, G., and Willgoose, G., 2001. The production of digital elevation models for experimental model landscapes. *Earth Surface Processes and Landforms*, **26**: 475–490.

Holman, R.A., and Sallenger, A.H., Jr., 1985. Set up and swash on a natural beach. *Journal of Geophysical Research*, **90**: 945–953.

Houwing, E.-J., 1999. Determination of the critical erosion threshold of cohesive sediments on intertidal mudflats along the Dutch Wadden Sea coast. *Estuarine, Coastal and Shelf Science*, **49**: 345–355.

Howell, G.L., 1992. A new nearshore directional wave gage. In *Proceedings Coastal Engineering 1992*. American Society of Civil Engineers, pp. 295–307.

Huntley, D.A., 1983. In situ sediment monitoring techniques: a survey of the state of the art in the USA. In *Proceedings Canadian Coastal Conference*. National Research Council of Canada, pp. 151–165.

Huntley, D.A., and Bowen, A.J., 1975. Comparison of the hydrodynamics of steep and shallow beaches. In Hails, J., and Carr, A. (eds.), *Nearshore Sediment Dynamics and Sedimentation*. London: John Wiley & Sons, pp. 69–110.

Irish, J.L., and White, T.E., 1998. Coastal engineering applications of high resolution lidar bathymetry. *Coastal Engineering*, **35**: 47–71.

Irish, J.L., Wozencraft, J.M., and Cunningham, A.G., 2001. Water wave measurement with lidar from a fixed platform. In *Proceedings Coastal Dynamics '01*. American Society of Civil Engineers, pp. 998–1006.

Irwin, H.P.A.H., 1980. A simple omnidirectional sensor for wind-tunnel studies of pedestrian level winds. *Journal of Wind Engineering and Industrial Aerodynamics*, **7**: 219–239.

Jackson, D.W.T., 1996. A new, instantaneous aeolian sand trap design for field use. *Sedimentology*, **43**: 791–796.

Kineke, G.C., and Sternberg, R.W., 1992. Measurements of high concentration suspended sediments using the optical backscatterance sensor. *Marine Geology*, **108**: 253–258.

Kineke, G.C., Sternberg, R.W., Cacchione, D.A., Krank, K., and Drake, D.E, 1991. Distribution and characteristics of suspended sediment on the Amazon shelf. *Oceanography*, **4**: 21–26.

Kirk, R.M., 1977. Rates and forms of erosion on intertidal platforms at Kaikoura Peninsula, South Island, New Zealand. *New Zealand Journal of Geology and Geophysics*, **20**: 571–613.

Konicki, K.M., and Holman, R.A., 2000. The statistics and kinematics of transverse sand bars on an open coast. *Marine Geology*, **169**: 69–101.

Lawler, D.M., 1992. Design and installation of a novel automatic erosion monitoring system. *Earth Surface Processes and Landforms*, **17**: 455–463.

Leatherman, S.P., 1978. A new aeolian sand trap design. Sedimentology, **25**: 303–306.

Lee, D.-Y., and Wang, H., 1984. Measurement of surface waves from subsurface gage. In *Proceedings Coastal Engineering*. American Society of Civil Engineers, pp. 271–286.

Maa, J.P.-Y., Wright, L.D., Lee, C.-H., and Shannon, T.W., 1993. VIMS sea carousel: a field instrument for studying sediment transport. *Marine Geology*, **115**: 271–287.

Masselink, G., and Hegge, B., 1995. Morphodynamics of meso- and macrotidal beaches: examples from central Queensland, Australia. *Marine Geology*, **129**: 1–23.

McKenna Neuman, C., Lancaster, N., and Nickling, W.G., 2000. Effect of unsteady winds on sediment transport intermittency along the stoss slope of a reversing dune. *Sedimentology*, **47**: 211–226.

Mikkelsen, O.A., and Pejrup, M., 2001. The use of a LISST-100 laser particle sizer for in-situ estimates of floc size, density and settling velocity. *Geo-Marine Letters*, **20**: 187–195.

Morang, A., Larson, R., and Gorman, L., 1997a. Monitoring the coastal environment; Part 1: waves and currents. *Journal of Coastal Research*, **13**: 111–133.

Morang, A., Larson, R., and Gorman, L., 1997. Monitoring the coastal environment; Part 111: geophysical and research methods. *Journal of Coastal Research*, **13**: 1064–1085.
Morris, B.A., Davidson, M.A., and Huntley, D.A., 2001. Measurements of the response of a coastal inlet using video moniotoring techniques. *Marine Geology*, **175**: 251–272.
Nickling, W.G., and McKenna Neuman, C., 1997. Wind tunnel evaluation of a wedge-shaped aeolian transport trap. *Geomorphology*, **18**: 333–345.
Nielsen, P., and Dunn, S.L., 1998. Manometer tubes for coastal hydrodynamics investigations. *Coastal Engineering*, **35**: 73–84.
Osborne, P.D., Vincent, C.E., and Greenwood, B., 1993. Measurement of suspended sediment concentrations in the nearshore: intercomparison of optical and acoustic backscatter sensors. *Continental Shelf Research*, **14**: 159–174.
Plant, N.G., Holman, R.A., Freilich, M.H., and Birkemeir, W.A., 1999. A simple model for interannual bar behaviour. *Journal of Geophysical Research*, **104**(C7): 15,755–15,776.
Ribe, R.L., and Russin, E.M., 1974. Ocean wave measuring instrumentation. In *Proceedings International Symposium on Ocean Wave Measurement and Analysis*. American Society of Civil Engineers, pp. 396–416.
Ridd, P.V., 1992. A sediment level sensor for erosion and siltation detection. *Estuarine, Coastal and Shelf Science*, **35**: 355–362.
Ruessink, B.G., van Enckvort, I.M.J., Kingston, K.S., and Davidson, M.A., 2000. Analysis of two- and three-dimensional nearshore bar behaviour. *Marine Geology*, **169**: 161–183.
Sallenger, A.H., Krabill, W., Swift, R., and Brock, J., 2001. Quantifying hurricane-induced coastal changes using topographic lidar. In *Proceedings Coastal Dynamics '01*. American Society of Civil Engineers, pp. 1007–1018.
Seymour, R.J. (ed.), 1989. Nearshore Sediment Transport. New York: Plenum.
Stockton, P., and Gillette, D.A., 1990. Field measurements of the sheltering effect of vegetation on erodible land surfaces. *Land Degradation and Rehabilitation*, **2**: 77–85.
Sunamura, T., 1992. *The Geomorphology of Rocky Coasts*. Chichester: John Wiley & Sons.
Sutherland, T.F., Lane, P.M., Amos, C.L., and Downing, J., 2000. The calibration of optical backscatter sensors for suspended sediment of varying darkness levels. *Marine Geology*, **162**: 587–597.
Timpy, D.L., and Ludwick, J.C., 1985. Bore height measurement with improved wave staff. *Journal of Waterways, Port, Coastal and Ocean Engineering*, **111**: 495–510.
Tolhurst, T.J., Black, K.S., Shayler, S.A., Mather, S., Black, I., Baker, K., and Paterson, D.M., 1999. Measuring the *in situ* erosion shear stress of intertidal sediments with the cohesive strength M (CSM). *Estuarine, Coastal and Shelf Science*, **49**: 281–294.
Traykovski, P., Latter, R.J., and Irish, J.D., 1999. A laboratory evaluation of the laser in situ scattering and transmissometry instrument using natural sediments. *Marine Geology*, **159**: 355–367.
Trudgill, S.T., High, C.J., and Hanna, F.K., 1981. Improvements to the micro erosion meter. British Geomorphology Research Group, *Technical Bulletin* **29**, p. 17.
Viles, H.A., and Trudgill, S.T., 1984. Long term remeasurements of micro-erosion meter rates, Aldabra Atoll, Indian Ocean. *Earth Surface Processes and Landforms*, **9**: 89–94.
Wang, P., and Kraus, N., 1999. Horizontal water trap for measurement of aeolian sand transport. *Earth Surface Processes and Landforms*, **24**: 65–70.
White, T.E., 1998. Status of measurement techniques for coastal sediment transport. *Coastal Engineering*, **35**: 17–45.
Willis, D.H., 1987. The Canadian coastal sediment study: an overview. In *Coastal Sediments '87*. American Society of Civil Engineers, pp. 682–693.
Wright, L.D., Nielsen, P., Short, A.D., and Green, M.O., 1982. Morphodynamics of a macrotidal beach. *Marine Geology*, **50**: 97–128.
Wright, L.D., Boon, J.D., Kim, S.C., and List, J.H., 1991. Modes of cross-shore sediment transport on the shorefaceof the Middle Atlantic Bight. *Marine Geology*, **96**: 19–51.

Cross-references

Airborne Laser Terrain Mapping
Erosion Processes
Geographic Information Systems
Global Positioning Systems
Monitoring, Coastal Geomorphology
Muddy Coasts
Photogrammetry
Sandy Coasts

BEACH CUSPS—See RHYTHMIC PATTERNS

BEACH DRAIN

Introduction

For over half a century, reporters have suggested a link between the elevation of beach groundwater and erosional or accretional trends of the beach face. Beach dewatering (the artificial lowering of the water table within beaches by a system of drains and pumps) is suggested by its proponents as a practical alternative to more traditional methods of coast-stabilization. Within the last 15–20 years several tests have been installed, and to date seven to eight commercial dewatering systems have operated. The following is a review of the origins and development of the dewatering concept from early work on beach face permeability and beach groundwater dynamics, to recent field and laboratory studies that have explicitly examined the effect of artificial groundwater manipulation on beach face accretion and erosion.

The origin of the beach drain

The beach drain (Figure B16) is not a new concept, but was revived in the last 20 years due to commercial interests (Turner and Leatherman, 1997).

The origins of the beach drain concept can be traced back 50 years to early work in two parallel fields of coastal research: the role of beach face permeability in controlling erosion or accretion (e.g., Bagnold, 1940); and the tidal dynamics of beach groundwater (e.g., Grant, 1948). The installation within the last 10 years of prototype beach dewatering systems in Europe (Vesterby, 1994) and the United States (Lenz, 1994) signified the transition of the beach dewatering concept from the hypothetical to the practical. The potential use of beach drain technology is beginning to be noted within the mainstream coastal engineering community (e.g., Abbott and Price, 1994, pp. 334–336), and in the last five years a limited number of journal articles (e.g., Weisman *et al.*, 1995; Li *et al.*, 1995) and more frequent papers presented at the coastal engineering conferences (e.g., Davis and Hanslow, 1991; Ogden and Weisman, 1991; Davis *et al.*, 1992, 1993; Oh and Dean, 1994) have served to raise the awareness of the beach dewatering concept. However, if beach dewatering technology is to meet the promise that its proponents claim, the answers to a number of fundamental questions must be addressed.

Counter to the impression that may be gained from publications and other materials produced by commercial players in the beach dewatering industry, the underlying physical mechanisms that may contribute to the success of the beach drain concept are *not* yet fully elucidated.

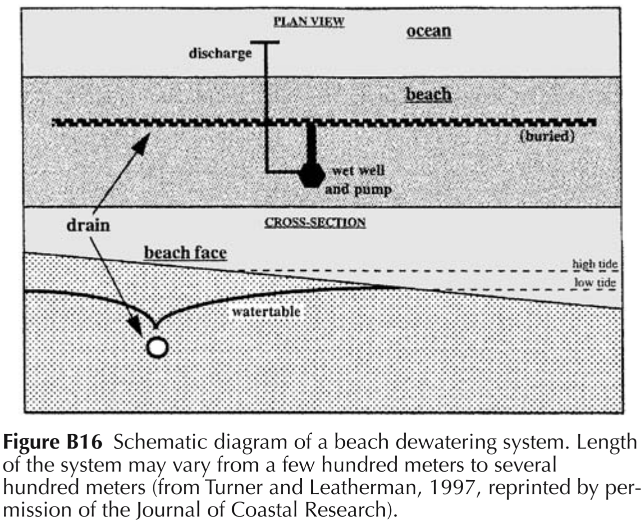

Figure B16 Schematic diagram of a beach dewatering system. Length of the system may vary from a few hundred meters to several hundred meters (from Turner and Leatherman, 1997, reprinted by permission of the Journal of Coastal Research).

Dewatering is a well-established practice in the excavation industry, but the inclusion of a highly dynamic land–ocean boundary where sediment motion is a function of both static inter-granular forces, and surf and swash zone hydrodynamics, makes the description of transport mechanisms across a dewatered beach face unique. On the more practical side to many coastal scientists and engineers, the field evidence from operating dewatered sites remains inconclusive. A comprehensive and independent assessment of the mid-to-long-term operation of a prototype installation is yet to be reported in the scientific literature, and until such a study is completed it is unlikely that the prevailing mood of healthy skepticism (e.g., Bruun, 1989) will be either validated or changed.

History and development

Bagnold's seminal laboratory investigations (Emery and Foster, 1948) undertook the first published study describing the dynamics of the water table in sandy beaches. They referred to prior unpublished work of Zinn (1942).

Several laboratory and field tests are described by Turner and Leatherman (1997) including Bagnold (1940), and Emery and Foster (1948). Ogden and Weisman (1991) undertook two-dimensional (2-D) tests using irregular waves ranging from erosive to accretive and concluded that for the range of conditions tested, the beach drain had no significant effect on the rate of erosion or accretion at the still water line, but did promote berm development and hence overall beach face steepening. A more recent study by the same researchers (Weisman *et al.*, 1995) examined the effectiveness of beach dewatering under the influence of the tides, and concluded that water table lowering maintains its effectiveness in promoting berm growth and beach face steepening for both tidal and nontidal cases. Heaton (1992) undertook a series of single and multiple wave experiments, and quantified a general trend that increasing water table elevation resulted in an increasing volume of sediment eroded from the beach face. Oh and Dean (1994) reported a set of three experiments where the water table was alternatively elevated, lowered, and equal to mean sea level, and concluded that an elevated water table resulted in the overall destabilization and erosion of previously marginally stable regions of the beach face. A simple seepage model (Oh and Dean, 1994) demonstrated that outflow across the beach face may act to reduce the effective weight (and hence stability) of surficial sediment.

Considerable activity has taken place in Australia. Davis *et al.* (1992, 1993) made field tests and found that for fair conditions the drain increases beach stability while storm conditions had the opposite effects. Nielsen (1990, 1992) did considerable testing including the infiltration effects on sediment mobility under recorded stabilizing as well as destabilizing forces and their relation to fluidization. Nielson's latest (2001) is an attempt to determine infiltration on effects on sediment mobility. Many field tests are described by Turner and Leatherman (1997) providing results or conclusions from each separate test programs.

First full-scale test—Thorsminde, Denmark

A test site at Hirtshals was not considered a success by the Danish Geotechnical Institute and was subsequently dismantled, but the results from the Hirtshals West site were deemed encouraging, and it was decided to undertake the first large-scale test of the dewatering concept at Thorsminde on the west coast of Denmark. Hansen (1986) provides details of the beach and installation, which are summarized to varying degrees by Ovesen and Schuldt (1992) and Vesterby (1991, 1994). The test site is located on the exposed North Sea coast, where the shoreline fluctuated seasonally by ± 15 m with a reported average erosion rate of 2–4 m/year.

The conclusion after year of operation was:

1. The usual seasonal fluctuation in shoreline position was halted and net recession ceased.
2. The southern drained region prograded seaward approximately 10 m and stabilized at a distance of 20–25 m in front of the drain line, while the northern drained region, after an initial period of recession, also stabilized at a distance of 20–25 m in front of the drain.
3. End effects appeared to extend the effective drain length by 100–200 m, particularly on the southern down-drift side of the dewatering system.

Continued tests with independent observers were not very successful. The report by the Coastal Directorate (Bruun, 1989) has the following conclusion:

1. Under mild wave conditions the coastal drain system stabilizes beach profiles and provides a wider, higher high-tide beach. The coastal drain system is useful under certain specific conditions as described.
2. The coastal drain does not stop beach or dune erosion during storms. It is in no way a substitute for artificial nourishment. Its effectiveness on an eroding shore will decrease with time.

First installation in the USA—Sailfish Point, Florida

In 1988, Coastal Stabilization, Inc. (a subsidiary to Moretrench American Corporation) installed a 180 m-long-beach dewatering system at Sailfish Point, near the southern end of Hutchinson Island, on the Atlantic coast of Florida, USA. The beach is composed of fine-grained, well-sorted sand; the most notable feature along this otherwise open Atlantic coast is the natural coastline protection provided by a rock reef located approximately 100–150 m offshore. It has been suggested that despite the presence of the reef, between 1972 and 1986 recession of the high-tide shoreline exceeded 2 m/year. It is important to note that this erosional trend is reported to have reversed and become accretionary prior to the installation of the beach drains in 1988 (Terchunian, 1989; Dean, 1989).

The dewatering system installed at Sailfish Point (referred to as "Stabeach" by Coastal Stabilization, Inc.) consisted of a 0.3–0.5 m diameter PVC pipe buried at an elevation of approximately -2.5 m, providing a collection drain for numerous 1.5 m long horizontal well points attached at approximately 3 m intervals along its length. Collected water traveled via a suction pipe to a pumping station located landward of the dune line (Lenz, 1994). An independent report prepared for Coastal Stabilization, Inc., by Dean (1989) after 11 months of monitoring concluded that it was not possible to separate natural beach changes from those induced by the dewatering system; but a second report by the same author (Dean, 1990) after approximately 20 months of operation provided the first independent evidence that the dewatering system was having a positive effect on the beach. From a straightforward analysis of time series of sand volumes and the position of the high-tide shoreline, Dean concluded that, while it remained difficult to separate natural beach changes and those caused by water table lowering:

1. The dewatering system appeared to have resulted in local moderate accretion, in contrast to a general erosional trend to the north and a relatively small accretionary trend to the south.
2. The system appeared to result in a considerably more stable high-tide shoreline relative to both control segments north and south.

Recent installations

Some beach drains have been installed in the United Kingdom, United States, and Denmark. The results, however, were generally nonconclusive (Turner and Leatherman, 1997).

Conclusions

This brief report provides an overview of the history and current status of beach dewatering as a potential practical alternative to more traditional methods of coastal stabilization. The specific findings are as follows:

1. A link between the elevation of coastal groundwater and erosion or accretion trends at the shore has been reported in the coastal literature for over 50 years. The origins of this work can be traced to parallel but initially unrelated strands of beach research in the 1940s that were simultaneously providing new insight into the role of swash infiltration in determining erosion and accretion at the beach face, and the dynamics of beach groundwater in controlling the saturation characteristics of the foreshore.

2. In the mid-1970s, the first laboratory investigations were reported that examined the artificial lowering of beach groundwater as a method to promote shore accretion and stability, and the results proved encouraging. By the late 1970s the results of the first field investigation of this approach were reported, but the results of this work were less conclusive.

3. Commercial interest in beach dewatering as a practical alternative to more traditional methods of shore stabilization was initiated in the early 1980s as the result of an unrelated engineering project on the Danish coast. The decreasing efficiency of a buried seawater filtration system was observed to correspond to the rapid build up of sediment in front of intake pipes.

4. A full-scale test of the dewatering concept on the open North Sea coast of Denmark was undertaken during the period 1985–91. Initial results proved encouraging, and for the first two and half years of the system's operation published data suggest that, relative to untreated control sites, the dewatered beach stabilized and showed a positive trend of shore accretion. During the ensuing four years, the published

monitoring results were less conclusive, and it was interpreted that the beach drain was having no discernible positive effect on enhancing net beach width. Relative to the eroding control sections of beach, it was tentatively concluded that the dewatering system reduced the rate at which the coastline was eroding.

New dewatering sites should at present be regarded as experimental, rather than a proven solution to erosion management. The main problem with the drain is that it does not produce sand. It only takes some sand away from adjoining beaches. Compared to artificial nourishment the drain is uneconomical. Coastal researchers must investigate further both the dynamics of coastal groundwater determining the time-varying saturation characteristics of the beach face; and the modification of sediment transport mechanisms at the beach face induced by groundwater infiltration and seepage. Only when a physical understanding of these processes is gained, will the mechanisms determining the success or failure of the dewatering concept be understood (Turner and Leatherman, 1997).

Per Bruun

Bibliography

Abbott, M.B., and Price, W.A. (eds.), 1994. *Coastal Estuarial and Harbor Engineers' Reference Book*. New York: E & FN Spon.

Bagnold, R.A., 1940. Beach formation by waves; some model experiments in a wave tank. *Journal of the Institute of Civil Engineers*, **15**: 27–54.

Bruun, P., 1989. The coastal drain: what can it do or not do? *Journal of Coastal Research*, **8**(1): 123–125.

Davis, G.A., and Hanslow, D.J., 1991. Beach stabilization by drainage of the watertable. *Proceedings of the 10th Australasian Conference on Coastal and Ocean Engineering.*

Davis G.A., Hansow, D.J., Hibbert, K., and Nielsen, P., 1992. Gravity drainage: a new method of beach stabilization through drainage of the watertable. *Proceedings of the International Conference on Coastal Engineering*. pp. 1129–1141

Davis, G.A., Hanslow, D.J., Hibbert, K., and Nielsen, P., 1993. Beach stabilization by gravity drainage. In Bruun, P. (ed.), *Proceedings of the Hilton Head Island South Carolina, USA International Symposium*, Volume 2, p. 707.

Dean, R.G., 1989. Independent analysis of beach changes in the vicinity of the Stabeach System at Sailfish Point, Florida. Report prepared for Coastal Stabilization Inc., 15 p.

Emery, K.O., and Foster, J.F., 1948. Water tables in marine beaches. *Journal of Marine Research*, **7**: 644–654.

Grant, U.S., 1948. Influence of the water table on beach aggradation and degradation. *Journal of Marine Research*, **7**: 655–660.

Hansen, H.K., 1986. Coastal drain system: full–scale test at Thorminditangen. *Summary Report* No. 170–83322, Lyngby, Denmark: Danish Geotechnical Institute.

Heaton, K.L., 1992. Local effects of groundwater table variation on on-shore off-shore sediment transport. Unpublished BE (Hons) thesis. Perth: University of Western Australia, 134 p.

Lenz, R.G., 1994. Beachface dewatering—a tool for coastal stabilization. Alternative technologies in beach preservation. *Proceedings of the 1994 National Conference on Beach Preservation Technology*, pp. 27–52.

Li, L., Barry, D.A., and Pattiaratchi, C., 1995. Numerical modelling of beach dewatering through drainage. *Journal of Waterways, Port, Coastal, and Ocean Engineering*.

Nielsen, P., 1990. Tidal dynamics of the water table in beaches. *Water Resources Research*, **26**(9): 2127–2134.

Nielsen, P., 1992. *Coastal Bottom Boundary Layers and Sediment Transport*. River Edge, NJ: World Scientific Co. Pte. Ltd.

Nielsen, P. *et al.*, 2001. Infiltration effects on sediment mobility under waves. *Coastal Engineering*, **42**(2): 105–114.

Ogden, M.R., and Weisman, R.N., 1991. Beach stabilization using drains; an experimental model study. In *Procedings of the Coastal Sediments '91*. American Society of Civil Engineers, pp. 1955–1969.

Oh, T.M., and Dean, R.G., 1994. Effects of controlled water table on beach profile dynamics. *Procedings of the 24th International Conference on Coastal Engineering*, pp. 234–235.

Ovesen, N.K., and Schuldt, J.C., 1992. Beach Management System-Documentation. Summary Report. Danish Geotechnical Institute.

Terchunian, A.V., 1989. Performance of the Stabeach System at Hutchinson Island, Florida. *Proceedings of the 2nd National Conference on Beach Preservation Technology, Florida Shore and Beach Preservation Association*, pp. 229–238.

Turner, I.L., and Leatherman, P.L., 1997. Beach Dewatering as a "soft" engineering solution to coastal erosion—a history and critical review. *Journal of Coastal Research*, **13**(4): 1050–1063.

Vesterby, H., 1991. Coastal drain system: a new approach to coastal restoration. *Proceedings of the GEO-Coast '91*, pp. 651–654.

Vesterby, H., 1994. Beach face dewatering—the European experience. Alternative Technologies in Beach Preservation: *Proceedings of the 1994 National Conference on Beach Preservation Technology*, pp: 53–68.

Weisman, R.N., Seidel, G.S., and Ogden, M.R., 1995. The effect of water table manipulation on beach profiles. *Journal of Waterways, Port, Coastal and Ocean Engineering*, **121**: 134–142.

Zinn, D.J., 1942. An ecological study of the interstitial microfauna of some marine sandy beaches, with special reference to Copepoda. Unpublished Ph.D. dissertation. New Haven: Yale University.

Cross-references

Beach Erosion
Cross-Shore Sediment Transport
Depth of Disturbance
Hydrology of the Coastal Zone

BEACH EROSION

Introduction

Beaches are loose accumulations of sand, gravel, or a mixture of the two that bound an estimated 30% of the world's coasts (Bird, 1996). Because they consist of more or less loosely packed noncohesive sediments, beaches act as buffers that absorb, reflect, and dissipate energy delivered to the shore by waves. By doing so, they shelter areas behind the beach, especially during storms, from wave attack and flooding. Such back-beach zones may be cliffs, dunes, or low-lying marshes and lagoons. On many coasts of the world, the beaches and these associated back-beach environments have been taken up by development (Nordstrom, 1994). A lot of this development has occurred over the last three to four decades, thriving on the worldwide growth in domestic and international tourism, and largely favored by the diversification of beach recreational activities. The boom in coastal development, especially on low-lying sandy coasts, has been matched by an increasing awareness that the beaches that form the foundations of prosperity of many communities are eroding in many places. An estimated more than 70% of the world's beaches are now eroding (Bird, 1996). Lack of foresight in construction and development planning has, in many cases, led to massive and irreversible urbanization of the coast that renders many communities vulnerable to the insidious effects of beach erosion. Erosion impairs the capacity of a beach to act as a buffer against storms. This means that beach erosion may have serious negative repercussions for low-lying island states, for shorefront communities, and for beach-based leisure activities on which depend many jobs and from which many coastal communities draw income.

Beach studies started mainly in connection with military activities, notably during the preparation of the World War II landings on the French coast. Since then, they have increased dramatically, especially over the last three decades, as beach erosion has become a critical issue in coastal zone management in many countries. Evaluating the implications of beach erosion necessitates a clear definition of what beach erosion is and how it is measured, notwithstanding the fact that the physical processes involved in the dynamics of erosion are still not well understood.

Perception of beach erosion and measurement of erosion rates

Proper coastal management requires a clear definition of beach erosion and accurate quantification of erosion rates. Although beach erosion has received great attention from coastal scientists, government agencies, local authorities, and beachfront owners, its perception and exact definition are controversial issues, mainly as a result of the diverse interests of the different parties involved in beaches and/or their management (Esteves and Finkl, 1998). This statement, made in reference to beaches in Florida, holds true for beaches in many developed countries. Beach erosion is a process whereby a beach loses its sediment, resulting in a depletion of its sediment budget. This process occurs where the beach can no longer balance energy produced by waves and by water piling up against it, leading to net sediment loss and lowering and retreat of the beach. Basically therefore, beach erosion may be viewed as resulting from an imbalance between, on the one hand, the energy inputs and, on the other, the resistance of the beach bed and sediment liable to be mobilized by the fluid forces. The erosion process itself is thus a way of eventually reestablishing balance through dissipation of energy. However, this is a scientific and objective view of beach erosion. Perception of the problem generally tends to be associated with developed shores in urban areas mainly where sandy beaches are important to the economy (Finkl and Esteves, 1998). As these authors have shown for the beaches of Florida, which account for about 25% of the total sandy shores in the United States, this bias of the erosion perception is shown in discrepancies in the delimitation of both erosion problem areas (EPA) and critically eroded areas (CEA) among different surveys. There are no common standards for objectively classifying beach erosion. Each party perceives beach erosion in its own way. Furthermore, beach erosion is not commonly perceived as a problem on undeveloped shores. In an effort at objective standardization, Esteves and Finkl (1998) and Finkl and Esteves (1998) propose a useful, comprehensive beach erosion classificatory scheme covering developed and undeveloped coasts.

There are also no common standards for quantifying rates of beach change (Moore, 2000) and for determining high-tide shoreline position (Galgano *et al*., 1998; Morton and Speed, 1998; Douglas and Crowell, 2000). Beach erosion is generally quantified through some statistical treatment of retreat rates and volumetric losses (e.g., Leatherman, 1983). The input data comes either from field surveys that have gained in accuracy with the advent of electronic stations and differential global positioning systems (GPS), or from numerically rectified aerial photographs, maps, and land-use documents. Other new methods include digital video imagery near ground level or from low-flying aircraft, and airborne scanning laser altimetry or light detection and ranging (LIDAR) (Mason *et al*., 2000; see also entry on *Mapping Beaches and Coastal Terrain*). Beach erosion rates and volumetric losses may also be estimated from the depletion of beach nourishment material where such nourishment is regularly carried out (e.g., Finkl, 1996, see entry on *Beach Nourishment*).

Rates of beach erosion may range from a net moderate loss of less than a meter a year to several meters following just one storm event. Such rates may also vary alongshore, decreasing from a maximum in "hot spots" or high-tide shoreline areas subject to the most severe perturbations, to "cold spots" where the effects of such perturbations are no longer felt and the high-tide shoreline is stable. Extreme rates of beach retreat in isolated "hot spots" along the southeast barrier-island coast of the United States approach 4 m yr^{-1}, causing substantial loss of land and oceanfront property (Finkl, 1993). Reliable determination of rates of beach retreat is important in coastal planning, especially as regards construction setbacks.

Beach erosion processes within the profile

The beach is a three-dimensional (3-D) sediment body that extends alongshore from the upper limits of wave run-up to the outer limits of wave action, the so-called closure depth, in the nearshore zone. However, while the upper limits may be relatively easy to identify using geomorphic features (Morton and Speed, 1998), the offshore limits are not, for obvious reasons. Beach erosion may be a short-term (order of hours to seasons) process that reflects adjustment to wave energy changes, or a longer-term (order of years) one that reflects an increasingly deficient beach sediment budget (Figure B17). On sandy beaches, short-term changes involving erosion are commonly part of a so-called morphodynamic cycle of adjustment of the beach profile to seasonal or nonseasonal changes in wave energy (Short, 1999). Seasonal changes commonly correspond to the classic winter profile flattened by storms and the summer profile that accretes under fair weather conditions. Beach profile adjustment generally leads to better absorption of the nearshore and incident wave energy, leading over a more or less long period of time (hours to months), to an equilibrium situation and shoreline stability. The period of adjustment depends on the wave energy inputs, the beach morphology, and the sediment volume. Rapid beach recovery is quite common, but recovery may sometimes take several years following major storms (Morton *et al*., 1994; Galgano *et al*., 1998). Sandy beach morphology and sediment volume are intricately related, defining profiles that are either short, steep, and reflective, commonly associated with coarse sediment, or wide, flat, and dissipative, commonly with fine sand (see entries on *Reflective Beaches* and *Dissipative Beaches*). In the former, much of the sand is locked up in the intertidal beach, forming especially a voluminous subaerial beach sometimes comprising an upper beach terrace called a berm. In the latter, much of the sand is stored in shallow intertidal to subtidal bars. High-energy waves impinging on steep reflective sandy beaches result in the fastest response times, resulting in erosion of the upper beach and berm, and seaward removal of the sand to form barred dissipative beaches. In such situations, erosion of the upper beach is therefore compensated by accumulation on the lower beach, without there being necessarily a net loss of sediment. However, coastal managers are sensitive to changes in subaerial beach volume, so that such short-term upper beach and berm erosion, especially where severe, may raise anxiety.

Beaches characterized dominantly by gravel differ in their behavior. Such coarse-grained beaches have generally steep, narrow reflective profiles that are commonly inert and unresponsive, to a certain degree, to increases in wave energy. This is either a result of the spatial organization of the constituent clasts and/or because of the capacity of these coarse-grained beaches to absorb wave energy through high percolation rates (Carter, 1988; Forbes *et al*., 1995; Orford *et al*., 1996). Macrotidal beaches, found in areas with large tidal ranges (>4 m at spring tides), also commonly show slow or moderate response to high-energy events, compared to their more common microtidal counterparts, because of their wide, dissipative profiles and the rapidity of migration of the wave domains that goes with the important tidal excursion.

On any sandy or gravelly beach profile, short-term morphodynamic changes may be embedded in longer-term changes involving net sediments gains or losses, the latter being synonymous with overall beach erosion throughout the profile. Whatever its origins, a net loss of beach sediment results in durable changes in beach morphology as the beach seeks to adjust to this situation of sediment deficit. Durable erosion generally results in a permanently scarped beach profile exhibiting an upper beach scarp (Figure B18). In some cases, erosion can lead

Figure B17 The erosion and retreat of a gravel beach in Picardy, France, has left a World War II blockhouse stranded on the beach.

Figure B18 Durable beach erosion is commonly manifested by an upper beach scarp. The erosion here, which threatens coastal settlements and has led to inland relocations of several villages and an international highway, has occurred downdrift of the port of Cotonou in Benin, West Africa on a coast subject to strong longshore drift.

to the total disappearance of a beach. The sediment lost accumulates elsewhere, either further alongshore, in other beaches, in estuarine and lagoonal sinks, or in offshore sinks.

In spite of a considerable amount of beach research over the past three decades, the sediment transport processes operating on beaches and involved in beach erosion are still poorly known (Butt and Russell, 2000; also see entries on *Beach Processes* and *Surf Zone Processes*). "Normal" waves and wave shoaling processes in the nearshore zone lead to net shoreward flows that may result in sediment drifting alongshore or from offshore working its way toward the beach. In essentially reflective beach systems, these shoreward flows are balanced by gravity-driven seaward return flows down the beach face. The net sediment budget of the beach would then depend on its ability to diminish the seaward return flow volume through processes such as infiltration and grain size and bedform adaptations to flow strength. These processes depend on beach slope and grain size, both interrelated, and on the water table on the beach. Low, dry season water tables on microtidal beaches composed of medium to coarse sand in West Africa favor exceptionally steep beach slopes (up to 18°) that result from berm buildup through sand deposition by swash infiltration. Higher rainy season water tables have the opposite effect of encouraging little infiltration, thus favoring sediment transport down the beach. Erosion processes on the more complicated dissipative beaches depend on the complex interplay of various modes of fluid motion near the beach (Komar, 1998). These include incident gravity waves, infragravity or low-frequency waves generated by transfer of energy from the former, alternations of high and low waves or wave groupiness, wave-induced currents, tidal currents, currents due to wind forcing, wave–current interactions, and patterns of energy concentration related to the way these various fluid forces interact with the morphology. On a few of the world's coasts subject to strong, sustained winds, eolian processes may also contribute in removing sand from the beach.

Gravel beaches show specific behavioral modes because of their coarse grain sizes. The dynamics of these beaches are fully discussed elsewhere (Forbes *et al*., 1995; Orford *et al*., 1996; see entry on *Gravel Barriers*). On sandy beaches subject to episodic high-energy erosive events, a variety of related meteorological and hydrodynamic factors combine to enhance beach stripping. These are setup of the water level close to the beach, due to waves, onshore wind forcing, and sometimes the direct exposure of the beach to the low pressure system that generates the storm waves, and durable saturation of the beach face through both rainfalls that commonly accompany stormy weather and enhanced swash run-up. Water pileup on the seaward-sloping beach must be balanced by seaward return flows, or in low-lying sand barrier systems, by overwash. In these high-energy events, liquefaction and removal of the beach sand may be accompanied by deposition of these sediments in offshore areas where the energy of seaward flows peters out or is balanced by shoreward-directed energy. The beach profile retreat is also a short-term mechanism of creating accommodation space for the temporary water pileup. The sediment transported seaward may initially travel via major rip current pathways generated by incident and infragravity waves. Sand may flow alongshore from these pathways, because of the commonly oblique incidence of waves and wind setup, feeding strong longshore currents. Subsequently, as larger waves and strong winds lead to more pileup of water on the beach, the seaward return flows may no longer be simply canalized in rip channels and mass balancing seaward flows may occur, resulting in generalized beach stripping and offshore sediment loss. The intensity of beach erosion depends on various factors such as the wave energy level, the antecedent beach morphology, orientation of winds relative to the coast, their strength, beach grain size, tidal range and tidal state, and the duration of high-energy conditions.

Longshore manifestations of beach erosion

A beach may comprise one or several sediment cells with bounding limits to longshore drift. Swash and drift-aligned beaches (see entry on *Drift and Swash Alignment*), respectively, designate beaches associated with weak and strong rates of longshore drift (Davies, 1980). In many cases of beach erosion, the process is a subtle, insidious one that does not require the high-energy events described above (although these may spectacularly enhance erosion rates) other than seasonal increases in wave energy to which the beach is generally well adapted. This is particularly the case on coasts subject to strong longshore drift rates on which depend the overall stability of the beach. Erosion functions essentially where major engineering structures block the sediment load drifting alongshore. Continuity of sediment transport downdrift by the longshore current is assured by beach erosion. The strongest drift rates, sometimes exceeding 1 million $m^3 a^{-1}$ of sand or gravel, are found where large swell waves impinge with marked obliquity on long, open beaches, as on the Gulf of Guinea coast in West Africa, in New Zealand, and the Kerala coast of India.

The longshore manifestations of beach erosion have received attention in the literature, both in terms of the plan shape of freestanding beaches (as opposed to short, headland-bound bay beaches) and of the effects of major engineering structures. The plan shape of freestanding beaches may change rapidly in response to sediment depletion. These changes basically reflect sediment cell divisions (Carter, 1988) that may also involve switches from drift to swash alignment, in an attempt by the beach to adjust to sediment deficit by diminishing longshore transport. Examples have been described from both sandy (e.g., Anthony, 1991) and gravelly beaches (e.g., Forbes *et al*., 1995; Orford *et al*., 1996; Anthony and Dolique, 2001). The large-scale changes in beach plan shape are also accompanied by beach textural and profile reorganizations. Downdrift of jetties, a major cause of beach erosion (see next section), the high-tide shoreline morphology in plan commonly defines a log-spiral curve (see entry on *Headland Bay Beach*) or a half-heart bay (Silvester and Hsu, 1993) that may extend several kilometers. This shape illustrates the more severe retreat that affects the beach just downdrift of such structures. Continuity of sediment transport by the longshore current after the jetty is assured by sometimes rapid and significant beach erosion. Erosion diminishes downdrift of this "hot spot" as the longshore current becomes increasingly charged with sediment, leading to a more linear high-tide shoreline. At some distance downdrift, erosion becomes nil and the high-tide shoreline may even show advance from the accumulation of sediment eroded from the beach updrift. These longshore changes are sometimes manifested by a fast "erosion front" and a slow "erosion front" separated by a salient, or "bump," that may exacerbate erosion downdrift (Bruun, 1995). The existence of such two fronts along any eroding beach probably reflects two sediment cells on either side of a central downdrift accumulation terminus fed by beach erosion within the more updrift cell. The accumulation "bump," or salient, influences wave incidence angles in such a way as to minimize drift and capture sediment, thus aggravating erosion within the following longshore cell, as examples from gravel barrier beaches have shown (e.g., Orford *et al*., 1996; Anthony and Dolique, 2001). According to Bruun (1995), the distance of downdrift migration of erosion fronts on the Atlantic shoreline of Florida is of the order of 30–40 km, the fronts migrating essentially from inlet to inlet. This distance is similar to that downdrift of the seaport of Lomé, in West Africa (Anthony and Blivi, 1999).

On some beaches, especially headland-bound bay beaches, it is not uncommon for seasonal or longer-term changes in the predominant direction of wave approach to induce changes in longshore drift. This process results in "beach rotation" (Short, 1999), which is the periodic lateral movement of sand towards alternating ends of the embayed beach. It results in erosion at one end of the beach, while the other accretes. In rare cases, beach rotation is due to short- to medium-term (order of a few years) changes in nearshore bathymetry that affect wave refraction and dissipation patterns. The massive mud banks delivered by the Amazon river to the muddy coast of South America migrate westward toward the Orinoco delta, inducing changes in incident wave energy levels by strongly modulating wave refraction and diffraction patterns. These generate lateral movement of sand in embayed beaches between bedrock headlands in Cayenne, French Guiana, resulting in alternations in erosion (Figure B19) and accretion over time, without net sediment loss (other than through illicit sand extraction). Similar effects on beaches elsewhere may be generated by changing sand bank configurations offshore, as in the case of the sandy beaches of northern France bounding the English Channel and the southern North Sea.

The causes of beach erosion

The sediment that accumulates on the shore to form a beach may come from various sources. Any poorly consolidated material on which waves and currents impinge may be a source. Such material may be an initial coastal and nearshore deposit of diverse origin such as glacial till or fluvial sediments, or may be delivered to the shore through landslides or by volcanoes. These sources are usually cut into coastal cliffs and underwater slopes that recede as they deliver sediments to the shore for beach construction. Dunes may also deliver sand to the beach but the beach and dunes, especially those immediately bounding the beach, should be considered as an interrelated system with sand interchanges. Some infilled estuaries and many sand- or gravel-rich deltas also supply sediment to beaches, especially at times of high river discharge.

Any natural or human action that affects the supply capacity of a given source and the cross-shore and longshore sediment transport processes on beaches may result in erosion. In many cases, especially on long open beaches, several factors, whose specific roles are difficult to

Figure B19 An example of a beach in Cayenne, French Guiana, affected by periodic rotation due to mudbanks migrating alongshore. The erosion presently affecting this end of the beach (concomitant with accretion at the opposite end) has been aggravated by illicit sand extraction. Note the massive rock protection on the upper beach.

Figure B20 Beach accumulation and erosion, respectively, updrift and downdrift of a jetty in Upper Normandy, France. Erosion on this coast has been exacerbated by the stabilization of cliffs that hitherto supplied flint clasts to the beaches.

disentangle, may jointly cause beach erosion. The most readily discernible causes of beach erosion are where identified human actions and activities perturb the beach sediment budget and the morphodynamic functioning of the beach. This cause of beach erosion dramatically developed in the 20th century with the multiplication of dams across rivers and large-scale urbanization of the coastal zone worldwide. On many coasts of the world, the construction of dams has, over the long run, led to coastal sediment starvation and beach erosion. The effects of artificial structures on the shore (see *Shore Protection Structures*), and especially beaches, have received considerable attention in the literature (Walker, 1988; Silvester and Hsu, 1993; Bird, 1996; Charlier and De Meyer, 1998). One important cause of beach erosion worldwide is the construction of jetties and ports (Figure B20). In many coastal communities, as along the eastern United States, the lagoons behind barrier islands are important economic waterways whose inlets need to be deepened by dredging and kept open permanently by groins and breakwaters. These impede the longshore drift of sand that is vital in nourishing beaches downdrift. Esteves and Finkl (1998) estimate that 90% of beach erosion in southeast Florida has been caused by human action, mostly the construction of deepened inlets with protective jetties. Deepwater ports constructed on open beach coasts subject to strong longshore drift have similar negative effects, as in the Bight of Benin in West Africa (Anthony and Blivi, 1999). Here, national seaports in Togo, Benin, and Nigeria have resulted in dramatic beach erosion downdrift of the port breakwaters, and in equally spectacular beach accretion updrift. The continual beach erosion on this coast (Figure B18) has led

Figure B21 Chronic beach and dune erosion in Wissant, a tourist and recreational resort in northern France.

to successive inland relocations of coastal communities and of the major international highway linking the three countries, at great cost to their already beleaguered economies.

Another source of perturbation of beach sediment budgets and a cause of beach erosion is coastal urbanization, which involves the development of urban fronts with high-rise condominiums and hotels on the upper beach. Some of the best examples include the US Atlantic and Gulf coasts (Nordstrom, 1994; Esteves and Finkl, 1998), and the Mediterranean rivieras (Anthony, 1997). In many cases, related dune systems have been flattened or severely degraded, and this has had a dramatic effect on beach stability. Dunes tend to be overlooked as the "savings account" of the shore while the beaches act basically as a "checking account." The dunes store important volumes of sand that help in balancing the beach budget. In the past, uncontrolled shorefront urbanization has commonly entailed narrowing of many beaches, diminishing in time their wave-energy buffering capacity, and leading to beach erosion (Figure B21). Beachfront urbanization also often requires defense structures, notably walls and revetments emplaced on the upper beach. Depending on their design, these structures may act as static barriers that reflect wave energy offshore, thus aggravating beach erosion, although some doubt has been recently cast on this negative effect of sea walls (Kraus and McDougal, 1996). In the past, urbanization and the development of road and rail networks has sometimes involved the direct quarrying of sand or gravel from beaches with fragile sediment budgets. This practice is still frequent in developing countries that lack awareness of the environmental consequences of such beach sediment depletion.

To stabilize already eroding beaches, groins are sometimes built across the beach with the aim of trapping sediment drifting past. Breakwaters are also sometimes built off the beach to dissipate some of the wave energy that erodes the beach. In playing an efficient role sometimes in alleviating beach erosion, these structures may be instrumental in simply transferring the erosion problem further downdrift, often to the detriment of another community. It is not uncommon to see groin fields sprouting downdrift, increasing in numbers as the problem goes from one community to the next. The gravel barrier beach in Upper Normandy (Figure B20) and Picardy, France, is a clear illustration of this downdrift "march" of erosion and of the attendant groin field. In Picardy, a groin field emplaced to stabilize an eroding gravel beach grew from 6 groins in 1976 to 96 groins in 2000 over a distance of 10 km (Anthony and Dolique, 2001). The initial erosion of this beach started with the construction of several jetties updrift in the 19th and 20th centuries, and was aggravated by the artificial consolidation and stabilization of several sectors of cliffs that hitherto liberated gravel flint clasts to the beach longshore drift cell (Figure B20). This practice of cliff stabilization has, in some cases like this one, led to beach sediment starvation and erosion.

Natural sediment depletion is considered as a major cause of worldwide beach erosion (Bird, 1996). On many of the world's coasts, especially in areas where sea level over the past 5–6,000 years has been relatively stable, the sand forming the beaches was derived from sediments on the inner continental shelf. These drowned nearshore deposits have been reworked by waves and driven onshore to form successive beach ridges and dunes sometimes several kilometers wide, as along large stretches of the Australian, West African, and Brazilian coasts.

This process, called progradation, has stopped in most areas as the nearshore sediment supply has petered out. The beaches bounding these prograded coasts are sensitive to any long-term changes in wave energy, resulting, for instance, from greater storminess or sea-level rise. Although exhaustion of nearshore sediment stocks is commonly invoked as a cause of beach erosion, it is hard to substantiate because of the lack of records of long-term beach and nearshore profile changes.

Beaches, as mentioned earlier, may show short-term changes in profile in response to storms and fair weather conditions. Apart from the various causes of sediment depletion evoked above, changes in the state of the sea also lead to durable beach erosion. These changes include greater storminess, short-term variations in sea level related to major changes in sea surface temperatures such as, those associated with El Nino events, and secular sea-level rise, commonly imputed to global warming. Changes in offshore wave energy are due to storms, such as the northeasters in the eastern United States, and cyclones, or may reflect more subtle increases in wave energy due to greater storminess and sea-level rise. Exceptional waves generated by submarine landslides or earthquakes may also lead to significant beach erosion. These events generate destructive high-energy waves that remove the beach sediments offshore. The seaward return flows may lead to losses of sand beyond the offshore limits of the beach profile, such that the sand cannot be returned to the beach during the following fair-weather wave conditions. On beaches bounding low-lying coasts, permanent losses of sediment may also occur inland as waves wash over the shore. Greater storminess implies more frequent episodes of higher incident wave energy often accompanied by strong wind setup of water level inshore. Many beaches do not have the available sediment stocks to adapt to such increases in wave power and to the currents resulting from wind and wave forcing. Sea-level rise, either on a short-term basis, due to short-term events such as El Nino, or of a secular nature due to global warming, would similarly favor wave energy impingement higher up the beach face (see entry on *Sea-Level Rise, Effect*). New sediment commonly does not move in from alongshore to balance the increase in wave energy, and the beach erodes as its sediment stocks are transferred seaward. Depending on the wave energy regime and the rate of sea-level rise, such sediment may be permanently trapped offshore as the base of wave action moves upward through sea-level rise. This pattern of beach erosion resulting from sea-level rise has been extensively debated in terms of what has become known as the Bruun rule (e.g., SCOR Working Group, 1991; Thieler *et al*., 2000).

Managing beach erosion

Good beach management requires both accurate bookkeeping on rates and patterns of beach change and implementation of the right strategies in the face of erosion. Beaches are a multiresource asset in many ways, involving huge sums of money in developed economies, both in terms of revenue and for design and implementation of management policies. As a result, the number of parties involved in beach management may be considerable, ranging from state legislators and engineering bodies, through recreational and tourist agencies, to scientists, beachfront home owners individually or as associations, and environmental and ecological pressure groups. Beaches are, as such, objects of conflicting interests. In many developed economies, beach erosion has become the fundamental coastal zone management problem, and a national issue in several countries bordered by long stretches of densely developed low-lying shores, such as the Netherlands and the United States. It has also become a cause of major concern for low-lying island states subject to sea-level rise (Leatherman, 1997). In the face of beach erosion, the management options are very few indeed. These include the determination of development setback lines in order to accommodate future erosion without endangering constructions. In the absence of precise determination of beach erosion rates, this strategy may fail, as on the Bight of Benin coast in West Africa (Figure B18). A second strategy is that of letting erosion take its course, generally in undeveloped areas where the process does not constitute a hazard. A third strategy is that of fighting beach erosion at all cost, especially where vital national, economic, or military interests are at stake. The finest example of such a policy is that of the Netherlands (Hillen and Roelse, 1995). On some developed shores such as parts of south Florida, the value of beach real estate and the revenue from beaches are such that the high-tide shoreline position has to be maintained, generally through the implementation of costly solutions such as efficient bypassing of inlets (see entry on *Bypassing at Littoral Drift Barriers*) and, especially, regular beach nourishment (Finkl, 1996). These are often, and increasingly, the only efficient ways of restoring the beach sediment volume. These "soft engineering" techniques have been discussed in numerous papers in scientific journals, especially the *Journal of Coastal Research* and *Shore and Beach*, as well as in regular newspaper commentaries in many countries. A specific comment needs to be made here on engineering practice in managing beach erosion. In many countries, including the United States, beach erosion has been managed using various assumed empirical relationships. Some of these relationships have been reviewed recently by Thieler *et al*. (2000) who draw attention to their oversimplified assumptions relative to the complex reality of beaches. In Europe, engineering practice in some countries, notably the Netherlands, has treated coastal management within a geomorphic systems approach, rather than simply in terms of deterministic engineering models. While the dependence on engineering models is still well entrenched in France, the tendency in Britain has shifted in recent years toward considering beach and, more generally, shore erosion management, in terms of a geomorphic systems approach (Hooke, 1999) that integrates local experience (Brunsden and Moore, 1999). The complexity of beach erosion and the large number of parties involved in its management should call for a sensible and balanced mix of science with a systems approach, engineering expertise, past and present experiences, and the specificities of the local context in which erosion occurs.

Many developed countries are today faced with minor to critical beach erosion problems, largely because of lack of foresight in coastal development patterns. While they may have the resources to combat beach erosion, the same is not true for developing countries which cannot divert much needed money toward beach management, often considered as a "low priority" area. These countries are increasingly subject to the pressures of an often rapid pace of economic development, and of beach-based tourist activities, while facing the threats of sea-level rise from global warming. It is perhaps reassuring that because of the still moderate level of coastal development in many of these countries, they have the opportunity of avoiding the mistakes made in the past by the developed countries by planning such development in a way as to ensure sustenance of the beach resource. This opportunity can be seized through active transfer of knowledge from the developed to the developing countries.

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Bibliography

Anthony, E.J., 1991. Beach-ridge plain development: Sherbro Island, Sierra Leone. *Zeitschrift fur Geomorphologie, Suppl.-Bd*, **81**: 85–98.

Anthony, E.J., 1997. The status of beaches and shoreline development options on the French Riviera: a perspective and a prognosis. *Journal of Coastal Conservation*, **3**: 169–178.

Anthony, E.J., and Blivi, A.B., 1999. Morphosedimentary evolution of a delta-sourced, drift-aligned sand barrier–lagoon complex, western Bight of Benin. *Marine Geology*, **158**: 161–176.

Anthony, E.J., and Dolique, F., 2001. Natural and human influences on the contemporary evolution of gravel shorelines between the Seine estuary and Belgium. In Packham, J.R, Randall, R.E., Barnes, R.S.K and Neal, A. (eds.), *The Ecology and Geomorphology of Coastal Shingle*. Otley, West Yorkshire: Westbury Academic and Scientific Publishers, pp. 132–148.

Bird, E.C.F., 1996. *Beach Management*. Chichester: Wiley and Sons.

Brunsden, D., and Moore, R., 1999. Engineering geomorphology on the coast: lessons from west Dorset. *Geomorphology*, **31**: 391–409.

Bruun, P., 1995. The development of downdrift erosion. *Journal of Coastal Research*, **11**: 1242–1257.

Butt, T., and Russell, P., 2000. Hydrodynamics and cross-shore sediment transport in the swash zone. *Journal of Coastal Research*, **16**: 255–268.

Carter, R.W.G., 1988. *Coastal Environments*. London: Academic Press.

Charlier, R.H., and De Meyer, C.P., 1998. *Coastal Erosion—Response and Management*. Berlin: Springer.

Davies, J.L., 1980. *Geographical Variation in Coastal Development*, 2nd edn. London: Longman.

Douglas, B.C., and Crowell, M., 2000. Long-term shoreline position prediction and error propagation. *Journal of Coastal Research*, **16**: 145–152.

Esteves, L.S., and Finkl, C., Jr., 1998. The problem of critically eroded areas (CEA): an evaluation of Florida beaches. *Journal of Coastal Research*, **26** (Special issue): 11–18.

Finkl, C., Jr., 1993. Pre-emptive strategies for enhanced sand bypassing and beach replenishment activities in southeast Florida: a geological perspective. *Journal of Coastal Research*, **18**(Special issue): 59–89.

Finkl, C., Jr., 1996. What might happen to America's shoreline if artificial beach nourishment is curtailed: A prognosis for southeastern

Florida and other sandy regions along regressive coasts. *Journal of Coastal Research*, **12**: iii–ix.
Finkl, C., Jr., and Esteves, L.S., 1998. The state of our shores: A critical evaluation of the distribution, extension, and characterization of beach erosion and protection in Florida. *Proceedings of the 11th National Conference on Beach Preservation Technology*. Tallahassee, FL, pp. 302–318.
Forbes, D.L., Orford, J.D., Carter, R.W.G., Shaw, J., and Jennings, S.C., 1995. Morphodynamic evolution, self-organization and instability of coarse-clastic barriers on paraglacial coasts. *Marine Geology*, **126**: 63–86.
Galgano, F.A., Douglas, B.C., and Leatherman, S.P., 1998. Trends and variability of shoreline position. *Journal of Coastal Research*, **26** (Special issue): 282–291.
Hillen, R., and Roelse, P., 1995. Dynamic preservation of the coastline in the Netherlands. *Journal of Coastal Conservation*, **1**: 17–28.
Hooke, J.M., 1999. Decades of change: contributions of geomorphology to fluvial and coastal engineering and management. *Geomorphology*, **31**: 373–389.
Komar, P.D., 1998. *Beach Processes and Sedimentation*. Upper Saddle River, NJ: Prentice Hall.
Kraus, N.C., and McDougal, W.G., 1996. The effects of seawalls on the beach: Part I, an updated literature review. *Journal of Coastal Research*, **12**: 691–701.
Leatherman, S.P., 1983. Shoreline mapping: a comparison of techniques. *Shore and Beach*, **51**: 28–33.
Leatherman, S.P. (ed.), 1997. Island states at risk. *Journal of Coastal Research*, **24**(Special issue).
Mason, D.C., Gurney, C., and Kennett, M., 2000. Beach topography mapping—a comparison of techniques. *Journal of Coastal Conservation*, **6**: 113–124.
Moore, L.J., 2000. Shoreline mapping techniques. *Journal of Coastal Research*, **16**: 111–124.
Morton, R.A., and Speed, F.M., 1998. Evaluation of shorelines and legal boundaries controlled by water levels on sandy beaches. *Journal of Coastal Research*, **14**: 1373–1384.
Morton, R.A., Paine, J.G., and Gibeaut, J.C., 1994. Stages and durations of post-storm beach recovery, southeastern Texas, USA. *Journal of Coastal Research*, **10**: 884–908.
Nordstrom, K.F., 1994. Developed coasts. In Carter, R.W.G., and Woodroffe, C. (eds.), *Coastal Evolution: Late Quaternary Shoreline Morphodynamics*. Cambridge University Press, pp. 477–509.
Orford, J.D., Carter, R.W.G., and Jennings, S.C., 1996. Control domains and morphological phases in gravel-dominated coastal barriers of Nova Scotia. *Journal of Coastal Research*, **12**: 589–604.
S.C.O.R. Working Group 89, 1991. The response of beaches to sea-level changes: a review of predictive models. *Journal of Coastal Research*, **7**: 895–921.
Short, A.D. (ed.), 1999. *Beach and Shoreface Morphodynamics*. Chichester: John Wiley & Sons.
Silvester, R., and Hsu, J.R.C., 1993. *Coastal Stabilization: Innovative Concepts*. Englewood Cliffs, NJ: Prentice Hall.
Thieler, E.R., Pilkey, O.H., Jr., Young, R.S., Bush, D.M., and Chai, F., 2000. The use of mathematical models to predict beach behavior for U.S. coastal engineering: a critical review. *Journal of Coastal Research*, **16**: 48–70.
Walker, H.J. (ed.), 1988. *Artificial Structures and Shorelines*. Dordrecht: Kluwer Academic.

Cross-references

Beach Nourishment
Beach Processes
Bypassing at Littoral Drift Barriers
Dissipative Beaches
Drift and Swash Alignments
Gravel Barriers
Mapping Shores and Coastal Terrain
Modes and Patterns of Shoreline Change
Reflective Beaches
Sea-Level Rise, Effect
Shore Protection Structures
Surf Zone Processes

BEACH FEATURES

Limits and formation of beach features

When discussing the types of features that can be observed along a beach, it is important to first consider the boundaries in the coastal zone that define the limits of a beach. In everyday usage and in the scientific literature, there are some differences in defining these limits, primarily with regard to the seaward limit. Recreational beach users will often consider the beach to extend no farther seaward than the shoreline, and thus limit the beach to an entirely emergent feature, having a width that varies with changing water level. Scientific usage typically extends the beach out to the maximum limit of low water regardless of the water level at any particular time. In some scientific usage, such as in the discussion of coastal sediment dynamics, the seaward limit of the beach may be considered to extend out to the breaker zone (Figure B22) well beyond the low-water shoreline. The most useful definition, and the one used here, is that the beach refers to the zone containing unconsolidated material that extends from the limit of ordinary low-water (or mean low-tide level) on its seaward side to the limit of influence by storm waves on its landward side (Figure B22) (Hunt and Groves, 1965; Baker *et al.*, 1966; Coastal Engineering Research Center, 1984).

Based on morphology, the beach is divisible into two zones. The *backshore* is the more landward and higher part of the beach and is typically a near-horizontal to gently landward-sloping surface. The backshore is not affected by the run-up of waves except during storm events, and so this is the typically dry part of the beach. The landward limit of the beach, which is the limit of influence of storm waves, generally is marked by a change in material, a change in morphology, or a change to a zone of permanent vegetation. Examples of such a landward limit include dunes, cliffs or bluffs, or even engineered structures such as bulkheads or revetments. The *foreshore*, also called the *beachface*, is the more seaward part of the beach. The foreshore has an overall seaward slope, but may include one or more ridges and troughs on its lower slope. Because the foreshore extends to the limit of ordinary low-water, at times of high-water the lower part of the foreshore is submerged.

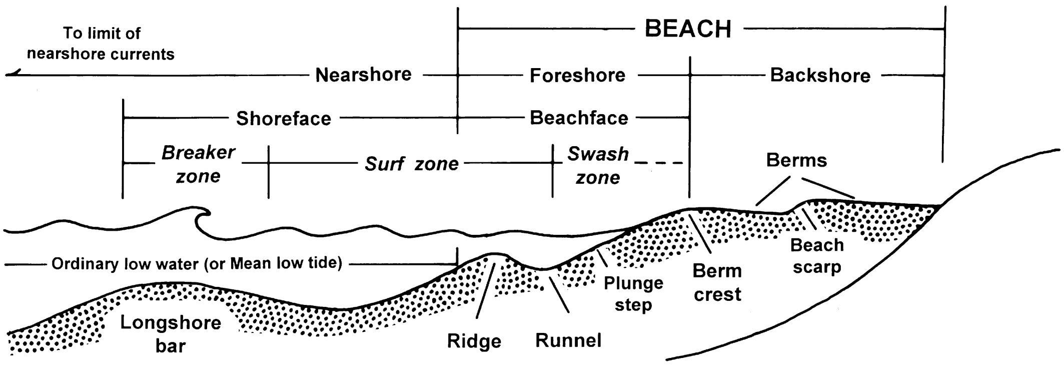

Figure B22 Generalized beach and nearshore profile showing names of major beach features and zones.

Critical in the definition of a beach is the presence of unconsolidated materials. These unconsolidated materials are what make a beach, and it is the erosion, transport, and deposition of these materials that results in beach features. Worldwide, the most common beach material is sand-size sediments composed of mineral, shell, or rock fragments. Coarser beach materials include gravel, cobbles, shingle, and even boulders. The beach will be made from whatever is locally available for the waves to rework. Along shores impacted by commercial or industrial activity, it is not uncommon to find beaches composed in part or completely of bricks, broken concrete, demolition debris, or any other material that may have been dumped along the shore and subjected to movement and redistribution by wave action.

The types of features that may occur along and across a beach vary in time, scale, and relative position. The primary agent in forming beach features is wave action (Davis, 1985). Other important agents are the rise and fall of water level, currents, wind, and ice in settings where coastal ice can form. Some beach features can form and persist indefinitely with minimal change in shape or location, but these are the exception. Because the beach is a dynamic setting, most beach features are ephemeral. Once formed, most beach features will only persist until new wave, water level, current, or wind conditions destroy them and replace them with new features.

Beach slope

Beach slope, which refers to the slope of the foreshore or beachface, deserves special mention as a beach feature because it is one of the characteristics used to distinguish different beaches. Slope is a dynamic feature that changes with changes in wave conditions as well as the gain or loss of different sediment sizes on the beach face. In general, the slope angle, measured relative to a horizontal plane, increases as the grain size increases; thus beaches composed of material such as pebbles or cobbles will tend to have a steeper beach face than ones made of sand. This slope difference relates to the greater permeability of the larger materials (Bagnold, 1940; Bascom, 1951; King, 1972). The wave run-up (or swash) can percolate downward through the interstitial spaces of the larger materials, and this minimizes the erosional influence of the run-back (or backwash). Storm conditions will flatten the beach slope as beach sediment is eroded and transported seaward. In the calmer wave conditions following the storm, beach slope recovers to a steeper slope as material is accreted to the beach.

Major beach features

Major beach features are here defined as those having large topographic expression or large areal extent. One of the beach features that can have the largest vertical expression on a beach is a *beach scarp*. Beach scarps are erosional features that occur when the slope of the beachface is lowered during storm events, and the beachface migrates landward by cutting into the backshore. The result is a near-vertical slope along the limit of this erosion (Figure B23). The height of a beach scarp may be just a few centimeters or a meter or more depending on the degree of wave action and the type of beach material. Beach scarps are commonly observed in areas where a new supply of sediment (i.e., beach nourishment) has recently been applied in an effort to replenish and build up the beach and wave action has cut into this nourishment and redistributed the sediment in the process of reestablishing an equilibrium beach profile.

Figure B23 Beach scarp resulting from recent storm erosion along a sand beach on the Illinois shore of Lake Michigan at Illinois Beach State Park. (photo by Michael Chrzastowski, Illinois State Geological Survey.)

One of the reasons that beach scarps are prominent beach features is that these near-vertical erosional features are cut into a near-horizontal area of the upper beach. This upper beach, in many cases, is a broad, near-horizontal to gently landward-sloping area called a *beach berm*, or simply a *berm* (Figure B22). Berms are depositional features formed from the wave-induced onshore accumulation of sediment. Local coastal conditions may preclude formation of a berm along some beach segments, while other beach segments may have two or more berms at different elevations. When more than one berm occurs, the lower berm(s) (sometimes called the *ordinary berm*) is a result of average or more typical waves, and the higher berm(s) (sometimes called the *storm berm*) is a result of the less frequent larger waves. A beach scarp may exist between two berms having different elevations. The seaward margin of the berm is typically defined by a rather abrupt change in slope from the near horizontal surface of the berm to the inclined surface of the beachface. The line defined by this change in slope is called the *berm crest* or *berm edge*. The berm crest is the distinguishing beach feature that divides the beach into the foreshore and backshore zones (Figure B22).

When low-water occurs, large-scale beach features are exposed that formed underwater and have a morphology influenced by waves, water-level changes, and associated currents. *Ridges and runnels* are the most common of these features. Ridges are elongate low mounds of beach material that are parallel or subparallel to the shore; runnels are the low areas or troughs that occur between the ridges and on the landward side of the shoremost ridge. A single ridge–runnel set may occur with the runnel on the landward side of the ridge or, if the lower foreshore is a broad, low-slope area, multiple sets of ridges and runnels may extend across this area. Such multiple sets of ridges and runnels form a washboard or corrugated topography across the lower beachface that contrasts with the smoother surface across the upper beachface. Another term for these features is "*ball*" referring to the ridge, and "*low*" referring to the runnel. The term "trough" is also sometimes applied to the runnel.

Ridges and runnels, when present in multiple sets, are one example of the types of repetitive patterns that can be observed in beach features. Another major beach feature with a repetitive nature is the *beach cusp*. Beach cusps are low mounds of beach material, separated by crescent-shaped troughs, occurring in a series along the shore. Yet another repetitive feature, and one of potentially large vertical scale, is the *beach ridge*. Beach ridges are depositional features, formed by the mounding of beach material by wave action, usually during storm events. Beach ridges are formed in the backshore zone, in some places at the most landward limit of wave influence, and they can extend as a nearly continuous linear feature for many kilometers along the shore. Wind transport may contribute sand to the tops of these ridges and form superimposed dunes (i.e., dune ridges). A single beach ridge may develop, persist for some time, and then be destroyed by renewed storm action. Sequential beach ridges may form through a series of depositional events and, with time, the juxtaposition of these ridges will contribute to the progradation of the coast.

A major beach feature common to barrier island beaches is the *washover fan*. During storm events, elevated water levels and large wave run-up can combine to transport large volumes of beach material across the beach to be deposited in broad, lobate accumulations on the landward margin of the beach. These deposits, called *washover fans*, essentially extend the landward limit of the beach and play an important role in the landward and upward migration of the beach during conditions of rising sea level (Kraft and Chrzastowski, 1985). On barrier island beaches, the formation of washover fans results in the burial of back-barrier marshes and filling along the barrier margin of lagoons.

Coasts that are subject to seasonal ice formation, such as the Great Lakes coasts of North America, can have various beach features that are related to the presence of coastal ice. A hummocky topography may develop along the upper foreshore as a result of sediment pushed into ridges and mounds by wave-thrusted ice. Shallow depressions may also develop across the upper foreshore where wave run-up may remove thin slabs of ice-cemented sand. Once an ice complex forms along the shore, a hummocky topography may develop in the lower foreshore by the action of grounded ice and the scour and fill by waves and currents around the edges of the ice. On the outer edge of the ice, an erosional trough may develop caused by the downward deflection of wave energy as waves impact the ice face. This trough can be a half-meter deep and 2–3 m wide (Barnes *et al*., 1994). The trough location will shift toward or away from shore as the ice margin shifts in these directions. These troughs and all other ice-related beach features are relatively short-lived. Once the ice conditions cease, any ice-induced modifications to the beach morphology are quickly eliminated by ice-free wave conditions.

Minor beach features

Minor beach features are defined here as those with minimal topographic expression or small areal extent. Although limited in height and area, some of these beach features can be visually prominent because of contrast in color, texture, or materials compared to the surrounding beach. Prime examples of such prominent, small-scale beach features are the *tidemark* which is the high-water mark left by tidal water, and *swash marks* formed along the landward limit of wave swash on the beach face. The tide mark is generally a nearly continuous wavy line defined by an accumulation of driftwood, seaweed, and other floatable debris collectively called *flotsam*, left on the beach by the previous highest tide level. Swash marks are a series of superimposed scalloped or fan-shaped patterns defined by fine sand, mica flakes, or bits of seaweed deposited along the most landward reach of the swash. Swash marks are beach features that are in a nearly continuous state of formation and destruction with each new swash event. So too are *backwash patterns* which are diagonal patterns formed on the beach by the dispersion of backwash flowing around small obstacles such as a shell or pebble. A falling tide or the lowering of water level after a storm can contribute to the formation of *rill marks* which are small, erosional furrows or channels across the beachface caused by the seaward flow of water as the water table in the beach lowers and water percolates out onto the beachface in a spring-like manner. *Air holes* may also occur on the beachface as water percolates and forces air from the pore spaces up to the surface.

Near the ridge and runnel on the lower foreshore slope, a subtle linear feature may occur called the *step* or *plunge step*. This is a subtle decline in the foreshore profile that is caused by the final plunge of waves before running up the beachface. The plunge step is best developed in settings of low tidal range and steep foreshore slope (Davis, 1985).

Because one or more of the berms of a beach are elevated above the influence of the swash, fine sand across these berms is typically dry and can be influenced by wind action. Although the berms are located on the beach, features can develop here that are common to dunes and desert settings. Sand ripples may develop, small dunes may form in wind shadows such as behind logs or other driftwood, and wind deflation areas may occur where the fine sand has been removed to lower the surface and leave a concentration of coarser particles similar to a desert pavement.

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Bibliography

Bagnold, R.A., 1940. Beach formation by waves; some model experiments in a wave tank. *Journal of the Institution of Civil Engineering*, **15**: 27–52.

Baker, B.B., Jr., Deebel, W.R., and Geisenderfer, R.D. (eds.), 1966. *Glossary of Oceanographic Terms*. Washington, DC: U.S. Naval Oceanographic Office, Special Publication 35.

Barnes, P.W., Kempema, E.W., Reimnitz, E., and McCormick, M., 1994. The influence of ice on southern Lake Michigan coastal erosion. *Journal of Great Lakes Research*, **20**: 179–195.

Bascom, W., 1951. The relationships between sand size and beachface slope. *American Geophysical Union Transactions*, **32**: 866–874.

Coastal Engineering Research Center, 1984. *Shore Protection Manual*, Volume 1. Washington, DC: U.S. Army Corps of Engineers.

Davis, R.A., Jr., 1985. Beach and nearshore zone. In Davis, R.A., Jr. (ed.), *Coastal Sedimentary Environments*. New York: Springer-Verlag, pp. 379–444.

Hunt, L.M., and Groves, D.G., 1965. *A Glossary of Ocean Science and Undersea Technology Terms*. Arlington, VA: Compass Publications.

King, C.A.M., 1972. *Beaches and Coasts*. London: Edward Arnold.

Kraft, J.C., and Chrzastowski, M.J., 1985. Coastal stratigraphic sequences. In Davis, R.A., Jr. (ed.), *Coastal Sedimentary Environments*. New York: Springer-Verlag, pp. 625–663.

Otvos, E.G., 1999. Rain-induced beach processes; landforms of ground water sapping and surface runoff. *Journal of Coastal Research*, **15**: 1040–1054.

Cross-references

Beach Nourishment
Beach Processes
Beach Ridges
Drift and Swash Alignments
Profiling, Beach
Rhythmic Patterns
Ripple Marks
Scour and Burial of Objects in Shallow Water

BEACH NOURISHMENT

Introduction

Beaches occur where there is sufficient sediment for wave deposition above water level along lakes, open ocean coasts, embayments, and estuaries. Beach nourishment most commonly takes place along marine beaches, which are among the most dynamic environments on earth. On a global scale, estimates of marine sandy beaches (see entry on *Sandy Coasts*) range from about 34% (170,000 km) (Hardisty, 1990) to 40% of the world's coastline (Bird, 1996). Beaches form essentially 100% of the coast of The Netherlands, 60% in Australia, and 33% in the United States (Short, 1999). Comprising a significant proportion of the world's coastline, beaches are important considerations for coastal recreation and storm protection, while others are used for residential, commercial, and industrial purposes. Although they serve as natural barriers to storm surge (*q.v.*) and waves (*q.v.*), today about 75% of the world's beaches are subject to erosion (Bird, 1985). In the United States, the percentage of eroded beachfront is somewhat greater than the world average and is estimated by some coastal researchers to approach 90% (e.g., Leatherman, 1988). During the last century, many erosion-control techniques were developed to mitigate the unwanted impacts of erosional events, especially those associated with accelerated rates of erosion in the vicinity of groins, seawalls, or jetties along developed shores. Traditionally, coastal armoring structures such as seawalls, breakwaters, and groins were relied upon to reduce wave energy approaching the shore or to catch sediment moving across or along the shore, and thus provide protection from coastline retreat. Engineering works, however, provide only partial protection and in some cases actually exacerbate the problem they were designed to cure. During the last century, beach nourishment was recognized as an environmentally friendly method of shore protection, especially along the coasts of the western world. Today, artificial beach nourishment is the method of choice for shore protection along many developed coasts with eroding beaches (Figure B24).

Despite the fact that beach nourishment has been used in many hundreds of locations under a wide variety of environmental conditions (e.g., Psuty and Moreira, 1990; Silvester and Hsu, 1993), and frequently integrated with hard structures as part of strategic shore protection efforts, there is much debate about whether the procedure is the best solution to problems of coastline retreat. Although there are many arguments against beach nourishment, artificial supply of beach-sand remains the most practical method of protecting against coastal flooding from storm surges, for advancing the shoreline seaward, and for widening recreational beaches.

Definitions, terminology, and concepts

The term *beach nourishment* came into general use after the first renourishment project in the United States at Coney Island in 1922 (Dornhelm, 1995). In engineering parlance, the terms *beach (re)nourishment*, *beach replenishment*, and *beach restoration* are often used more or less interchangeably in reference to the artificial (mechanical) placement of sand along an eroded stretch of coast where only a small beach, or no beach, previously existed. There are, however, subtle connotations in the application of each term. Sediments that accumulate along the shore in the form of beaches are naturally derived from a variety of sources such as fluvial transport in rivers to deposition of sediments in deltas, from preexisting sediments on the offshore seabed, from chemical precipitates (e.g., oolites on carbonate banks in tropical and subtropical environments), or from organisms living along the shore (e.g., shells and exoskeletons from marine organisms). When the natural sediment supply is interrupted, beaches become sediment-starved and the shoreline retreats landward due to volume loss. Efforts to artificially maintain beaches that are deprived of natural sediment supply thus attempt to proxy nature and (re)nourish the beach by mechanical placement of sand. The beach sediment is thus *replenished* by artificial means. *Beach restoration* implies an attempt to restore the beach to some desired previous condition. A nourished or *constructed beach*

A

Figure B24 a.

could be placed along a previously beach-less shore, whereas a restored beach is revitalized by the mechanical placement of sediment.

Beach nourishment projects involve placement of sand on beaches to form a *designed structure* so that an appropriate level of protection from storms is achieved. The placement of sand is commonly by methods such as dredging sand from borrow areas on the seafloor (e.g., Finkl *et al*., 1997), bypassing sand around deepwater inlets or other obstructions (e.g., groins) along the coast that interrupt the littoral drift, or overland delivery of sand from inland quarries to the coast. Although pumping of sand from offshore is the most widespread method of application, due to the large volumes of sediment that are required for most projects, other developments feature placement of sand by trucking or barging from quarries or construction sites, as well as removal of sand from dunes, or relocation of sediment on the berm via *beach scraping* (e.g., Bird, 1990; Healy *et al*., 1990; McLellan, 1990; Verhagen, 1996). It is now known, however, that removal of sand from dunes is not an appropriate option for sand supply because dune and beach sediment budgets are inextricably interlinked (Psuty, 1988). Although most beach nourishment projects deal with sand-sized particles on low to moderate energy coasts, shore protection efforts are also undertaken in very high-energy environments where gravel beaches are featured (e.g., Zenkovich and Schwartz, 1987; Peshkov, 1993).

Beach nourishment, which entails the construction of new beaches where none existed before or restoration of degraded beaches, usually takes place on developed shores. Along undeveloped shores, beaches provide natural habitat (e.g., nesting grounds for sea turtles and shorebirds) but on developed shores beaches additionally protect coastal infrastructure from storms and are important recreational sites for a globally expanding tourist industry. When beaches are degraded by decreased width and lowering of berms (see entry on *Beach Features*), many communities choose to replace lost sediment by pumping beach-quality sand from offshore to selected renourishment sites onshore.

Arising from reviews of replenishment activities in the United States, a new terminology was developed to describe the ruggedness or persistence of sand placed along the shore to form artificial beaches. Beach durability defines how well the beach performed under a variety of conditions. The definition *of beach durability* by Leonard *et al*. (1990a) states that "... the time between placement and loss of at least 50% of

B

Figure B24 b.

the fill volume ...” represents the half-life of a beachfill. The identification of *profile evolution*, performance of the beachfill after placement, is an important consideration of durability and longevity.

Historical background to the deployment of beach nourishment for shore protection

The beach at Coney Island, New York, was the first to benefit from a concerted effort at beach nourishment when, in 1922, more than $1 \times 10^6\ m^3$ of material were dredged from New York Harbor and transported to Coney Island (Hall, 1953; Dornhelm, 1995). Based on the apparent success of this new shore-protection measure, there soon followed a number of other projects along eroding shores elsewhere in New York and along coastlines in New Jersey and southern California. As in the case of Coney Island, beachfills in these early artificial renourishment projects were dredged from sediments in harbors and ship channels. In 1939, Waikiki Beach, Hawaii, was artificially nourished as a recreational beach.

Some other early beach nourishment projects included efforts in Africa and Europe. The construction of jetties for the Durban, South Africa, harbor entrance in 1850 initiated erosion of an adjacent downdrift beach. Groins were built along the shore but they did not stop the aggressive beach erosion. Based on the recommendations of a Belgian engineer, additional long, low groins, in combination with sand bypassing, were added. This was the first attempt at *beach stabilization* using renourishment techniques in South Africa (Swart, 1996). Around 1850, seawalls and groins were deployed along the North Sea coast of Norderney in an effort to stabilize this eroding German barrier island. Although these engineering structures prevented dune erosion, they did not stop the loss of sediments from the beach. In an attempt to alleviate problems associated with coastline retreat due to beach erosion, the first large-scale beach nourishment project in Europe was initiated on Norderney in 1951. By 1989 the beach had been renourished an additional six times (Kunz, 1993).

Beach nourishment projects have been carried out in many other counties including Australia, Belgium, Brazil, Cuba, Denmark, France, Great Britain, Japan, New Zealand, Portugal, Russia (see discussion in

C

Figure B24 c. Beach renourishment on the Gulf Coast of western Florida. (A) Renourishment on Longboat Key, west-central Florida coast near Sarasota showing an eroded beach partly protected by rock revetments and seawalls. In the central foreground there is no beach at high tide. Placement of beachfill is advancing from north to south, as shown in the top of the photo. (B) The restored beach, now 100 m wide, provides a degree of protection from storm surge while providing a much enlarged recreational beach area. (C) Small coastal suction, cutterhead dredger obtaining sand from an offshore borrow near Captiva Key, southwestern Florida. Beach-quality sand is pumped ashore in a slurry via a floating pipeline. About 75% of the renourished beaches on the Florida west coast have a half-life longer than 5 years (Dixon and Pilkey, 1991) (photos: Courtesy of Coastal Planning and Engineering, Inc., Boca Raton, FL).

Walker and Finkl, 2002). Even though the basic goal of beach nourishment is to elevate the beach and advance the shoreline in order to realize all of the consequent benefits of multiple use but especially increased storm protection, the techniques of sand transfer to the shore and design parameters differ among national approaches. In the United States, the State of Florida has a long and distinguished record of beach nourishment along the southeast coast that involves such notable achievements as: (1) the first sand bypassing weir jetty in the world (Hillsboro Inlet, Broward County), (2) the longest continuously operating fixed sand bypass plant (South Lake Worth Inlet, Palm Beach County) in the world, (3) longest half-life of any renourished beach in the United States (Miami Beach, Miami-Dade County), and (4) the first successful groin-aragonite beachfill project in the United States (Fischer Island, Miami-Dade County) (Finkl, 1993; Balsillie, 1996).

Needs for beach nourishment

Although beach erosion (see entry on *Erosion Processes*) is common along most coastlines, it is often difficult to recognize in the field unless there are obvious indications of sediment removal. The development of beach scarps in the berm, dune breaches with overwash, presence of tree stumps or marsh muds on the beachface, and location or damage of buildings precariously close to uprush levels are all signs that beaches are moving landward due to sediment loss. Young *et al*. (1996) describe these features as *geoindicators* that are helpful for evaluating coastline change along beaches. Such indications of beach erosion are important parameters for estimating the sensitivity and extent of the beach-erosion problem and remediation. The removal of beach materials is by wave action, tidal or littoral currents, or wind. Ranges of countermeasures provide protection from beach erosion, foremost among them during the last quarter of the 20th century, being *artificial nourishment* (i.e., the mechanical placement of sand on the beach).

Beach protection measures are necessary because beaches are important natural resources that support multipurpose activities. When well maintained, beaches provide storm surge protection, flood control, recreational activities, and habitat for numerous species of plants and animals (Wiegel, 1988). Lack of proper coastal maintenance may allow beach erosion to reduce dunes and other natural upland protection, increase loss of natural habitats, degrade a major source of tourist revenue, and shrink the overall economy (Strong, 1994; Finkl, 1996). Beaches thus need to be protected because they reduce vulnerability to coastal development in high-velocity areas (see entry on *Global Vulnerability Analysis*).

Although beaches provide a measure of protection to the shore from damage by coastal storms and hurricanes (typhoons and tropical cyclones), their effectiveness as natural barriers against surge flooding depends on their size and shape and on the duration and severity of storms. Beaches are also highly valued as recreational resources that contribute to the economic well-being of many coastal regions in the world. The trend of increasing beachfront development since World War II has resulted in the replacement of dune systems with buildings. This practice has increased exposure of buildings to damage from natural forces (Figure B25), especially high-energy secular events. The presence of buildings close to an eroding coastline enhances the reduction in beach width because the stabilized shore cannot move landward as it would under natural conditions (see entry on *Human Impact on Coasts*). Fixation of the coastline by construction in turn adversely impacts both natural storm protection and recreational quality of affected beaches (NRC, 1995).

The deterioration or degradation of beaches is regarded as undesirable because beaches provide natural protection from storms and have economic value. The unwanted effects of beach erosion commonly place life and property at risk, usually from flooding, and decrease a community's ability to maintain a viable tourist-based economy. Commercial and residential development on upland areas behind beaches and in close proximity to eroding beachfronts are mainly jeopardized by decreasing (eroding) beach widths (e.g., Wiegel, 1988, 1994). Increased potential for economic loss and safety concerns for human life thus drive desires to remediate beach erosion by artificially replacing sand that is lost to erosion.

In addition to the use of beach nourishment for combating coastal erosion, the procedure has been advocated because it: (1) tends to be less expensive and easier to construct than hard structures, (2) is aesthetically desirable, "user friendly" (e.g., Nelson, 1993) and "environmentally green" (Finkl and Kerwin, 1997), (3) provides a source of sand for wind-created or artificially created dunes which add to the protection of inshore areas (e.g., Psuty, 1988; Malherbe and LaHousse, 1998), (4) utilizes "waste products" from dredging or construction projects (e.g., Hillyer *et al*., 1997), (5) contributes to the littoral sediment budget and may benefit down-drift locations (e.g., Lin *et al*., 1996), (6) capitalizes on natural processes (e.g., Charlier and De Meyer, 2000) and thus is more acceptable to society, and (7) restores habitat for biota (e.g., Finkl, 1993; Verhagen, 1996).

Causes of beach erosion

Artificial beach nourishment became necessary only when beachfronts were developed for recreational, urban, industrial, and military uses. It is often difficult to understand at once the causes of shore erosion because they can be natural or introduced by human activity along the shore. When induced or accelerated by engineering structures the process is sometimes referred to as *structural erosion* (Pilarczyk, 1990). Beach erosion is influenced by numerous factors such as uplift (e.g., neotectonism) or subsidence (e.g., groundwater withdrawal, compaction of sediments) of the land surface, change in climate patterns (especially storm frequency, deviation of prevailing wind direction), interruption of sediment supply to the coast, eustatic fluctuations of the sea surface, blockage of littoral drift, and construction on the coast. An increase in relative sea level (i.e., drowning of the coast and landward retreat of coastlines) is, however, often cited as the primary geophysical cause of beach erosion (e.g., Leatherman, 1988; Douglas *et al*., 2000) but many other factors are involved. Construction of dams on major exorheic rivers withholds delivery of sediment to the coast. In the case of the Mississippi River, the sediment load today is about half what it was in pre-dam construction days. Further, sediments that bypass a dam are usually fine-grained and therefore more likely to be carried out to sea and lost to coastal accumulation. Dredging of deep inlets, navigational entrances (see entry on *Tidal Inlets; Navigation Structures*), and the construction of shore protection structures such as jetties are other

Figure B25 Example of an overdeveloped coastal segment along Balneário Camboriú Beach (Santa Catarina State, southern Brazil), where construction of condominiums and tourist facilities restricts the natural dune–beach interaction with phases of seasonal storminess. During perigean spring tides, and storm surges resulting from the passage of atmospheric cyclonic fronts, the beach is under water and the beachfront road and adjacent shops are flooded. Periodic beach replenishment is required to widen the dry-beach width and add height to the berm. Without beach nourishment, developed coastal segments such as this one lose socioeconomic amenities associated with a wide recreational beach and become increasingly vulnerable to flooding.

interrelated factors that contribute to the degradation of natural beach systems. This list is by no means comprehensive and yet it must be concluded that the causes of beach erosion are manifestly complicated and often interrelated.

Although a range of natural processes contributes to the landward retreat of coastlines, urban development along the shore necessitates placement of sand on eroding beaches in an effort to protect infrastructure. The essence of the problem is not the dynamic adjustment of coastlines to fluctuating ambient conditions but construction too close to the water. Most coastline development is deliberate for reasons of access, proximity, or aesthetics. Whatever the initial impetus for developing coastlines, the result has been an expensive exercise in what is usually nationally funded coastal protection. Coastline retreat in Australia, for example, is not as problematical as it is in the United States because the Commonwealth government established Crown lands along most of the coast keeping urban development some distance inland. In other countries such as The Netherlands where land has been reclaimed from the sea, large dikes and other engineering structures (see Walker, 1988) such as dunes and renourished beaches are part of efforts to hold back the sea. Although these areas have multipurpose uses (e.g., storm protection, conservation, recreation, water catchment), they are not open to intensive urban uses (Pilarczyk, 1990).

Coastal development often includes the dredging of ports and harbors and the navigational channels that serve them. Jetties that provide protection from waves in the channel are used to stabilize the geographic migration or orientation of many entrances. Jetties and deep inlets, as well as other shore protection structures such as groins, interrupt the natural littoral drift by impounding sediment or causing it to be jetted offshore beyond the longshore sediment transport system. It is now widely recognized that the interruption of sediment transport along the shore by artificial structures causes downdrift shores to become sediment-starved, which in turn results in shore erosion and retreat of the coastline (Figure B26). Large littoral drift blockers (e.g., deep navigational entrances, groin fields) can initiate downdrift erosion that propagates for several tens of kilometers (Bruun, 1995). In a study of 1,238 km of Florida coastline, distributed among 25 coastal counties and covering about 95% of the state's beaches, Finkl and Esteves (1998) concluded that littoral drift blockers on Florida's Atlantic and Gulf coasts accounted for 72% of the statewide beach erosion. Along the southeastern coast where there are numerous stabilized inlets, and the volume of sediment transported in the littoral drift is relatively small (e.g., $<$50,000–100,000 $m^3 a^{-1}$), littoral drift blockers appear to cause about 90% of the beach erosion. Most of the beach erosion here is thus anthropogenically induced and is, at least theoretically, quite preventable from a technical point of view. In practice, however, remediation of the beach erosion problem is politically recalcitrant.

The causes of beach erosion are complicated interacting processes and it is emphasized that beach nourishment only treats erosional symptoms and does not eliminate the causes. Beachfills are sacrificial in the sense that they are not permanent solutions to the beach erosion problem; they thus provide only temporary protection and it is anticipated that replenishment will be repetitive.

Design of beach nourishment projects

Beach nourishment projects are designed to: (1) increase dune and berm dimensions (i.e., height, length, and width), (2) advance the coastline

Figure B26 Hillsboro Inlet, Broward County, southeast Florida. Erosion of the downdrift coast, seen in the coastal offset (photo center), was caused by stabilization of this inlet by jetties. Subsequent construction of a weir jetty on the updrift side of the inlet permitted sand to overwash into an interior sand trap on the landward side of the jetty. A floating dredge sucks the sand from the trap and pumps it to the downdrift side of the inlet via a submerged pipeline. Dredging is mostly conducted during storms when sediment is overwashed through the weir because the trap quickly fills with sediment. If the sand trap is not cleared as it fills, excess sediment spills into the navigation channel where it becomes a hazard to boaters. This sand bypassing arrangement moves 100% of the estimated net littoral drift to the downdrift beach which is thus maintained in a healthy state.

seaward, (3) reduce storm damage from flooding and wave action, and (4) widen the recreational beach area. Beach nourishment projects are complicated technical procedures that require careful preparation for successful execution of site-specific engineering design (Finkl and Walker, 2002). The scale of mechanical sediment supply is quite variable, ranging from large enterprises that involve federal and local partnerships where sometimes tens of millions of cubic meters of sand are involved to small-beach restorations that may require less than 50,000 m^3 of sand. Equipment for obtaining, transporting, and placement of sand on the beach varies with the scale of the project. Large projects require robust equipment that can handle large volumes under high-energy conditions for open ocean dredging (Figure B27). Placement of sand on small, protected beaches (e.g., pocket beaches, embayed shores) (see also entry on *Bay Beaches*) can often be achieved with small dredges during fair weather conditions (Figure B28). In either case, sediment is often redistributed along the shore by front-end loaders, graders, and tractors to achieve the design profile (see also Figures B24 and B30).

Emergency repair of *erosional hot spots* (localized coastal segments where there is increased erosion and rapid shoreline retreat that dramatically exceeds background rates of erosion, as described by Finkl and Kerwin, 1997) that develop during storms may require only a few thousand cubic meters of sediment until more thorough corrective action can be initiated. Beach nourishment the world over is based on the application of natural sediments, mostly beach-quality sands derived from offshore dredging. Many developed nations now recycle glass products and volumes of recycled glass cullet are increasing yearly. The State of Florida, for example, annually produces in excess of 130,000 tonnes of surplus glass cullet that could be made available for beach nourishment (Finkl and Kerwin, 1997). Glass cullet, a chemically inert form of silica, can be graded (mechanically ground) to desired colors and grain sizes to perfectly match native beach sands. For small projects, costs per cubic meter of placed cullet are usually less than beachfill sands.

The design process specifies the quantity, configuration, and timing of sediment distribution along a specific coastal segment to emulate natural storm protection or recreational area, or both. The design must consider rates of long-term (background) erosion as well as temporal impacts of storms and wave climate to adequately address variables associated with the quantity, quality, and placement of beachfill along the shore. As a general rule, sediment comprising the nourished beach is anticipated to erode at least as fast as the background rate of the pre-nourished coastline. It is usually observed in practice that sediment volume loss rates and coastline retreat for artificial beaches are significantly greater than historical rates for the natural beach (e.g., Dean, 2000), even when differences in grain size and sorting are taken into account (e.g., Ashley *et al*., 1987). Although an allowance for continued erosion of beachfill is part of the design assessment, the purpose of beachfill design is to maximize the longevity of artificial beaches. The designs can only be optimized by changing the morphological configuration of the beachfill or by the choice of the fill material. The grain-size of borrow material was traditionally considered to be the most important factor for optimizing beachfill. Studies by Eitner (1996), however, indicate that grain-size has little effect on beachfill longevity because grain-size influences the critical threshold stress to a lesser extent than does the grain

Figure B27 Large ocean-going dredge operating off the coast of southeast Florida. The *Illinois*, owned and operated by the Great Lakes Dock and Dredge Company (Oak Brook, IL, USA), is one of the largest suction cutterhead dredges in the United States. The 98-m long dredge, which digs to a nominal depth of 32 m, has a 760 mm discharge diameter and can pump sand 7,635 m without booster pumps. With total installed power of 8,400 kW, and 1,400 kW cutter power coupled with 662,000 L fuel capacity, the *Illinois* is ideal for working offshore for long periods in rough weather conditions. The dredge is not self-propelled and must be moved to projects by tugboats. When on site, the dredge has a large swing radius using a system of anchors and cables to maneuver within about a 300 m range without resetting the anchors. This dredge can work in swell up to 2 m high and usually operates 24 h a day when possible. Beach-quality sand from the borrow area is pumped in a pipeline to the beach. Because the dredge might operate up to 2 km offshore, the pipeline is floated on the surface near the dredge for ease of maneuverability and submerged near the shore for safety of boaters and beach users.

A

B

Figure B28 Small-scale beach renourishment at Alegre Beach, Santa Catarina, Brazil. Local erosion of a beach on the downdrift side of an inlet required sand renourishment for shore protection, recreation, and marine fisheries (beach launching sites for local fisherman). (A) The small coastal dredge pumps sand from the seabed offshore to the beach via a submerged pipeline. (B) Sediment is pumped up onto the upper beach-face in a slurry, as shown in the subaerial plume in the photo center. Extra pipe is stored on the berm for future use as the dredge moves about in the offshore borrow area.

density. Only a coarser material such as gravel, which also has a significantly higher critical threshold stress, may effectively extend beachfill longevity. In general, however, most researchers agree that coarser grain sizes produce steeper, more stable, and longer-lived fills (e.g., Bruun, 1990; Pilkey, 1990; Smith, 1990; Dean, 1991). Controversy does, however, surround the kinds of methods used to estimate erosion rates. Houston (1990), for example, emphasizes that designer extrapolations or predictions of replenished beach life of one to several decades is not advisable because beach conditions are too variable and especially vulnerable to cycles of storminess.

There are various methods of *beach nourishment design* that are complimentary in the overall process of optimizing *project performance*. Potential designs are initially evaluated at a preliminary level in which the anticipated project performance is predicted using simple and relatively inexpensive methods. After the performance characteristics are compared with project objectives, the design is refined until the performance predictions confirm an optimal design. For sites without complex boundaries (i.e., straight beaches without terminal groins, inlets, or headlands), prediction tools correctly estimate the time required for renourishment to within approximately 30% of actual project performance (NRC, 1995). Subsequent to establishing the preliminary design, more sophisticated predictive methods are used to optimize the design. This bimodal approach checks preliminary and advanced methods of design, facilitates a rapid and efficient convergence to final design, and provides a clear perspective of how well the design parameters fit project requirements. If the predicted volumetric losses, based on preliminary and advanced methods, differ by more than 50%, the essential elements of the design procedure are reviewed and revised, where necessary.

The design beach

The *design profile* is the shore-normal cross section that an *equilibrated beach* is anticipated to assume. The best estimate of this profile is obtained by the seaward transfer of the natural beach profile by the amount of beach widening that is required (USACE, 1992). Estimates of beachfill volumes are generally increased if the borrow material is finer grained than the native sand. The *construction profile* is the cross section that the contractor is required to achieve. Because the constructed beach, which contains design fill and the advanced-fill volumes, is often steeper than the design cross section due to construction limitations, it is also usually significantly wider than the design profile. Wave action adjusts the construction cross section to a flatter dynamically equilibrated slope within the first few months to a year after placement of the beachfill (cf. Figure B31). Because the dynamically adjusted profile contains *design and advanced fills*, it is wider than the design width during the *nourishment interval* (the time elapsed between nourishment episodes). At the time of renourishment, the design and equilibrium profile are theoretically equal (NRC, 1995).

Mechanical deposition of sediment along a beach nourishment site, during initial construction or renourishment, may not correspond to the natural profile of the beach at the time of placement (Figure B29). In the United States, use of a construction profile rather than a natural profile is the normal placement practice. It is customary for nourishment designs in the United States to establish uniform beach width along a project's length. It is also standard practice to provide sufficient sand to nourish the entire profile from the dune to the depth of significant sand removal, the so-called *depth of closure* (DoC). The DoC is a term used by engineers to define the *depth of active sediment* movement on the seabed (see also entry on *Depth of Closure*). Other terms that are related to this critical concept include profile pinch-off depth, critical depth, depth of active profile, maximum depth of beach erosion, seaward limit of nearshore eroding wave processes, and seaward limit of constructive wave processes. The DoC in beachfill design is defined as "The depth of closure for a given or characteristic time interval is the most landward depth seaward of which there is no significant change in bottom elevation and no significant net sediment transport between the nearshore and the offshore" (Kraus *et al*., 1998). This definition applies to the open coast where nearshore waves and wave-induced currents are

Figure B29 Emergency beach renourishment at Gravatá Beach, Santa Catarina, Brazil. A coastal highway and commercial infrastructure was threatened by beach erosion during a strong southeaster that brought heavy surf conditions to the coast during the Southern Hemisphere winter of 1999. Removal of the beach by wave and current action, undermined part of the coastal highway interrupting coastal access. Although of finer grain size than the native beach sand, emergency fill was trucked to the site for immediate shore protection.

the dominant sediment-transporting mechanisms. The definition of the DoC infers or stipulates that: (1) the landward water depth at which no sediment change occurs can be reliably identified, (2) there is an estimate of no significant change in bottom elevation and no significant net cross-shore sediment transport, (3) a time frame is related to the renourishment interval or design life of the project, and (4) at the DoC cross-shore transport processes are effectively decoupled from transport processes occurring farther offshore.

Estimates of fill requirements are based on the geometric transfer of the active cross-shore profile seaward by the design amount. If the beachfill grain size matches the native sand and there are no rock outcrops, seawalls, or groins, the design profile (shore-normal cross section) at each alongshore range marker (permanent locations of cross-shore survey sites are typically spaced every 330 m along Atlantic and Gulf coast beaches in Florida) should ideally match the dynamically stable shape of the native beach profile. Cross sections may be more closely spaced in beach nourishment project areas for better engineering control. Enough sediment is included in the design to nourish the entire profile (Hanson and Lillycrop, 1988).

The total sediment volume is independent of the cross-shore profile because the shape of *the renourished profile* is parallel and similar to the existing natural profile. Extra fill is required, however, in front of seawalls in order to achieve the proposed berm elevation. After these seawall volumes are calculated, estimates of nourishment fill volumes are based on seaward translocation of the entire profile. It is emphasized that sand is usually needed along the entire profile, both above and below the water because the beach, by definition, retains subaerial (berm) and submarine (beachface) sections. Placement of the required extra fill volume in front of seawalls typically moves the high-tide shoreline farther seaward than adjoining non-seawall segments. This design requirement, however, causes alongshore gradients in littoral drift that tend to become erosional hot spots. An alternative to providing the additional seawall volumes is to build narrower berms in front of seawalls. Narrower berms are advantageous because they reduce littoral drift gradients that are set up by overly wide sections of nourished beaches in front of seawalls. Similar levels of storm protection (for uplands) are provided by narrower berms when they are backed by seawalls compared to wider berms without them. In many instances, however, beach nourishment in front of seawalls can become problematic, especially where coastline retreat extends landward along coastal segments adjacent to the seawall and where there is deep wave scour in front of the wall (see discussions in Kraus and Pilkey, 1988).

Coastal engineers attempting to predict beach washout and profile response seaward of seawalls often employ beach and dune recession models. Commonly used approaches include EDUNE (Kriebel, 1986), SBEACH (Larson and Kraus, 1990), and GENESIS (Hanson, 1989; Hanson and Kraus, 1989). The numerical models predict the evolution of the cross-shore beach profile toward the so-called equilibrium storm profile (NRC, 1995). Both models are based on principles related to the disparity between actual and equilibrium (theoretical) wave-energy dissipation per unit volume of water within the surf zone. For convenience of calculations, the models assume that sand eroded from the upper beach deposits offshore, with no loss or gain of material to the profile. It is well-known, however, that beach sediment is often transported offshore and is lost from the littoral drift system (e.g., Pilarczyk, 1990). Estimates of storm surge used in coastline recession models, and for calculating run up, are based on USACE (1984, 1986, 1989) engineering manuals. Storm-surge frequencies and extents of coastal flooding are also deployed by the Federal Emergency Management Agency (FEMA) and the National Oceanographic and Atmospheric Administration (NOAA). Storm hydrographs are thus obtained from FEMA, NOAA, and universities to generate probabilities of storm-induced shoreline recession. Wave statistics can be obtained from wave gauge records, published summaries of observations, or wave hindcast estimates such as the Wave Information Study (USACE, 1989). Other methods such as the Empirical Simulation Technique (EST) (Borgman *et al.*, 1992) develop joint probability relationships between various multiple parameters contained in historical data records. Using historical storm records as input, the EST statistically develops a much larger storm-response database while maintaining statistical similarities to original data and it is thus possible to achieve estimates of storm-induced beach recession (Howard and Creed, 2000).

Protocols for overfill on design beaches

Beachfill is usually dispersed out of the nourishment area (i.e., away from the replenished or artificial beach) to adjacent shores or deeper water. The process leading to a decrease in beachfill volume is referred to as "loss," although this sand still temporarily contributes to the stability of the shore in general, but not at the original location. From the point of view of sediment transport, the sand is not "lost" because it is partly retained in the littoral system. From the perspective of the beach manager, however, migration of sediment away from the beachfill represents a tangible decrease of dry beach area.

Erosion of nourished beaches feature two distinct components: (1) the linear regression of the volume of sand in the coastal profile and (2) additional erosion arising from the newly nourished shoreline which becomes more exposed (lying more seaward) than adjacent shore up- and downdrift (Verhagen, 1996). Sediment losses alongshore as well as adjustment or equilibration of the constructed cross-shore profile are responsible for the so-called "additional erosion." In cases where the volume loss associated with the coastal erosion is large compared to the quantity of the beachfill and where the previous rate of erosion is known, a multiplier is used to compensate for all sand loss. Verhagen (1996) suggests a value of 40% extra fill. The presence of structures such as seawalls, due to their interaction with coastal processes, may also require additional fill.

The term *advanced fill* refers to the eroded part of the beach profile before nourishment becomes necessary. The volume and areal distribution of advanced fill is estimated from analysis of the historical rate of erosion and shoreline change. The potential impact of *project fill* on coastal processes is an additional consideration that is taken in account. Procedures used to make these estimates include the historical coastline change method (USACE, 1991) or numerical methods (Hanson and Kraus, 1989). The *historical shoreline change method* assumes that the nourished beach will erode at the same rate as the prenourished beach. This method is commonly employed by beach designers (based on survey results) but can yield a significant underestimate of nourishment requirements (NRC, 1995).

Most long-term erosion (as opposed to episodic storm erosion or development of erosional hot spots) of a nourished beach is initiated by increased gradients of littoral drift along the project length. Major littoral drift gradients affecting the stability of nourished beaches are the preexisting background (regional) rates or historical erosion of the prenourished shore and stresses associated with the high-tide shoreline salient that was advanced seaward by the project fill. These perturbations of normal coastal processes are the cause of *end losses* and *spreading* of the fill. All of these littoral drift gradients interact with the nourished beach causing a progressive loss of fill. Exclusive consideration of the background erosion rate neglects end (and spreading) losses, which causes an underestimate of nourishment volume and overestimate of project life. Although losses from the project due to spreading cause accretion on adjacent beaches, they must be included in the advanced-fill design in order to achieve performance objectives (NRC, 1995).

Nourishment profiles

Models of beachfill placement depend on renourishment design schemes that are selected by considering static and dynamic peculiarities of the site, fill requirements, temporal and spatial distribution of natural habitats, and costs. Some of the more common approaches include: (1) placing all of the sand in a dune behind the active beach, (2) using the nourished sand to build a wider and higher berm above mean water level, (3) distributing fill material over the entire beach profile (above and below water), or (4) placing sand offshore in an artificial bar (NRC, 1995). The approach taken partly depends on the location of the source material and the method of delivery to the beach. If the borrow site is a quarry on land and the sand is transported by trucks to the beach (cf. Figure B28), placement on the berm or in a dune is generally most economical. If the material is pumped shoreward from offshore ocean-going dredges (cf. Figure B26), it is usually more practical to place the sand directly on the beach, in the nearshore zone, or to build an artificial bar. If pumped onshore in a sand-water slurry, the sand is subsequently redistributed by grader or bulldozer across the shore to form a more natural beach profile (Figure B30).

The use of large dunes (i.e., man-made dikes) fronted by renourished beaches as an effective coastal protection measure has long been recognized in The Netherlands (Verhagen, 1990; Watson and Finkl, 1990). These constructed dune–beach systems are designed to withstand the 1-in-10,000-year condition of wave intensity and storm surge flooding. This extreme level of protection is justified because entire cities lie behind the coastal defenses.

Bruun (1988) advocates nourishing the entire beach profile, which he terms *profile nourishment*. The main advantage of this approach is that the sand is placed in approximately the same configuration as the existing profile, so that drastic initial adjustments are mostly avoided, especially

Figure B30 Restoration of Captiva Beach in 1996 on Sanibel Island, southwestern Florida, showing beach-quality sand pumped from offshore via a submerged pipeline to the shore. The sand–water slurry debauches from the pipe end (lower center foreground) to form an alluvial fan that builds up along the shore. Shell hunters often congregate about the proximal part of the sediment fan hoping to find collector's specimens. Shore birds search the distal portions of the fan for crustaceans brought up from the seafloor in the offshore borrow area. Road graders and bulldozers redistribute the pumped sediment into the beachfill design shape that often incorporates overfill that will be sacrificed to the littoral system as the beach equilibrates to the ambient wave climate. The beachfill is sometimes initially somewhat darker than the native beach sand, as shown here, due to a small percentage of organic content. Organic matter in the fill bleaches out within a few weeks and the native-colored sands imperceptibly grade into the fill material. Pipe extensions lie at the foot of the foredunes (photo, lower right) and mark the approximate position of the back berm (photo: Courtesy of Coastal Planning and Engineering, Inc., Boca Raton, FL).

the rapid erosion of the nourished berm. When wave action undermines the newly constructed berm, a beach scarp frequently forms along the length of the project fill. These scarps can pose hazards to beachgoers trying to gain access to the water from the berm (Figure B31). In some cases, foot traffic across the scarp tramples the steep slope to a flatter one that cuts into the beachfill. These cuts or beach tracks can provide ingressive pathways for surge and run-up which in turn can accelerate erosion of the project fill volume.

Artificial nourishment of eroded beaches can be indirectly achieved by placing dredged sand in the offshore zone (McLellan, 1990). Dredged material is deposited in shallow water, typically using a split-hull barge either as a mound or shaped as a long liner ridge that simulates a shore-parallel sand bar. It is anticipated that the sand deposited in the offshore mound or artificial bar will migrate onto the beach. Prior to welding onto the beachface, the bar causes waves to break farther offshore, a process that reduces the wave energy on the beach in the lee of the bar. The disposal depth of the offshore nourishment should be shallower than the seaward boundary of active sediment transport (as defined by normal to moderately elevated energy conditions) so that sediment quickly moves onto the subaerial beach.

Mechanical bypass systems

Shore protection structures (*q.v.*) or other coastal construction works can interrupt littoral drift flow patterns and trap sediment near structures, within navigational entrances to port and harbors, and in flood- or ebb-tidal deltas. Sediment trapping by littoral drift blockers causes downdrift beach erosion (e.g., Finkl, 1993; Bruun, 1995). In order to mitigate the downdrift effects of sand starvation along the coast, it is necessary to move sand around barriers in order to supply beaches with sediment. Due to losses of sediment offshore (see previous discussion), the quantity needed for downdrift beach nourishment may be greater than the trapped sediment volume. Some bypassing systems that are geared for normal use may be overwhelmed during large storms while others function best during or immediately after storms when sediment is brought to the *sand trap* (a dredge pit that collects sediment).

Fixed bypassing systems generally are less effective and more expensive to run than mobile floating systems (Bruun, 1993). Most bypassing plants work at less than 50% efficiency, and some at 30%, which means that less than half of the drift is bypassed to the downdrift beaches. The combination of periodic beach replenishment and innovative bypassing techniques is an option that can restore longshore sediment transport and greatly reduce beach erosion (Bruun, 1996). Suggested new alternatives include mobilization of the bypass intakes on rails or cranes, implementation of jet pumps, or seabed fluidizers (Bruun and Willekes, 1992).

Several different kinds of mechanical bypassing systems are used effectively in a variety of coastal settings: (1) mobile dredges in the harbor and or entrance (e.g., Santa Cruz, CA), (2) movable dredge in the lee of a detached breakwater that forms an updrift sand trap (e.g., Channel Islands and Port Hueneme, CA), (3) floating dredge within an entrance using a weir jetty on the updrift side (e.g., Hillsboro Inlet, FL; Boca Raton, FL; Masonboro Inlet, NC; Perdido Pass, AL), (4) fixed pump with dredge mounted on a movable boom (Lake Worth Entrance, FL; South Lake Worth Inlet, FL), (5) jet pumps (eductor) mounted on a movable crane, with main water supply and booster pumps in a fixed building (e.g., Indian River Inlet, Delaware) (NRC, 1995). These, and other installations, and their operational performances are described in engineering and design manuals (e.g., USACE, 1991) which provide guidance for the design and evaluation of sand bypassing systems.

Veneer beachfills

Veneer fills are beach-quality sands that are placed over a relatively large volume of material that is generally not suitable for beach nourishment. The unsatisfactory materials, which may be either grossly coarser or finer

Figure B31 Example of a large beach-replenishment project involving more than 6 million m^3 of sand that was pumped from offshore in southeast Florida near Port Everglades (Fort Lauderdale). As shown in the photo, the beachfill has been eroded back to the native sand beach (left) and the built beach is perched about 2 m above the eroded beach. The beach scarp that developed during erosion of the fill was graded by dragging a heavy I-beam across the sand. The more gradual slope facilitated access to the water by reducing the hazardous vertical face of the scarp.

than normal beach sand, remain as an underlayer beneath the thin surface sand veneer. Veneer beachfills are thus used in situations where beach-quality sand is not available in sufficient quantities to economically undertake a nourishment project. The usual reason for placing a veneer fill is based in economics where the cost is prohibitive if a cross section is totally built of beach-quality sand. Veneer fills are of two basic types: (1) fills where the underlying materials are coarser than typical beach sands (e.g., boulders, coral, rocks) and (2) fills where the underlying materials are finer than typical beach sands (e.g., silts or silty sands where the median grain size is much smaller than native sand). In the United States, veneer beachfills have been used in Corpus Christi, TX; Key West, FL; and Grand Isle, LA (NRC, 1995). A fundamental design problem associated with veneer fills involves selection of a veneer that is thick-enough, so that it will not erode away and expose the underlayer during storms or before scheduled replenishment. Although variable, depending on the local conditions, the thickness of the veneer must provide a sedimentary envelope that incorporates profile variations without compromise.

Pros and cons

In the United States, there are two schools of thought regarding beach nourishment; the larger group favors artificial placement of sand on eroding beaches as part of shore protection measures while the other smaller group discourages the practice on the grounds of environmental, economic, sociological, or political grounds. A recent study by the US National Research Council (NRC) examined the diversity of viewpoints about the success or failure of nourishment projects (Table B2). It was concluded that the factors involved include a large number of interest groups who have different "... viewpoints, objectives, needs, and ideas ..." (NRC, 1995, p. 41).

In the 1980s and early 1990s, there was much debate about the pros and cons of beach nourishment with many illuminating facts coming from both sides of the issue (Pilkey, 1990). As persuasive as many arguments were, the end result was that federal support for new beach nourishment was largely withdrawn when the US Congress removed the USACE from many new projects by reducing or eliminating funding (e.g., Finkl, 1996). If beach renourishment is to continue as a shore-protection measure in the United States, local funds will have to support the practice along many stretches of the coast.

Proponents of beach nourishment favor continuance of the engineering practice for many reasons, the most important of which feature shore protection (mainly flood control against storm surge) and economic value in terms of income from recreational use. The arguments for beach nourishment are legion and include those factors already indicated as part of the needs for shore protection.

Antagonistic to views of beach nourishment are concepts that focus on environmental impacts of offshore dredging (e.g., Nelson, 1993), especially near sensitive environments such as coral reefs and sea-grass beds, and placement of sand along the shore which buries meiofauna and infauna, or which may adversely impact rare species such as sabellariid worm reefs. Divergent opinions also focus on expenditures of public funds for coastal segments that do not provide public access to the beach or interpretations of coastal management practices that call for retreat from the coast. Other issues that are sometimes raised concern beachfill performance (i.e., durability, longevity, half-lives of replenishments) which is keyed into the design life of renourished beaches (e.g., Houston, 1991; Davis *et al.*, 1993; Farrell, 1995). The USACE, for example, often estimated and advertised life spans of a decade or so for many proposed projects. Studies subsequently found that, on average, the life spans of renourished beaches were usually less than anticipated. Durabilities of renourished Atlantic beaches, for example, were found to have a half-life of about 4–5 years. (Pilkey and Clayton, 1989). The percentage of renourished beaches lasting more than 5 years along Atlantic coastal barriers averaged about 65% while those on Gulf and Pacific coasts, respectively, averaged about 75% and 65% (Leonard *et al.*, 1990b; Dixon and Pilkey, 1991; Trembanis and Pilkey, 1998).

Conclusions

Among the different approaches to beach nourishment the world over, are various techniques that essentially boil down to methods of placing suitable sediment along the shore to: (1) maintain an existing but eroding beach, (2) create a new beach where none existed before, or (3) to improve a seriously degraded former beach. No matter the actual method of beach nourishment design that involved acquisition of suitable sediment and placement along the shore, this soft engineering approach to erosion mitigation must be regarded as a temporary solution to a chronic problem. In spite of the fact that in the United States, for example, there has been more than a half-century of experience, beach nourishment remains a procedure with unclear universal application. Experience has shown that there are no simple rules that work everywhere because it is now widely appreciated that local site characteristics must be important criteria in successful design. Peculiarities of local conditions related to bathymetry, sediment grain size or shape and composition, exposure and orientation of the beach to prevailing and storm wind patterns, wave climate, and para- and diabathic sediment

Table B2 Evaluation of beach nourishment projects, based on objectives, interpretation of criteria for success, and various measures of performance (modified from NRC, 1995)

Objective	Criteria for success	Measures of performance
Create, improve, or maintain a recreational beach	A viable recreational asset (acceptable dry-beach width and carrying capacity) during the tourist season	Periodic surveys of beach width using quantitative techniques. Assessment of beach visits and carrying capacity via such methods as aerial photography
Protect coastal infrastructure from wave attack and flooding by storm surge	Sufficient sand, gravel, or cobbles remaining in a configuration that blocks or dissipates wave energy and surge. Hard structures may be included in the solution	Evaluation of structural flooding damage following storms that do not exceed the project design
Maintain an intact dune seawall system	No overtopping during a storm that does not exceed the design water level and wave height limits	Verification of stabilized shore line position
Create, restore, or maintain beach habitat	Episodic erosional extremes do not exceed the design profile. Structures, if allowed, remain intact. Postfill erosion rates comparable to historical values	Profile surveys showing that the sedimentary volume and configuration meet or exceed the design profile
Protect the environment	Sediment volume, areal extent, and condition plus vegetation of the back beach or dune meeting environmental needs	Observation and survey of habitat characteristics and conditions
Avoid long-term ecological changes in affected habitats	Return to pre-nourishment (native beach) conditions within an acceptable time frame	Periodic monitoring of faunal assemblages of critical concern

flux pathways can all affect shore erosion and beach stability. Intricacies of shore processes and their interactions with engineering works such as jetties, dredged channels, groins, and breakwaters, for example, can exacerbate natural shore erosion. Fortunately, it is now realized that many shore protection structures are themselves the main causes of accelerated beach erosion. In southeast Florida, for example, stabilized navigational entrances (i.e., jettied tidal inlets) are responsible for about 90% of the beach erosion problem (e.g., Finkl and Esteves, 1998). As formidable as this figure may seem, it is now evident that improved sediment bypassing at littoral drift blockers, as described by Bruun (1995), can significantly enhance beach nourishment efforts by prolonging what are relatively short life spans of placed sediments. Using a combination weir jetty, interior sand trap, and floating suction cutterhead dredge, the Hillsboro Inlet in Broward County (southeast Atlantic coast of Florida), is able to annually bypass 100% of the estimated net littoral drift (Finkl, 1993). Thus, in many instances, improved sand bypassing at inlets can maintain sediment transport alongshore to supply downdrift beaches with incremental sand.

Shore protection via beach nourishment comes, however, with a high price tag but there often are few options that are practical. Retreat from the shore in highly developed coastal infrastructures is not possible nor is a passive approach where structures or facilities are threatened by beach erosion or coastal flooding. The Dutch, for example, have taken an aggressive approach by actually reclaiming land from the seabed by diking and poldering (see Walker, 1988). Elsewhere, most of the world's developed shores thus face the promise of attempting to maintain present coastlines via beach nourishment.

Even though beach nourishment is the shore protection measure of choice for many coastal managers, the future of the procedure in the short-term (less than 50 years beyond today) may seem bright but in the long term (more than 100 years from today) the prognosis would be poor. If the natural rise in mean sea level (see entry on *Sea-Level Rise*) continues to be exacerbated by human action to the point that relative sea level continues to increase, many coastal areas will experience inability to locate suitable beachfill materials in sufficient quantity to support artificial nourishment. As beachfill materials become scarcer due to increased demands, project costs will escalate, but cost/benefit ratios will probably be favorably maintained because of higher property values per length of coastal segment. Further, if the general rise in eustatic sea level accelerates as some researchers predict, renourished and constructed beaches will be no match for increased vulnerabilities from erosion and storm surges. The problem is, unfortunately, growing as more and more stretches of shore are developed.

Recommendations to improve performance (i.e., life span, durability of beach nourishment projects) include mapping of the shore zone (both subaerial and submarine features) to better understand the topographic features and sediment transport pathways that are related to coastal stability in the local area. Post-project monitoring is another important step that can help assimilate factors that are related to the degradation of beachfill project life. For now, beach nourishment projects meet the needs of many coastal communities that require protection of beaches.

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Bibliography

Ashley, G.M., Halsey, S.D., and Farrell, S.C., 1987. A study of beachfill longevity: Long Beach Island, NJ. In Kraus, N.C. (ed.), *Coastal Sediments '87*. New York: American Society of Civil Engineers, pp. 1188–1202.

Balsillie, J.H., 1996. Florida's history of beach nourishment and coastal preservation: the early years, 1910–1974. *Proceedings of the 1996 National Conference on Beach Preservation Technology*. Tallahassee, FL: American Shore and Beach Preservation Association, pp. 350–368.

Bird, E.C.F., 1985. *Coastline Changes*. Chichester: Wiley-Interscience.

Bird, E.C.F., 1990. Artificial beach nourishment on the shores of Port Philip Bay, Australia. *Journal of Coastal Research*, **6**(Special issue): 55–68.

Bird, E.C.F., 1996. *Beach Management*. Chichester: Wiley.

Borgman, L.E., Miller, M.C., Butler, H.L., and Reinhard, R.D., 1992. Empirical simulation of future hurricane histories as a tool in engineering and economic analysis. College Station, TX: *ASCE Proceedings, Civil Engineering in the Oceans V*.

Bruun, P., 1988. Profile nourishment: its background and economic advantages. *Journal of Coastal Research*, **4**(2): 219–228.

Bruun, P., 1990. Discussion of: Leonard, L., Clayton, T., and Pilkey, O.H. An analysis of replenished beach design parameters on U.S. East Coast barrier islands. *Journal of Coastal Research*, **6**(1): 15–36. *Journal of Coastal Research*, **6**(4): 1037–1039.

Bruun, P., 1993. An update on sand bypassing procedures and prices. *Journal of Coastal Research*, **18**(Special issue): 277–284.

Bruun, P., 1995. The development of downdrift erosion. *Journal of Coastal Research*, **11**(4): 1242–1257.

Bruun, P., 1996. Navigation and sand bypassing procedures at inlets: Technical management and cost aspects. *Journal of Coastal Research*, **23**(Special issue): 113–119.

Bruun, P., and Willekes, G., 1992. Bypassing and backpassing at harbors, navigation channels, and tidal entrances: Use of shallow-draft hopper dredges with pump-out capabilities. *Journal of Coastal Research*, **4**(4): 687–701.

Charlier, R.H., and De Meyer, C.P., 2000. Ask nature to protect and build-up beaches. *Journal of Coastal Research*, **16**(2): 385–390.

Davis, R.A., Jr., Inglin, D.C., Gibeaut, J.C., Creaser, G.J., Haney, R.L., and Terry, J.B., 1993. Performance of three adjacent but different nourishment projects, Pinellas County, Florida. In Magoon, O. T. (ed.), *Coastal Zone '93*. New York: American Society of Civil Engineers, pp. 379–389.

Dean, R.G., 1991. Equilibrium beach profiles: characteristics and applications. *Journal of Coastal Research*, **7**(1): 53–84.

Dean, R.G., 2000. Beach nourishment design: consideration of sediment characteristics. *Proceedings of the 2000 National Conference on Beach Preservation Technology*. Tallahassee, FL: Florida Shore and Beach Preservation Association, pp. 41–63.

Dixon, K.L., and Pilkey, O.H., 1991. Summary of beach replenishment on the U.S. Gulf of Mexico shoreline. *Journal of Coastal Research*, **7**(1): 249–256.

Dornhelm, R.B., 1995. The Coney Island public beach and boardwalk improvement of 1923. *Shore and Beach*, **63**(1): 7–11.

Douglas, B.C., Kearney, M.S., and Leatherman, S.P., 2000. *Sea Level Rise: History and Consequences*. New York: Academic.

Eitner, V., 1996. The effect of sedimentary texture on beach fill longevity. *Journal of Coastal Research*, **12**(2): 447–461.

Farrell, S. 1995. Beach nourishment at Avalon, New Jersey: A comparison of fill performance with and without submerged breakwaters. In Tait, L.S. (ed.), *Sand Wars, Sand Shortages and Sand-Holding Structures*. Tallahassee, FL: Florida Shore and Beach Preservation Association, pp. 149–164.

Finkl, C.W., 1993. Pre-emptive strategies for enhanced sand bypassing and beach replenishment activities in southeast Florida: A geological perspective. *Journal of Coastal Research*, **18**(Special issue): 59–89.

Finkl, C.W., 1996. What might happen to America's shorelines if artificial beach replenishment is curtailed: a prognosis for southeastern Florida and other sandy regions along regressive coasts. *Journal of Coastal Research*, **12**(l): iii–ix.

Finkl, C.W., and Esteves, L.S., 1998. The state of our shores: a critical evaluation of the distribution, extension, and characterization of beach erosion and protection in Florida. Tallahassee, FL: Florida Shore and Beach Association, pp. 302–318.

Finkl, C.W., and Kerwin, L., 1997. Emergency beach fill from glass cullet: an environmentally green management technique for mitigating erosional 'hot spots' in Florida. *Proceedings of the 10th National Conference on Beach Preservation Technology*. Tallahassee, FL: Florida Shore and Beach Association, pp. 304–319.

Finkl, C.W., and Walker, H.J., 2002. Beach nourishment. In Chen, J., Eisma, D., and Walker, J. (eds.), *Engineered Coasts*. Dordrecht, The Netherlands: Kluwer.

Finkl, C.W., Khalil, S.M., and Andrews, J.L., 1997. Offshore sand sources for beach replenishment: Potential borrows on the continental shelf of the eastern Gulf of Mexico. *Marine Georesources and Geotechnology*, **15**: 155–173.

Hall, J.V., Jr., 1953. Artificially nourished and constructed beaches. *Proceedings, of the 3rd Conference on Coastal Engineering*. Cambridge, MA: pp. 119–136.

Hanson, H., 1989. Genesis—a generalized shoreline change numerical model. *Journal of Coastal Research*, **5**(1): 1–27.

Hanson, H., and Kraus, N.C., 1989. *GENESIS: Generalized Model for Simulating Shoreline Change. Report 1: Reference Manual and Users Guide*. Vicksburg, MS: Coastal Engineering Research Center.

Hanson, M.E., and Lillycrop, W.J., 1988. Evaluation of closure depth and its role in estimating beach fill volume. *Proceedings of Beach Preservation Technology '88*. Tallahassee, FL: Florida Shore and Beach Preservation Association, pp. 107–114.

Hardisty, J., 1990. *Beaches, Form and Process*. London: Unwin Hyman.

Healy, T.R., Kirk, R.M., and deLange, W.P. 1990. Beach renourishment in New Zealand. *Journal of Coastal Research*, **6**(Special issue): 77–90.

Hillyer, T.M., Stakhiv, E.Z., and Sudar, R.A., 1997. An evaluation of the economic performance of the U.S. Army Corps of Engineers Shore Protection Program. *Journal of Coastal Research*, **13**(l): 8–22.
Houston, J.R., 1990. Discussion of: Pilkey, O.H. A time to look back at beach nourishment (editorial) and Leonard, L., Clayton, T., and Pilkey, O.H. An analysis of beach design parameters on U.S. East Coast barrier islands. *Journal of Coastal Research*, **6** (4): 1023–1036.
Houston, J.R., 1991. Beachfill performance. *Shore and Beach*, **59**(3): 15–24.
Howard, S.C., and Creed, C.G., 2000. The empirical simulation technique in predicting storm-induced beach damage. *Proceedings of the 2000 National Conference on Beach Preservation Technology*. Tallahassee, FL: Florida Shore and Beach Preservation Association, pp. 147–157.
Kraus, N.C., and Pilkey, O.H. (eds.), 1988. The effects of seawalls on the beach. *Journal of Coastal Research*, **4**(Special issue): 146p.
Kraus, N.C., Larson, M., and Wise, R., 1998. Depth of closure in beach-fill design. Vicksburg, MS: US Army Engineer Waterways Experiment Station, Coastal Hydraulics Laboratory, CETN II–40 (3/98).
Kriebel, D.L., 1986. Verification study of a dune erosion model. *Shore and Beach*, **54**(3).
Kunz, H., 1993. Sand losses from an artificially nourished beach stabilized by groynes. In Stauble, D.K. and Kraus, N.C. (eds.), *Beach Nourishment Engineering and Management Considerations*. New York: American Society of Civil Engineers, pp. 191–205.
Larson, M., and Kraus, N.C., 1990. *SBEACH: Numerical Model for Simulating Storm-Induced Beach Change. Report 2: Numerical Foundation and Model Tests*. Vicksburg, MS: Coastal Engineering Research Center, *Technical Report* CERC-89-9.
Leatherman, S.P., 1988. Beach response strategies to accelerated sea-level rise. *Proceedings of the 2nd North American Conference on Preparing for Climate Change*. Washington, DC: The Climate Institute, pp. 353–358.
Leonard, L.A., Clayton, T., and Pilkey, O.H., 1990a. An analysis of beach design parameters on U.S. East Coast barrier islands. *Journal of Coastal Research*, **6**(1): 15–36.
Leonard, L.A., Dixon, K.L., and Pilkey, O.H., 1990b. A comparison of beach replenishment activities on U.S. Atlantic, Pacific, and Gulf of Mexico coasts. *Journal of Coastal Research*, **6**(Special issue): 127–140.
Lin, P.C.-P., Hansen, I., and Sasso, R.H., 1996. Combined sand bypassing and navigation improvements at Hillsboro Inlet, Broward County, Florida: The importance of a regional approach. In Tait, L.S. (ed.), *The Future of Beach Nourishment*. Tallahassee, FL: Florida Shore and Beach Preservation Association, pp. 43–59.
Malherbe, B., and LaHousse, B., 1998. Building coastal protection with sand in Belgium. *Journal of Coastal Research*, **26**(Special issue): 101–107.
McLellan, T.N., 1990. Nearshore mound construction using dredged material. *Journal of Coastal Research*, **7**(Special issue): 99–107.
National Research Council (NRC), Committee on Beach Nourishment and Protection, 1995. *Beach Nourishment and Protection*. Washington, DC: National Academy Press.
Nelson, W.G., 1993. Beach restoration in the southeastern US: environmental effects and biological monitoring. *Ocean and Coastal Management*, **19**(2): 157–182.
Peshkov, V.M., 1993. Artificial gravel beaches in the coastal protection. In Kosiyan, R., and Magoon, O.T. (eds.), *Coastlines of the Black Sea. Coastal Zone '93*. New York: American Society of Civil Engineers, pp. 82–102.
Pilarczyk, K.W., 1990. *Coastal Protection*. Rotterdam, The Netherlands: Balkema.
Pilkey, O.H., 1990. A time to look back at beach renourishment. *Journal of Coastal Research*, **6**(1): iii–vii.
Pilkey, O.H., and Clayton, T.D., 1989. Summary of beach replenishment experience on U.S. East Coast barrier islands. *Journal of Coastal Research*, **5**(1): 147–159.
Psuty, N.P., 1988. Sediment budget and dune/beach interaction. *Journal of Coastal Research*, **3**(Special issue): 1–4.
Psuty, N.P., and Moreira, M.E.S.A., 1990. Nourishment of a cliffed coastline, Praia da Rocha, the Algarve, Portugal. *Journal of Coastal Research*, **6**(Special issue): 21–32.
Short, A.D. (ed.), 1999. *Handbook of Beach and Shoreface Morphodynamics*. Chichester: Wiley.
Silvester, R., and Hsu, J.R.C., 1993. *Coastal Stabilization: Innovative Concepts*. Englewood Cliffs, NJ: Prentice-Hall, 578 p.
Smith, A.W.S., 1990. Discussion of: Pilkey, O.H. A time to look back at beach renourishment. *Journal of Coastal Research*, **6**(1): iii–vii; and Leonard, L., Clayton, T., and Pilkey, O.H. An analysis of beach design parameters on U.S. East Coast barrier islands. *Journal of Coastal Research*, **6**(1): 15–36. *Journal of Coastal Research*, **6**(4): 1041–1045.
Strong, W.B., 1994. Beaches, tourism and economic development. *Shore and Beach*, **62**(2): 6–8.
Swart, D.H., 1996. The history of coastal engineering in South Africa. In Kraus, N.C. (ed.), *History and Heritage of Coastal Engineering*. New York: American Society of Civil Engineers, pp. 429–464.
Trembanis, A.C., and Pilkey, O.H., 1998. Summary of beach nourishment along the U.S. Gulf of Mexico shoreline. *Journal of Coastal Research*, **14**(2): 407–417.
U.S. Army Corps of Engineers (USACE), 1984. *Shore Protection Manual*. Washington, DC: U.S. Government Printing Office, U.S. Army Corps of Engineers Publication No. 008-002-00218-9.
U.S. Army Corps of Engineers (USACE), 1986. *Storm Surge Analysis*. Washington, DC: U.S. Government Printing Office. Engineer Manual No. EM 1110-2-1412.
U.S. Army Corps of Engineers (USACE), 1989. *Water Level and Wave Heights for Coastal Engineering Design*. Washington, DC: U.S. Government Printing Office. Engineering Manual 1110-2-1414.
U.S. Army Corps of Engineers (USACE), 1991. *Sand Bypassing System, Engineering and Design Manual*. Washington, DC: U.S. Government Printing Office. Engineer Manual No. EM 1110-2-1616.
U.S. Army Corps of Engineers (USACE), 1992. *Monitoring Coastal Projects*. Washington, DC: U.S. Government Printing Office, Engineer Regulation ER 1110-2-8151.
Verhagen, H.J., 1990. Coastal protection and dune management in The Netherlands. *Journal of Coastal Research*, **6**(1): 169–180.
Verhagen, H.J., 1996. Analysis of beach nourishment schemes. *Journal of Coastal Research*, **12**(l): 179–185.
Walker, H.J. (ed.), 1988. *Artificial Structures and Shorelines*. Dordrecht, The Netherlands: Kluwer Academic Publishers.
Walker, H.J., and Finkl, C.W., 2002. Beach nourishment: case studies. In Chen, J., Eisma, D., and Walker, J. (eds.), *Engineered Coasts*. Dordrecht, The Netherlands: Kluwer Academic Publishers.
Watson, I., and Finkl, C.W., 1990. State of the art in storm-surge protection: The Netherlands Delta Project. *Journal of Coastal Research*, **6**(3): 739–764.
Wiegel, R.L., 1988. Keynote address: some notes on beach nourishment, problems and advancement in beach nourishment. *Proceedings of Beach Preservation Technology '88*. Tallahassee, FL: Florida Shore and Beach Preservation Association, pp. 1–18.
Wiegel, R.L., 1994. Ocean nourishment on the USA Pacific coast. *Shore and Beach*, **62**(1): 11–36.
Young, R.S., Bush, D.M., Pilkey, O.H., and Neal, W.J., 1996. Evaluating shoreline change and associated risk from coastal hazards: an inexpensive qualitative approach. In Berger, A.R., and Iams, W.J. (eds.), *Geoindicators: Assessing Rapid Environmental Changes in Earth Systems*. Rotterdam: Balkema, pp. 193–206.
Zenkovich, V.P., and Schwartz, M.L., 1987. Protecting the Black Sea-Georgian S.S.R. gravel coast. *Journal of Coastal Research*, **3**(2): 201–210.

Cross-references

Beach Erosion
Beach Processes
Bypassing at Littoral Drift Barriers
Cross-Shore Sediment Transport
Dredging of Coastal Environments
Erosion Processes
Management (see Coastal Zone Management)
Natural Hazards
Net Transport
Sandy Coasts
Sediment Budget
Shore Protection Structures
Storm Surge

BEACH PROCESSES

The continuous changes taking place in the coastal zone constitute *beach processes*. A *beach* is one part of the *coastal zone*, which is the transitional area between terrestrial and marine environments. The coastal zone comprises the beach; an underwater region that extends seaward to the depth where waves no longer effect the sea bottom; and inland to features such as *sea cliffs*, *dune fields*, and *estuaries* (*q.v.*). Beach studies focus on understanding spatial and temporal changes in alongshore and cross-shore geomorphic features of the beach (*Beach Features: q.v.*) and in the size and composition of beach sediment. Over time, the coastal zone changes in character in response to changes in wave climate and other physical processes.

Comprehensive knowledge of beach processes is crucial to society because the majority of the world's coastlines are eroding (Thornton *et al.*, 2000). Moreover, *sea-level rise* (*q.v.*) from *global warming* could accelerate coastal land loss, resulting in an increasing rate of loss of coastal habitat and structures. Coastal land loss has a large negative societal impact because approximately two-thirds of the world's population lives adjacent to the ocean or large inland bodies of water. In the United States, more than half of the population lives within 50 miles of the shore, while about 85% of the sandy shore is eroding from a combination of damming of rivers, inlet improvements, sand mining in the coastal zone, sea-level rise, and large storms (*q.v.*). Understanding the processes that cause land loss on sandy coasts is particularly important because beaches are a popular recreational area; are essential to commerce; and protect coastal cliffs, dunes, and structures.

Beaches continually change in response to changes in wave conditions. The changes occur over both short- and long-terms, reflecting both subtle changes associated with daily or weekly variations in tidal level or wave climate and gross changes associated with seasonal variations in wave climate. Beaches usually shift back and forth between a wide, built-up berm with a barless nearshore zone and a small-to-absent berm with one or more well-developed bars in the nearshore. These fluctuations often occur on top of an equilibrium profile that exhibits no net long-term change.

Textbooks that discuss beaches, the coastal zone, and the processes that affect them include Johnson (1919), Guilcher (1958), Shepard (1963), Shepard and Wanless (1971), Bascom (1980), Komar (1983, 1998), Bird (1984), Carter (1988), Carter and Woodroffe (1994), and van Rijn (1998). A report by Thorton *et al.*, (2000) summarizes the state of coastal processes research as of 1999.

Beaches

Many classifications exist to describe different types of coast. Van Rijn's (1998) classification consists of mud, sand, gravel/ shingle, and rocky coasts. In this classification, approximately 10% of the coasts are mud and 15% are terrigenous sand and carbonate sand coasts. Because of their large societal importance, most studies of coastal change have been conducted on sand coasts, which typically are wave-dominated environments (*q.v.*).

The beach is the most prominent visual feature of sand coasts. It is the area of unconsolidated material that extends landward from the low-water line to the place where there is a definite change in material or physiographic form (e.g., a cliff or dune), or to the line of permanent vegetation (*Coastal Boundaries: q.v.*). Beach material can be any combination of sand (grain size = 0.0,625–2 mm), granules (2–4 mm), pebbles (4–64 mm), cobbles (64–256 mm), and boulders (>256 mm) (*Sediment Classification: q.v.*). The grain size on most beaches is in the sand range.

Grain size and composition

In temperate climates, beaches typically consist of quartz and feldspar grains derived from the weathering of terrestrial rocks. Commonly, denser minerals (*heavy minerals*) that are specific to also occur in small percentages. On volcanic islands, the sand frequently includes lava fragments and associated minerals. Worldwide, many beaches have a calcium carbonate fraction from the breakup of shells, concentrations of foraminiferans, and nearby coral reefs (*q.v.*). For example, on the island of Hawaii, beach sand can be black, green, or white. The black sand comes from the erosion of lava beds and decomposition of hot lava flowing into the ocean; the green sand comes from the mineral olivine, which crystallizes when magma cools; and the white sand consists of calcium carbonate. Although calcium carbonate beaches are usually associated with the tropics, beaches in other climes also can have a large shell fraction.

Because the heavy-mineral fraction in beach sand is indicative of the provenance of that sand, heavy minerals can be used to trace the movement of sand along the beach. For example, Trask (1952) showed that the sand reaching the harbor at Santa Barbara, CA comes from the area around Morro Bay, CA, more than 160 km up coast. He established this by using the mineral augite as a tracer. Along the stretch of coast between the two cities, augite occurs in all beach sands, but the only source area is in the vicinity of Morro Bay. The concentration of augite decreases between Morro Bay and Santa Barbara as non-augite sources (e.g., local streams and cliffs) add to the sand moving along the coast.

Beach profile

Figure B32 shows a typical shore-normal cross section, or *profile*, of the coastal zone. The coastal zone can be divided into four major subzones: *backshore*, *foreshore*, *inshore*, and *offshore*. The locations of the beach subzones depend on whether the beach has accreted or eroded (*Beach Erosion: q.v.*). The offshore and inshore are seaward of the low-water line. The former is a relatively flat part of the profile seaward of the breaker zone. The latter includes the breaker and surf zones (*q.v.*). The foreshore is the sloping portion of the profile (typically 2–10°) and encompasses the normal intertidal part of the beach. On an eroding beach, the foreshore could cover the entire beach. The *beach face* is the upper portion of the foreshore that is normally exposed to wave *uprush*; it is often synonymous with the foreshore. On an accreting beach, the *berm crest* is the transitional area between the foreshore and backshore; often it is a striking feature several meters above low water, especially on cobble and boulder beaches (Guilcher, 1958). The backshore extends landward from the berm crest to the edge of the beach; the nearly horizontal part is the *berm*, which forms from the deposition of lower-foreshore sediment that is transported over the berm crest. There can be more than one berm on a backshore. A nearly vertical escarpment (*scarp*) caused by erosion of an earlier berm crest can separate berms on multi-berm beaches.

The loss of beach sand usually corresponds to a gain of sand in the nearshore, and *vice versa.* Beach sand also can be lost to sanddunes, estuaries, and submarme canyons Human activities can result in both beach loss and gain. *Pocket beaches* can be an exception where the sand often shifts alongshore with changes in wave climate. Visually, the result is a shift in the profile of the beach in both the horizontal and vertical. Although such changes are commonly associated with winter

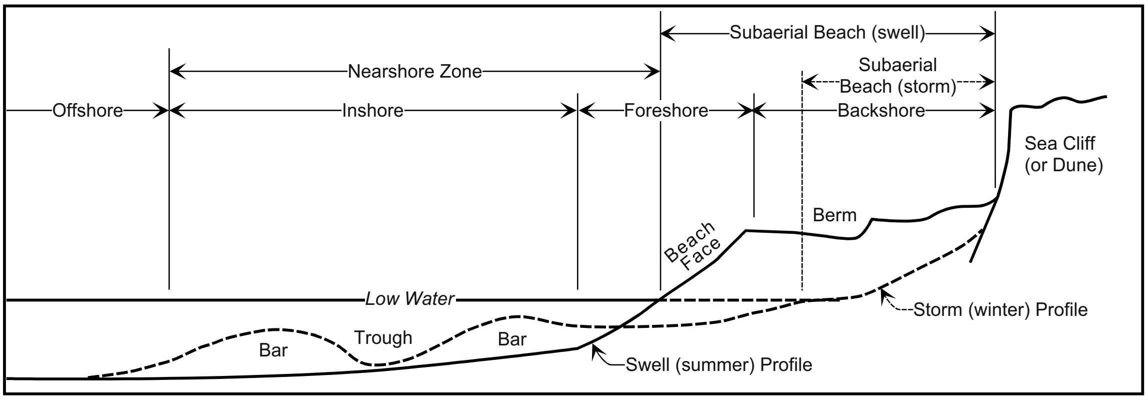

Figure B32 Terminology for the coastal zone along a shore-normal profile.

(erosion) and summer (accretion), in reality they occur any time of year in response to stormy and fair weather. *Storm waves* erode the berm and shift the shoreline (defined as the high-water line or wet–dry boundary) landward. *Swell waves* build up the berm and shift the shoreline seaward.

For sandy beaches, Wiegel (1964) developed a relationship between median grain size, slope of the beach face, and wave climate using data from many US beaches. Grain size and beach slope were measured at mean tide level where the correlation was best (Bascom, 1951). The relationship shows that that the slope depends on two factors—grain size and wave exposure (Figure B33). For any given wave climate, slope increases with increasing grain size. Correspondingly for a given grain size, the smaller the waves, the steeper the beach face. Thus, a beach face becomes flatter when eroding (larger, storm waves) and steeper when accreting (smaller, swell waves).

Beach processes

A complete understanding of beach change requires an understanding of the processes active throughout the coastal zone. Accordingly, beach processes is a subset of *coastal processes*, or *coastal morphodynamics* (morpho-: form, structure; dynamics: motivating or driving forces). Thus, coastal processes involve investigations of the interactions of coastal-zone features and hydrologic, meteorologic, and fluvial forces by means of sediment transport (*q.v.*). Coastal geomorphology and fluid dynamics couple at a continuum of temporal and spatial scales such that the fluid dynamics produces sediment transport, which produces geomorphic change. Progressive modification of the geomorphic features in turn alters boundary conditions for the fluid dynamics, which evolve to produce further changes in sediment-transport patterns and consequently the geomorphic features (Cowell and Thom, 1994).

Coastal processes happen over a wide range of spatial and temporal scales; the upper limits are generally set at 10 km and 1 yr, respectively. For example, the properties of waves entering the coastal zone from deep water and interacting with the nearshore profile determine the overall characteristics of nearshore waves and flows. However, small-scale processes control the turbulent dissipation of breaking waves, bottom boundary layer, and bedform processes that determine the local sediment flux. Cross- and longshore variations in waves, currents, and bottom slope cause spatial gradients in sediment fluxes resulting in changes due to erosion or accretion. Traditionally, the study of coastal processes has been restricted to small and intermediate scales (Thornton *et al.*, 2000); making it but one of several influences on the coastal zone (Figure B34).

The rate of response of geomorphic features to the fluid dynamics depends on scale; larger features take relatively longer to change (temporal scale in Figure B35). Hence, equilibrium is almost instantaneous for small-scale processes, and quasi-equilibrium becomes more noticeable as the geomorphic scale increases. For example, bedform scale and forcing history link the rate of transition between bedform types inside and just outside of the surf zone. Under large waves, significant changes in small-scale bedforms can occur within a single wave cycle, but changes in large-scale bedforms can exhibit significant hysteresis.

Wind, waves, tides, storms, and stream discharge are important driving forces in the coastal zone. Streams transport sediment from the

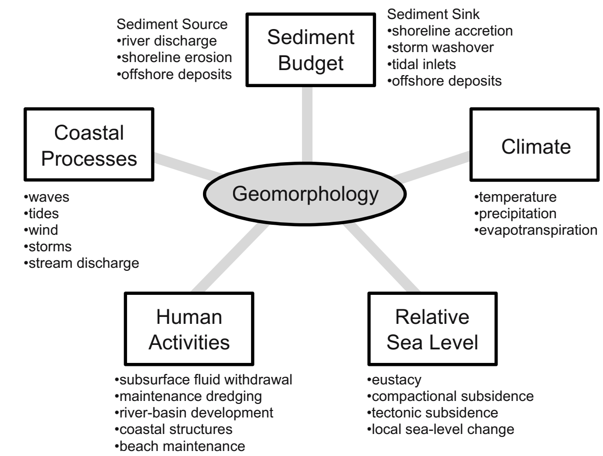

Figure B34 Processes that influence the geomorphology of the coastal zone (Thornton *et al.*, 2000).

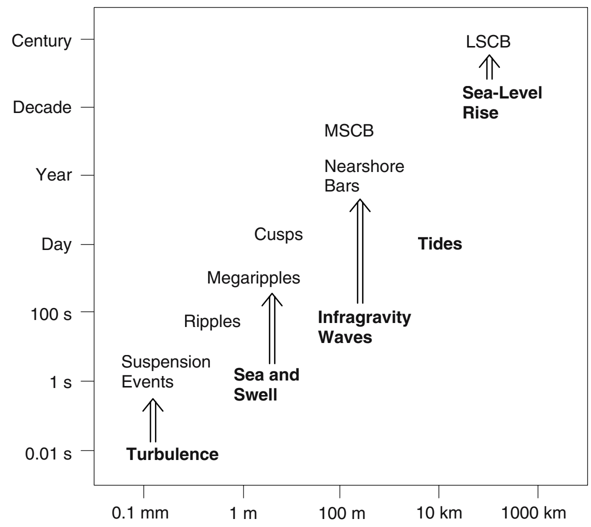

Figure B35 Temporal and spatial scales for coastal processes (Ruessink, 1998). Fluid forces are boldfaced.

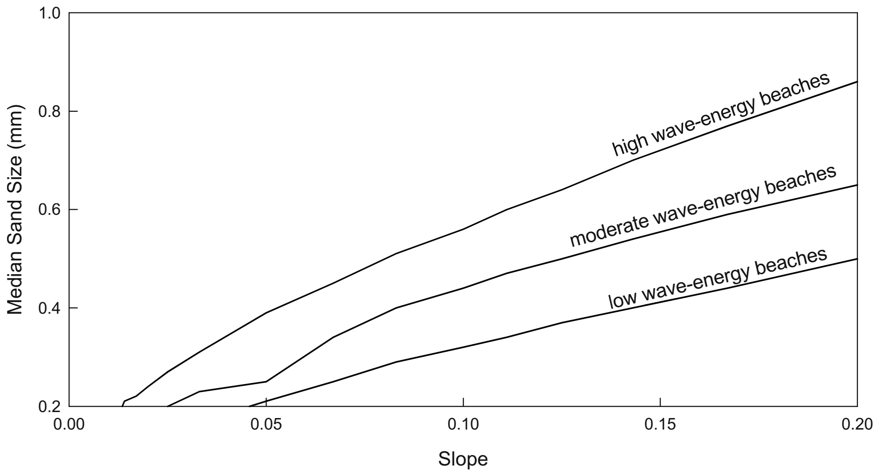

Figure B33 Beach slope as a function of sand size and wave energy.

hinterland to the coastal zone for the other forces to distribute. Wind directly transports beach sand and generates waves (*Meteorological Effects: q.v.*). Waves produce currents that transport sand cross-shore and longshore. Tides play a supporting role by exposing different parts of the beach face to waves and currents. Storms produce strong wind, waves, and elevated sea level, which can cause extensive coastal erosion; movement of sand to the nearshore or across barrier islands (*q.v.*) and spits (*q.v.*); and coastal flooding, as well as intensifying small- and intermediate-scale processes.

Waves are the major source of energy driving beach change in the nearshore. As a wave approaches the coast, it reaches a water depth where it begins to interact with the bottom (*shoal*). That depth occurs when the ratio of water depth to deep-water wavelength is less than 0.5, which is commonly in the depth range of 10–20 m. *Wave base* is the term Bradley (1958) gave to the depth at which normal wave erosion begins (also see Dietz, 1963). *Depth of Closure* (*q.v.*) is the maximum depth of extreme bottom changes; it is a function of the nearshore storm-wave height exceeded 12 h per year and the associated wave period (Hallermeier, 1981). Shoreward, a wave undergoes a systematic transformation (*shoaling*) where wavelength (L) decreases and height (H) increases (H/L is the wave *steepness*). When the ratio of the wave height to water depth reaches 0.73–1.03 (Galvin, 1972), the wave breaks, producing a beachward flow (*bore*). The result is an upward (*run-up*) and downward (*backwash*) flow of water on the beach face (the *swash zone*). If run-up reaches the back of the beach, it can erode cliffs, dunes, and structures.

Sand transport begins soon after waves begin to shoal. Transport volume and velocity increase shoreward in proportion to increasing intensity of the wave-bottom interaction. Whether there is beach erosion or beach accretion depends on wave height and period. Storms raise sea level by piling up water against the coast (*storm surge*) (*q.v.*), and greatly increase wave height and steepness (*storm* waves). Such waves tend to produce beach face erosion with the sediment being moved to the nearshore. Lower, less-steep waves (*swell* waves) produce accretion by moving nearshore sand onto the beach face. Storm and swell waves can occur any time of the year, although the former are more common in the winter and the latter in the summer. Consequently, the belief that erosion is a winter phenomenon (season of storm waves) and accretion a summer phenomenon (season of swell waves) is not completely correct. Basically, the wave climate continuously varies, and the beach face never reaches an equilibrium state where the volume of sand moving up the beach face equals that moving down.

Breaking waves create a circulation system where the water driven shoreward across the surf zone returns to the offshore via a strong, narrow flow called a *rip current* (Figure B36) whose spacing ranges from tens to hundreds of meters. Velocities in rip currents are often strong enough to carry sediment and swimmers bathers alongshore through the breakers; in those cases, a distinct watercolor demarcates the rip current. Often the velocity is too strong to swim against, so people caught in a rip current must swim parallel to shore to escape.

If a wave enters the coastal zone at an angle to the bathymetric contours, its crest bends to align with those contours (*refraction*). If the wave breaks at an angle to the beach, a longshore current develops. Because sand grains move with the flow, longshore sediment transport (*q.v.*) occurs (*littoral drift*). Because the sand grains are subject to both run-up and littoral drift, they follow a zigzag path along the beach face.

Geomorphic features

Based on the temporal and spatial scales of sediment transport and geomorphic change, the coastal zone can be divided into two cross-shore subzones—the upper and middle shoreface—and three longshore subzones—micro, meso, and macro cell. In the upper shoreface, breaking waves and bores generate active sand transport and rapid geomorphic response. In the middle shoreface, slow sand transport rates result in slow geomorphic change. Micro cells include smaller geomorphic features such as ripples and small beach-face features that change in times of a day or less. Meso cells include geomorphic features such as sand bars (*q.v.*), beach profiles, and beach cusps that change in a year or less. Macro cells extend kilometers and include large coastal geomorphic features (Figure B35).

Micro-scale beach features

When waves begin to shoal, there is a back-and-forth water motion at the seafloor with onshore and offshore excursions being equal. The sediment moves the same way, and symmetric *oscillation ripples* form normal to the direction of wave advance. Nearer to shore, the water motion at the seafloor becomes asymmetric because the onshore component of the wave orbits is larger than the offshore one. These *current ripples* have a gentle seaward facing slope and a steep onshore one. Near the breaker zone and in much of surf zone, the bed is flat because of intense water motion. However, current ripples can form in the seaward flowing *rip currents*.

At the upper swash limit, deposition of debris forms scalloped, marginal lines known as *swash marks*. Backwash flowing around small obstacles form seaward opening "V"s, and in some cases, rhombic patterns develop as a result of the minor currents generated in the backwash. Seepage of interstitial water down the foreshore slope at low tide cuts miniature channels termed *rill marks*.

Meso- and macro-scale beach features

Generally, sediment eroded from the foreshore and transported offshore forms one or more longshore bas with a trough shoreward of each one. The bars can extend alongshore for kilometers except for breaks caused by rip currents. Studies by Evans (1940), Keulegan (1948), King and Williams (1949), and Shepard (1950) concluded that there is a strong relationship between the bars and breaking waves. Keulegan, who studied bars formed in a laboratory wave channel, found that wave height and steepness govern the bar position. An increase in wave height moves the bar seaward (deeper water). Holding the wave height constant and reducing the wave period moves the bar shoreward. Starting with a smooth profile, the bar initially forms just shoreward of the breaker position. As it grows, both the bar and breakers migrate landward. When the bar is fully developed, it modifies energy transfer in the surf

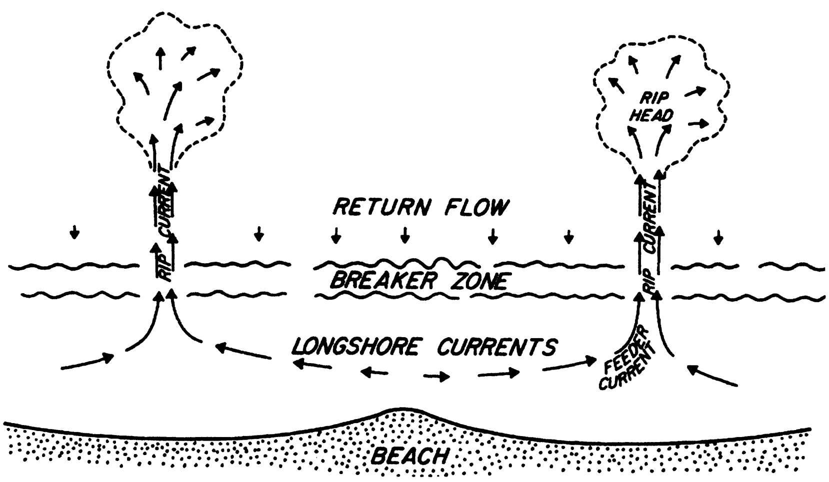

Figure B36 The nearshore circulation cell. Breaking waves create a circulation system where the water-driven shoreward across the surf zone travels alongshore (*longshore current*) and returns to the offshore via a strong, narrow flow (*rip current*) (after Komar, 1998).

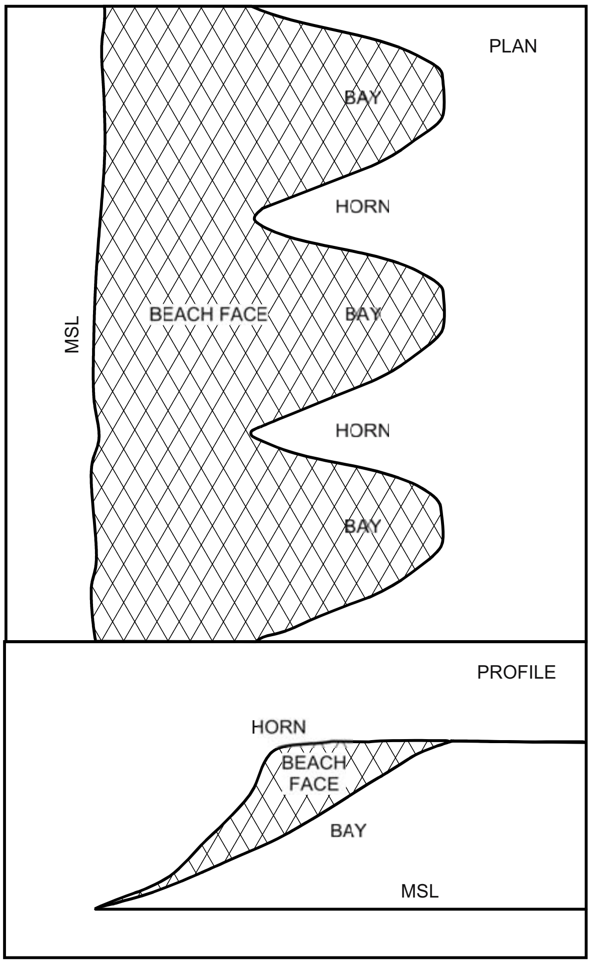

Figure B37 Plan and profile views of a cuspate beach.

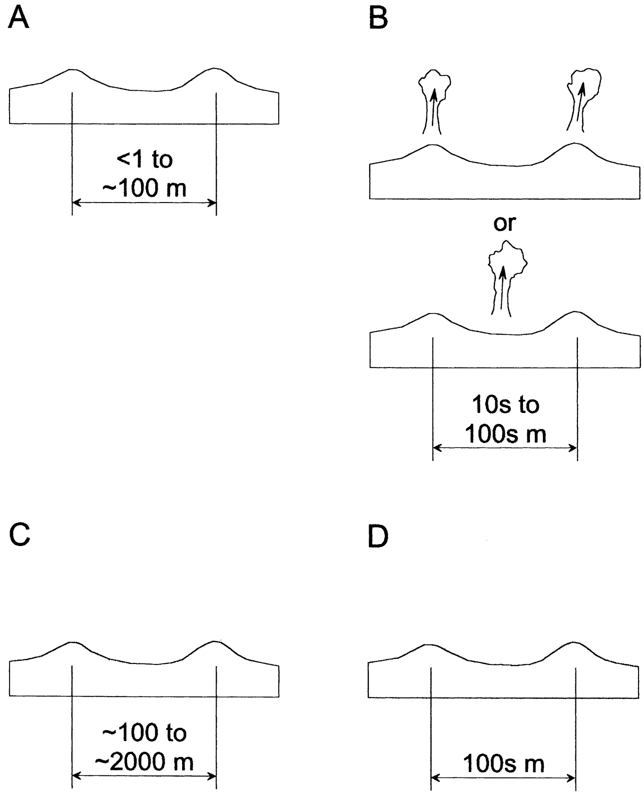

Figure B38 Classification of rhythmic shoreline forms: (A) reflective beach cusps; (B) rip current embayment cusps; (C) crescentic bay cusps; (D) transverse and oblique bars (after Komar, 1998).

zone by reducing the amount of energy that the breaking waves can impart to the waves that reform in the surf zone. Shepard (1950) found that there can be no bar and trough for short-period storm waves because there is not a well-defined breaker zone.

Beaches can be two- or three-dimensional based on the linearity of the berm and beach face. On two-dimensional beaches, the berm crest and foreshore contours are straight and parallel. On three-dimensional beaches, the berm crest and foreshore have rhythmic, crescentic features (*Beach Cusps: q.v.*) that vary greatly in height and length. All cusps are characterized by seaward facing ridges or *horns* separated by *embayments* or *bays* (Figure B37). Attempts to classify these features by size (e.g., Dolan and Ferm, 1968; Dolan *et al*., 1974) have been unsuccessful because there is a large overlap in sizes of cusps formed by different mechanisms. Komar (1983) developed a classification scheme with four types of cusps whose origins can be attributed to different processes of waves and currents within the nearshore (Figure B38). These are reflective beach cusps, rip current embayment-cusp systems, crescentic bar-cusp systems, and transverse and oblique bars (see Figure B38 for wavelengths).

There has been much disagreement as to the origin of reflective beach cusps. One recent theory is that *edge waves*, waves trapped at the shore with net motion in the longshore direction and decreasing amplitude offshore (Holman, 1983), play a role in the formation of beach cusps. In a laboratory wave tank, Guza and Inman (1975) found that beach cusps developed in response to edge waves. In the field, Huntley and Bowen (1978) observed the formation of cusps in the presence of edge waves. Werner and Fink (1993) proposed that beach cusps form through strong, positive feedback between wave run-up and beachface topography.

A rip current embayment-cusp system develops when rip currents erode the beach face creating embayments. The cusps are midway between the embayments (Bowen and Inman, 1969). The cusps correspond to positions of zero longshore sediment transport produced by waves combining with the feeder currents that flow alongshore toward the rip currents (Figure B36). In some cases, rip current erosion can be so extensive that the embayments cut across the beach, exposing foredunes and cliffs to wave attack (Komar and Rea, 1976; Komar and McDougal, 1988). Beach sediment can be deposited in the lee of a rip current so that the cusps correspond to the rip locations. This appears to occur most commonly on steep beaches under relatively low wave conditions (Komar, 1971).

Crescentic bars are rhythmic lunate features with uniform spacing primarily found underwater, commonly on long straight beaches (Shepard, 1952; Hom-ma and Sonu, 1963). They appear to be confined to regions of small- to medium-tidal range (Bowen and Inman, 1971) and form best where the beach slope is low. The generally accepted mechanism for their formation is by edge waves in the infragravity range (Bowen and Inman, 1971).

Transverse and oblique sandbars are nearshore features that do not parallel the beach, and they are welded to the beach face. There are various explanations for their formation that include the rotation of rip-current segmented bars (Sonu, 1972), processes akin to the migration of river bars or the development of river meanders (Bruun, 1954; Sonu, 1969; Dolan, 1971), and the superposition of edge waves (Holman and Bowen, 1982).

Beach change

Beach cycles

Various cycles affect the beach, depending essentially on changes in wave steepness and effective sea level. In regions of tidal action, the swash–backwash zone and breaker zone shift landward and seaward with the flood and ebb tide. The range of the tides and the slope of the foreshore determine the distance over which the shift takes place. With other parameters remaining constant, breaker height will be greater at high than at low tide. At high tide, the prevailing waves approach less hindered over the relatively deeper water of the nearshore, whereas at low tide, approaching waves are modified by the shoaler depths of the gently sloping nearshore bottom. Under this changing regime, scour may be slightly increased at high tide.

At approximately 7.5-day intervals, the tidal range changes from minimum (neap) to maximum (spring). The change from neap to spring tides can produce upper-foreshore erosion with lower-foreshore and nearshore deposition (LaFond, 1938; Inman and Filloux, 1960). LaFond and Rao (1954) postulate the cumulative result of higher effective sea level, higher waves, and a time lag in recovery during the low spring tide cause the redistribution of sand. The opposite sand movement can occur during neap tides.

Effects of beach erosion

Erosion strips sediment from the beach face and moves it to the nearshore or redistributes it alongshore. When enough sand is removed, there is no longer a high, wide beach, and waves can attack coastal features such as cliffs, dunes, and anthropogenic structures. In some places, storms transport beach sand across spits and barrier islands and deposit the sand in adjacent lagoons. If the sand returned to the beach is less than the volume eroded, the beach narrows, and if possible, the shoreline shifts landward.

Shoreline retreat is a natural process that is of little or no concern in unpopulated areas. However, in populated areas, shoreline retreat is a major issue. Several years can pass between storms severe enough to cause significant damage to a stretch of coast. Consequently, many people build or buy homes and other facilities on the coast with the idea that the adjacent beach is permanent. Later they watch storm waves remove the beach sand and directly attack their property or the coastal cliffs and dunes that protect them. Then, affected communities quickly want to know how to save their beaches and protect their homes and facilities. Although beaches usually rebuild after storms, a beach does not always return to its pre-storm position, and the community must take remedial measures to reverse long-term shoreline retreat.

Because storm waves threaten coastal facilities when fronting beaches lack sand, post-storm beach accretion is essential to minimize economic loss in the coastal zone. In areas where there is insufficient beach to protect coastal structures, there are several procedures to prevent or mitigate shoreline retreat. Traditionally, these procedures require building a protective structure (*q.v.*) on the beach or at the landward edge of the backshore. These include *seawalls*, *revetments*, *groins*, and *breakwaters*. While these structures often protect the property behind them, the fronting beach typically does not return because increased water turbulence at the structure prevents sand deposition during swell conditions (see Dean, 1999 for examples). The result is a section of coast with no beach, and if longshore sand transport is not properly taken into account, the shoreline downdrift of a structure also can lose its sand and retreat. Furthermore, structures often fail if improperly designed, allowing coastal retreat to resume.

Beach nourishment (*q.v.*) is another technique used to prevent shoreline retreat by augmenting the native beach sand with sand imported from other areas. Although beach nourishment creates wide beaches, this technique may not provide a long-term solution to beach loss especially where erosion rates are high or there is a persistent problem. A major problem is that the cost of importing sand can be high, especially since the sand should be similar in character to the native sand and because more sand is frequently needed after a storm season. When this technique is successful, there will be a year-around beach for public use and shoreline protection.

Other techniques include relocating coastal structures to allow for shoreline retreat and defining setback lines for coastal development. Shoreline retreat permits nature to take its course, but often is infeasible in populated areas. Setback lines, which are based on historical shoreline retreat rates, need to be implemented before coastal development begins.

Sediment budget and littoral cell

The budget of littoral sediments is simply an application of the principle of conservation of mass to the littoral sediments—the time rate of change of sand within the system depends upon the rate at which sand enters the system versus the rate at which it leaves. An analysis, therefore, involves evaluations of the relative importance of various sediment sources and losses to the nearshore zone, and a comparison of the net gain or loss with the observed rate of beach erosion or accretion.

The budget of littoral sediment involves making assessments of the sedimentary contributions (credits) and losses (debits) and equating these to the net gain or loss (balance of sediments) in a given sedimentary compartment or littoral cell (Bowen and Inman, 1966; Komar, 1996). The balance of sediments between the credits and debits should be approximately equivalent to the local beach erosion or accretion.

Table B3 The budget of littoral sediments

Credit	Debit
Longshore transport into the area	Longshore transport out of the area
River transport	Wind transport away from the beach
Sea cliff erosion	Offshore transport
Biogenous deposition	Solution and abrasion
Hydrogenous deposition	Mining
Wind transport onto beach	
Beach nourishment	

Table B3 summarizes the possible credits and debits of sand for a littoral sedimentary budget, while some of the more important components are diagrammed in Figure B39. In general, the longshore movement of sand into a littoral compartment, river transport, and sea-cliff erosion provide the major natural credits; longshore movement out of the compartment, offshore transport (especially through submarine canyons), transport into estuaries, and wind transport shoreward to form sand dunes are the major debits. Included in Table B3 are the major human-induced credits and debits, including beach nourishment, which is increasingly used to rebuild lost beaches, and mining (*q.v.*), which directly removes sediment from the nearshore.

Research techniques

More is known about the geomorphology of the coastal zone than about the processes that modify the geomorphology. For example, it has only been during the past decade that the coupling between waves, circulation, and changes in nearshore bathymetry has begun to be observed and modeled. In addition, research is now focusing on fluid velocities and particle flux profiles in the bottom boundary layer and in the surface boundary layer under breaking waves. Studies of velocity and sediment concentration measurements in the swash zone are moving forward with the development of new instruments.

At present, it is not possible to forecast the effect of an upcoming storm season on a section of coast. However, it is possible to ascertain both the ultimate storm profile and the rate at which a beach returns to its original profile or shifts to a new equilibrium profile. The principle method for obtaining quantitative data on beach change is to repeatedly survey a beach. Comparing the resulting profiles will give erosion and accretion rates for the time encompassed by the surveys.

Geomorphology

Various surveying techniques are available to determine the geomorphic character of the coastal zone. Offshore, a boat-mounted depth sounder can be used to measure the bottom profile (*Bathymetric Surveys: q.v.*), and a side-scan sonar can be used to collect oblique views of the bottom. Both of these techniques are limited to depths where boats can safely operate, which does not include the breaker and surf zones. Furthermore, their accuracy is limited because of boat movement by waves and currents.

Vehicles have been developed to measure beach profiles across the nearshore. These include remotely controlled tractors (Seymour *et al.*, 1978; Dally *et al.*, 1994), sleds (Sallenger *et al.*, 1983), and an 11-m high motorized tripod that can drive across the beach and into the nearshore (Birkemeier and Mason, 1978). MacMahan (2001) mounted an echosounder and a *global positioning system* (GPS) (*q.v.*) unit onto a waverider to profile from deep water through the surf zone.

Surveying with a rod and transit is the most common method used to obtain beach profiles. Techniques range from the "Emery Board" procedure that uses two wooden rods separated by a rope, to sophisticated instruments that use light beams to measure the distance to a prism. The vertical accuracy of the latter instruments can be less to than 1 cm. With all these instruments, however, a closely spaced grid of points can be difficult to achieve except in small beach areas. When a beach is two-dimensional, a single cross-shore survey is sufficient to characterize it, but a beach with cusps and other three-dimensional features requires multiple cross-shore and, in some cases, alongshore surveys. With temporal surveys along a fixed shore-normal line, the existence of cusps can cause errors in comparing beach volume and beach-face location. At present, kinematic GPS units mounted on survey rods and various kinds of vehicles are being used to rapidly survey large sections of beach (Kaminsky *et al.*, 1998).

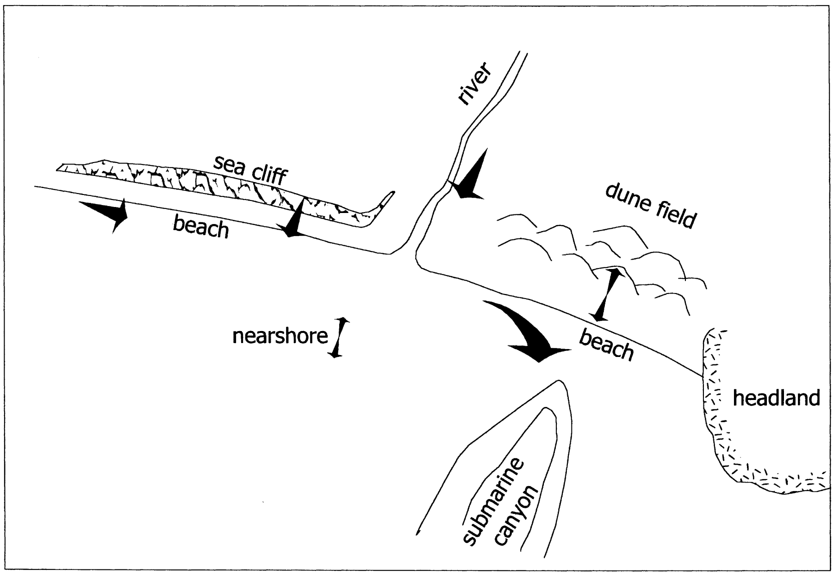

Figure B39 Principle constituents of a littoral zone sediment budget in a natural setting. Size of arrows is a rough indication of relative importance, though they can vary for a given situation. Not shown are anthropogenic impacts such as sand mining (after Komar, 1998).

Remote sensing techniques can be used to study large sections of the subaerial part of the coastal zone. Some of the techniques also can be used to look at geomorphic features in the nearshore. Air photos provide a qualitative look at the geomorphology, and quantitative measurements are possible with images that can be orthorectified. Such images are especially useful for measuring cliff retreat. Plant and Holman (1997) measured intertidal beach shape using a combination of time exposures and differential GPS. Lippmann and Holman (1989) investigated the locations and forms of offshore bars by taking several-minute time exposures from a camera mounted high above the beach and looking longshore. Recently, LIDAR (light detection and ranging) (*q.v.*), an airborne scanning instrument, has been used to map the subaerial part of the coastal zone (Brock *et al.*, 1999) and shallow parts of the nearshore (Irish and Lillycrop, 1999). These instruments are capable of rapidly estimating elevations to within a couple of centimeters approximately every 3 m^2 over regional scales.

Physical processes

Many instruments are available to measure water and sediment motion at all scales throughout the coastal zone. Many of them are sturdy enough to withstand the forces generated in the breaker and surf zones. Pressure sensors and bidirectional current meters have a sampling rate fast enough to measure water depth and wave-generated currents, respectively. Optical sensors measure sediment concentration. Sonic devices measure rapid changes in bottom elevation and, in some cases, the height of sediment suspension.

Future research in coastal processes

The goal of future research in coastal processes includes developing predictive models for:

- bedload and suspended sediment transport under combined wave and current forcing;
- turbulent wave/current boundary layers over 3-D small-scale morphology;
- effects of moving sediment on boundary layer;
- contribution to sediment transport by bedform migration;
- effects of grain-size distribution on sediment transport (Thornton *et al.*, 2000).

Morphology and its variability are important end products of predictive models. However, because sediment transport is not well understood, prediction of morphological change is inadequate at all scales. For example, at smaller scales, ripples and megaripples are observed to be ubiquitous, but have not been incorporated into models even though their effect on the flow field (as roughness elements) and sediment transport may be significant. Complex patterns in long-term, large-scale morphology have also been observed. However, models for morphology change have predictive skill only over the short term, whereas long-term, large-scale predictions are not yet possible. Research issues include:

- predicting morphology across the spectrum of length scales;
- free versus forced large-scale morphology models;
- understanding feedback between morphology and the flow field;
- coupling between length scales.

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Bibliography

Bascom, W.N., 1951. The relationship between sand size and beach face slope. *American Geophysical Union Transactions*, **32**: 866–874.

Bascom, W.N., 1980. *Waves and Beaches*. New York: Doubleday & Company.

Bird, E.C.F., 1984. *Coasts: An Introduction to Coastal Geomorphology*. Oxford, England: Basil Blackwell.

Birkemeier, W.A., and Mason, C., 1978. The CRAB: A unique nearshore surveying vehicle. *Journal of Survey Engineering*, **110**(1): 1–7.

Bowen, A.J., and Inman, D.L., 1966. Budget of littoral sands in the vicinity of Point Arquello, California. U.S. Army Coastal Engineering Research Center, *Technical Memorandum* No. 19.

Bowen, A.J., and Inman, D.L., 1969. Rip currents, 2: laboratory and field observations. *Journal of Geophysical Research*, **74**: 5479–5490.

Bowen, A.J., and Inman, D.L., 1971. Edge waves and crescentic bars. *Journal of Geophysical Research*, **76**(36): 8662–8671.

Bradley, W.C., 1958. Submarine abrasion and wave-cut platforms, *Geological Society of America Bulletin*, **69**: 957–974.

Brock, J., Sallenger, A., Krabill, W., Swift, R., Manizade, S., Meredith, A., Jansen, M., and Eslinger, D., 1999. Aircraft laser altimetry for coastal processes studies. *Coastal Sediments '99*, 2414–2428.

Bruun, P.M., 1954. Migrating sand waves and sand humps, with special reference to investigations carried out on the Danish North Sea coast. *Proceedings of the 5th Coastal Engineering Conference, on American Society of Civil Engineers*, pp. 269–295.

Carter, R.W.G., 1988. *Coastal Environments*. San Diego, CA: Academic Press.

Carter, R.W.G., and Woodroffe, C.D., 1994. *Coastal Evolution*. New York: Cambridge University Press.

Cowell, P.J., and Thom, B.G., 1994. Morphodynamics of coastal evolution. In Carter, R.W.G., and Woodroffe, C.D. (eds.), *Coastal Evolution*. New York: Cambridge University Press, pp. 33–86.

Dally, W.R., Johnson, M.A., and Osiecki, D.A., 1994. Initial development of an amphibious ROV for use in big surf. *Marine Technology Society Journal*, **28**(1): 3–10.

Dean, C., 1999. *Against the Tide*. New York: Columbia University Press.
Dietz, R.S., 1963. Wave base, marine profile of equilibrium, and wave built terraces: a critical appraisal. *Geological Society of America Bulletin*, **74**: 971–990.
Dolan, R., 1971. Coastal landforms: crescentic and rhythmic. *Geological Society of America Bulletin*, **82**: 177–180.
Dolan, R., and Ferm, J.C., 1968. Crescentric landforms along the Atlantic coast of the United States. *Science*, **159**: 627–629.
Dolan, R., Vincent, L., and Hayden, B., 1974. Crescentic coastal landforms. *Zeitschrift für Geomorphologie*, **18**: 1–12.
Evans, O.F., 1940. The low and ball of the East Shore of Lake Michigan. *Journal of Geology*, **48**: 476–511.
Galvin, C.J., Jr., 1972. Waves breaking in shallow water. In Meyer, R.E. (ed.), *Waves on Beaches*. New York: Academic Press, pp. 413–456.
Guilcher, A., 1958. *Coastal and Submarine Morphology*. New York: John Wiley & Sons.
Guza, R.T., and Inman, D.L., 1975. Edge waves and beach cusps. *Journal of Geophysical Research*, **80**(21): 2997–3012.
Hallermeier, R.J., 1981. A profile zonation for seasonal sand beaches from wave climate. *Coastal Engineering*, **4**: 253–277.
Holman, R.A., 1983. Edge waves and the configuration of the shoreline. In Komar, P.D. (ed.), *CRC Handbook of Coastal Processes and Erosion*. Boca Raton, FL: CRC Press, pp. 21–34.
Holman, R.A., and Bowen, A.J., 1982. Bars, bumps, and holes: models for the generation of complex beach topography. *Journal of Geophysical Research*, **87**(C1): 457–468.
Hom-ma, M., and Sonu, C.J., 1963. Rhythmic patterns of longshore bars related to sediment characteristics. *Proceedings of the 8th Coastal Engineering Conference on American Society of Civil Engineers*, pp. 248–278.
Huntley, D.A., and Bowen, A.J., 1978. Beach cusps and edge waves. *Proceedings of the 16th Coastal Engineering Conference on American Society of Civil Engineers*, pp. 1378–1393.
Inman, D.L., and Filloux, J., 1960. Beach cycles related to tide and local wind wave regimes. *Journal of Geology*, **68**: 225–231.
Irish, J.L., and Lillycrop, W.J., 1999. Scanning laser mapping of the coastal zone: the SHOALS system. *ISPRS Journal of Photogrammetry and Remote Sensing*, **54**(2/3): 123–129.
Johnson, D.W., 1919. *Shore Processes and Shoreline Development*. New York: John Wiley & Sons.
Kaminsky, G., Ruggiero, P., and Gelfenbaum, G., 1998. Monitoring coastal change in southwest Washington and northwest Oregon during the 1997–98 El Nino. *Shore and Beach*, **66**(3): 42–51.
Keulegan, G.H., 1948. An experimental study of submarine sand bars. U.S. Army Corps of Engineers. *Beach Erosion Board Technical Report No. 3*.
King, C.A.M., and Williams, W.W., 1949. The formation and movement of sand bars by wave action. *Geographical Journal*, **107**: 70–84.
Komar, P.D., 1971. Nearshore cell circulation and the formation of giant cusps. *Geological Society of America Bulletin*, **82**: 3593–3600.
Komar, P.D., 1983. Rhythmic shoreline features and their origins. In Gardner, R., and Scoging, H. (eds.), *Mega-Geomorphology*. Oxford, England: Clarendon Press, pp. 92–112.
Komar, P.D., 1998. *Beach Processes and Sedimentation*, 2nd edn. Upper Saddle River, NJ: Prentice-Hall.
Komar, P.D., and McDougal, W.G., 1988. Coastal erosion and engineering structures. The Oregon experience. *Journal of Coastal Research*, **4**: 77–92.
Komar, P.D., and Rea, C.C., 1976. Erosion of Siletz Spit, Oregon. *Shore and Beach*, **44**: 9–15.
Komar, P.D., 1996. The budget of littoral sediments—concepts and applications. *Shore and Beach*, **64**: 18–26.
La Fond, E.C., 1938. Relationship between mean sea level and sand movements. *Science*, **88**: 112–113.
LaFond, E.C., and Rao, R.P., 1954. Beach erosion cycles near Waltair on Bay of Bengal. *Andhrea University Mem. Oceanography*, **1**: 63–77.
Lippmann, T.C., and Holman, R.A., 1989. Quantification of sand bar morphology: a video technique based on wave dissipation. *Journal of Geophysical Research*, **95**(C7), 11,575–11,590.
MacMahan, J., 2001. Hydrographic surveying from a personal watercraft. *Journal of Surveying Engineering*, **127**(1): 12–24.
Plant, N.G., and Holman, R.A., 1997. Intertidal beach profile estimation using video images. *Marine Geology*, **140**(1–2): 1–24.
Ruessink, B.G., 1998. Infragravity waves in a dissipative multiple bar system. Doctoral Thesis, Department of Physical Geography, University of Utrecht, The Netherlands.
Sallenger, A.H., Jr., Howard, P.C., Fletcher, C.H., and Howd, P.A., 1983. A system for measuring bottom profile, waves and currents in the high-energy nearshore environment. *Marine Geology*, **51**: 63–76.
Seymour, R.J., Higgins, A.L., and Bothman, D.P., 1978. Tracked vehicle for continuous nearshore profiles. *Proceedings of the 16th Coastal Engineering Conference American Society of Civil Engineers*, pp. 1542–1554.
Shepard, F.P., 1950. Beach cycles in Southern California. U.S. Army Corps of Engineers. *Beach Erosion Board Technical Memorandum* No. 20.
Shepard, F.P., 1952. Revised nomenclature for depositional coastal features. *Bulletin of American Association of Petroleum Geologists*, **36**: 1902–1912.
Shepard, F.P., 1963. *Submarine Geology*, 2nd edn. New York: Harper & Row.
Shepard, F.P., and Wanless, H.R., 1971. *Our Changing Coastlines*. New York: McGraw–Hill.
Sonu, C.J., 1969. Collective movement of sediment in littoral environment. *Proceedings of the 11th Coastal Engineering Conference on American. Society of Civil Engineers*, pp. 373–400.
Sonu, C.J., 1972. Field observation of nearshore circulation and meandering currents. *Journal of Geophysical Research*, **77**(18): 3232–3247.
Thornton, E., Dalrymple, R.A., Drake, T.G., Elgar, S., Gallagher, E.L., Guza, R.T., Hay, A.E., Holman, R.A., Kaihatu, J.M., Lippmann, T.C., and Ozkan-Haller, H.T., 2000. State of nearshore processes research, II. *Naval Postgraduate School Technical Report* NPS-OC-00-001.
Trask, P.D., 1952. Sources of beach sand at Santa Barbara, California, as indicated by mineral grain studies. U.S. Army Corps of Engineers. *Beach Erosion Board Technical Memorandum* No. 28.
van Rijn, L.C.,1998. *Principles of Coastal Morphology*. Amsterdam, The Netherlands: Aqua Publications.
Werner, B.T., and Fink, T.M., 1993. Beach cusps as self-organized patterns. *Science*, **260**: 968–971.
Wiegel, R.L., 1964. *Oceanographical Engineering*. Englewood Cliffs, NJ: Prentice-Hall.

Cross-references

Barrier Islands
Bars
Beach Cusps (see Rhythmic Patterns)
Beach Erosion
Beach Features
Beach Nourishment
Cliffed Coasts
Coastal Boundaries
Coral Reefs
Cross-Shore Sediment Transport
Depth of Closure on Sandy Coasts
Dune Ridges
Erosion Processes
Global Positioning Systems
Greenhouse Effect and Global Warming
Longshore Sediment Transport
Meteorologic Effects on Coasts
Mining of Coastal Materials
Muddy Coasts
Nearshore Geomorphological Mapping
Profiling, Beach
Rhythmic Patterns
Ripple Marks
Rock Coast Processes
Sandy Coasts
Sea-Level Rise, Effect
Sediment Budget
Shore Protection Structures
Spits
Surf Zone Processes
Tides
Wave-Dominated Coasts
Waves

BEACH PROFILE

The beach profile is one the most studied features of coastal morphology. The shape of the beach profile determines the vulnerability of the coast to storms, the extent of usable beach for habitat and recreation, and the legal boundary distinguishing public and private ownership of land (Shalowitz, 1962, 1964; Anders and Byrnes, 1991). The first modern studies of the beach profile were motivated to understand its shape and variability in support of amphibious operations during World War II, when personnel and supply boats had to cross the beach profile from offshore to the dry beach (Bascom, 1980).

Beach profile terminology

The term "beach profile" refers to a cross-sectional trace of the beach perpendicular to the high-tide shoreline and extends from the backshore cliff or dune to the inner continental shelf or a location where waves and currents do not transport sediment to and from the beach. The profile shape is variable, depending on the time of year within the annual beach cycle and, also, the elapsed time after a storm. Waves, water level, and sediment grain size are the main controlling factors of beach profile shape.

Terminology associated with the beach profile is shown in Figure B40. The backshore runs from the seaward-most dune or the cliff to the land and water intersection. One or more berms may appear on a beach, depending on seasonal changes in water level. Berms are flat areas created during times of accretionary wave conditions, typically during summer. The beach intersects the water at the foreshore, and the foreshore is typically a plane slope that extends over a water level range from low tide to high tide. During a storm, a vertical step or scarp may form on the berm. The inshore covers the surf zone from the seaward end of the foreshore to past the seaward-most longshore sand bar, joining to the offshore. Several bars and associated troughs may appear on the beach profile.

Approximations of beach profile shape

As a first approximation, it is often possible to represent the profile of a gravel, pebble, or sandy beach (Here, the profile is assumed to have an unlimited supply of sand and that no "hard bottom" is present such as limestone reefs, coral reefs, and other non-erodable (hard) substances.) by a straight line of constant slope as,

$$h = x \tan \beta, \qquad \text{(Eq. 1)}$$

where h is the still-water depth, x the distance from shoreline, and $\tan \beta$ is the beach slope. This expression, defining a "plane beach" or planar beach slope has convenience for making simple calculations. Typically, however, the foreshore is the only area of the beach profile well represented by a straight line.

A more realistic representative profile shape was introduced by Bruun (1954) and studied extensively by Dean (1977, 1991). This profile is called the "equilibrium" or "x to the two-thirds profile" and is given as,

$$h = Ax^{2/3}, \qquad \text{(Eq. 2)}$$

where A is the shape parameter and will be discussed below. For water with temperature of about 20°C and typical sand sizes with sediment fall speed varying between about 1 and 10 cm s^{-1}, Kriebel *et al.* (1991) found that A could be related to fall speed w by,

$$A = 2.25\left(\frac{w^2}{g}\right)^{1/3}, \qquad \text{(Eq. 3)}$$

where g is the acceleration due to gravity. Moore (1982) was the first to study the functional dependence of A and found it to be an increasing function of the median grain size d_{50} for a wide range of materials. The empirical curve can be approximated by,

$$\begin{aligned} A &= 0.41(d_{50})^{0.94}, & d_{50} &< 0.4 \\ A &= 0.23(d_{50})^{0.32}, & 0.4 \le d_{50} &\le 10 \\ A &= 0.23(d_{50})^{0.28}, & 10 \le d_{50} &\le 40 \\ A &= 0.46(d_{50})^{0.11}, & 40 &\le d_{50} \end{aligned} \qquad \text{(Eq. 4)}$$

for which d_{50} is expressed in millimeters. The equilibrium profile can encompass a large range in grain size, as seen by the values in equation 4. Because the shape parameter A increases with increasing sediment grain size or fall speed, finer-grained beaches have gentle slopes and coarser-grained beaches are steeper, in accord with observations (Bascom, 1980).

Bars and troughs

Bars and troughs, also called longshore bars and longshore troughs, are the major perturbations from the equilibrium profile. Typically, areas with small tide range possess the most prominent bars because the wave breaker line remains in one position longer. Multiple-barred beaches are common on the Great Lakes, bays, and the Gulf of Mexico coast of the United States, for example, where the tide range is small. Some of these bars are formed by various predominant waves, such as typical waves and storm waves. Likewise, if the wave conditions occupy a relatively narrow range of height and period, such as on the north shore of Long Island, New York (facing the Long Island Sound), bars tend to be more prominent as compared to bars on the south shore of Long Island, because the Atlantic Ocean has a much more variable wave climate and smears out such bottom features.

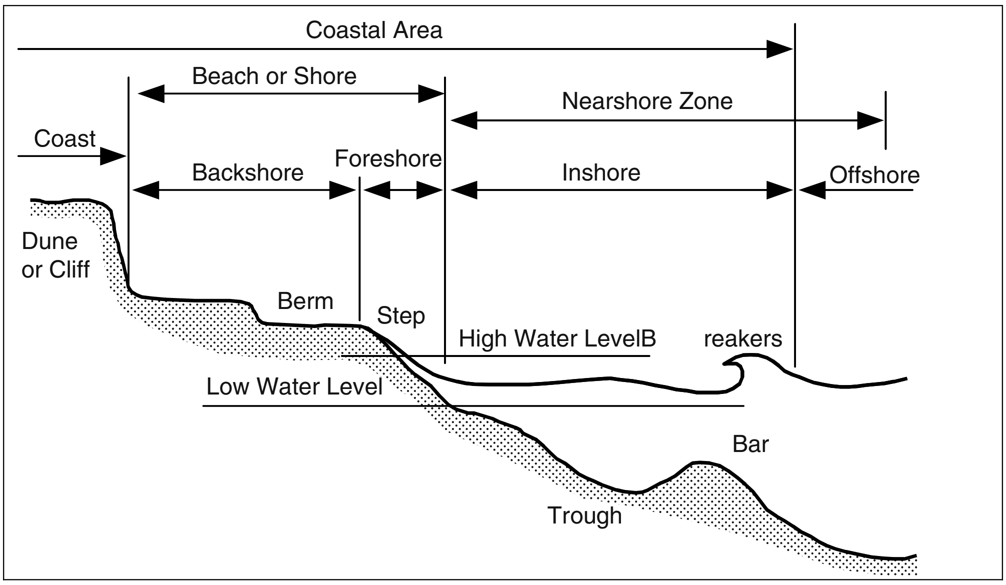

Figure B40 Terminology associated with the beach profile.

Larson and Kraus (1989) analyzed large-scale laboratory data for breaking waves and sandy beach beaches and found that the depth over the crest of a bar h_c was related to the breaking wave height H_b as

$$h_c = 0.66\ H_b. \quad \text{(Eq. 5)}$$

Conversely, if the depth over the crest of a well-established bar is measured, the breaking wave height that created the bar may be inferred from equation 5 to be $H_b = 1.5\ h_c$.

The beach profile at North Padre Island, TX, located along the Gulf of Mexico, was surveyed in the mid-1970s in one of the earliest applications of a sea sled, and then again in the mid-1990s with a sled, assuring high accuracy (see Profiling, Beach). Figure B41 plots a time series of surveys made at the same location on the beach. In both eras, profile elevation was referenced to mean sea level (MSL) at a local ocean tide gauge. Although the beach may have advanced or receded during the two decades, the shape of the profile can be compared because of the common vertical datum.

Figure B41 indicates that one to three (occasionally four) bars can appear on the profile at North Padre Island. It can be estimated through equation 5 that these bars are related to different classes of waves as outer bar—severe storm waves; middle bar—typical storm waves; and inner bar—waves under normal Gulf of Mexico conditions. In Figure B42, the average of 18 beach profile surveys taken alongshore at North Padre Island in 1996 is plotted together with the equilibrium ($x^{2/3}$) profile, equation 2, determined by the median grain size in the surf zone (0.18 mm). Sediment sampling performed during the surveys demonstrated a decrease in grain size with distance offshore, and such a decrease in size of surficial sediments is typical along the beach profile, with the coarsest material located at the beach face and bars, and the finer material located offshore. Coarse material is also found at the landward sides of longshore bars, because the finer material is transported away from this hydrodynamically energetic area.

Seasonal characteristics of the beach profile

During winter and the occurrence of seasonal storms (periodic northeasters, tropical storms, hurricanes), waves and cross-shore currents remove material from the beach and deposit it in bars far offshore. Large volumes, for example, 30–100 $m^3\ m^{-1}$ width of beach, can be removed from the beach berm and dune in a single large storm. Whether a beach will erode or accrete and the bar move onshore or offshore can be estimated with a dimensionless parameter called the Dean number formed as $N = H/wT$, where H is the wave height in deeper water, w is the beach sediment fall speed, and T is the wave period. For $N > 3.2$, erosion is

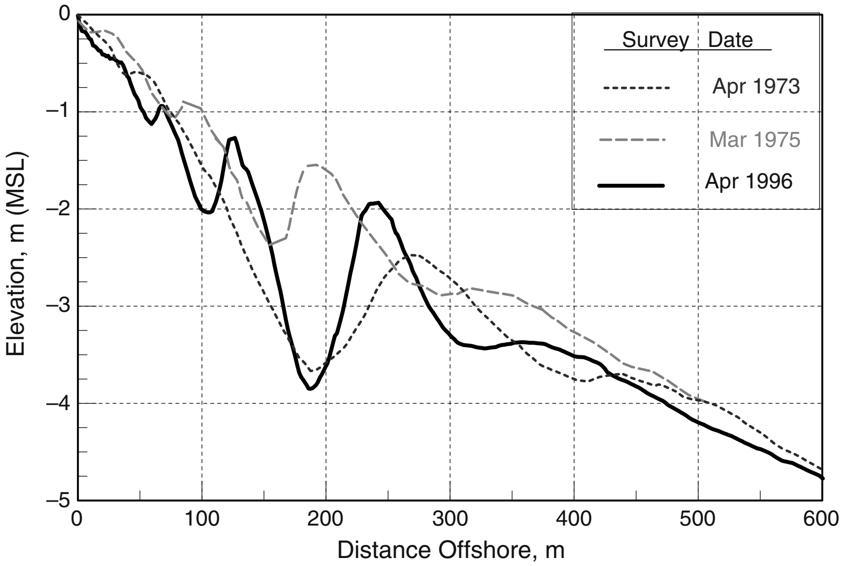

Figure B41 Beach profile surveys taken by sled two decades apart, North Padre Island, TX.

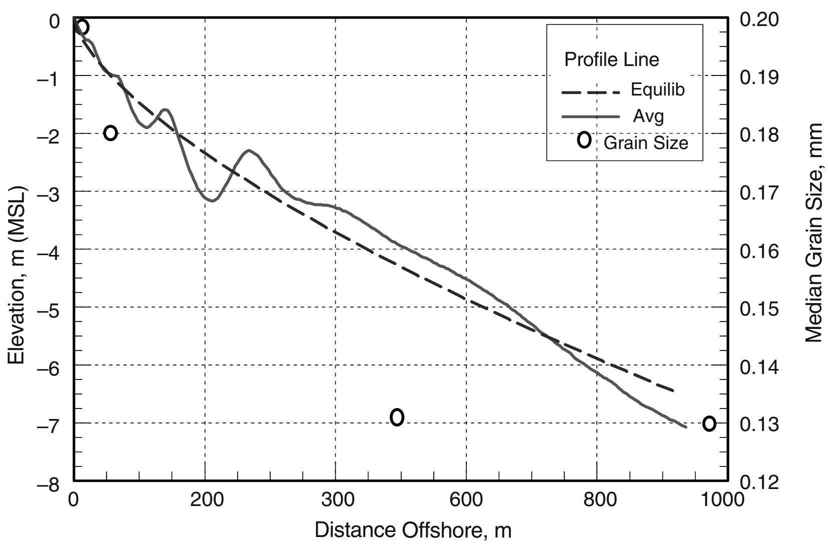

Figure B42 Average beach profile and equilibrium profile, North Padre Island, TX, 1996.

probable, whereas for $N < 3.2$, accretion is probable (Kraus *et al.*, 1991). Large values of the Dean number tend to occur during storms, when wave heights are large, and small values tend to occur in the summer, when wave heights are small or after storms, when the wave period of the swell waves becomes long. During the following milder accretionary waves of summer, material gradually moves onshore and returns to the beach, deflating the bar(s) and building the berm. Wind-blown sand then gradually rebuilds the dunes.

Sometimes, storms can be so severe that the beach does not recover on the timescale of human lifetime or engineering projects. Such is probably the case on the south shore of Long Island for the "Great New England Hurricane" of September 1938, which weakened the barrier islands and caused many breachings or cutting of temporary inlets. Sand taken offshore by such strong storms lies is such deep water that summer or "recovery" waves cannot readily transport it back to the beach.

The seasonal averages of a large number of surveys made on the same cross-shore transect (Line 62) are plotted in Figure B43. The surveys were made at the US Army Corps of Engineers' Field Research Facility (FRF), in Duck, NC, located on the "Outer Banks" barrier island chain. The profile is surveyed every two weeks as routine monitoring or more frequently for specific research goals by means of a large motorized tripod, estimated to have a vertical accuracy of ± 2 cm. The National Geodetic Vertical Datum (NGVD) is close to MSL at the FRF. Bars are absent from the plots because the average is taken over a large number of surveys. Sand moves offshore in winter (arbitrarily defined as the interval January to March) and returns in summer (June–August). A broad hump in the winter average at about 2–4 m depth indicates the presence of storm bars during that season. During summer months, the steep profile in shallower water created by the winter waves is gradually replenished and becomes shallower.

One property of the beach profile observed in Figure B43 is that the spring (April–June) and fall (October–December) average profiles almost plot on top of one another, and in between the two terminal states of summer and winter (Larson and Kraus, 1994). The regularity in profile response to waves indicates that the processes should be predictable with relatively simple techniques. Although not shown, the average of all profiles corresponds well with the equilibrium profile with a median grain size of 0.2 mm.

The seasonal response of the profile is shown in another way in Figure B44, which plots the average change in depth irrespective of sign (absolute value) as a function of average depth for the winter and summer profiles. The change in depth is less in summer than in winter, which is intuitively reasonable because waves are smaller in summer. The average maximum depth change occurs in winter, near the shoreline, and is

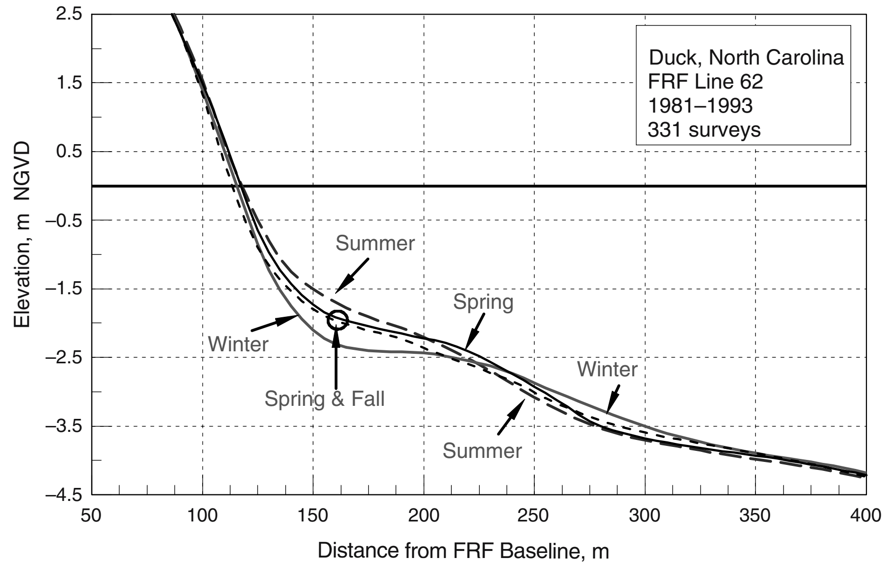

Figure B43 Average seasonal profiles, Duck, NC.

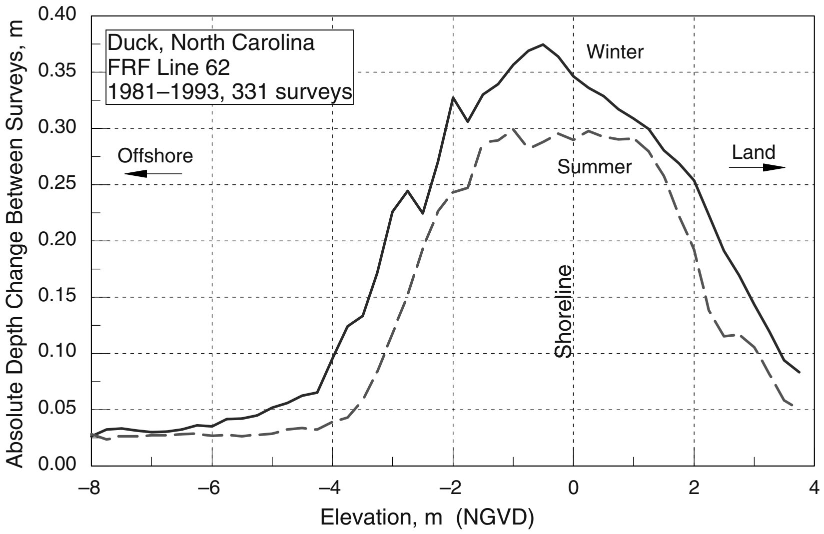

Figure B44 Average absolute depth change for summer and winter surveys, Duck, NC.

about 0.5 m (1.5 ft). The location where the profile changes little through time, delineating the area where sediment in the offshore is no longer exchanged with the beach, is called of the depth of closure (Kraus *et al.*, 1999). Knowledge of how much the profile elevation changes is required for modeling coastal processes, for designing beach fills, for placing pipes, outfalls, and cables, across the surf zone, and for placement of instruments so that they do not become buried.

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Bibliography

Anders, F.J., and Byrnes, M.R., 1991. Accuracy of shoreline change rates as determined from maps and aerial photographs. *Shore and Beach*, **59**(1): 17–26.

Bascom, W., 1980. *Waves and Beaches, The Dynamics of the Ocean Surface*, Revised and Updated Edition. Garden City, NY: Anchor Books.

Bruun, P., 1954. Coast erosion and development of beach profiles. U.S. Army Corps of Engineers. *Beach Erosion Board, Technical Memorandum* No. 44.

Dean, R.G., 1977. Equilibrium beach profiles: U.S. Atlantic and Gulf coasts. Department of Civil Engineering, *Ocean Engineering Report* No. 12. Newark, DE: University of Delaware.

Dean, R.G., 1991. Equilibrium beach profiles: principles and applications. *Journal of Coastal Research*, **7**(1): 53–84.

Kraus, N.C., Larson, M., and Kriebel, D.L., 1991. Evaluation of beach erosion and accretion predictors. *Proceedings of Coastal Sediments '91*. American Society of Civil Engineers, pp. 572–587.

Kraus, N.C., Larson, M., and Wise, R.A., 1999. Depth of closure in beach-fill design. *Proceedings of the 12th National Conference on Beach Preservation Technology*. Tallahassee, FL: Florida Shore and Beach Preservation Association, pp. 271–286.

Kriebel, D.L., Kraus, N.C., and Larson, M., 1991. Engineering methods for predicting beach profile response. *Proceedings of Coastal Sediments '91*. American Society of Civil Engineers, pp. 557–571.

Larson, M., and Kraus, N.C., 1989. SBEACH: numerical model for simulating storm-induced beach change. Report 1: empirical foundation and model development, Technical Report CERC-89–9, U.S. Army Engineer Waterways Experiment Station, Vicksburg, MS: Coastal Engineering Research Center.

Larson, M., and Kraus, N.C., 1994. Temporal and spatial scales of beach profile change, Duck, North Carolina. *Marine Geology*, **117**: 75–94.

Moore, B.D., 1982. Beach profile evolution in response to changes in water level and wave height. MCE thesis, Department of Civil Engineering. Newark: University of Delaware.

Shalowitz, A.L., 1962. *Shore and Sea Boundaries. Volume 1: Boundary Problems Associated with the Submerged Lands Cases and the Submerged Lands Acts*. Publication 10–1, U.S. Department of Commerce, Coast and Geodetic Survey, 420 pp.

Shalowite, A.L., 1964. *Shore and Sea Boundaries. Volume 2: Interpretation and Use of Coast and Geodetic Survey Data*. Publication 10–1, U.S. Department of Commerce, Coast and Geodetic Survey.

Cross-references

Accretion and Erosion Waves on Beaches
Bars
Coastal Warfare
Depth of Closure on Sandy Beaches
Monitoring, Coastal Geomorphology
Meteorologic Effects on Coasts
Storm Surge
Surf Zone Processes
Tidal Datums

BEACH RATING—See RATING BEACHES

BEACH RIDGES

Definitions

Johnson (1919) defined beach ridges as depositional features constructed by waves at successive shore positions. Reineck and Singh (1975) characterized beach ridges as having been formed at high-tide level of "rather coarse sediment" and related to storms or exceptionally high water stages. Bates and Jackson (1980) designated beach ridges as low mounds of beach and beach-and-dune material, heaped up by waves over the backshore beyond the present limit of storm waves or ordinary tides, but there is a risk of confusing wave-deposited and wind-deposited ridges. Referring to relict strandplain dunes as beach ridges, Carter (1986) used the term very broadly to cover "all large constructional forms of the upper beach, capable of preservation," applying it also to landward-shifting offshore bars, already welded swash bars, and prograded berm ridges.

Davis *et al*. (1972), Fraser and Hester (1977), Carter (1986, 1988) and others referred to onshore-migrating swash bars and/or to the stranded end products as "beach ridges." For some time after welding to the shore, the prograding sandy berm ridges may continue to be impacted by daily beach processes. Most North American and Australian authors considered stabilized onshore beach ridges as either of predominantly wave-built or of wave and wind-constructed, composite origin (e.g., Price, 1982). Hesp (1984, 1985), Mason (1990), and Mason *et al*. (1997), thus distinguishing between low profile "smooth, terrestrial," squat berm ridges and steep dune ridges that often overlie and bury them.

Beach ridge presently is defined as a relict shore ridge that is more or less parallel with the coastline and with other landward-adjacent ridges. It is built by wave swash (a berm ridge) that may be surmounted by wind-deposited sediment (a foredune). Once such a ridge becomes isolated from daily active beach processes by coastal progradation, which may lead to the construction of one or more new ridges to seaward, it becomes a beach ridge. On wide eolian backshore plains, shore-parallel dune ridges may also form behind active foredunes. Regardless of their dimensions, shapes, and origin, active beach/shore ridges impacted and modified almost daily by shore processes are excluded from the designation.

Associated landforms

A berm was originally defined as a narrow, scarp-backed, and wave-cut horizontal surface in the beach foreshore (e.g., Komar, 1976). Subsequently, it came to mean a wedge-shaped ridge, between an upper foreshore slope and a landward-inclined berm top surface (King, 1972). Its base is the horizontal plan that intersects the foreshore slope at the level of the backshore plain. Hine (1979) defined a berm as a shore-parallel linear body of triangular cross section with a horizontal to slightly landward-dipping surface (berm top) and a steeper seaward-dipping slope (beach face). Berms are short-lived and frequently reforming landforms, often absent from a beach.

Swash currents deposit sediment that builds the landward-sloping high-tidal berm above the level of the adjacent backshore. The ephemeral high-tidal sand berms are of aggradational origin, with secondary indications of erosional scarping. Increased onshore winds during falling tides briefly stabilize the water level. Intermediate-level berms with vertical scarplets may form during these stillstands.

Berm ridges, occasionally sizable and more permanent than berm surfaces are wave-built intertidal-supratidal landforms. The lithosomes are composed of intertidal and high tidal (swash-overwash) deposit, bounded by the backshore plane and the berm surface along its foreshore margin (Figure B45). After becoming isolated by progradation from the daily effects of beach processes, these inactive landforms attain the status of wave-constructed beach ridges. Formed on mainland or island beaches or on shore-parallel sand spits, berm ridges are bracketed between the foreshore and the landward (or lagoonward) margin of the backshore. On the landward side they may follow the shoreline of an elongated lagoon or beach pond ("cat's eye pond"; Coastal Research Group, 1969), enclosed by a sand spit (Figure B46). Shore-parallel inter-ridge swales bracket each ridge. Sets of prograding berm ridges form on beaches of limited sand supply. Short *et al*. (1989) reported on a nearly exclusively swash-built beach ridge plain in Australia, and in Egypt, Goodfriend and Stanley (1999) described a shelly sandridge plain, composed of 20–30 cm high, wave-built ridges without eolian cover.

Several authors have regarded high-tidal sand berms as incipient (wave-built) beach ridges during the Australian "berm debate" (Davies, 1957; Bird, 1960; Hails, 1969). Later, Bird and Jones (1988) proposed that if a berm survived a 15-day tidal cycle, it becomes a beach ridge. Berms have also been credited with providing a foundation for the development of embryonic foredunes that develop into full-sized ones (Davies, 1957; Bird and Jones, 1988). However, foredunes often form along the seaward margin of the backshore plain as well. The presence of berms, often absent from beaches, especially from dissipative and high-energy beaches (e.g., Short, 1984), is not an indispensable precondition for foredune development (Hesp, 1984, 1985).

Berm formation by wave action on the Tabasco shore of the Gulf of Mexico has been attributed to alternating "cut-and-fill" cycles of erosion and aggradation (Psuty, 1966). This berm-shaping process,

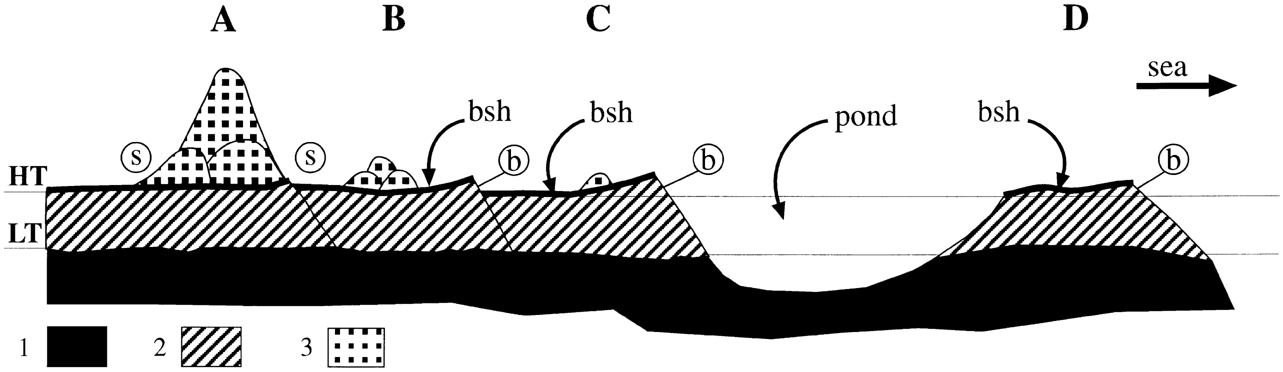

Figure B45 Beach ridge-associated depositional facies and landforms on a prograding strandplain. Depositional facies: (1) subtidal; (2) wave-built, intertidal-to-supra-high tidal; (3) eolian. Landforms: b, berm; bsh, backshore plain; s, swale; A, foredune; B, accreting embryonic dunes on backshore plain in the early foreshore stage; C, single embryonic dune on berm ridge (backshore-berm) surface; pond: spit-growth-enclosed beach pond; D, pond-isolated berm ridge (intertidal-supratidal sand spit interval), without embryonic dunes. HT, high-tide level; LT, low-tide level. (Otvos, 2000) Reprinted with permission from Elsevier Science.

however, appears to be a localized and ephemeral phenomenon without bearing on long-term strandplain development. The adjacent, 20-km wide Tabasco strandplain, for example, has been receiving abundant eolian sand supply and represents a foredune ridge plain, underlain by backshore and berm ridge deposits.

Truncation lines that separate mainland and island beach ridge sets

Where beach ridge progradation has been abruptly terminated by shore erosion, then followed by renewed beach growth, there are cross-cutting truncation lines that separate generations of beach ridges in mainland and island barriers of Quaternary age. This process was historically documented on several Mississippi coast strandplain islands (Otvos, 1981). The St. Joseph Bay-area barrier spit and small mainland strandplains in NW Florida provide good illustrations (Figure B47).

Gravel-boulder ("storm") ridges. Storm-associated high tides and waves build gravel ramparts as high as 6 m (Clapperton, 1990). Gravel-boulder ridges, associated with storm surges therefore rise well over their associated sea (lake) levels. Coarse clastic sediments, including shelly material resist backwash erosion and become stranded on these shore ridges. Permanent shingle emplacement at superelevated tide levels is aided by backwash percolation into the permeable gravelly substrate (Carter, 1988). For a given still-water level, the height of wave-built ridges built during winter storms may vary by as much as 2–2.6 m (Adams and Wesnousky, 1998). These "storm" ridges are common on glaciated and bedrock shores; also on tectonically or isostatically raised marine and lacustrine terraces. Examples abound in Canada's Maritime Provinces, New England, and on high Pacific shores between Alaska and Mexico.

Coarse clastic beach ridges or bedrock terrace veneers accompany raised strandlines of pluvial and glacial lakes in the North American interior basins (Fulton, 1989; Morrison, 1991). Coarse clastic sediments were delivered by high-gradient streams, alluvial fans, mass wasting, occasionally even fluvioglacial processes. Adjacent bedrock areas that served as sediment sources have undergone intensive physical weathering under periglacial and cold-temperate conditions. Pluvial Lake Bonneville and its successor, early stage Great Salt Lake in Utah and Nevada; as well as Lake Lahontan in Nevada and California provide the best examples (Morrison, 1965, 1991; Currey, 1980; Adams and Wesnousky, 1998, 1999). Gravel-boulder ridges, deposited on wave-cut bedrock terraces form discontinuous tabular and tabular cross-stratified bodies, several meters thick (Adams and Wesnousky, 1998). Playa beach ridges that contain carbonate nodules, include paleosols and incorporate secondarily calcreted and gypsum-creted grit in the arid Lake Eyre basin, Australia, one of the world's largest internally drained regions (Nanson *et al.*, 1998).

Bouldery-gravelly coarse clastic sediments that often dominate ice-dammed glacial lake shorelines along the fluctuating glacial margin in North America have originated from reworked moraine till, fluvioglacial delta, periglacial colluvium, and in particular ice-contact (esker, kame, outwash delta) deposits. Major examples include relict shore features on glacial Lakes Agassiz, Algonquin, and Ojibway (Fulton, 1989, pp. 144–145, 257, 343, 362–364). Due to sand scarcity, the short life span of a given strandline, and erosive wave regimes, instead of regular beach ridges gravelly-bouldery shore zone lags frequently veneer wave-cut lake terrace surfaces.

Sediment types and ridge forms of sandy beach ridges. Depending on the wide range of wave and current conditions and source sediments on a given marine or lake shore sector, the upper beach deposits (intertidal–supratidal intervals) may be represented by fine-to-coarse, even gravelly sands in the berm ridges. Whereas, sorting tends to be good in uniformly sandy beaches, it becomes moderately sorted when coarser clastic fractions are also included (e.g., Thompson, 1992). Chappell and Grindrod (1984) described a rare transition between sandy ridges of a regular beach ridge plain and a small adjacent chenier plain, composed of shell-rich ridges. Reflecting the low relief and gentle seaward and landward slopes of the berm ridges, basinward-dipping parallel laminae and low-angle (3–5°) cross lamination tend to characterize the upper foreshore slope.

Subhorizontal or gently landward-dipping laminations occur on the landward beach ridge surfaces. The highly variable beach ridge dimensions, the height above still water level and slope angles depend on wave conditions, local tidal, or lake level ranges, including wind-induced rise in sea and lake-levels along given shore sectors.

"Pebble-armored ridges." Pebble sheets plastered onto sand dunes during storms were designated as gravelly ramp barriers (Orford and Carter, 1982; Mason, 1990). The hydraulic ratios and shapes of shell bioclasts result in higher transport and dispersal rates than with regard to larger, denser silicate rock clasts. Whereas, gravel/boulder ridges accumulate during direct storms, their impact tends to flatten and disperse already existing ridges, composed of sand and lighter, platy shell clasts (Greensmith and Tucker, 1969; Rhodes, 1982, p. 217). Higher-energy events enable accumulation even of bioclastic rudites (Woodroffe *et al.*, 1983; Meldahl, 1993).

Beach ridges, strandplain versus terrace development

Strandplain formation may be a continuous process with grain-by-grain addition of sand to the widening foreshore. Continuous progradation of the neap berm at mid-to-high level results in a gently undulating, almost level beach plain. On mesotidal foreshores where neap high tide remains below the highest foreshore levels, continuously accreting neap berms are uninterrupted by inter-berm swales (Hine, 1979, Figure 17(A)). Increased sand supply along the low-microtidal Gulf of Mexico beaches leads to steady outbuilding of the foreshore, accompanied by consequent progradation of narrow, closely spaced beach ridges. Discontinuous beach ridge progradation involved either the stranding or remolding of landward-migrated mesotidal swash bars on the foreshore (Hine, 1979; Carter, 1986) or spit growth from and downdrift reattachment (Figure B46) to the beach in microtidal settings (Otvos, 1981). Both processes isolate elongated ponds. Fronted seaward (lakeward) by newly formed active foredunes, and slowly filled by eolian and washover sands, such ponds may gradually become wide supratidal inter-ridge swales (Figure B46).

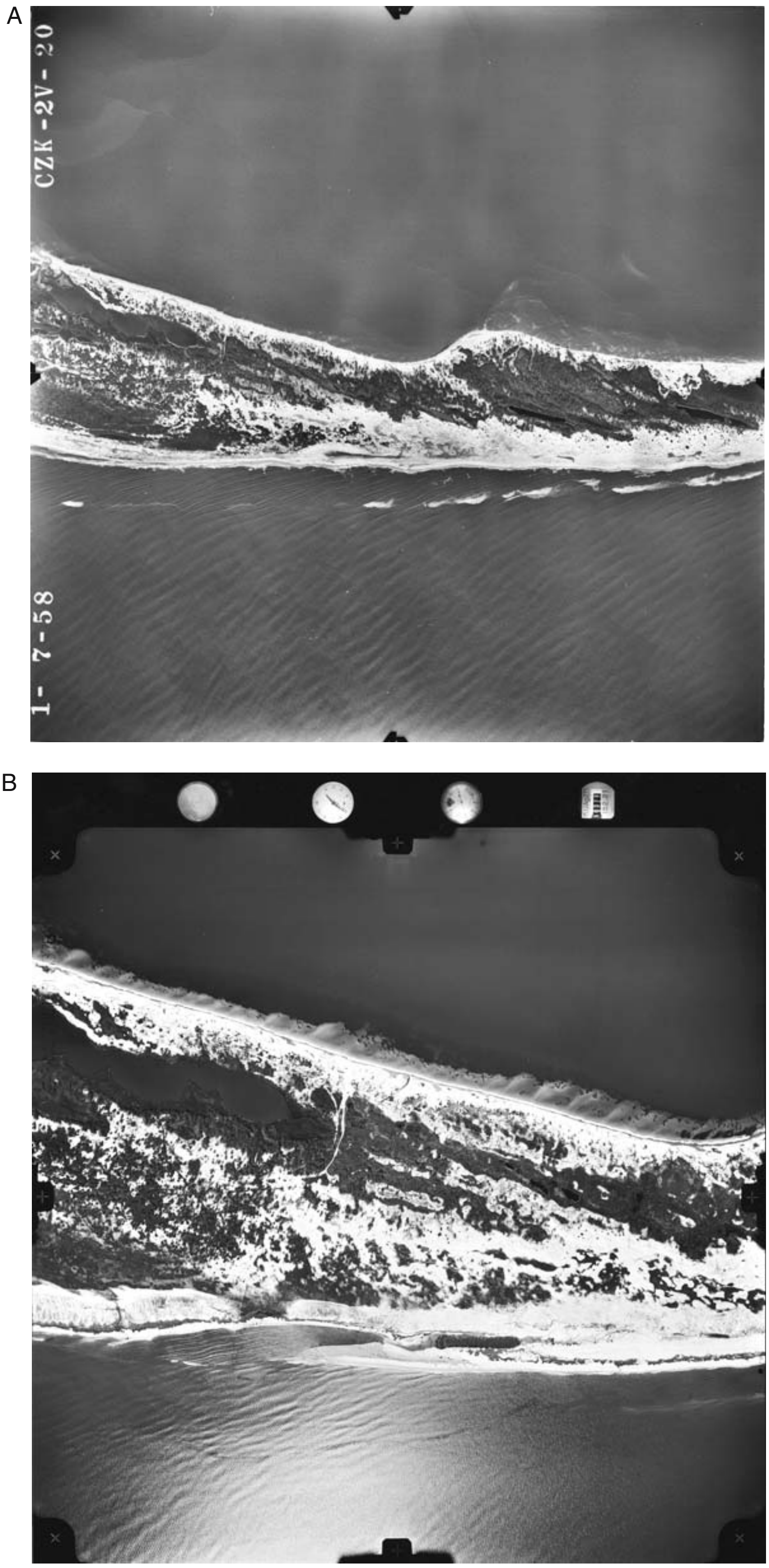

Figure B46 Prograding strandplain, southeastern Horn Island, MS. (A) Inter-swale ponds between old beach ridge sets in wooded interior. Elongated shore-parallel ponds of different orientation isolated from Gulf of Mexico (south) by barren narrow strip of backshore and berm ridges (USDA aerial photo, January, 1958. Width of image: 4.56 km). (B) Eleven years later: western half of previous image. Shore-parallel westward spit and spit-platform growth about to form new cat's eye pond (bottom). Already isolated beach pond to east. Eroding (white) and forested (dark) old interior beach ridge sets, with swale ponds in the island interior (USCGS aerial photo, October, 1969. Width of image: 2.3 km).

A

B

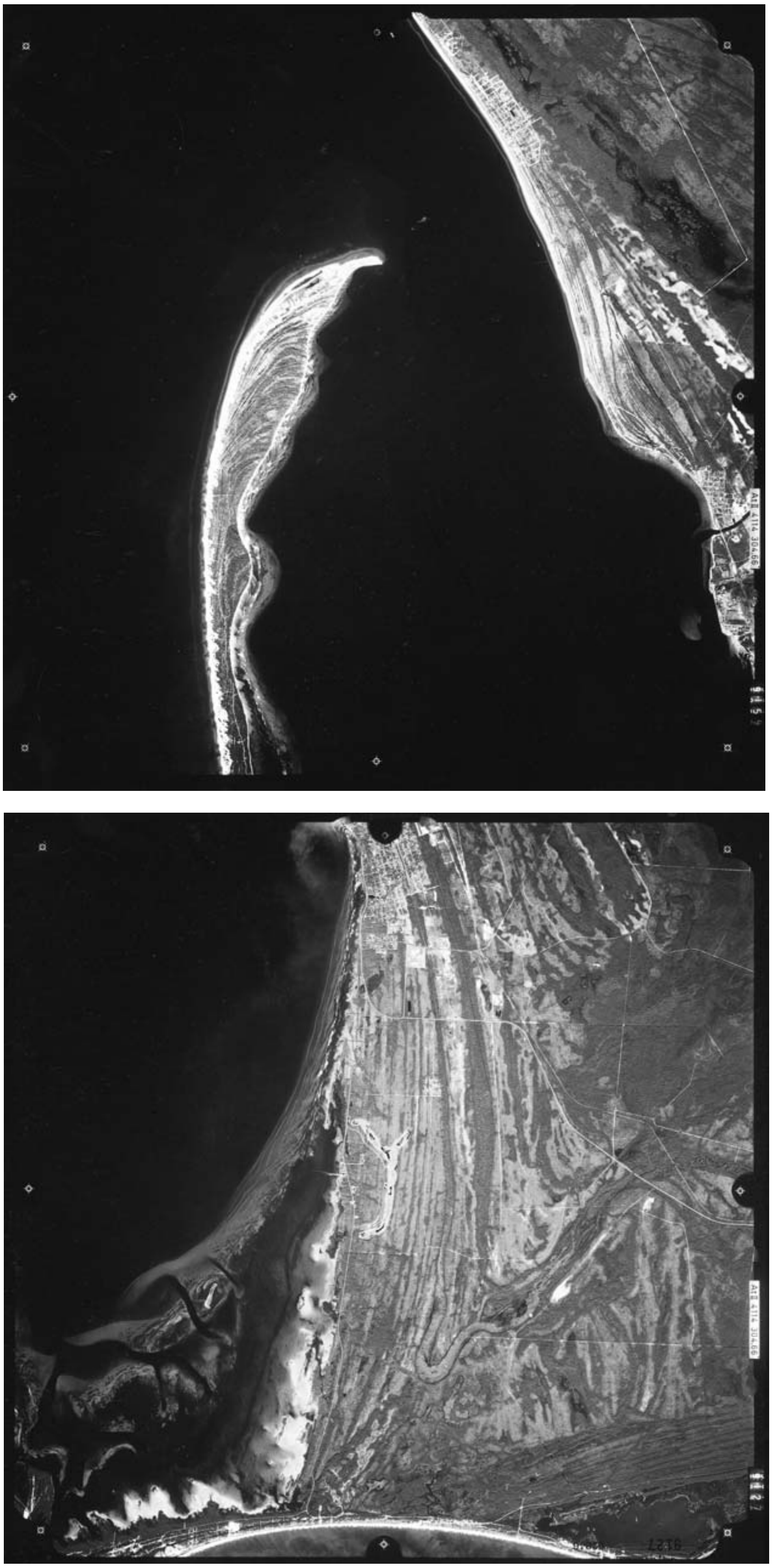

Figure B47 (A) Generations of narrow Late Holocene strandplain fans, northern St. Joseph Bay, NW Florida. Left: north tip of St. Joseph Peninsula (barrier spit). Right: small Holocene Palm Point mainland strandplain. Upper right corner: wide Late Pleistocene (Sangamon) beach ridges and partially filled swales represent erosion-impacted strandplain sectors. (B) St. Joseph Bay, NW Florida. Quaternary strandplains. Wide beach ridges of N–W-trending Late Pleistocene (Sangamon) strandplain (center field of photo) again contrast with narrow, crisply outlined Late Holocene strandplain ridges (lower right corner) (USGS aerial photo, October, 1978. Width of image: ca. 15 km). (Otvos, in press).

A comparison of Late Pleistocene strandplain ridges with the sharply outlined, narrow Late Holocene beach ridges illustrates the fact that prolonged infilling and erosional modification of Pleistocene strandplain ridges result in more subdued, more gently sloping beach ridges, separated by wider swales (Figure B47).

Instead of strandplains, gently undulating, nearly level eolian sand terraces form when sand supply and beach progradation does not keep pace with rapidly growing and sand-trapping beach vegetation. Beach progradation and/or eolian sand supply rates under these conditions are relatively low (e.g., Ruz and Allard, 1994).

Rates of beach ridge development

Depending on ridge dimensions, sediment supply rates, hydrodynamic and vegetative conditions, beach ridge development may proceed slowly or quickly. Thus, Nanson *et al.* (1998) report on a beach ridge along an Australian playa lake that during a high lake stage, formed in less than one year. Development rates in a number of other calculated examples from worldwide locations ranged between 1.8 and 3.3 yr/ridge, at other sites the rates were as low as 30–150 yr/ ridge (in: Otvos, 2000, pp. 90–91).

Beach ridges as ancient sea/lake levels markers

Former sea (lake) levels may be identified when a hortizontal interface is recognizable between the wave-built foreshore and the overlying eolian lithosome in a given ridge. This was the case in Lake Michigan strandplain ridges (Fraser and Hester, 1977; Thompson, 1992) where low-angle sand and gravel cross beds and trough-cross-bedded lacustrine sands of wave-built origin underlie land snail-bearing, cross-bedded, in part massive, structureless dune sands. On pure sand beaches that lack granule and pebble clasts due to the very short transport distance from the immediately adjacent source of sediment, distinctions between wave- and wind-deposited lithosomes often are difficult or impossible to make on granulometric grounds alone.

At times of superelevated lake and sea levels, associated with storm-related temporary rise of the water levels wave-built sandy, shelly, or gravelly beach ridges may aggrade significantly above normal high-tide levels (e.g., Mason, 1990; Mason and Jordan, 1993). Precise reconstruction of former sea levels from beach ridges therefore may be problematical.

Similar to beach ridge summits, coarse clasts are useful markers of reference surfaces to document postdepositional tectonic and isostatic changes, including vertical displacement, tilting, and warping of the land surface. In their absence, wave-cut bedrock terraces and lag clast-veneered strandlines may also serve this purpose. Strandline stairsteps were documented on isostatically uplifted marine and glacial lake shores in subarctic North America and Scandinavia (e.g., Hudson Bay, Tyrell Sea; Fulton, 1989). Flights of raised beaches characterize former pluvial/playa lake shores in western North America, Australia, and other presently arid and semiarid regions. Bounding surfaces of lower foreshore Lepidopthalmus (formerly, Callianassa) ghost shrimp-burrowed barrier ridge deposits and correlative adjacent lagoonal-saltmarsh surfaces in six Late Pliocene-Pleistocene coastal terrace sequences were similarly utilized on the Georgia-northeast Florida seaboard (Hoyt, 1969).

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Bibliography

Adams, K.D., and Wesnousky, S.G., 1998. Shoreline processes and the age of the Lake Lahontan highstand in the Jessup embayment. *Geological Society of America Bulletin*, **110**: 1318–1332.

Adams, K.D., and Wesnousky, S.G., 1999. The Lake Lahontan highstand: age, surficial characteristics, soil development, and regional shoreline correlation. *Geomorphology*, **30**: 357–392.

Bates, R.L., and Jackson, J.A., 1980. *Glossary of Geology*, 2nd edn. Falls Church, VA: American Geological Institute.

Bird, E.C.F., 1960. The formation of beach ridges. *Australian Journal of Science*, **22**: 349–350.

Bird, E.C.F., and Jones, D.J.B., 1988. The origin of foredunes on the coast of Victoria, Australia. *Journal of Coastal Research*, **4**: 181–192.

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 Press.

Chappell, J.C., and Grindrod, J., 1984. Chenier plain formation in northern Australia. In Thom, B. G. (ed.), *Coastal Geomorphology in Australia*. New York: Academic Press, pp. 197–232.

Clapperton, C.M., 1990. Quaternary glaciations in the southern ocean and Antarctic Peninsula area. *Quaternary Science Reviews*, **9**: 229–252.

Coastal Research Group, 1969. *Coastal Environments, NE Massachusetts and New Hampshire, Guidebook, SEPM Field Trip*. Boston: University of Massachusetts.

Currey, D.R., 1980. Coastal geomorphology of Great Salt Lake and vicinity. Great Salt Lake—an overview. *Utah Geological and Mineral Survey Bulletin*, **116**: 69—82.

Davies, J.L., 1957. The importance of cut and fill in the development of sand beach ridges. *Australian Journal of Science*, **20**: 105–111.

Davis, R.A., Jr., Fox, H.T., Hayes, M.O., and Boothroyd, J.C., 1972. Comparison of ridge-and-runnel systems in tidal and non-tidal environments. *Journal of Sedimentary Petrology*, **42**: 401–412.

Fraser, G.S., and Hester, N.C., 1977. Sediments and sedimentary structures of a beach-ridge complex, southwestern shore of Lake Michigan. *Journal of Sedimentary Petrology*, **47**: 1187–1200.

Fulton, R.J. (ed.), 1989. Quaternary geology of Canada and Greenland. Geology of Canada, No.1, The Geological Society of America, Geology of North America Series (DNAG), K–1.

Goodfriend, G.A., and Stanley, D.J., 1999. Rapid strand-plain accretion in the northeastern Nile Delta in the 9th century AD and the demise of the port of Pelusium. *Geology*, **27**: 147–150.

Greensmith, J.T., and Tucker, E.V., 1969. The origin of Holocene shell deposits in the chenier plain facies of Essex (Great Britain). *Marine Geology*, **7**: 403–425.

Hails, J.R., 1969. The origin and development of the Umina-Woy ridge system, Broken Bay, NSW. *The Australian Geographer*, **11**: 1–12.

Hesp, P.A., 1984. Foredune formation in southeast Australia. In Thom, B.G. (ed.), *Coastal Geomorphology in Australia*. Sydney: Academic Press, pp. 69–97.

Hesp, P.A., 1985. The formation of sand "beach ridges" and foredunes. *Search*, **15**: 289–291.

Hine, A.L., 1979. Mechanism of berm development and resulting beach growth along a barrier spit complex. *Sedimentology*, **26**: 333–351.

Hoyt, J.H., 1969. Late Cenozoic structural movements, northern Florida. *Gulf Coast Association of Geological Societies Transactions*, **19**: 1–9.

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

King, C.A.M., 1972. *Beaches and Coasts*, 2nd edn. London: Arnold.

Komar, P.D., 1976. *Beach Processes and Sedimentation*. Englewood Cliffs, NJ: Prentice-Hall.

Mason, O.K., 1990. Beach ridge geomorphology of Kotzebue sound: implications. Paleoecology and archeology. Ph.D. dissertation. Anchorage: University of Alaska.

Mason, O.K., and Jordan, J.W., 1993. Heightened North Pacific storminess during synchronous Late Holocene erosion of northwest Alaska beach ridges. *Quaternary Research*, **40**: 55–69.

Mason, O.K., Hopkins, D.M., and Plug, L., 1997. Chronology and paleoclimate of storm-induced erosion and episodic dune growth across Cape Espenberg Spit, Alaska. *Journal of Coastal Research*, **18**: 770–798.

Meldahl, K.H., 1993. Geographic gradients in the formation of shell concentrations. Plio-Pleistocene marine deposits. Gulf of California. *Paleogeography, Paleoclimatology, Paleoecology*, **101**: 1–25.

Morrison, R.B., 1965. New evidence on Lake Bonneville stratigraphy and history from southern Promontory Point, Utah. *U.S. Geological Survey Professional Paper* 525-C, pp. C110–C119.

Morrison, R.B., 1991. Quaternary stratigraphic, hydrologic, and climatic history of the Great Basin, with emphasis of Lake Lahontan, Bonneville, and Tecopa. Geological Society of America, The Geology of North America, K-2, pp. 283–320.

Nanson, G.C., Callen, R.A., and Price, D.M., 1998. Hydroclimatic interpretation of Quaternary shorelines on South Australian playas. *Paleogeography, Paleoclimatology, Paleoecology*, **144**: 281–305.

Orford , J.D., and Carter, R.W.G., 1982. Crestal, overtop and washover sedimentation on a fringing sandy gravel barrier coast. *Journal of Sedimentary Petrology*, **52**: 265–278.

Otvos, E.G., 1981. Barrier island formation through nearshore aggradation-stratigraphic and field evidence. *Marine Geology*, **43**: 195–243.

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

Otvos, E.G., in press. Holocene coastal barriers, Gulf of Mexico. Chronology, evolution, and sea-level relationships. *Journal of Coastal Research*.

Price, W.A., 1982. Beach ridge plain. In Schwartz, M.L. (ed.), *Encyclopedia of Beaches and Coastal Environments*. Stroudsburg, PA: Hutchinson Ross Publishing, pp. 159–160.
Psuty, N.P., 1966. The geomorphology of beach ridges in Tabasco. *Coastal Studies Institute Technical Report*, 30. Baton Rouge, LA: Louisiana State University.
Reineck, H.E., and Singh, I.B., 1975. *Depositional Sedimentary Environments*. Berlin, New York: Springer Verlag.
Rhodes, E.G., 1982. Depositional model for a chenier plain, Gulf of Carpentaria, Australia. *Sedimentology*, **29**: 201–222.
Ruz, M.-H., and Allard, M., 1994. Foredune development along a subarctic emerging coastline, eastern Hudson Bay, Canada. *Sedimentology*, **42**: 725–734.
Short, A.D., 1984. Beach and nearshore facies: southeast Australia. *Marine Geology*, **60**: 261–282.
Short, A.D., Buckley, R.C., and Fotheringham, D.G., 1989. Preliminary investigations of beach ridge progradation on Eyre Peninsula and Kangaroo Island. *Marine Geology*, **90**: 289–296.
Thompson, T.A., 1992. Beach-ridge development and lake-level variation in southern Lake Michigan. *Sedimentary Geology*, **80**: 305–318.
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.

Cross-references

Barrier
Barrier Islands
Beach Processes
Cheniers
Dunes and Dune Ridges
Meteorological Effects on Coasts
Rock Coast Processes
Sea-Level Indicators, Geomorphic
Spits
Tectonics and Neotectonics

BEACH SAFETY—See LIFESAVING AND BEACH SAFETY

BEACH SEDIMENT CHARACTERISTICS

Beach sediments are derived from a wide variety of sources, including cliff erosion, rivers, glaciers, volcanoes, coral reefs, sea shells, the Holocene rise in sea level, and the cannibalization of ancient coastal deposits. The nature of the source and the type and intensity of the erosional, transportational, and depositional processes in a coastal region determine the type of material that makes up a beach. In turn, the characteristics of the sediments strongly influence beach morphology and the processes that operate on it (Trenhaile, 1997).

Grain size

The grain size of pebbles and other large clastic material can be measured with callipers, and sieves are used for sand and other coarse beach sediments. A number of techniques are used to determine the size of finer sediments including Coulter Counters, pipettes, hydrometers, optical settling instruments, and electron microscopes. The grain size can be expressed using the Wentworth scale, which is based on classes that are separated by factors of two, so that each is twice the size of the one below. A $\log_2$ transform can be used to provide integers for each of the Wentworth class limits:

$$D_\phi = -\log_2 (D_{mm})$$

where D_ϕ is the grain diameter in phi units (ϕ) and D_{mm} is the corresponding diameter in millimeters (Table B4). Unfortunately, the term "grain diameter" can refer to several different things (Sleath, 1984):

(1) the mesh size of the sieve through which the grains are just able to pass;
(2) the diameter of a sphere of the same volume;
(3) the length of the long, short, or intermediate axes of the grain, or some combination of these lengths; or
(4) the diameter of a smooth sphere of the same density and settling velocity as the grains.

Table B4 Sediment grain size classification

Type	ϕ units	Wentworth (mm)
Boulder	>-8	>256
Cobble	-8 to -6	256–64
Pebble	-6 to -2	64–4
Granule	-2 to -1	4–2
Sand		
Very coarse sand	-1 to 0	2–1
Coarse sand	0–1	1–0.5
Medium sand	1–2	0.5–0.25
Fine sand	2–3	0.25–0.125
Very fine sand	3–4	0.125–0.0625
Silt		
Coarse silt	4–5	0.0625–0.0312
Medium silt	5–6	0.0312–0.0156
Fine silt	6–7	0.0156–0.0078
Very fine silt	7–8	0.0078–0.0039
Clay		
Coarse clay	8–9	0.0039–0.00195
Medium clay	9–10	0.00195–0.00098

The weight-percentages of the sediment can be plotted against the diameter in phi units in the form of histograms or frequency curves. Grain-size distributions are most frequently represented, however, by plotting the grain size data on a probability, cumulative percentage ordinate, and the phi scale on an arithmetic abscissa. The percentiles on the cumulative size distribution can be used to estimate the mean, standard deviation, and other simple descriptive statistical measures, although the calculations can also be made by computer. For comparative purposes, sediment samples can be represented by the mean or median grain size, or by the size of the grain that is coarser than some percentage of the sample.

There have been many attempts to identify the transportational processes and the depositional origin of sediments based on their sediment-size distributions. The grain-size distributions of beach sediments often consist of three straight-line segments, rather than the single straight line of a normal distribution plotted on a Gaussian probability axis. The three segments have been variously interpreted as representing: coarse bed load, fine suspended load, and intermediate-sized grains that move in intermittent suspension; the effect of packing controls on a grain matrix, the larger grains being a lag deposit, with the finest grains resting in the spaces between grains of median size; and different laminae in the beach, representing several depositional episodes. A further possible explanation is that the segmentation of grain-size distributions on log-normal cumulative probability paper may reflect the use of an inappropriate probability model. The log-normal model poorly represents the extremes of natural grain-size distributions, which may conform much better to a hyperbolic probability function (Trenhaile, 1997). Some workers believe that the four parameters of a logarithmic hyperbolic distribution are more sensitive to sedimentary environments and dynamics than the statistical moments of the normal probability function, but others have found that there is little difference (Sutherland and Lee, 1994). Grain sizes may also be fitted to a skew log-Laplace model, a limiting form of the log-hyperbolic distribution which is essentially described by two straight lines, and is defined by three parameters (Fieller *et al*., 1984).

Grain shape

The shape of beach grains can be expressed in various ways. The roundness of a grain, which refers to the smoothness of its surface, has been defined as the ratio of the radius of curvature at its corners to the radius of curvature of the largest inscribed circle. Grain sphericity describes the degree to which its shape approaches that of a sphere with three equal orthogonal axes. The shape of a grain can range from spherical, to plate, to rod-like forms, according to the relationship between the three axes, which can be depicted in the form of a ternary diagram. Grain shape can be measured and defined using a variety of indices. They include the E shape factor (ESF):

$$\mathrm{ESF} = D_s\left[\frac{D_s^2 + D_i^2 + D_l^2}{3}\right]^{-0.5}$$

and the Corey shape factor (CSF):

$$\mathrm{CSF} = \frac{D_\mathrm{s}}{(D_\mathrm{i}\,D_\mathrm{l})^{0.5}}$$

where D_l, D_s, and D_i are the long, short, and intermediate axes of the grain, respectively.

The shape of coarse clasts can be determined fairly easily by direct measurement, but this is usually impossible or too time-consuming for sand and other small grains. Therefore, the roundness and sphericity of sand grains has often been estimated by visual comparison with a set of standard grain images of known roundness, although Fourier analysis is increasingly being used (Powers, 1953; Thomas *et al.*, 1995). Winkelmolen's (1971) "rollability" index, the time taken for a grain to roll down the inside of a revolving, slightly inclined cylinder, is easier to measure than other shape parameters, and the shape distribution factors, obtained by plotting grain rollability against grain size, may be more characteristic and indicative of the mode of origin of coastal sediments.

Grain density

The density of a grain is determined by its mineralogy (Table B5). In temperate regions, most beach sediment originated from the granitic rocks of continents, and they largely consist of quartz and, to a much lesser extent, feldspar grains, but carbonates may dominate in the tropics, especially where there are coral reefs. The sediments in pocket beaches enclosed between prominent headlands, and in beaches derived from other restricted source areas, however, can be strongly influenced by the mineralogy of the local geological outcrops, or by the accumulation of shelly carbonate material. Beaches can consist almost entirely of heavy minerals in volcanic areas, and the usually small amounts of heavy minerals in continental beach sediments, such as magnetite, hornblende, and garnet, help to identify the source rocks, their relative importance, and the direction of longshore transport.

Bulk density and packing

Bulk density reflects the way the grains are arranged or packed together. Spherical grains of uniform size can be packed in four ways. The centers of grains in unstable cubic packing describe the corners of a cube, whereas a tetragonal arrangement is formed by moving the upper layer of grains so that they occupy the hollows between the grains below. With orthorhombic packing, the centers of the lower layer of grains form a diamond pattern, with the centers of the grains in the upper layer directly above. A rhombohedral arrangement is created by moving the upper layer of grains into the hollows created by the lower layer. The porosity of the sediments is 48%, 30%, 40%, and 26% with cubic, tetragonal, orthorhombic, and rhombohedral packing of spherical grains, respectively.

The shape of the grains exerts an important influence on the bulk properties of a sediment, including its packing geometry, stability, porosity, and permeability. Small cavities are created in a deposit by shell fragments and other flat, flaky, or plate-like particles, which greatly increase its porosity. Differences in the size of the grains also affect packing density and porosity. Smaller grains occupy the spaces between larger grains, increasing the packing density and decreasing the porosity. Grains that are less than about one-seventh the size of the larger grains can pass down through the voids between the larger grains. Packing is also influenced by deposition rates. Cubic arrangements develop when there are high depositional rates and grain collisions, and rhombohedral packing, when slow deposition allows grains to settle into their optimum positions. Grains settling onto the bed with high fall velocities jostle and vibrate the underlying layers, increasing the packing density and reducing the porosity. Suspended grains settling out in still water are also less densely packed that those that are deposited by waves and currents.

Table B5 The mean density of some minerals found in beach sands

Mineral	Density (kg m^{-3})
Aragonite	2,930
Augite	3,400
Calcite	2,710
Foraminifera shells	1,500
Garnet	3,950
Hornblende	3,200
Magnetite	5,200
Microcline	2,560
Muscovite	2,850
Orthoclase	2,550
Plagioclase	2,690
Quartz	2,650
Rutile	4,400
Zircon	4,600

Grain sorting

Grains are sorted or separated according to their shape, size, and density (Table B6). Beach sediments are generally better sorted than river sediments, but less well than dunes. Beach grain-size distributions are occasionally positively skewed, but the skew is generally negative. Although the presence of a tail of coarse grains has been attributed to the removal of fine grains, or the addition of coarse clasts or shells, skewness can also arise from a single sedimentary event, and it is not necessarily symptomatic of the mixing of two or more sediment populations (McLaren, 1981).

Cross-shore and longshore changes in beach sediment characteristics can result from mechanical and chemical breakdown, differential transport of grains according to their size, longshore variations in wave energy, the addition or loss of sediment, or the mixing of two or more distinct sediment populations. Sorting occurs through selection, breaking, and mixing (Carter, 1988). Rejection and acceptance phenomena play an important role in the selection process, and in perpetuating sorted grain distributions on beaches. Rejection accelerates the transport of coarse grains over finer grains, whereas shielding impedes the movement of fine grains over coarser grains. Grains moving over material of similar size have a high probability of being assimilated or accepted by the underlying material.

Erosion of a source material produces a lag deposit that is coarser, better sorted, and more positively skewed than the original sediment. If all the transported sediment is deposited, the deposit will be finer, better sorted, and more negatively skewed than the source. If the transported sediment is only selectively deposited, the deposit will be better sorted and more positively skewed than the source. The deposit will be finer than the source if only material finer than the mean size of the source is eroded, but it may be coarser if sediment larger than the mean size is removed from the original deposit (McLaren, 1981).

The mean grain size of beach sediments depends on the characteristics of the source and the nature of the sedimentary processes. Mean grain size varies according to differences in wave energy along beaches and on the exposed and sheltered sides of islands, and it also changes through time as gently sloping, storm-eroded beaches recover to their steeper, fully accreted states. In the cross-shore direction, the coarsest sediments are generally found on a beach at the plunge point of the breaking waves, and the grains tend to become finer seawards and

Table B6 Factors controlling sediment sorting (Steidtmann, 1982; with permission of Blackwell Science)

Rate of sediment accumulation
- Slow—allows reworking of grains
- Rapid—allows little or no reworking of grains
- None—scour

Nature of the sediment surface
- Size distribution of grains
- Packing and arrangement of grains
- Type of bedforms present

Style of grain motion
- Traction, including sliding and rolling
- Saltation
- Suspension

Fluid characteristics
- Velocity or shear velocity
- Turbulence
- Depth

Grain characteristics
- Size
- Shape
- Density

landwards of this point. There are often coarser sediments on the upper part of the beach, however, which could either have been stranded over the berm crest by large swash events, or it could be a deflation lag deposit, resulting from the aolian removal of finer grains to form dunes. It is not known whether larger or smaller grains move most easily alongshore, and therefore whether examples of beach sediments becoming coarser downdrift represent anomalous or normal situations. In any case, whereas there is often longshore grading on beaches in essentially closed embayments, it is generally lacking or poorly developed where there are large amounts of sediment moving alongshore, or where active sediment throughput does not allow enough time for it to develop (Carter, 1988). The degree of grain-size sorting normal to the beach is also a contentious issue. Some workers have found that the poorest sorting occurs in the breaker and surf zones and the best in the swash zone, whereas others have found that the degree of sorting declines on either side of the breaker zone.

Beach sediments are also sorted according to grain density, and particularly to the abundance and mineralogy of the heavy mineral component. Small heavy mineral grains occupy the spaces between the larger and less dense quartz and feldspar grains, shielding them from the flow so that they are less easily entrained. The lighter quartz grains are transported alongshore more rapidly than the heavy minerals, even when both types of grains have the same settling velocity—presumably because the smaller size of the heavy mineral grains inhibits entrainment during each brief suspension episode. Selective longshore transport of quartz grains may therefore result in heavy minerals becoming concentrated in erosive lag deposits.

There are often concentrations of heavy minerals on beaches in the form of bands or streaks near the high tide or upper swash zones, in the troughs of ripples, or where there are shells, coarse clasts, or other flow obstructions. The upper swash zone may consist of dark layers of fine, heavy mineral grains grading upwards into light colored layers of coarser, quartz–feldspar grains. The alternating layers are between about 1 and 25 mm in thickness, and they typically extend along the beach for a few tens of meters (Clifton, 1969). The formation of swash laminae has been attributed to shear sorting in the downrush, which causes the coarser grains to migrate upwards into the zone of lower shear, while the finer and heavier grains move downwards, into the zone of maximum shear at the bed. An alternate explanation is that the smaller particles tend to fall into the spaces between the larger grains, thereby displacing coarser grains toward the surface.

Heavy mineral concentrations in the cross-shore direction have either been attributed to wave asymmetry, the heavy minerals being carried onshore by high current velocities, but not by the weaker offshore flows, or to beach erosion and offshore transport of the quartz–feldspar grains. There may be poor separation under vigorous wave conditions, however, and the heavy and light minerals can be entrained and transported together.

There has been little research on the effect of sand grain shape on longshore and cross-shore sorting patterns. The proportion of angular grains increases in the direction of longshore transport between Delaware and Chesapeake Bays, possible because their lower settling velocities allow them to remain in suspension longer, so that they are carried further and at higher rates than more rounded grains. On Long Island, however, grains become rounder with longshore transport. In a laboratory and field study, grains of similar size and mineralogy (quartz) were differentially transported and sorted within the swash zone, with the more rounded grains being deposited near the top of the uprush (Trenhaile *et al*., 1996).

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Bibliography

Carter, R.W.G., 1988. *Coastal Environments*. London: Academic Press.

Clifton, H.E., 1969. Beach lamination: nature and origin. *Marine Geology*, **7**: 553–559.

Fieller, N.R.J., Gilbertson, D.D., and Olbricht, W., 1984. A new method for environmental analysis of particle size distribution data from shoreline sediments. *Nature*, **311**: 648–651.

McLaren, P., 1981. An interpretation of trends in grain size measures. *Journal of Sedimentary Petrology*, **51**: 611–624.

Powers, M.C., 1953. A new roundness scale for sedimentary particles. *Journal of Sedimentary Petrology*, **23**: 117–119.

Sleath, J.F.A., 1984. *Sea Bed Mechanics*. Chichester: Wiley.

Steidtmann, J.R., 1982. Size-density sorting of sand-size spheres during deposition from bedload transport and implications concerning hydraulic equivalence. *Sedimentology*, **29**: 877–883.

Sutherland, R.A., and Lee, C.-T., 1994. Application of the log-hyperbolic distribution to Hawaiian beach sands. *Journal of Coastal Research*, **10**: 251–262.

Thomas, M.C., Wiltshire, R.J., and Williams, A.T., 1995. The use of Fourier descriptors in the classification of particle shape. *Sedimentology*, **42**: 635–645.

Trenhaile, A.S., 1997. *Coastal Dynamics and Landforms*. Oxford: Oxford University Press.

Trenhaile, A.S., Van der Nol, L.V., and LaValle, P.D., 1996. Grain roundness and beach sorting. *Journal of Coastal Research*, **12**: 1017–1023.

Winkelmolen, A.M., 1971. Rollability, a functional shape property of sand grains. *Journal of Sedimentary Petrology*, **41**: 703–714.

Cross-references

Cross-Shore Sediment Transport
Cross-Shore Variation of Grain Size on Beaches
Longshore Sediment Transport
Sediment Suspension by Waves
Surf Modeling
Surf Zone Processes

BEACH STRATIGRAPHY

A beach is the boundary between the land and water bodies such as oceans and lakes that develops on wave-dominated coasts. It is defined as a shore consisting mainly of unconsolidated materials extending from the low-water line to where marked changes in physiographic form and/or materials are observed or to the permanent vegetation line. The zone between the low-water and high-water levels, which has a concave topography and slopes gradually seaward, is known as the foreshore or beach face. The area landward from the crest of the most seaward berm of a beach is called the backshore.

The slope gradient of the beach face varies according to material, particularly the grain size, and wave intensity (Carter, 1988; Hardisty, 1990). In general, beaches consisting of coarse-grained materials and high-energy beaches have steeper slopes. Waves and currents continuously change the slope gradient and materials of beaches, resulting in the formation of characteristic sediment facies (Harms *et al*., 1975; McCubbin, 1981).

Succession of coastal sediments

At accumulating or progradational beaches, the succession of coastal sediments consists of lower shoreface, upper shoreface, foreshore, backshore, and dunes in ascending order. This is a typical succession on a wave- or storm-dominated sandy coast. The shoreface, located in the nearshore zone, has a concave topography formed by waves. The upper shoreface, also called the inshore, is a zone with bar and trough topography constantly influenced by waves and wave-induced currents. The migration of bars landward or seaward and rip currents result in the tabular cross-stratification and trough cross-stratification that characterize the upper shoreface sediments. Two-dimensional (2-D) and three-dimensional (3-D) wave ripple structures are also commonly found. These sedimentary facies reflect mostly fair-weather wave conditions. The upper shoreface sediments overlie the lower shoreface sediments, which are characterized by swaley cross-stratification (SCS) or hummocky cross-stratification (HCS). HCS is characterized by low-angle ($<15°$) erosional lower set boundaries with subparallel and undulatory laminae that systematically thicken laterally and by scattered lamina-dip directions (Harms *et al*., 1975). SCS is amalgamated HCS with abundant swaley erosional features. These sedimentary structures are thought to be formed by the oscillatory currents of storm waves with offshore-directed currents. During storms, beaches are eroded and longshore bars migrate seaward. Strong oscillatory currents caused by storm waves agitate sea-bottom sediments at the shoreface. Some of the sediments are transported offshore by bottom currents caused by coastal set-up and gravity currents. Oscillatory currents related to calming storm waves produce HCS/SCS in the shoreface to inner shelf region overlain by wave ripple lamination. HCS/SCS is found only in sediments of coarse silt to fine sand. Because lower shoreface sediments are mainly deposited during storms, there is a sharp boundary formed by bar migration between upper and lower

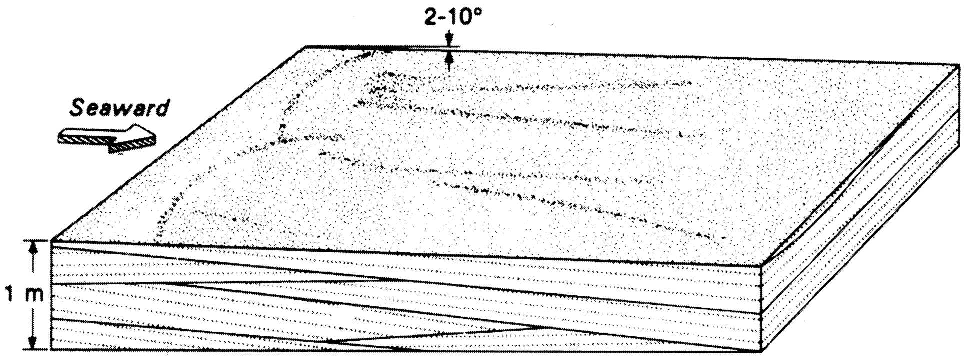

Figure B48 Swash cross-stratification. Stratification and set boundaries are formed parallel to changing slope of beachface and dip generally seaward (after McCubbin, 1981).

shoreface sediments. The lower shoreface topography depends on inner-shelf topography. Because typical shoreface topography can form only on a gentle/flat basal surface, no clear shoreface topography can form in the steep shelf regions of active plate margins. Thus, sometimes only the upper shoreface is referred to as the shoreface.

The uppermost part of the upper shoreface sediments is a step zone sediment characterized by slightly coarser materials, which are overlain by foreshore sediments. The foreshore sediments are characterized by gently seaward-dipping (2–10°) parallel lamination and wedge-shaped set boundaries. This structure is called swash cross-stratification or wedge-shaped cross-stratification (Figure B48). The essential characteristics of this stratification are 1–30 cm-thick bedsets, low-angle dips of laminae and set contacts, an average dip direction toward the sea or lake, mostly erosional set contacts, and laminae lying parallel to set contacts.

Backshore sediments overlie foreshore sediments with a gradual contact and are characterized by low-angle landward-dipping parallel lamination, current ripples, plant remains such as rootlets, and heavy mineral concentrations. Light minerals are removed by winds and form eolian coastal dunes behind the backshore. Heavy mineral concentrations are also a characteristic feature of erosional beaches, where they are residuals of the eroded beach sediments.

The coastal succession and sedimentary facies reflect the intensity of current velocity under fair-weather and storm conditions and seaward-decreasing energy conditions. Under fair-weather conditions, from the foreshore to the upper and lower shoreface, the bedforms (sedimentary structures) found are upper plane beds (parallel lamination), 3-D and 2-D subaqueous dunes (trough and tabular cross-bedding), and 3-D and 2-D ripples (ripple lamination), respectively. On the other hand, under storm conditions, beaches are eroded and the lower shoreface resembles an upper flow regime resulting in the formation of HCS. Ripples are formed in shelf regions.

Foreshore sediments

There are three hierarchies of foreshore sediments: lamination, tide-controlled structures, and storm wave/current-controlled structures.

Foreshore sediments are characterized by parallel lamination formed by the combined processes of wave swash (uprush) and backwash. Each lamina shows reverse grading from fine to coarse with thicknesses of a few millimeters to 2 cm related to each swash and backwash event as a result of either downward filtering of fine particles, or Bagnoldian dispersive pressure resulting from shear between the grains in the flow (Clifton, 1969; Allen, 1984). The fabric of the foreshore sediments shows elongated grains that orient themselves normal to the shoreline, and both landward-imbricated and seaward-imbricated grains are reported. However, these imbricated structures are influenced by the combination of waves and tides.

Reversals of the imbrication dip are thought to result from a predominance of swash transport during the flood stage and backwash transport during the ebb stage. The tidal pattern also influences the depositional thickness of the foreshore sediments. The thick layers are deposited during cycles of higher tidal range, and the thin layers are deposited during cycles of smaller tidal range (Yokokawa and Masuda, 1991). Grain size is also influenced by tides. Water-level changes by tides cause the breaker zone of waves and swash/backwash to shift. Allen (1984) showed that coarser sediments are deposited during flood stages, and finer sediments are deposited during ebb stages.

Storm waves and storm-induced currents have an erosional impact on beaches. During subsequent waning and fair-weather conditions, beaches recover as a result of sediment accretion by waves. This cycle results in an upward-fining succession from a basal erosional surface with coarse-grained materials to finer sandy materials. The coarse deposits formed under high wave energy just after the storm show a remarkable dominance of seaward-dipping imbrication, independent of tidal cycles (Yokokawa and Masuda, 1991). By regarding major erosional surfaces in beach sediments as sequence boundaries according to the sequence stratigraphic model, the depositional zone of foreshore sediments and their stacking pattern can be analyzed. A bedset with a thickness of tens of centimeters bounded by major erosional surfaces is regarded as a depositional sequence, and a lamina set with a thickness of several centimeters to ca. 20 cm is regarded as a parasequence. The depositional pattern of lamina sets shows a landward shift of the depositional zone (onlap) in the lower part of the bedset and a seaward shift of the depositional zone (downlap/progradation) in the upper part. Moreover, bedsets also form a higher order sequence (Masuda *et al.*, 1995).

Changes in waves from seasonal changes in wind direction and wave strength and type produce seasonal beaches. For example, high, strong waves may create high-level beaches with coarse sediments and a steep beachface, or occasionally erosional beaches with residual coarse sediments and heavy minerals, at the high-water level in winter; and calm waves may make gentle, accretional beaches in summer, depending on the location of the beach.

Beaches are distributed not only along wave-dominated coasts but also along tide-dominated coasts influenced by waves. In general, tide-dominated coasts have muddy or sandy tidal flats in the intertidal zone. However, waves create narrow beaches in the upper part of the intertidal zone, occasionally with beach ridges landward from the beach. A typical example is the coast of the Mekong River delta, which is a meso-tidal coast with waves. Beaches and well-developed beach ridges are found in the upper part of the intertidal zone to the supratidal zone (Ta *et al.*, 2002).

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Bibliography

Allen, J.R.L., 1984. *Sedimentary Structures: Their Character and Physical Basis*. Amsterdam: Elsevier.
Carter, R.W.G., 1988. *Coastal Environments*. London: Academic Press.
Clifton, H.E., 1969. Beach lamination: nature and origin. *Marine Geology*, **7**: 553–559.
Hardisty, J., 1990. *Beaches: Form and Process*. London: Unwin Hyman.
Harms, J.C., Southard, J.B., Spearing, D.R., and Walker, R.G., 1975. *Depositional environments as interpreted from primary sedimentary structures and stratification sequences*. SEPM, Short Course No. 2, 161p.
Masuda, F., Okazaki, H., Yokokawa, M., Murakoshi, N., and Sakai, T., 1995. Mini-sequence stratigraphy for Pleistocene beach deposits. In Saito, Y., Hoyanagi, K., and Ito, M. (eds.), *Sequence Stratigraphy,*

Towards a New Dynamic Stratigraphy. Memoir of the Geological Society of Japan, No. 45, pp. 101–109.
McCubbin, D.G., 1981. Barrier-island and strand-plain facies. In Scholle, P.A., and Spearing, D. (eds.), *Sandstone Depositional Environments*. AAPG Memoir 31, pp. 247–279.
Ta, T.K.O., Nguyen, V.L., Tateishi, M., Kobayashi, I., Saito, Y., and Nakamura, T., 2002. Sediment facies and late Holocene progradation of the Mekong River delta in Bentre Province, southern Vietnam: an example of evolution from a tide-dominated to a tide- and wave-dominated delta. *Sedimentary Geology*, **152**: 313–325.
Yokokawa, M., and Masuda, F., 1991. Tidal influence on foreshore deposits, Pacific coast of Japan. In Smith, D.G., Reinson, G.E., Zaitlin, B.A., and Rahmani, R.A. (eds.), *Clastic Tidal Sedimentology*. Canadian Society of Petroleum Geologists. Memoir 16, pp. 315–320.

Cross-references

Beach Erosion
Beach Features
Beach Processes
Rhythmic Patterns
Shelf Processes

BEACH USE AND BEHAVIORS

Introduction

Beaches comprise only 9% of the total conterminous coastline in the United States (Ozmore, 1976). Unfortunately, while no national census of beach visits exists, several studies rank beach recreation as one of the most popular outdoor recreational activities in the United States. It is, therefore, surprising that so little research has been undertaken that addresses the socioeconomic aspects of these activities. This anomaly is particularly noticeable when contrasting the volume of physical and biological research undertaken dealing with beaches and the nearshore environment. Historically, beach recreational activities have centered on the following three activities: bathing, shore-based fishing, and beachcombing. During the past 20 years, many new activities have emerged, several of which use beaches primarily as a staging area. Such activities include surfing, windsurfing, boogie boarding, and a host of shallow-water boating activities including kayaking, canoeing, personal water crafts (PWC), and surfboarding. Many of these activities are incompatible with the more traditional uses of the beach, resulting in user conflicts. Some of these have been managed through the introduction of local, state, and federal legislation, while others have been adjudicated in the courts. Finally, the absolute increase in the number of users of beaches, as well as the diversity of activities occurring, have resulted in a growing demand for both access and accessibility to the nation's beaches.

This entry begins with a historical overview of beach uses, followed by a discussion of three related concerns: increased beach density; the demands this has placed on beach access and accessibility; and how this problem has been addressed. The entry concludes with a discussion about the increasing threat to beachgoers from pollutants in coastal waters.

Overview of beach uses and the factors affecting beach activities

It is likely that beach recreation owes its origin to the perceived value of beaches as healthy environments capable of relieving serious medical conditions (Goodhead and Johnson, 1996). In Britain, during the early part of the 1800s, many people visited beaches with the belief that immersion in, and the drinking of, seawater was healthy and would result in the relief of a number of physical ailments (Meyer-Arendt, 1986). Half a century later, these activities had evolved into resorts generally located within a day's travel of major European and North American cities. Newport, and to a lesser extent Narragansett, Rhode Island, became well-known resorts in New England and were connected by rail to both Boston and New York. In England, Brighton served the same function. More recently, the Hamptons on Long Island have become important beach destinations point for the wealthy. However, nearly all of the research dealing with the early history of beach recreation is descriptive.

One of the few examples where geographers have sought to move from purely descriptive studies to nomothetic research can be found in the extensive writings of Meyer-Arendt, who built on the early work of British geographers with an interest in beach recreation. These studies centered on the concomitant urbanization of coastal areas. Meyer-Arendt studied a number of beach resorts along the northeast Gulf coast, and, based on these efforts, developed the Coastal Resort Morphology Mode (Meyer-Arendt, 1986). This is a spatiotemporal, five-stage recreational land use model. The initial stage is characterized by easy beach access that has attracted limited residential developments, which, in turn, support a small recreational business district (RBD). Toward the end of this first stage, increased day visitation takes place. This second phase is referred to as the "Take Off" stage and is characterized by increased recreational development, extending outward from the RBD. Most of this development is along the coastline on both sides of the RBD. Sometimes a recreational fishing pier is constructed, usually at the foot of the RBD. The third phase is dominated by further development and urban expansion. True central business district (CBD) land uses characterize the area immediately surrounding the RBD. Residential developments continue to expand outward, and most of the early structures located closest to the CBD undergo rapid demolition or conversion to more up-scale recreational developments. Most of the structures still cater to a seasonal clientele, but with a small core of year-round residents. If the resort is located on a barrier beach, developments will have reached the bay-shore. As a result, much of the wetlands located there will have been destroyed by canalization, or filled in, resulting in significant environmental impacts. The fourth stage is consolidating the developments characterized in the previous stage, except that condominium developments now cater to those who no longer can afford to buy (let alone build) single-family homes. The final phase is characterized by complete saturation, where lower income residents and those on fixed incomes are forced to sell out, in part because of high property values and property taxes. Dolan and his co-workers analyzed the rise and decline of religious sea camps during the 19th century, only vestiges of which exist today.

Following the end of World War II, beach visitation became one of the most popular outdoor recreational activities that cut across all socioeconomic groups, although significant social and ethnic discrimination was still in evidence. Furthermore, the popularity of beach visitation continued to increase in concert with a general population migration from inland to coastal areas (Kimmelman *et al*., 1974). Perhaps a contributing factor to the popularity of beach recreation related to the low cost associated with bathing, where transportation often represented the only cost of engaging in the activity. During this period, the predominant activities were bathing, sunning, and socializing, attractions that are as popular today as they were then. The only difference is that today many beachgoers are experiencing significant competition due to other outdoor recreational activities that utilize the beach as a staging area for other water-based activities. These include shore-based fishing and the launching of a variety of light vessels that can be trailered or car-topped, including kayaks, canoes, surf and sailboards, and PWCs.

For many beachgoers, the beach represents a place on which a host of activities can be undertaken, including bathing, sunbathing, ball playing, and socializing. While some degree of specialization appears to take place on certain beaches, most beach visitors tend to participate in several different activities during a day-on-the-beach.

Most early studies concerned with beach uses sought to describe and classify beach users and beaches based on perceived preferences. Several of these studies emerged from the Chicago School of Geography; under the direction of Gilbert White (1973) and his students, where the resource users' perceptions of the environment were seen as the primary factors influencing behavior. The initial research thrust dealt with perceived flooding risks, but these studies soon expanded to include all kinds of perceived environmental factors influencing behavior, including those affecting beach visitation. Few studies have analyzed the activities and social interactions occurring on the beach (Gerlach, 1987). Examples of these include Hecock who concluded that beachgoers were attracted to certain beaches based on their physical characteristics (Hecock, 1966). This study suggested that younger beachgoers preferred beaches with a stronger wave environment where bodysurfing could be undertaken. Conversely, families with small children preferred beaches where the wave environment was more gentle and where the beach slope was less steep, allowing children to play safely in the shallows (Jubenville, 1976).

One area that most coastal recreational planners and resource managers have addressed concerns the number of beach visitors that a given beach can accommodate. While no overall accepted standard exists on the number a given site can accommodate before the perceived value of a visit begins to decline, some efforts were made to address this issue nationally. The Outdoor Recreational Resources Review Commission suggested that 2,000 bathers could be accommodated per one mile of beach (Rockefeller, 1962). One problem with this measure is that no distinction is being made on the basis of the width of the beach. Jackson (1972), citing a California study, suggested that each bather in lakes required a minimum of 50 square feet of water. Other factors play a role in the decision-making process leading to a person, family, or group deciding to visit a given beach. In a study conducted in the New York–New Jersey Metropolitan area, West (1973) found that access and especially accessibility were considered more important factors compared to water quality.

By far, most of the social studies conducted on beaches have dealt with density and crowding (Boots, 1979). In this context, several authors identified "crowding" as a factor influencing beach use (Sowman, 1987). De Ruyck *et al*. (1987) identified two types of densities, one of which defined overall density as a number of visitors per unit area. He also defined "patch densities," a term he referred to as "social carrying capacity" on three beaches in South Africa. These researchers found that density tolerance was influenced by the size of the beach (the smaller the beach, the greater the willingness to accept more people (greater densities). He also found that "crowd attracting beach activities," such as impromptu ball games and other sports events, resulted in higher crowding tolerance by the visitors.

In an unrelated study, West (1974) also found that beachgoers' perception of beach density varied depending upon the respondent's residence. Those beach visitors living in urban areas were willing to tolerate greater beach crowding compared with those living in suburban areas.

Beach access and accessibility

Access and accessibility are terms often used interchangeably, however, in this entry access refers to the ability to move from an existing "right-of-way," such as a road or public parking lot, to a public beach. Accessibility refers to the obstacles that a beach visitor may encounter in traveling from his or her home to the beach. Such obstacles may include a lack of parking facilities, high entry fees, or in an urban context, a lack of public transportation to the beach.

Physical access to the shore is governed by two sets of law, one related to common law, the other by legislation. The common law principles concerning beach use originate from old Roman Law, which held that beach resources (seaweed, fish, and shellfish) were held by the sovereign, who then allowed the citizens to fish and collect seaweed from the shore. This principle was adopted in Britain during the Roman reign and eventually transferred to North America during the Colonial Period where, following independence, the concept of the "sovereign" was replaced with the general public. This meant that the government held the submerged lands seaward of the mean high water line (MHWL) in trust for the general public. In most US states, the legal definition of the public domain is seaward of the MHWL (Anon, 1988). The MHWL, in turn, is defined on the basis of the location of the average high-tide shoreline during a full metonic cycle.

A legislative approach to increasing public access was initially implied in the Coastal Zone Management Act (1972), and in the subsequent amendments. The 1986 amendments were identified as an area of special interest. Most coastal states have made some efforts to increase public access to the nation's beaches, although accomplishments vary widely. One of the aims of California's and Oregon's, and to a lesser extent Washington State's Coastal Management Program has been to increase physical access at certain intervals along their respective coastlines. Along California's rural coast, the aim is to provide coastal access every three miles. This goal is comparable to those formulated in Oregon and Washington. The objective of providing access to the shore at regular intervals has been more problematic along the Eastern Seaboard, in large part because of much higher population densities, less land in public ownership, and overall higher land prices. Together these factors have made eminent domain acquisition much more difficult and costly. Some states have attempted to increase coastal access using the principle of perfecting public right-of-ways. Rhode Island has undertaken a statewide search to identify existing and abandoned right-of-ways, largely through legal research. This effort has significantly increased public access to the state's coastal areas.

The absence of physical access to the beach is only one of the many constraints that a potential beach visitor is likely to encounter. Lack of accessibility may at times be a greater hindrance to visiting the beach. Such factors may be deliberate attempts by local cities and towns to limit or outright prohibit out-of-town visitors on local beaches. In other instances, impeded accessibility is unintentional or unavoidable (Heatwole and West, 1980).

Limiting or prohibiting beach access to out-of-town citizens on facilities owned and operated by local municipalities may vary from outright prohibition to charging unreasonably high entry or parking fees. Many of these instances have been adjudicated in the courts, which have generally ruled that where higher entrance fees have been levied against nonresidents, such fees may be permitted as long as the increased fees cover the additional costs resulting from accommodating the increased number of nonresidents. The courts have generally assumed that a portion of a resident's property tax is designated to the operation of recreational facilities, including beaches, and that opening such beaches to nonresidents often means increased expenditures to insure the health and welfare of the visitors. This may mean higher costs to cover the costs of additional guards, beach patrols, cleanups, and other services. The courts have generally felt that such additional expenditures could be recovered by charging the nonresidents a higher fee compared with those levied on residents (Neptune v Burrough of the City of Avon, 1972).

The popularity of beaches and beach uses has increased significantly during the latter part of the 20th century, a development that is likely to increase for the foreseeable future. This increased demand has raised two concerns: use conflict and water quality declines.

Conflict resolution

As mentioned in the introduction, many additional beach uses now exist. Some of these are incompatible with traditional recreational beach activities. Examples include shore-based fishing, various boating activities, including water skiing, use of personal water crafts and surfing. Most of these conflicts have been dealt with on the local level, while a few have been adjudicated in a court of law. Of the management procedures that have been introduced on the local level, zoning procedures are probably the most common. Zoning, as it was first conceptualized in New York City in 1916 (Haar, 1977), was originally intended to control building height. Zoning maps later followed with zoning ordinances specifying restricted or prohibited uses.

Recreational applications of zoning have been attempted both on land and on the water in an attempt to reduce conflicts between and among different recreational pursuits. On the water, zoning has been used by a number of municipalities to segregate swimmers and bathers from boaters—especially powerboaters, surfers, and PWCs. Two versions of zoning have been used: permanent zones and space/time zoning ordinances. In the case of permanent zones, a protected activity (e.g., swimming or bathing) is protected from all other activities by prohibiting those from entering the designated area. A less common practice is sometimes referred to as time zoning. In this instance, the competing uses are assigned different periods when each activity can take place, thereby eliminating any conflicts between competing uses. If a given beach is sought by both swimmers and surfers, the beach may be restricted to one user group while the other use may be permitted during different periods. A coastal municipality may allow surfers access to the beach during the early morning and again in the late afternoon. Swimmers and bathers may have exclusive use of the beach and adjacent nearshore during the period from mid-morning to late afternoon.

The same procedures may be utilized on land in areas where users compete for the same area. Sunbathing and shore-based fishing are both legitimate recreational activities that sometimes may compete for the same stretch of beach area. Shore-based fishing may be restricted to the early morning and late afternoon, while sunbathing may be permitted from mid-morning to late afternoon.

Environmental impacts on beach use

Socioeconomic factors are not the only variables influencing beach recreation quantitatively as well as qualitatively. There are at least two additional variables that increasingly have played a role in this nation's beach recreational activities. One concerns the increased outbreaks of algal blooms, in particular, those classified as harmful. The other concerns the impact that beach activities may have on endangered species and the restrictions imposed on beach visitors to protect threatened and endangered biological resources.

Algal blooms have occurred along the nation's coasts at least since the Spanish first settled Florida. However, there is growing evidence that these incidents are increasing quantitatively and qualitatively. The number of harmful algal bloom (HAB) incidents have increased significantly in recent years as have the impacts on marine life, swimmers, bathers, and people handling fish and shellfish affected by these incidents. While the cause for these events has not yet been determined, there is growing evidence that land-based pollution is partially responsible (Anon, 2000). The effects of HAB events range from the discoloration of large patches of waters to fish kills, die-offs of manatees in Florida, and possibly the deaths of small marine mammals in the United States, Scandinavia, and the Mediterranean. So-called "red tide" incidents in Florida, have significantly affected swimmers and bathers. Toxins released from these HABs can also become airborne, resulting in respiratory irritation, coughing, and sneezing by people who are not even in direct contact with the affected waters (Luttenberg, 2001).

The second factor that has influenced swimming and bathing in recent years is the potential conflict between the Endangered Species Act and all types of beach recreational activities. During the 1980s and 1990s, large stretches of barrier beaches on Cape Cod were closed to fishing, overland vehicular traffic, and bathing in an effort to protect the Piping Plover nests and fledglings from being trampled. In 1988, it was estimated that only 20 pairs of piping plovers were nesting within the Cape Cod National Seashore (Lopez, 1998). Largely because of the severe restrictions placed on beach traffic (both pedestrian and vehicular), a substantial increase in nesting pairs has been noted throughout the seashore (Lopez, 1998). These accomplishments, however, have not been made without impacts on beach recreation in general, shore-based fishing or bathing. Within the Cape Cod National Seashore, less than 10 miles of the Atlantic shore are now open to ORV traffic during the nesting season (from March through July). Similar restrictions have been imposed on bathing and beachcombing in piping plover nesting areas.

Conclusions

Beaching and bathing continue to be two of the most popular outdoor recreational activities both here and abroad, yet with few exceptions, not many studies have addressed the behavior and motivation of the beach-going public. Studies conducted on or adjacent to beaches generally fall into two groups. The first has sought to analyze the reaction of the beach visiting public to deteriorating water quality. The second group of studies has concentrated on infrastructure changes that have taken place in the areas immediately inland from many popular bathing beaches. While these studies are important both socially and economically, it is suggested that many additional findings would enhance our understanding of the factors motivating the beach visitor and the management of beaches. Answers that are needed include studies dealing with crowding and density tolerances and better understanding of beach preferences by different user groups. Are some beaches attracting certain population groups simply because they are more accessible, or because the amenities found on the beach attract specific user groups interested in participating in activities (e.g., surfing) that may not be readily available on all beaches? The role of physical access is still an issue in many communities, notwithstanding that access along the shore is recognized by most states as a public right, Public beaches constitute less than 10% of all the beaches in the United States. This increasingly scarce resource may be better managed if we had a better understanding of the factors that attract and detract the public to certain beaches.

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Bibliography

Anon, 1988. *Managing the Shoreline for Water Dependent Uses: A Handbook of Legal Tools*, Marine Law Institute, University of Maine School of Law, Portland, ME.

Anon, 2000. National Assessment of Harmful Algal Blooms in U.S. Waters, Washington: National Science and Technology Council Committee on the Environment and Natural Resources.

Boots, B.N., 1979. Population Density, Crowding and Human Behavior. *Progress in Human Geography*, **3**(1): 13–63.

De Ruyck, M.C., Soares, A.G., and McLachlan, A. 1987. Social carrying capacity as a management tool for sandy beaches. *Journal of Coastal Management*, **13**(2): 822–830.

Gerlach, J., 1987. Spring Break at Padre Island: A New Type of Tourism. *Paper presented at the Annual Meeting of the AAG,* Portland, OR.

Goodhead, T. and Johnson, D., 1996. Maritime Leisure. In Goodhead, T., and Johnson, D. (eds.), *Coastal Recreation Management*. London: E & FN Spoon Publishers.

Haar, C.M., 1977. *Land-Use Planning: A Casebook on the Use, Misuse and Re-use of Urban Land*, Boston, Little Brown and Company.

Heatwole, C.A., and West, N. 1980. Mass transit and beach access in New York City. *Geographical Review,* **70**: 210–217.

Hecock, R.D., 1966. Public beach recreation opportunities and patterns of consumption on Cape Cod. unpublished Ph.D. dissertation, Worcester: Graduate School of Geography, Clark University.

Jackson, R., 1972. 'Recreational zoning and lake planning'. *Town Planning Review,* **43**(1): 41–55.

Jubenville, A., 1976. *Outdoor Recreation Planning*, Philadelphia: W.B. Saunders.

Kimmelman, B. *et al*., 1974. *Studies in Environment: Outdoor Recreation and the Environment*, Vol. V. Washington: Office of Research and Development, EPA, Publication EPA 600/5/73-012.

Lopez, J.J. 1998. *The use of negotiated rulemaking in piping plover/off-road vehicle management at Cape Cod national seashore*. unpublished Masters of Arts thesis, Kingston: Department of Marine Affairs, University of Rhode Island.

Luttenberg, D. 2001. Managing harmful algal blooms in U.S. waters: the evolution and implementation of monitoring and management. unpublished Paper, Kingston: Department of Marine Affairs, University of Rhode Island.

Meyer-Arendt, K.J., 1986. 'Barrier island settlement and landuse evolution: a gulf coast model'. *Proceedings* of *The Coastal Society's 10th National Conference*, New Orleans, LA.

Neptune City v Borough of Avon-By-the-Sea 1972:294 A.2d47 {N.J.1972}

Ozmore, K., 1976. 'Texas—A pioneer in open beaches, In Susan H. Anderson (ed.), *Recreation—Marine Promise: Proceedings, Discussion, and Overview of the National Conference on Marine Recreation*. Newport Beach, California, October 2–4, 1975.

Rockefeller, L. 1962, outdoor Recreation for America: A Report to the President & to Congress by the outdoor Recreation Resources Review Commission, U.S. Gov. Printing Office, Washington DC.

Sowman, M.R. 1987. A Procedure for assessing recreational carrying capacity of coastal resort areas. *Landscape and Urban Planning*, **14**.

West, N., 1973. An analysis of factors influencing beach activities in the New York-New Jersey metropolitan estuarial region. unpublished Ph.D. dissertation, New Brunswick: Department of Geography, Rutgers, the State University, New Brunswick.

West, N., 1974 Perceived density as a factor in selecting beaches. unpublished New Brunswick: Study, Department of Geography, Rutgers, the State University.

White, G.F., 1973. *Natural Hazard Perception and Choice*, Cambridge: Oxford University Press.

Cross-references

Beach Processes
Cleaning Beaches
Coastal Boundaries
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Human Impact on Coasts
Lifesaving and Beach Safety
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Tourism and Coastal Development
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BEACHROCK

Formation and distribution of beachrock

Beachrock is defined by Scoffin and Stoddart (1987, p. 401) as "the consolidated deposit that results from lithification by calcium carbonate of sediment in the intertidal and spray zones of mainly tropical coasts." Beachrock units form under a thin cover of sediment and generally overlie unconsolidated sand, although they may rest on any type of foundation. Maximum rates of subsurface beachrock cementation are thought to occur in the area of the beach that experiences the most wetting *and* drying—below the foreshore in the area of water table excursion between the neap low and high tide levels (Amieux *et al*., 1989; Higgins, 1994). Figure B49 shows a beachrock formation displaying typical attributes.

There are a number of theories regarding the process of beach sand cementation. Different mechanisms of cementation appear to be responsible at different localities. The primary mechanisms proposed for the origin of beachrock cements are as follows:

(1) physicochemical precipitation of high-Mg calcite and aragonite from seawater as a result of high temperatures, $CaCO_3$

Figure B49 Multiple unit beachrock exposure at barrio Rio Grande de Aguada, Puerto Rico. The sculpted morphology, development of a nearly vertical landward edge, and dark staining of outer surface by cyanobacteria indicate that this beachrock has experienced extended exposure. Landward relief and imbricate morphology of beachrock units define shore-parallel runnels that impound seawater (photo: R. Turner).

supersaturation, and/or evaporation (Ginsburg, 1953; Stoddart and Cann, 1965);

(2) physicochemical precipitation of low-Mg calcite and aragonite by mixing of meteoric and fresh groundwater with seawater (Schmalz, 1971);

(3) physicochemical precipitation of high-Mg calcite and aragonite by degassing of CO_2 from beach sediment pore water (Thorstensen *et al*., 1972; Hanor, 1978); and

(4) precipitation of micritic calcium carbonate as a byproduct of microbiological activity (Taylor and Illing, 1969; Krumbein, 1979; Strasser *et al*., 1989; Molenaar and Venmans, 1993; Bernier *et al*., 1997).

Although most beachrock cement morphologies suggest an inorganic origin, physicochemical mechanisms operating alone do not adequately account for the discontinuous distribution of beachrock formations. As Kaye (1959, p. 73) put it, "the problem hinges more on an adequate explanation for the absence of beachrock from many beaches than on its presence in others." The discontinuity of beach cementation, along with the complex assemblage of cement types found in adjacent samples of beachrock led Taylor and Illing (1971) to propose that the microenvironment exerts a greater influence on the cementation process than does the macroenvironment.

Several beachrock researchers concur with this assessment and support the theory that *initial* cementation in beach sands is controlled by the distribution and metabolic activity of bacteria because: (1) dark, organic-rich micritic rims have been identified around cemented grains in most petrographic studies of beachrock (Krumbein, 1979; Beier, 1985); (2) microbially mediated precipitation of carbonates has been repeatedly demonstrated in both marine and terrestrial environments (Buczynski and Chafetz, 1993); and (3) bacterial populations are particularly large and productive in the intertidal zone of water table fluctuation where beach lithification occurs. Once biologically mediated cryptocrystalline cements are established as nucleation sites, larger crystals precipitated via physicochemical processes can grow and bridge the sediment grains.

Rates of beachrock formation are undoubtedly variable but are generally believed to be quite rapid, on the scale of months to years (Frankel, 1968). For example, Hopley (1986) reported that beachrock formed within six months on Magnetic Island near Townsville, Australia, while Moresby (1835) wrote that Indian Ocean natives made an annual harvest of beachrock for building stone and within a year they had a new lithified crop.

Several Pleistocene and older beachrock formations have been identified. However, the dynamic nature of sandy coastlines and a historically fluctuating sea level necessitate that most occurrences of *intertidal* beachrock are less than 2,000 years old. This is commonly supported by the incorporation of modern man-made artifacts in beachrock formations rather than by ^{14}C dates, as beachrock is poorly suited for radiocarbon dating.

The majority of recent beachrock is formed on beaches in the same regions that favor coral reef formation. This is generally below 25° latitude where there is a well-defined dry season and "the temperature of ground water at a depth of about 76 cm in beaches remains above 21°C for at least 8 months of the year" (Russell, 1971, p. 2343). However, beachrock can also form at higher latitudes. For example, beachrock exposures are common throughout the Mediterranean and have been reported along portions of the coasts of Norway, Denmark, Poland, Japan, New Zealand, South Africa, the Black Sea, and the northern Gulf of Mexico. Beachrock formations have also been reported on lakeshores in Pennsylvania, Michigan, Africa, New Zealand, southeast Australia, and the Sinai Peninsula.

Subaerially exposed beachrock units constitute only a small proportion of the cemented sediments in the intertidal zone. For example, Emery and Cox (1956) found beachrock *exposures* on only 24% of the predominantly calcareous beaches of Oahu, Kauai, and Maui, whereas jet-probing conducted by Moberly (1968, p. 32) revealed that "exposed or covered beachrock appears to be present at all calcareous beaches in the state" of Hawaii. In the event of continued sea-level rise and human activities that exacerbate coastal erosion, much more beachrock will be exhumed.

Morphology of beachrock formations

Beachrock formations typically consist of multiple units, representing multiple episodes of cementation and exposure. Beachrock that forms below the foreshore has an upper surface slope that tends to mimic that of the seaward dipping (4–10°) internal beach bedding. However, beach sand cementation has also been found to occur below the berm and foredune of a beach (Russell, 1971; Hopley and MacKay, 1978). Those authors found that the beachrock forming below the backshore had a nearly horizontal upper surface that corresponded to the groundwater table and truncated the original beach bedding.

Most intertidal beachrock formations are detached from subaerial and subtidal cemented sediments. Beachrock is laterally discontinuous as well, usually exposed for only short distances before disappearing under loose sand or ending entirely. It is likely that the formation and preservation of beachrock on a given section of beach is negatively correlated to alongshore increase in wave energy and frequency of beach erosion.

The reported thickness of beachrock formations ranges from a few centimeters up to 5 m, with approximately 2 m being most common. Variations in degree of cementation within a beachrock unit can be controlled by variability in porosity, permeability, and composition of different sand layers (Molenaar and Venmans, 1993). Generally, precipitation of cements is most rapid near the top of a beachrock unit. Accordingly, young beachrock units are better cemented at the top and noticeably less so near the base. This attribute makes them more susceptible to scour at their base upon exposure, commonly resulting in undercutting and slumping. It is this undercutting that fosters the development of nearly vertical landward edges on beachrock units. In areas where a chronic deficit in sand supply or erosive conditions have exhumed the seaward edge of a beachrock formation, it is frequently observed to be steep as well.

Long-term exposure of beachrock will radically change the ecology of a sandy shoreline by providing a hard substrate that can support an increased diversity of animal and plant life. The reader is referred to the papers of McLean (1974), Jones and Goodbody (1984), and Miller and Mason (1994) to learn more about the ecology and biophysical modification of intertidal beachrock exposures.

Beachrock and coastal evolution

Although beachrock, as defined, forms in the intertidal zone, it does not always remain there. On prograding coasts, a series of beachrock units may form at depth, leaving older units stranded well behind the active beach. On retreating coasts, outcrops of beachrock may be evident offshore where they may serve as a hard substrate for coralgal reef growth. If the strike of the beach changes over time, then the strike of the beachrock units will reflect that change.

Armed with the knowledge that beachrock is formed in the intertidal zone, many geologists have related beachrock outcrops to changes in sea level for particular coasts. Semeniuk and Searle (1987) demonstrated that beachrock formation can keep pace with slow shore recession, resulting in a wide, continuous band of beachrock, but that rapid shore recessions (or periods of high wave energy and foreshore instability) are represented by gaps (unconsolidated sediment) in a sequence of beachrock units. Assuming a nearly constant rate of sea-level rise, these gaps may indicate that beachrock can temporarily stabilize the position of the shore under erosive conditions until sea level has risen enough to cause the shore to jump back (Cooper, 1991). Many other researchers have asserted that beachrock outcrops will protect a beach from erosion, as well as control the plan configuration of a coastline.

Research by Turner (1999) has demonstrated that the influence beachrock has on beach processes will largely depend on the extent and morphology of the exposure, both of which evolve over time. Cumulative exposure and erosion of a beachrock formation over a period of years to decades can foster a gradual increase in the landward and seaward relief of the beachrock units and the development of shore-parallel runnels and shore-perpendicular breaches in the beachrock. The high seaward relief of such a beachrock unit effectively attenuates incident wave energy and retards onshore sediment transport. The high landward relief of the beachrock unit can act as the seaward wall of a runnel that blocks offshore return of backwash and forces impounded seawater and entrained sand to flow laterally on the foreshore to low spots and shore-normal breaches in the beachrock formation. Beachrock breaches and runnels are erosionally enlarged over time, locally increasing onshore inputs of wave energy and longshore sediment transport rates on the foreshore.

On a beach on Puerto Rico's west coast, beach width and volume were found to be least stable where the seaward beachrock unit was breached and most stable away from the breaches behind high relief beachrock. Sections of foreshore most protected by a high relief beachrock ridge exhibited the lowest volumes of subaerial sand storage, unusually narrow beach widths, and the slowest beach erosion recovery rates. In short, a beach with a high relief intertidal beachrock exposure is more likely to be sediment deficient and out of synch with the wave regime. This puts the backshore of a beachrock beach at risk of catastrophic retreat following the development of a breach in the beachrock or in the event of a high energy wave event coupled with a storm surge or spring high tide.

Conclusions

The transformation of sandy beaches to rocky beachrock beaches is increasingly common in the tropics and subtropics. Where beachrock is exposed by erosion, it acts as a natural breakwater or revetment, decelerating further shoreline and backshore retreat. However, it also tends to retard beach buildup and is poorly suited to recreational use, both major issues in the tropics where tourism is often the primary source of income. The potential for beachrock to significantly alter the evolution of a coast justifies additional research on its influence on beach processes. In particular, the characteristics and effects of beachrock on dissipative beaches have received little attention and are likely to be significantly different than those observed on more reflective beaches.

Despite many petrographic investigations of beachrock cements, the processes responsible for beachrock formation are still poorly understood. Given the likelihood of cement diagenesis in the beach environment, there is a need to pursue other research methods. For example, the subsurface formation of beachrock should be tracked on a variety of beaches over an extended period. The processes affecting beach sand cementation should also be examined under controlled conditions in a laboratory setting. Preliminary experiments conducted by Turner (1995) indicate that the addition of dissolved nitrate or organic carbon to beach sand microcosms stimulates bacterial growth and the precipitation of intergranular calcium carbonate. This leads to the question as to whether coastal discharges of groundwater contaminated with fertilizers or human wastes are increasing the rate and geographic range of beachrock formation.

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Bibliography

Amieux, P., Bernier, P., Dalongeville, R., and Medwecki, V., 1989. Cathodoluminescence of carbonate-cemented Holocene beachrock from the Togo coastline (West Africa): an approach to early diagenesis. *Sedimentary Geology*, **65**: 261–272.

Beier, J.A., 1985. Diagenesis of Quaternary Bahamian beachrock: petrographic and isotopic evidence. *Journal of Sedimentary Petrology*, **55**: 755–761.

Bernier, P., Guidi, J.B., and Bottcher, M.E., 1997. Coastal progradation and very early diagenesis of ulramafic sands as a result of rubble discharge from asbestos excavations (northern Corsica, western Mediterranean). *Marine Geology*, **144**: 163–175.

Buczynski, C., and Chafetz, H.S., 1993. Habit of bacterially induced precipitates of calcium carbonate: examples from laboratory experiments and Recent sediments. In Rezak, R. and Lavoie, D.L. (eds.), *Carbonate Microfabrics*. New York: Springer-Verlag, pp. 105–116.

Cooper, J.A.G., 1991. Beachrock formation in low latitudes: implications for coastal evolutionary models. *Marine Geology*, **98**: 145–154.

Emery, K.O., and Cox, D.C., 1956. Beachrock in the Hawaiian Islands. *Pacific Science*, **10**: 382–402.

Frankel, E., 1968. Rate of formation of beachrock. *Earth and Planetary Science Letters*, **4**: 439–440.

Ginsburg, R.N., 1953. Beachrock in south Florida. *Journal of Sedimentary Petrology*, **23**: 85–92.

Hanor, J.S., 1978. Precipitation of beachrock cements: mixing of marine and meteoric waters vs. CO_2-degassing. *Journal of Sedimentary Petrology*, **48**: 489–501.

Higgins, C.G., 1994. Subsurface environment of beaches—temperature and salinity. Geologic Society of America, Abstracts with Programs, Seattle Meeting, A–364.

Hopley, D., 1986. Beachrock as a sea-level indicator. In van de Plassche, O., (ed.), *Sea-level Research: A Manual for the Collection and Evaluation of Data*. Norwich: Geo Books, Regency House, pp. 157–173.

Hopley, D., and Mackay, M.G., 1978. An investigation of morphological zonation of beach rock erosional features. *Earth Surface Processes*, **3**: 363–377.

Jones, B. and Goodbody, Q.H., 1984. Biological alteration of beachrock on Grand Cayman Island, British West Indies. *Bulletin of Canadian Petroleum Geology*, **32**: 201–215.

Kaye, C.A., 1959. Shoreline features and Quaternary shoreline changes, Puerto Rico. *U.S.G.S. Professional Paper*, **317**(B): 49–140.

Krumbein, W.E., 1979. Photolithotropic and chemoorganotrophic activity of bacteria and algae as related to beachrock formation and degradation (Gulf of Aqaba, Sinai). *Geomicrobiology*, **1**: 139–203.

McLean, R.F., 1974. Geologic significance of bioerosion of beachrock. *Proceedings of the 2nd International Coral Reef Symposium, Brisbane*, **2**: 401–408.

Miller, W.R., and Mason, T.R., 1994. Erosional features of coastal beachrock and aeolianite outcrops in Natal and Zululand, South Africa. *Journal of Coastal Research*, **10**(2): 374–394.

Moberly, R., 1968. Loss of Hawaiian littoral sand. *Journal of Sedimentary Petrology*, **38**(1): 17–34.

Molenaar, N., and Venmans, A.A.M., 1993. Calcium carbonate cementation of sand: a method for producing artificially cemented samples for geotechnical testing and a comparison with natural cementation processes. *Engineering Geology*, **35**: 103–122.
Moresby, R.M., 1835. Extracts from Commander Moresbys' report on the northern atolls of the Maldives. *Journal of the Royal Geography Society of London*, **5**: 398–404.
Russell, R.J., 1971. Water-table effects on seacoasts. *Geology Society of America Bulletin*, **82**: 2343–2348.
Schmalz, R.F., 1971. Formation of beach rock at Eniwetok Atoll. In Bricker, O.P., (ed.), *Carbonate Cements*. Baltimore: Johns Hopkins University Press, pp. 17–24.
Scoffin, T.P., and Stoddart, D.R., 1987. Beachrock and intertidal cements. In Scoffin, T.P. (ed.), *An Introduction to Carbonate Sediments and Rocks*. Glasgow: Blackie Publishing Company, pp. 401–425.
Semeniuk, V. and Searle, D.J., 1987. Beach rock ridges/bands along a high-energy coast in southwestern Australia—their significance and use in coastal history. *Journal of Coastal Research*, **3**(3): 331–342.
Stoddart, D.R. and Cann, J.R., 1965. Nature and origin of beach rock. *Journal of Sedimentary Petrology*, **35**(1): 243–273.
Strasser, A., Davaud, E., and Jedoui, Y., 1989. Carbonate cements in Holocene beachrock: example from Bahiret el Biban, southeastern Tunisia. *Sedimentary Geology*, **62**: 89–100.
Taylor, J.C.M., and Illing, L.V., 1969. Holocene intertidal calcium carbonate cementation, Qatar, Persian Gulf. *Sedimentology*, **12**: 69–107.
Thorstenson, D.C., Mackenzie, F.T., and Ristvet, B.L., 1972. Experimental vadose and phreatic cementation of skeletal carbonate sand. *Journal of Sedimentary Petrology*, **42**(1): 162–167.
Turner, R.J., 1995. Bacteria and algae-mediated precipitation of calcium carbonate in Puerto Rico sand and seawater suggests that nutrient-rich ground water discharges enhance coastal sand cementation. Geologic Society of America, Abstracts with Programs, New Orleans Meeting, **27**(6): A-346.
Turner, R.J., 1999. Morphodynamic relationship between beachrock exposure and littoral zone processes on the west coast of Puerto Rico. Unpublished dissertation, University of North Carolina, Chapel Hill, 395 p.

Cross-references

Beach Features
Coral Reef Coasts
Eolianite
Rock Coast Processes
Sea-Level Indicators, Geomorphic

BEAUFORT WIND SCALE

The Beaufort scale of wind velocity relates wind speed to the physical appearance of the sea surface by considering such factors as apparent wave height and the prominence of breakers, whitecaps, foam and spray. It is the oldest method of judging wind force. Originally devised by Admiral Sir Francis Beaufort of the British Navy in 1805 to simplify the signaling of wind and weather conditions between sailing vessels, it has since been repeatedly modified to make it more relevant to modern navigation. Table B7 gives an updated modern version of the Beaufort scale, adapted from British Admiralty (1952), Thomson (1981), and US Army Coastal Engineering Research Center (1984). Meyers *et al*. (1969) presented an elaborate version of the wind scale based on British Admiralty (1952), McEwen and Lewis (1953), and Pierson *et al*. (1953). Wind speed measured at 11 m (36 feet) above sea surface is usually applied to use the scale. The wave heights are approximate and represent fully arisen sea state. As with any subjective judgment method, the Beaufort Scale is far from perfect. Similar subjective scales have been proposed to assess tornado and hurricane damages. The Fujita scale (F-scale) was proposed in 1951 by Tetsuya Fujita for rating the severity of tornadoes as a measure of the damage. The Saffir–Simpson scale is used for rating the severity of damages by a hurricane.

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Bibliography

British Admiralty, 1952. *Manual of Seamanship*. London: H.M. Stationary Office.
McEwen, W.A. and Lewis, A.H., 1953. Encyclopedia of nautical knowledge. Cambridge, MD: Cornell Maritime Press.
Meyers, J.J., Holm, C.H., and McAllister, R.F., 1969. Handbook of ocean and underwater engineering. New York: McGraw-Hill Book Company.
Pierson, W.J., Neuman, G., and James, R.W., 1953. Practical methods for observing and forecasting ocean waves. New York: New York University, College of Engineering.
Thomson, R.E., 1981. *Oceanography of British Columbia Coast*. Sidney, B.C.: Department of Fisheries and Oceans.
US Army Coastal Engineering Research Center, 1984. *Shore Protection Manual*, 4th edn., Washington, D.C.: US Government Printing Office.

Cross-references

Climate Patterns in the Coastal Zone
Coastal Climate
Meteorologic Coastal Wind Effects on Coasts
Nearshore Wave Measurement
Wave Hindcasting

BIOCONSTRUCTION

The term, *bioconstruction*, usually refers to a bioconstructed limestone that has been built-up by colonial and sediment-binding organisms including algae, corals, bryozoans, and stromatoporoids. The term, bioconstructed limestone, was introduced by Carozzi and Zadnick (1959) in their study of the Silurian Wabash reef in southern Indiana. The word, bioconstructed, was used to distinguish the limestones and dolomites which were found in a reef from the dolomitic calcarinites preserved in the reef flanks and the dolomitic shales in the country rock (Carozzi and Zadnick, 1959). The term, bioconstruction, was next applied to Devonian stromatoporoid reefs in the Beaverhill Lake Formation, Upper Devonian, Alberta Canada (Carozzi, 1961).

European use of the word bioconstruction

The word, *bioconstruction*, was widely accepted and used in European geologic journals, but has not appeared in any North American journals since 1961. The European use of the term, *bioconstruction*, includes what the North American geologists would refer to as reefs, bioherms, and biostromes. Based on living coral reefs, Ladd (1944) defined a reef as a rigid, wave-resistant framework constructed by large skeletal organisms. A broader definition of a reef as "a discrete carbonate structure formed by in-situ organic components that develops topographic relief upon the seafloor" has been proposed by Wood (1999, p. 5). Cumings (1930) defined a bioherm as a mound-like, dome-like, lens-like, or reef-like mass of rock built-up by sedentary organisms (such as corals, algae, foraminifera, mollusks, gastropods, and stromatoporoids), composed almost exclusively of their calcareous remains and enclosed or surrounded by rock of different lithology. A biostrome is defined as a distinctively bedded and widely extensive lenticular, blanket-like mass of rock built by and composed mainly of the remains of sedentary organisms and not swelling into a mound-like or lens-like form; an organic layer, such as a bed of shells, crinoids, or corals, or a modern reef in the course of formation, or even a coal seam (Cumings, 1930).

Types of bioconstructions

Examples of several different types of bioconstructions, which would fall into the categories of reef, bioherms, and biostromes, are included to show how the term bioconstruction is used in the European literature. In Spain, rugose corals and calcareous algae bioconstructions are also called biostromes (Rodrigues and Sanchez, 1994). In Jurassic and Cretaceous strata in Germany, Rehfeld (1996) describes different forms of sponge bioconstructions which comprise bioherms, biostromes, and sponge meadows. The wave resistant calcisponge and algal reefs of the Capitan reef facies, partially wave resistant reef mounds and non-wave resistant skeletal mounds in the Guadalupe Mountains of New Mexico, are described as Permian bioconstructions (Noe, 1996). Therefore, bioconstruction is a general term for limestone and dolomite deposits formed by colonial and sediment binding organisms which include reefs, bioherms, and biostromes.

Table B7 Beaufort wind scale

Beaufort No.	Name	Wind speed knot	m/s	Effect of wind at sea surface	Significant wave height (m)	Effect of wind on land
0	Calm	<1	0.0–0.2	Like a mirror	0	Still, smoke rises vertically
1	Light air	1–3	0.3–1.5	Ripples form with the appearance of scales, but without foam crests	0.1–0.2	Smoke drifts, vanes remain motionless
2	Light breeze	4–6	1.6–3.3	Small wavelets, crests appear glossy but no breaking	0.3–0.5	Leaves rustle, vanes move, wind can be felt on face
3	Gentle breeze	7–10	3.4–5.4	Larger wavelets begin to break, some scattered white horses	0.6–1.0	Constant movement of leaves and small twigs, flags begin to stream
4	Moderate breeze	11–16	5.5–7.9	Small waves predominant but fairly frequent white horses	1.5	Dust and loose paper are lifted, thin branches move
5	Fresh breeze	17–21	8.0–10.7	Moderate waves, distinctly elongated, many white horses, chance of spray	2.0	Small trees in leaf begin to sway
6	Strong breeze	22–27	10.8–13.8	Long waves with extensive white foam, breaking crests, spray likely	3.5	Large branches move, power lines whistle, stop lights sway, umbrellas difficult to control
7	Moderate gale	28–33	13.9–17.1	Sea heaps up and white foam from breaking waves begins to be blown in streaks, spindrift begins to be seen	5.0	Entire trees sway, some resistance to walkers, car feels force of wind
8	Fresh gale	34–40	17.2–20.7	Moderately high waves of greater lengths, edges of crests break into spindrift, foam is blown into well-marked streaks	7.5	Twigs break off trees, difficult walking against wind
9	Strong gale	41–47	20.8–24.4	High waves, rolling sea, dense streaks of foam, spray may affect visibility	9.5	Roof tiles lifted off, windows may be blown in, trees may topple
10	Whole gale	48–55	24.5–28.4	Very high waves with long overhanging crests, foam in great patches blown in dense white streaks downwind, heavy rolling sea causes ships to slam, visibility reduced by spray, sea surface takes on whitish appearance	12.0	Trees uprooted considerable structural damage to some buildings
11	Storm	56–66	28.5–32.7	Exceptionally high waves, sea covered with long white patches of foam blown downwind, wave crests blown into froth everywhere, visibility impeded by spray	15.0	Widespread damage, extensive flooding in low lying areas if wind is directed onshore
12	Hurricane	>66	>32.7	Air filled with foam and spray, sea completely white, visibility seriously impaired	>15.0	Severe structural damage to buildings, widespread devastation and flooding

Conclusions

Bioconstruction is distinctly a European term for a limestone which has been built-up by colonial and sediment binding organisms such as algae, corals, bryozoans, and stromatoporoids. It combines what North American geologists would refer to as reefs, bioherms, and biostromes.

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Bibliography

Carozzi, A.V., 1961. Reef petrography in the Beaverhill Lake Formation, Upper Devonian, Swan Hills area, Alberta, Canada. *Journal of Sedimentary Petrology*, **31**: 497–513.

Carozzi, A.V. and Zadnik, V.E., 1959. Microfacies of Wabash reef, Wabash, Indiana. *Journal of Sedimentary Petrology*, **29**: 164–171.

Cumings, E.R., 1930, List of species from the New Corydon, Kokomo, and Kenneth Formations of Indiana, and from reefs in the Mississineaw and Liston Creek formations. *Proceedings of the Indiana Academy of Science*, **39**: 204–212.

Ladd, H.S., 1944, Reefs and other bioherms. National Research Council, Division of Geology and Geography, *Annual Report 4*, Appendix K, pp. 26–29.

Noe, S.U., 1996, Late-stage reef evolution of the Permian Reef Complex; shelf margin and outer-shelf development of the Tansill Formation (Late Permian), northern Guadalupe Mountains, New Mexico, USA. In Reitner, J., Neuweiler, F., and Gunkel, F. (eds.), *Global and Regional Controls on Biogenic Sedimentation; 1, Reef Evolution, Research Reports*. Goettinger Arbeiten zur Geologie und Palaeontologie. Sonderband. SB2, pp. 249–258.

Rehfeld, U., 1996, Mediating and limiting processes during the development of spongiolitic bioconstructions in Jurassic and Cretaceous strata; a paleontological, facial and geochemical analysis. In Reitner, J., Neuweiler, F., and Gunkel, F. (eds.), *Global and Regional Controls on Biogenic Sedimentation; 1, Reef Evolution, Research Reports*. Goettinger Arbeiten zur Geologie und Palaeontologie. Sonderband. SB2, pp. 249–258.

Rodriguez, S. and Sanchez, C.F., 1994. Rugose corals and calcareous algae bioconstructions in the Torreon section, Visean, Badajoz. *Coloquios de Paleontologia*, **6**: 61–75.

Wood, R.A., 1999, *Reef Evolution*. Oxford: Oxford University Press.

Cross-references

Atolls
Bioerosion
Bioherms and Biostromes
Coral Reefs
Reefs, Non-Coral
Tidal Environments

BIOENGINEERED SHORE PROTECTION

In an effort to arrest shore erosion at many coastal locations and to provide protection to marinas and harbors, it may be necessary to construct structures in high wave energy zones. Current practice involves utilization of structures constructed using large armor stones, concrete and steel walls, and a variety of other "hard" engineering techniques. Quite often, these structures do not add to the aesthetic and recreational attributes of a site and may impact significantly on the local environment. Integration of bioengineered components into the design of breakwaters and shore protection systems can be utilized, in certain cases, to enhance the project by providing better biological habitat and ancillary water quality improvement. Thus the goal of a project changes to include not only the stabilization of the eroding area or the provision of "quiet" waters, but to increase the quantity and quality of habitat available to fish and waterfowl communities, while providing an effective and aesthetic control of natural environment.

Background

In both the engineered and natural environment, the flow of water often causes erosion. The causes must be understood before the problem can be addressed. In the coastal zone, the flow of water results from wave action, the associated runup and backwash, wave breaking, alongshore currents, and the natural flow of water along side and overtop of the high-tide shoreline. In addition to the interaction of the high-tide shoreline or lakeward structure with water, a considerable amount of animal and human activity create additional stresses on the high-tide shoreline. Bioengineering methods of shore protection offer a practical solution that can also create an aesthetically pleasing and environmentally beneficial "buffer zone." Bioengineering, in this context, is the utilization of vegetation, either by itself or in combination with other defense mechanisms, depending upon the local environment. The other defense mechanisms may include the use of rock lining, offshore islands, wave screens, and submerged shoals that limit the wave energy reaching a site. Quite often, these defense structures can be designed to provide significant enhancement to the environment, particularly in providing suitable fish habitat for spawning, feeding, and hiding from predators.

The value of vegetation for protecting the soil depends on the combined effects of roots, stems, and foliage. Roots and rhizomes reinforce the soil. Immersed foliage elements absorb and dissipate energy and may cause sufficient interference with the flow to prevent scour. In a sediment-laden environment, they may also promote deposition.

A coastline requiring protection can be considered as two separate areas and thus habitat enhancement can be geared toward two communities; the high energy nearshore environment and the onshore environment, which can be suitably modified to ensure low wave energy levels. Enhancement of the nearshore zone can include construction of rock revetments as reef habitat, inclusion of submerged offshore structures to reduce wave energy levels reaching the shore and primary wave defense structures which provide habitat enhancement potential by the nature of their design (Figure B50). Selection of stone and design of its placement is developed in a manner to provide a reef like habitat beyond minimum stone placement required for the minimal shore

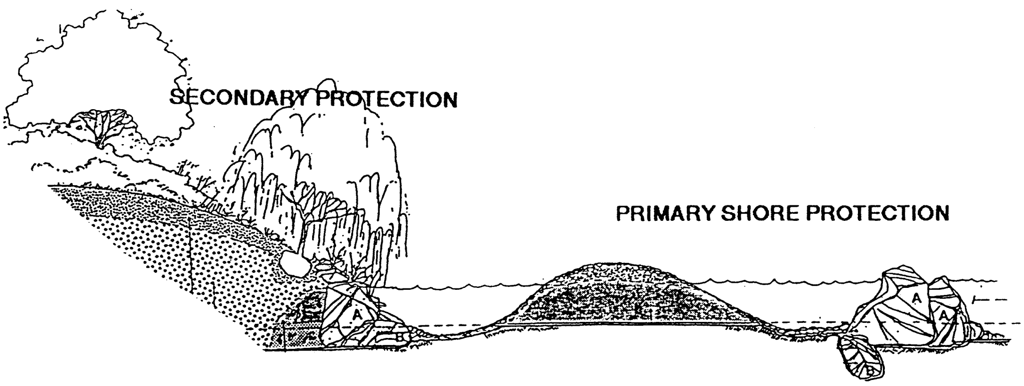

Figure B50 Utilization of shoals.

stabilization. The effectiveness of both natural and artificial reef like habitats as fish community habitats has been well documented. Proper design and installation of rock will provide protective cover and feeding areas and will supply the needs for small aquatic and benthic organisms by providing protection from the high energy wave action and from larger predators.

Low wave energy areas can be created behind a primary defense such as an offshore rock structure or wave screen, or through the creation of lagoons behind stable control structures (Figure B51). The development of constructed wetland pockets and areas for other shallow water plants can occur in these lagoons. This can be promoted by the establishment of "biological" riprap in the form of brush and woody plant debris. These materials will provide a setting that will foster the accumulation of shore plants from wind blown seed banks. As the brush decomposes, it provides a limited release of nutrients to the developing plant community and is eventually replaced by living plants. The establishment of new habitat will provide the opportunity for colonization by wetland plant and animal species that require quiescent waters. Transient use of the habitats by a variety of aquatic and migratory waterfowl is an additional potential for these environments.

Goals and objectives

The designer is encouraged to consult specialists in the fields of coastal hydraulics, fisheries, geomorphology, biology, landscape architecture, or any field that could make the project a success. The design of bioengineered breakwaters and shore protection that functions environmentally requires a multidisciplined approach. Usually, no individual has all the expertise required to ensure successful implementation.

The following geomorphologic, hydraulic, and biological changes may occur as a result of modification of the shore, which would occur from the creation of a marina or harbor, or from local erosion protection schemes:

- Loss or elimination of aquatic vegetation
- Loss or elimination of backshore vegetation
- Removal of specific nearshore bathymetrical features
- Modified substrate conditions
- Modified hydrodynamic, flow, sediment, and water quality regimes
- Changes in nutrient conditions and reductions in food organisms
- Aesthetic degradation
- Reductions in habitat diversity and environmental stability
- Increased water temperatures.

The shore is a dynamic system where impacts are difficult to predict. Engineered structures, when properly designed and constructed, can provide both species and habitat diversity and thereby mitigate potential adverse changes. However, the goals and objectives of the shore protection design must be correctly identified early in the design process. The designer must be aware of the design goals and objectives to correctly identify, size, and locate the various functional elements within the system. Biodiversity within and adjacent to the shore is interrelated with the quality in updrift and downdrift areas. Changes to any one of these components may adversely impact on others.

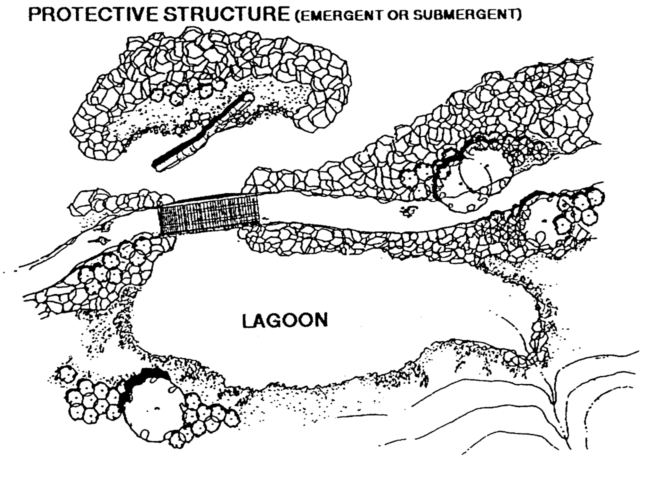

Figure B51 Development of a lagoon cell.

Habitat requirements

Aquatic life generally requires a habit that contains the following:

1. Sufficient water depth and volume for each life stage.
2. Adequate water quality with preferred ranges of temperature, dissolved oxygen, PH, etc.
3. A variety of continuous hydrodynamic conditions varying from deep water to shallow water for breeding and cover. Also flow conditions that sort bed load materials to provide a good environment for bottom dwelling organisms are advantageous.
4. Adequate cover to provide shade, concealment, and orientation.
5. Adequate food to maintain metabolic processes, growth, and reproduction.

Shore improvements should be designed for the individual fish species. Specific requirements for reproduction, juvenile rearing, and adult rearing with regard to feeding location, concealment from predators and competitors, and sanctuary from flow extremes and ice formation varies between species. Loss of the natural bathymetric features, which are utilized by particular species as a result of implementation of shore protection, could eliminate many of the requirements necessary to sustain significant biodiversity along the nearshore area. In addition, removal of existing shore vegetation, in either the emergent of submergent zones would significantly reduce or eliminate the potential to sustain a fish population.

Utilizing vegetation

In certain low wave energy environments, vegetation may be used by itself to provide suitable protection to an eroding shore. Reeds and

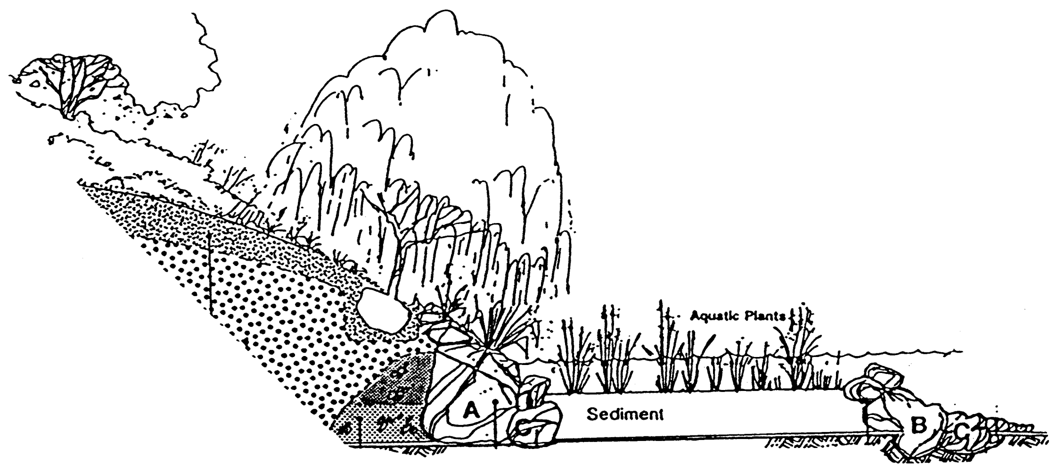

Figure B52 Emergent vegetation used in conjunction with stone.

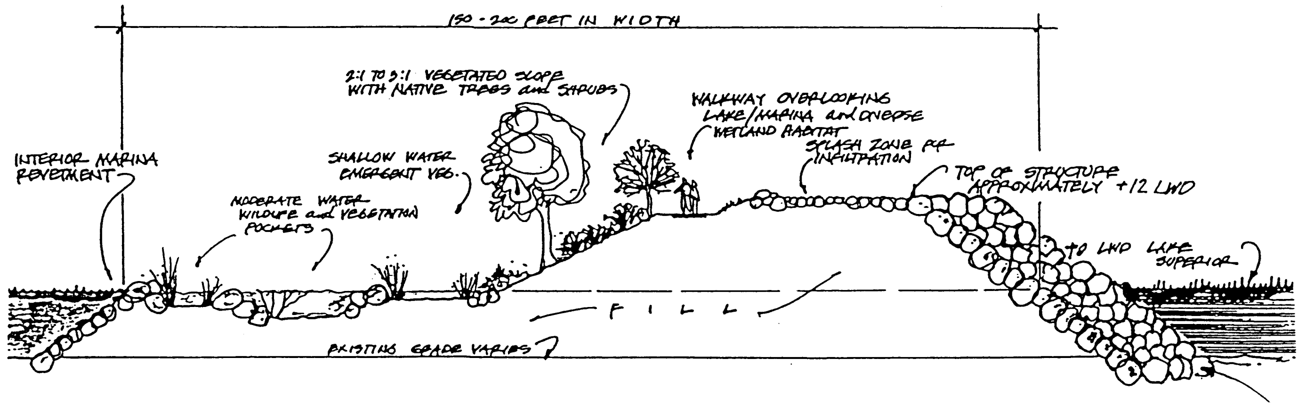

Figure B53 Control structure/lagoon.

other marginal plants can form an effective buffer zone by absorbing wave energy and restricting the alongshore flow velocity adjacent to the shore. They therefore have a protective value. Specific functions that they can perform include:

1. Absorbing and dissipating wave-wash energy.
2. Interference and protection of the shoreline bank from the flow.
3. Reinforcement of the surface soil through the root mat and prevention of scour of the bank material.
4. Sediment accumulation brought about by the dense plant stems.

Marginal plants require very wet ground and generally will not survive in water that is more than 0.5 m deep for long period of time. They flourish in conditions of low flow velocity and their integrity is weakened by wave action in excess of 0.5–0.75 m. Different species offer different levels of protection with regard to wave energy dissipation. For incident wave conditions under 0.5 m, reed beds having a width of 2–2.5 m may dissipate 60–80% of the incoming wave energy. In areas with higher levels of wave energy, riprap and geotextiles may be used in conjunction with vegetation to provide effective bank protection (Figure B52). In areas of high incident wave energy, an area of low wave energy can be created behind a primary defense such as an offshore rock structure or wave screen, or through the creation of lagoons behind stable control structures (Figure B53), as described above.

Natural methods of protection generally have low capital cost in comparison with conventional engineering methods. However, they may well have higher recurrent cost due to regular inspection, trimming and cutting, and repair. In areas where a combination of conventional and bioengineered structures are required, recent experience at several sites on the Great Lakes has established that these techniques may cost 20–30% more than conventional techniques alone.

Possible disadvantages are that natural protection schemes take time to mature and to become fully effective. Depending on the type, natural protection may take several growing seasons to reach the desired standard of protection.

Bioengineering differs from other conventional forms of engineering in two key respects, which strongly influence the design approach:

1. Bioengineering involves considerable practical experience and judgment, as opposed to the application of quantitative design theory or rules.
2. Careful management is required not only in the establishment of vegetation, but also in its aftercare over the initial growing seasons.

Use of vegetation requires the following points to be considered

The principal plant groups that can be used are aquatic plants, grasses, shrubs, and trees. Selection is based on consideration of the different roles to be performed by the vegetation, taking into account the physical and chemical properties of the soil, the climatic conditions, and the soil/water regime under which the plant must survive. Vegetation establishment may take several growing seasons and is a seasonal activity that must be managed and maintained. The engineer must prepare and agree to specific management objectives and a management program with the owner/client. This is in order to ensure that the vegetation is maintained in a fit condition to perform its intended roles.

Zones and horizons of natural protection

With natural methods of protection, and particularly methods involving the use of live material, the effectiveness of different materials is strongly dependent on their location in relation both to the dominant external water level and to the subsoil soil/water regime. To achieve effective protection using natural materials, the designer will almost inevitably need to use different methods of protection in different zones and horizons of the shore (Coppin and Richards, 1989).

Use of reeds

The emergent and marginal types of aquatic plants, such as the common reed, bulrush, and great pond sedge, are frequently used for interference and protection purposes to form a protective margin along the shore at the waterline. They also encourage siltation by absorbing current flow energy, and thus reducing the sediment-carrying capacity of the flow. Reeds can be easily weakened by erosion and loosening of the soil around the rhizomes due to wave energy. It is therefore necessary to protect the zone containing roots from high-velocity flow or significant wave attack. Provided this is done, the stems and leaves will protect the shore bank above.

Uses of shrubs and trees

A limited range of trees are water-tolerant and can be used in bioengineering structures for bank protection in both the aquatic and damp zones. The willow, alder, and black poplar are the principal water-tolerant species. In particular, a dense root structure is able to provide some protection as well as substantial reinforcement effect to enhance the stability of the shore both above and below the mean water level. The willow and poplar are particularly useful for bioengineering because they can be propagated from cut limbs. The cut limbs can be placed such that secondary root growth develops and shoots sprout from dormant buds. Trees, which are not water-tolerant, do not have any major direct function in shore stabilization, although they may provide shade to control the growth of aquatic life as discussed earlier.

Use of grasses

Grass is used very extensively in bank protection in the zones above the high water level. Grass roots cannot tolerate prolonged submergence periods. A wide variety of grass species and mixtures therefore are appropriate to satisfy the functional, environmental, and management requirements for a protection scheme. The principal functions which grass fulfills are those of interference, protection, root reinforcement, and soil restraint. The surface root structure forms a composite soil/root mat, which enhances the erosion resistance of the bare subsoil,

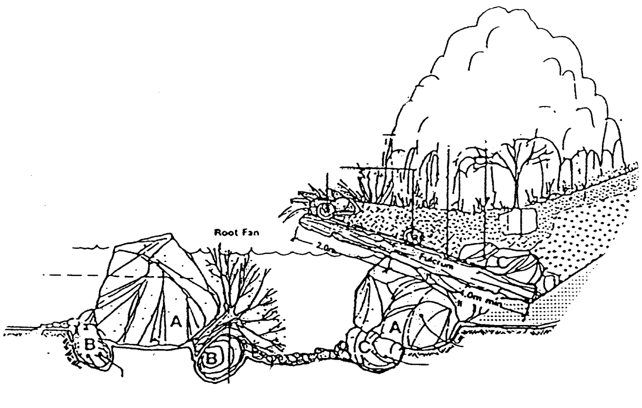

Figure B54 Timber (CANTILEVER) to provide habitat and shade.

and which is anchored into the subsoil by deeper roots. The engineering function of grass may be augmented by the use of geotextile or cellular concrete reinforcement to form composite protection. With both types of reinforcement, the visual effect of grass is retained. Erosion of grass cover by wave runup generally occurs by the scouring of soil from around the roots of a plant, thereby weakening its anchorage until the plant itself is removed by the drag of the flowing water. The effectiveness of grass protection can also be seriously reduced by any localized patches of bare soil or poor grass cover.

The rate of growth of different grasses varies considerably. Complete grass cover should normally be achieved by the middle of the first growing season while full protective strength of the sward is reached during the second season. Provision should be made for aftercare including mowing, fertilizing, and weed control.

Use of timber and woody material

A variety of timber and other dead woody materials can be used in the shore protection scheme usually fulfilling reinforcement, protection, and sometimes drainage functions (see Figure B54). Natural hardwoods will retain their integrity for 5–10 years if built into the bottom of a bank below the water level. Out of the water they can last longer but the worst environment for timber is the alternately wet and dry zone around mean water level.

Monitoring

As part of the project design for the shore stabilization enhancements, a monitoring program is required. The purpose of the monitoring program is to measure the success and applicability of the enhancement methods to other shore projects.

Baseline habitat conditions should be assessed by observation and characterization of existing conditions. A plant survey and macroinvertebrate sampling of the nearshore benthic environment and a terrestrial plant survey should be performed to document existing plant and animal populations. Incidental observations of birds should be made as part of fieldwork. Sampling of nearshore fish populations should be coordinated with local regulatory agencies. Post-construction monitoring of the establishment of biological communities should be completed to evaluate the success of a particular scheme.

Conclusions

Shore protection enhancements similar to those described in this entry have been successfully implemented at numerous sites on the Great Lakes, most notably in Canada at Red Rock Marina, Lake Superior; Thunder Bay Harbor, Lake Superior; Kingston, Lake Ontario; and various reaches of the St. Lawrence Seaway, and at Bender Park, Lake Michigan; Silver Bay, Lake Superior; and in Louisiana (Gulf of Mexico) in the United States. The range of design wave conditions range from 0.75 to 4.5 m at these various sites. Many other projects are in the process of implementation.

Utilization of bioengineered shore protection, in concert with virtually transparent offshore protection (submerged breakwaters, wave screens, etc.) can provide for significant levels of protection while maintaining the natural beauty of an area, and, in most circumstances, providing significant opportunities for habitat enhancement and increased biodiversity.

Further suggested reading may be found below.

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Bibliography

Coppin, N.J. and Richards, I.G., 1989. *The Use of Vegetation in Civil Engineering*. CIRIA / London: Butterworths.
Hemphill, R.W. and Bramley, M.E., 1989. *Protection of River Canal Banks. CIRIA* / London: Butterworths.
Fisheries and Aquatic Biota, 1986. Volume 3F, Environmental Reference Book. Toronto: Ontario Ministry of Transportation.
Stream Enhancement Guide, 1980. Victoria: Canada Fisheries and Oceans and Province of British Columbia.

Cross-references

Beach Erosion
Coastal Zone Management
Geotextile Applications
Monitoring, Coastal Ecology
Shore Protection Structures
Vegetated Coasts
Wetlands Restoration

BIOEROSION

In his study of the erosion of steep cliffs around Huntington Sound in Bermuda by excavating sponges, Neumann (1966) defined the term bioerosion as the removal of consolidated material or lithic substrate by direct action of organisms. Soon after the term, bioerosion, was introduced, geologists and biologists described many different types of bioeroding organisms including algae, bacteria, foraminifera, sponges, bryozoa, annelid worms, barnacles, gastropods, bivalves, echinoderms, fish, and mammals. The process of bioerosion was also reported from many different marine and non-marine environments ranging from mountain slopes to the tops of deep sea knolls, and from rocky intertidal zones and coral reefs to the flanks of continental shelves. Bioerosion has also been reported from climatic zones extending from tropical and subtropical to the subarctic and arctic. Several different types of experiments have been devised for studying the rates of bioerosion by different types of organisms and in different environments. Although bioerosion was first recorded for living sponges in the intertidal zone, evidence for bioerosion was also found in the ancient rocks extending back at least to the Silurian.

Bioeroding organisms

Several different types of microbial borers have been described from modern environments and ancient rocks. Microbial borers including cyanobacteria and chlorophytes were found in modern reef environments at depths between 0 and 230 m, and boring hetertrophs are present between 100 and 300 m (Vogel *et al*., 1996). Evidence for boring algae (cyanophyta) has been preserved in Silurian bivalves and may be responsible for the silicification of their shells (Liljedahl, 1986). Twenty species of foraminifera, ranging age from Jurassic (Callovian) to Recent, are known to make cavities in hard substrates (Venec, 1996). The bioeroding foraminifera were found in turbulent, warm, shallow-water environments.

A wide variety of living and fossil invertebrates have been identified as bioeroders. Several species of boring sponges have been reported from reef areas in Bermuda (Neumann, 1966) and Grand Cayman Island in the British West Indies (Acker and Risk, 1985), and on a deep sea knoll at depths of 1,600 to 1,800 m (Boerboom, 1996). The bioeroding mollusks include chitons, gastropods, and bivalves. *Chiton pelliserpentis* removed hardened mudrock during feeding at Mudrock Bay in Kaikoura, New Zealand (Horn, 1984). The spawn of the gastropod, Nerita, settled on the sea bottom and eroded carbonate rocks at Cathedral Point in Costa Rica (Fischer, 1980). The bivalve genus *Lithophaga* was an active chemical borer in reefs from the Carboniferous through the Eocene (Krumm, 1992). Rock-boring echinoids excavated large cavities in reefs in the South Florida keys (Kues and Siemers, 1974) and on Enewetak Atoll in the Marshall Islands (Russo, 1980). The rock-boring barnacle, *Lithotyra*, eroded the rock face while grazing in the intertidal zone (Ahr and Stanton, 1973). The polychaete annelid, *Eunice*, burrowed into the carbonate rocks along the shore of the Gulf of California.

Vertebrates including fish and mammals play an important role in bioerosion on reefs and mountain slopes. Parrotfish have been observed feeding on coral in reef environments and their rates of bioerosion were measured (Frydl and Stearn, 1978). Recolonization experiments on coral reef communities near Aquaba on the Red Sea demonstrated that herbivorous fish were a major factor in structuring coral reef communities (van Treeck *et al*., 1996). In the Pyrenees of Spain, the indirect effect of digging by small mammals was considered more significant than the direct detachment of soil cover (Martines and Pardo, 1990).

Rates of bioerosion

Several different field experiments have been used to estimate the rates of bioerosion by different organisms and in different environments. On the carbonate coastline of Bermuda, experiments show that the sponge *Cliona lampa* is capable of removing 6–7 kg of material from 1 sq. m of carbonate substrata in 100 days, corresponding to an erosion rate of calcarenite of more than 1 cm per year (Neumann, 1966). In Kaikoura, New Zealand, *Chiton pelliserpentis* removed mudrock from the surface at a rate of 47.3 g/sq. m on the high shore and 173 g/sq. m on the low shore (Horn, 1984). This was equivalent to about 2% of total on the high shore and 5.5% on the low shore. On Moorea reef barrier flat in French Polynesia, bioerosion rates for echinoids was estimated at 4.5 kg/sq. m per year and for scarid fish at 1.7 kg/sq. m per year (Peyrot *et al*., 1996).

Environments of bioerosion

Although most examples of bioerosion have been studied from tropical reefs and intertidal zones, bioerosion also has been reported from high-latitude environments, the outer continental shelf and deep-sea knolls. Algae borings were found in gastropod shells and echinoderm tests in the high-latitude, low-energy environments in the firths of Clyde and Lorne, Scotland (Akpan and Farrow, 1985). Boring sponges were dredged up from Newfoundland from depths of approximately 1,600–1,800 m on top of Orphan Knoll, 550 km northeast of Saint John's (Boerboom, 1996). Evidence for bioerosion was also found in the clastic sediments on the outer continental shelf around the Hudson Canyon off the eastern coast of the United States (Twichell *et al*., 1984). A workshop on bioerosion convened by Bromley (1999) has reviewed several different aspects of bioerosion ranging from the style of bioerosion in Late Jurassic reefs to the role of bioerosion in carbonate budgets in Indo-Pacific reefs.

Conclusions

Bioerosion by microorganisms, invertebrates, and vertebrates is widespread throughout many different carbonate environments from the early Paleozoic to the Recent. Bioeroding organisms have been reported from mountain slopes to deep-sea knolls, from the rocky intertidal zone to coral reefs and from the tropics to the arctic circle. The rates of bioerosion vary from a few grams per square meter to several kilograms per square meter depending on the organisms involved and the depositional environments.

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Bibliography

Acker, K.L. and Risk, M.J., 1985. Substrate destruction and sediment production the boring sponge Cliona caribbaea on Grand Cayman Island. *Journal of Sedimentary Petrology*, **55**: 701–711.

Ahr, W, M. and Stanton, R.J. Jr., 1973. The sedimentologic and paleoecologic significance of Lithotrya, a rock-boring barnacle. *Journal of Sedimentary Petrology*, **43**: 20–23.

Akpan, E. and Farrow, G., 1985. Shell bioerosion in high-latitude low-energy environments; firths of Clyde and Lorne, Scotland. *Marine Geology*, **67**: 139–150.

Bromley, R.G., 1999. Bioerosion; a collection of workshop papers. *Bulletin of the Geological Society of Denmark Marine Geology*, **44**: 129–180.

Boerboom, C.M., 1996. Bioerosion and micritization in the deep sea; a look at the coral *Desmophyllum cristagalli*. Bachelor's thesis, McMaster University. Hamilton, ON.

Fischer, R., 1980. Bioerosion by spawn of *Nerita funiculate* and *Nerita scabricosta*, Gastrapoda. *Neues Jahrbuch fuer Geologie und Palaeontologie*, Monatshefte, **5**: 287–292.

Frydl, P. and Stearn, C.W., 1978. Rate of bioerosion by parrotfish in Barbados reef environments. *Journal of Sedimentary Petrology*, **48**: 1149–1157.

Horn, P.L., 1984. Beachrock erosion due to feeding by Chiton (*Sypharochiton pelliserpentis*) at Mudstone Bay, Kaikoura, New Zealand. *Mauri Ora*, **11**: 57–61.

Krumm, D.K., 1992. The paleoecology of the chemically boring bivalve genus Lithophaga. *Annual Report*—Western Society of Malacologists, 24, p. 16.

Neumann, A.C., 1966. Observations on coastal erosion in Bermuda and measurements of the boring rate of the sponge, *Cliona lampa*. *Limnology and Oceanography*, **11**: 92–108.

Kues, B.S., and Siemers, C.T., 1974. The geobiological significance of the modern echinoid Echinometra as an agent of bioerosion, southern Florida keys. *Abstracts with Programs—Geological Society of America*, **6**: 834.

Liljedahl, L., 1986. Endolithic microorganisms and silicification of a bivalve fauna from the Silurian of Gotland. *Lethaia*, **19**: 267–278.

Martines, R., and Pardo, A.M.P, 1990. First data on bioerosion due to small Mammalia in Spanish Pyrenees. *Ekologiya*, **1**: 27.

Neumann, A.C., 1966. Observations on coastal erosion in Bermuda and measurements of the boring rate of the sponge, *Cliona lampa*. *Limnology and Oceanography*, **11**: 92–108.

Peyrot, C.M., Le-Campion, A.T., Harmelin, V.M, Romano, J.C., Chazottes, V., Pari, N., and Le-Campion, J.,1996. Bioerosion in the carbonate cycle; quantification of the process in French Polynesia. Memoires de la Societe 36. at Mudstone Bay, Kaikoura, New Zealand. *Mauri Ora*, **11**: 57–61.

Russo, A., 1980. Bioerosion by two rock boring echinoids (*Echinometra mathaei* and *Echinostrephus aciculatus*) on Enewetak Atoll, Marshall Islands. *Journal of Marine Research*, **38**: 99–110.

Twichell, D.C., Grimes, C.B., Jones R.S., and Able, K.W., 1984. The role of bioerosion in shaping the outer continental shelf around Hudson Canyon, offshore Eastern United States. *Abstracts with Programs—Geological Society of America*, p. 16.

van Treeck, P., Schuhmacher, H., and Paster, M., 1996. Grazing and bioerosion by herbivorous fishes; key processes structuring coral reef communities. *Goettinger Arbeiten zur Geologie und Palaeontologie*, Sonderband. SB2, pp. 133–137.

Venec, P.M.T., 1996. Bioeroding foraminifera; a review. *Marine Micropaleontology*, **28**: 19–30.

Vogel, K., Kiene, W., Gektidis, M., and Radtke, 1996. Scientific results from investigations of microbial borers and bioerosion in reef environments. *Goettinger Arbeiten zur Geologie und Palaeontologie*, Sonderband. SB2, pp. 139–143.

Cross-references

Atolls
Bioconstruction
Cliffs, Erosion Rates
Coral Reefs
Erosion Processes
Karst Coasts
Tidal Environments

BIOGENOUS COASTS—See VEGETATED COASTS

BIOGEOMORPHOLOGY

Biogeomorphology is a discipline that combines ecology and geomorphology. Geomorphology is the study of landforms and their formation. Ecology is the study of the relationships between biota and their environment. The environment is defined as factors that affect biota. These factors can be abiotic (physical, chemical), biotic (other organisms), or anthropogenic (humans). Abiotic geomorphological processes may affect biota and biota may in turn affect geomorphological processes. The interaction between both defines the discipline of biogeomorphology. *Biogeomorphology is the study of the interaction between geomorphological processes and biota.*

Essential concepts

The term biogeomorphology was first used in the 1980s (Viles, 1988), although earlier studies have been conducted that were focused on biogeomorphology without using this term. Biogeomorphology is studied in terrestrial as well as in aquatic systems. In coastal systems biogeomorphological interactions are clearly demonstrated in the shallow, productive waters, and in various sedimentary environments. Examples of biogeomorphological interrelationships include sand dune development, tidal flats, salt marshes, mangrove systems, and coral reefs.

Relevant geomorphological factors in coastal systems are bathymetry, bed composition (rock, gravel, sand, silt), and the transport of sediment. It also includes factors that drive morphological processes, such as water flow and wave energy. The biota involved in coastal biogeomorphology include plants and animals, ranging from very small (algae) to very large (whales).

The geomorphological influence on biota is in its most direct form the influence on habitats (living environments) of flora and fauna. The coastal morphology and geomorphological processes define the gradients between high and low, between wet and dry, and between sedimentation and erosion. These gradients and the processes that cause them are determinative for gradients in grain size of the sediment, nutrient levels, organic matter levels, and moisture. Plants and animals are tuned to specific conditions and will therefore be abundant in specific locations.

The biological influence on geomorphological processes is the influence of biota to create, maintain, or transform their own geomorphological surroundings. This is demonstrated by the influence of vegetation on the hydraulic resistance, erodability and sedimentation, or by the influence of fauna on sediment characteristics through bioturbation and biostabilization.

In some cases morphological processes are dominant over biological processes and therefore the biota have to adjust to their environment. In other cases biological processes are dominant. The most interesting are those cases where there is a mutual interaction that leads to feedback coupling of processes. When looking for these cases, it is important to examine the temporal and spatial scales of the mutually interacting processes. Biogeomorphological interrelationships can be found in several coastal environments, for both hard and soft substrates.

Biogeomorphology for hard substrates

On rocky shores and coral reefs a typical community of organisms thrives that affects the erosion rates of its substrate. Influenced by abiotic factors such as wave energy, splash water, inundation frequency and -period, depth, desiccation and substrate type, a clear zonation can be found of various cyanobacteria, (macro-)algae, fungi, lichens, molluscs, sponges, worms, sea urchins, fish, etc. Some of these organisms dwell on the surface of the substrate, while others live within the substrate. Their effect on erosion of the substrate is divided in "biological corrosion," processes that modify the substrate but provides no erosion product, and "biological abrasion" (see *Bioerosion*), processes that do generate an erosion product. Grazing, burrowing and boring on or in the substrate carries out biological abrasion, and is most significantly found in coral reef systems.

Biogeomorphology for soft substrates

In soft coastal systems, the interrelationships between geomorphological factors and biota can mainly be noticed for benthic fauna and flora. The presence of benthic species is affected by hydraulic and morphologic conditions, such as depth, current velocity, salinity, and grain size. The effect of soft substrate communities on geomorphology is divided into biostabilization and biodestabilization. Biostabilization leads to an increase in soil resistance, preventing erosion, while biodestabilization leads to an increased erodability.

Biostabilization by plants

On tidal flats, small algae (diatoms) are capable of affecting the geomorphology. These diatoms can form extensive algal mats and excrete EPS mucus, which is a sticky substance made of polysaccharides that glues the sediment together and therefore protects the sediment against erosion. Sea grass is dependent on clear water, it needs sunlight to grow. A sea grass meadow slows down the current velocity near the bed and therefore sand and silt will not resuspend in the water, which otherwise would lead to turbid water. Furthermore, their root system binds the substrate. Ultimately, deposition of suspended sediment is encouraged in a sea grass meadow, which leads to the supply of organic material with nutrients, needed for growth.

Seaweeds are also capable of adjusting their physical environment by damping down wave energy; and salt marshes also play an important role in stabilizing sediments. Salt marsh vegetation makes fine sediment settle down resulting in a continuous heightening of the marsh. The higher the marsh gets, the more vegetation can grow and the better the marsh is protected against erosion. Other stabilizing effects result from cementation of beachrock by cyanobacteria and stromatolite formation by algae.

Biostabilization by animals

Some macrozoobenthos can actively catch sediment particles from the water column and bring them to the bed. The presence of a mussel bank, for example, will alter the bed in different ways. Mussels slow down the water flow and they protect the bed against erosion. Mussels also actively catch small particles from the water column by filterfeeding and subsequently excrete these as pseudofeces. This results in a change in the soil composition to finer sediments.

Animal tube fields are also believed to stabilize the sediment, because there is a clear accumulation of fine particles and organic matter between the tubes. The tube itself may affect small-scale turbulence and therefore have a stabilizing effect, however, a great deal may be attributed to the community of microorganisms between the tubes that excrete mucus. Other stabilizing effects result from large banks of dead shells and mucus binding by meio- and macrofauna.

Biodestabilization

Benthic fauna may destabilize the substrate by their digging and feeding activities (bioturbation). The constant mixing and recycling of sediment in the top centimeters of the bed results in a characteristic vertical particle-size profile. The selective uptake and excretion of preferred particle sizes results in sorting and pelletizing sediments. Together with the digging of burrows and the constant movement within the substrate, these activities lead to the generation of a surface micro-relief that has a higher hydraulic roughness and is more prone to erosion. Furthermore, bioturbation also affects the sediment water content, porosity, and sediment cohesion.

Scale interactions in biogeomorphology

Different physical and biological processes can have dynamic interactions when they operate on the same spatial and temporal scales. Processes that act on a very small scale may appear as noise in the interactions with processes on larger scales. Their effect can be accounted for by proper averaging procedures (e.g., for turbulence). Processes that act on a large-scale may be treated as slowly varying or even constant boundary conditions when studying their effects on processes on smaller scales (e.g., sea-level rise due to climate change). Techniques for scale interactions are reasonably well established in geomorphology (De Vriend, 1991) and are based on scale linkage via sediment transport. In biology, however, population and community dynamics give rise to spatial and temporal structures that are not easily linked. In recent years, the importance of scale has been increasingly recognized (Legendre *et al*., 1997) as an essential aspect of understanding the biotic and abiotic processes that affect the biogeomorphology of coastal systems.

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Bibliography

De Vriend, H.J., 1991. Mathematical modelling and large-scale coastal behaviour, Part 1: Physical processes. *Journal of Hydraulic Research*, **29**(6): 727–740.

Legendre, P., Thrush, S.F., Cummings, V.J., Dayton, P.K., Grant, J., Hewitt, J.E., Hines, A.H., McArdle, B.H., Pridmore, R.D., Schneider, D.C., Turner, S.J., Whitlatch, R.B., and Wilkinson, M.R., 1997. Spatial structure of bivalves in a sand flat: Scale and generating processes. *Journal of Experimental Marine Biology and Ecology*, **216**: 99–128.

Viles, H.A. (ed.), 1988. *Biogeomorphology*. Oxford: Basil Blackwell Ltd.

Cross-references

Algal Rims
Beachrock
Bioconstruction
Bioerosion

Bioengineered Shore Protection
Bioherms and Biostromes
Coral Reefs
Reefs, Non-Coral
Rock Coast Processes
Shore Platforms
Vegetated Coasts

BIOHERMS AND BIOSTROMES

History

Originally coined by Cumings (1932), the word bioherm along with its brother term biostrome have been widely used in reef literature, but their proper stratigraphic definition is often misunderstood.

In the original meaning (Chevalier, 1961) a bioherm was defined as a mound or lens-shaped organic build-up, edified by the skeletons of various organisms and lying unconformably inside a stratigraphic series of different lithology. Conversely, a biostrome was a flat, layered reef structure, wide or narrow in shape and causing no stratigraphic disturbance inside its sedimentary environment.

Discussion

Both words "bioherm" and "biostrome" were obviously coined for fossil build-ups, whose stratigraphic position in the sedimentary sequence can be studied; and they were also commonly used for the description of living or subfossil structures, whether the sedimentary environment of the latter is accessible or not to study.

Definitions vary according to authors: In the Encyclopaedia Britannica a bioherm is defined as an "ancient organic reef of moundlike form built by a variety of marine invertebrates ... (and coralline algae). A structure built by similar organisms that is bedded but not moundlike is called a biostrome."

Many geologists, however, extend these definitions to gravity deposited mounds or layers of skeletal remains, such as shells or broken coral, including reworked or transported material, as illustrated by Roger Suthren in his on-line lectures in Sedimentology, a second year Geology module at Oxford Brookes University: "Bioherms: (are) mound or lens-shaped (biological build-ups). Some are in-place organic structures (reefs), others are banks of loose, transported carbonate sediment consisting largely of shells or skeletons. Biostromes: (are) laterally extensive beds, sheets or ribbons of carbonate material. Some have grown in-place (reefs); others consist of transported shells and skeletons."

For Battistini *et al.* (1975) a bioherm is a: "lens shaped organic reef ... embedded *in situ* inside sedimentary layers of different lithological nature ... it may be surrounded by a peripheral talus of biodetrital sediments," whereas a biostrome is a "layered, bank like organic reef of variable extension, creating no discontinuity inside the embedding sedimentary layers."

There is, therefore, no general agreement upon a complete definition taking into account at one and the same time such different characters as: age, stratigraphic conformity or unconformity, along with the autochtonous or allochtonous nature of deposited organisms.

Furthermore, many authors (notably among biologists and geographers) tend to use "bioherm" as a general term not only for major biological build-ups such as extensive algal rims or coral reefs (e.g., see Adey and Burke, 1976) but also for small-scale organic build-ups, for which the word "biostrome" would better fit. Bosence and Pedley, who had first used "bioherm" in a preliminary publication (1979) dealing with Miocene layers of calcareous algae in Malta, appropriately dropped it for "biostrome" in their final paper (1982).

It is, therefore, difficult for an actualist (whether geologist or not) to find criteria sufficiently precise and reliable to distinguish between the alternate notions of bioherm and biostrome. For example, an algal rim growing on the outer edge of a coral reef is indeed a bioherm, or a part of a bioherm since it takes an active part in the sedimentary processes of the latter, but the same kind of formation thinly coating a limestone or a volcanic shore, or on a vertical cliff, without altering sedimentation should be called a biostrome even if both formations are in continuity with one another.

Further difficulty lies in the fact that, for actualists, detrital accumulations of dead shells and broken skeletal material (generally mud-supported) are considered as something very different from a true build-up or reef, since the latter is fundamentally made of an *in situ* developed formation, resulting in boundstone or framestone lithologies *sensu* Bathurst (1971).

Conclusions

Unless bio-accumulated detrital mounds and layers are taken out of the definition of bioherms and biostromes (a revision that only geologists can decide), and the status of small-scale build-ups is settled, the use of the latter words should preferably be restricted to the stratigraphic study of the fossil formations for which they were first coined (their associated detrital facies, and other types of detrital formations being included or not). Students of living reefs are conversely encouraged to prefer more general terms (such as "biological build-up," "reef-like structure," or "biogenic construction") instead.

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Bibliography

Adey, W.H. and Burke, R.B., 1976. Holocene bioherms (algal ridges and bank barrier reefs) of the eastern Caribbean. *Geological Society of America Bulletin*, **87**: 95–109.

Bathurst, R.G.C., 1971. *Carbonate Sediments and their Diagenesis. Developments in Sedimentology*, **12**. Amsterdam: Elsevier.

Battistini, R., Bourrouilh, F., Chevalier, J.P., Coudray, J., Denizot, M., Faure, G., Fischer, J.C., Guilcher, A., Harmelin-Vivien, M., Jaubert, J., Laborel, J., Montaggioni, L., Masse, J.P., Mauge, L.A., Peyrot-Clausade, M., Pichon, M., Plante, R., Plaziat, J.C., Plessi, Y.B., Richard, G., Salvat, B., Thomassin, B.A., Vasseur, J., and Weydert, P., 1975. Eléments de terminologie récifale indo pacifique. *Téthys* **7**: 1–111.

Bosence, D.W. and Pedley, H.M., 1979. Palaeoecology of a Miocene coralline algal bioherm, Malta. *Bulletin du Centre de Recherches d'Exploration et de Production d'Elf-Aquitaine*, **3**: 463–470.

Bosence, D.W. and Pedley, H.M., 1982. Sedimentology and palaeoecology of a miocene coralline algal biostrome from the Maltese Islands. *Paleogeography, Paleoclimatology, Paleoecology*, **38**: 9–43.

Chevalier, J.P., 1961. *Recherches sur les Madréporaires et les formations récifales du Miocène de la Méditerranée occidentale*. Thèse, Paris, Société Géologique de France, vol. 2, 562 pp.

Cumings, 1932. Reefs or bioherms? *Geological Society of America Bulletin*, **43**: 331–352.

Cross-references

Algal Rims
Coral Reefs
Reefs, Non-Coral
Sea-Level Indicators, Biologic

BLACK AND CASPIAN SEAS, COASTAL ECOLOGY AND GEOMORPHOLOGY

Coastal zone of the Black Sea

The coasts of the Black Sea are rather uniform and slightly embayed. The Crimea is the only large peninsula protruding offshore. The wide opened bays facing the sea (Odesskii, Kalamitskii, Tendrovskii, Karkynitskii, Yarylgachskii, Burgasskii) as well as the above mentioned Crimean Peninsula are located in the northern part of the region. The southern, eastern, and western coasts are smooth and uniform with small bays. The total extent of the coastline exceeds 4,000 km (Figure B55).

Zenkovich (1958, 1959) contributed much to the study of the Black Sea coasts. In the two-volume monograph, he described coasts of the former Soviet Union and analyzed dynamics and morphology of certain regions. Diverse coastal areas were described by investigators from different countries (Nevesskii, 1967; Shuiskii, 1974; Simeonova, 1976; Kiknadze, 1977; Zenkovich and Schwartz, 1987; Shuiskii and Schwartz, 1988; Kaplin *et al.*, 1991, 1993). The American Society of Civil Engineers has recently published a collection of articles concerning the Black Sea coasts (Kos'yan, 1993).

The environmental problems of the coasts have been discussed in many publications. The most complete summaries were given in the monographs of Sapozhnikov (1992) and Kuksa (1994).

Large-scale investigations were carried out in the frame of the international INEP program "Black Sea Environmental Program." Due to these activities about 2,000 analytical maps of the Black Sea natural environment were compiled, among them the map of the main sources

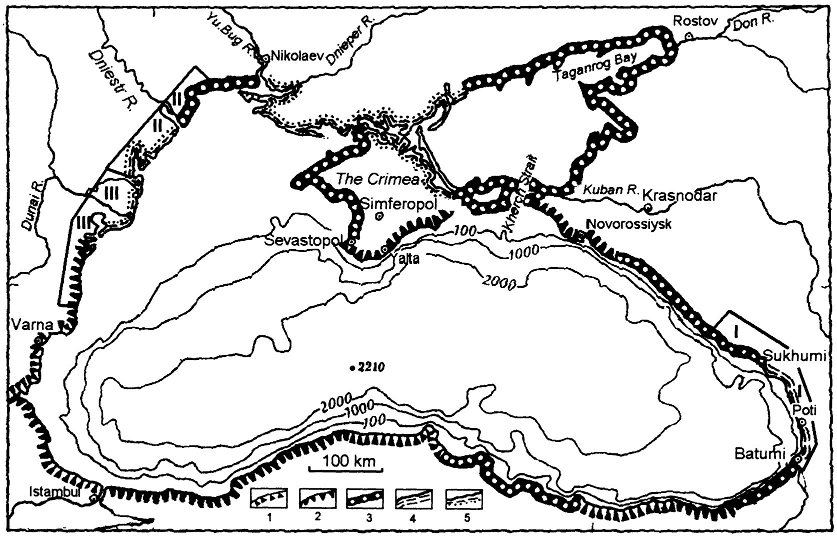

Figure B55 Types of coasts of the Black Sea and the Sea of Azov. 1, straight faulted; 2, erosional bight; 3, graded erosional and depositional; 4, graded depositional; 5, liman and lagoon, erosional and depositional. Key study areas are shown by numbers (I–III).

of pollution in the nearshore zone with subsequent entry to the geoecological information system (Berlyant *et al*., 1999). The Geographic Information System (GIS) was processed at the Geographical Faculty of the Moscow State University. The users of this GIS may receive not only maps, but also the tables with the data on the amount of pollutants and other information concerning the sources of pollution and natural reserves of the Black Sea. A compact-disc "Black Sea GIS" was published by INOPS/ENVP in 1998.

Environmental problems of the coast

Two main problems could be outlined among the environmental problems of the coasts: (1) influence of the rising sea level upon coastal processes and intensification of erosion related to it; (2) increasing anthropogenic impact.

Anthropogenic impact is mainly manifested by water pollution. Water contamination by pesticides leading to degradation of bottom vegetation was revealed in shallow bays (Kuksa, 1994). It is the result of disposal of freshwater from irrigation systems of Southern Ukraine. Water pollution caused a 3-fold decrease in the phytoplankton biomass in the nearshore zone and a 1.5-fold decrease in the zooplankton and zoobenthos biomass. Considerable pollution of the sea and especially its nearshore zone is determined by the influx of freshwater from the largest river of the region—the Danube. Its influence is noticed along the coasts of Ukraine, Romania, Bulgaria, and even Turkey. The Danube discharges enormous amount of oil-products, heavy metals, pesticides, and other pollutants. Pollutants are mainly accumulated in bottom sediments and biota. For instance, water plants of the Danube coast contain 0.007–0.020 mg/kg of mercury.

The concentration of pollutants discharged by the Danube decreases eastward (near Odessa and Sevastopol) and southward (in Romania and Bulgaria). Other rivers, the Dnieper, Inguri, Rioni, Chorokh and others, contribute much to the contamination of the nearshore waters.

Due to pollution of nearshore waters the role of biogenic sediments (mainly shells) in coastal dynamics decreases. At the end of the 1940s shelly sediments constituted 40–50% of coastal accumulative forms on the northwestern coast (Zenkovich, 1982), while in the 1980s its contribution was less than 10% (Shuiskii, 1974).

Another important ecological factor of anthropogenic origin is the influence of economic activity on the sediment budget in the coastal zone. Regulation of the rivers causes a sharp decrease in the solid river runoff and, hence, less sediments are supplied to beaches. Mass removal of sediments (sand, pebbles, gravel) directly from beaches, quarries, the nearshore zone, and river mouths considerably damaged the coastal zone. In the Caucasian coastal region this process started at the end of the last century when beach sediments were taken for construction of railroads. Mass sediment removal continued in the 1950s–1960s, when ports and other economic objects were built. During 1945–55, 100 million m^3 of beach pebbles were removed from the Tuapse-Adler coast (Kiknadze, 1977). As a result of this action, many beaches of the Caucasian coast became one-half smaller during two or three decades. This caused intensive coastal erosion. Of the 312-km-long Georgian coastline, 220 km were subjected to coastal erosion due to its retreat at a rate of 1–3 m/year. Active coastal erosion manifested by beach destruction was also recorded in the Crimea (Zenkovich, 1982).

During the last few decades many countries have been taking efforts to protect their shores. However, many hydrotechnical constructions such as seawalls, groins, breakwaters, and others have intensified an adverse effect of the sea on the coast. Construction of artificial beaches appeared to be the most effective method. During 1981–86 in Georgia, about 8 million m^3 of sediment was taken from subaerial quarries that facilitated creating artificial beaches with a total area of about 60 ha. As a result, a recreation zone was formed and the problem of shore protection in Georgia was practically solved (Kiknadze, 1977; Zenkovich and Schwartz, 1987). Creation of artificial beaches or additional sediment supply to existing natural ones was undertaken in other regions as well (Odessa, Crimea, Bulgaria).

Coastal geomorphology

In general, erosional coasts predominate along the Black Sea. Elevated mountainous coasts predominate in the eastern and southern parts of the Black Sea. This is a zone of young Alpine orogenesis. Graded and erosional accumulative coasts are typical of the western and northern parts of the sea. Geologically this zone is dominated by hard blocks protruding from the ancient Russian platform and remains of the Baikalian orogenesis. In the Eastern Black Sea erosional processes are especially active due to an extremely narrow continental shelf which sometimes nearly coincides with the coastline as in the Caucasus. Thus, the submarine slope has steep gradients allowing large storm waves to attack the coast.

Slopes of the Great Caucasian Ridge form the largest part of the Caucasian coast, since the axis of the ridge is subparallel to the coastline. This is the reason why cliffy coasts up to 200 m high prevail between Anapa and Sukhumi. The cliffs are cut in the steeply sloping flysch beds and its ridges are noticed in the submarine bench. In the southern part of the Caucasian coast, the Batumi region, foothills of the Little Caucasian Ridge reach the shore. The Colchis Lowland lies between the Great and Little Caucasian Ridges. It follows the large Alpine flexure. The lowland is swamped and its flanks are only slightly higher than the sea level. The lowland experiences a prolonged tectonic submergence. Many rivers flowing from the slopes of both ridges drain onto the Colchis Lowland. Despite this, sandy coasts do not migrate seaward. The heads of submarine canyons are located close to the mouths of the large rivers such as the Inguri, Rioni, Supsa, and others. The alluvial material is removed to the canyons instead of being accumulated on the beaches. Moreover, in many places the shores of the Colchis Lowland are eroded (up to 3 m/year).

The presence of large promontories near Adler, Pitsunda, Sukhumi, Burup-Talii are typical of the Caucasian coast. They are located near

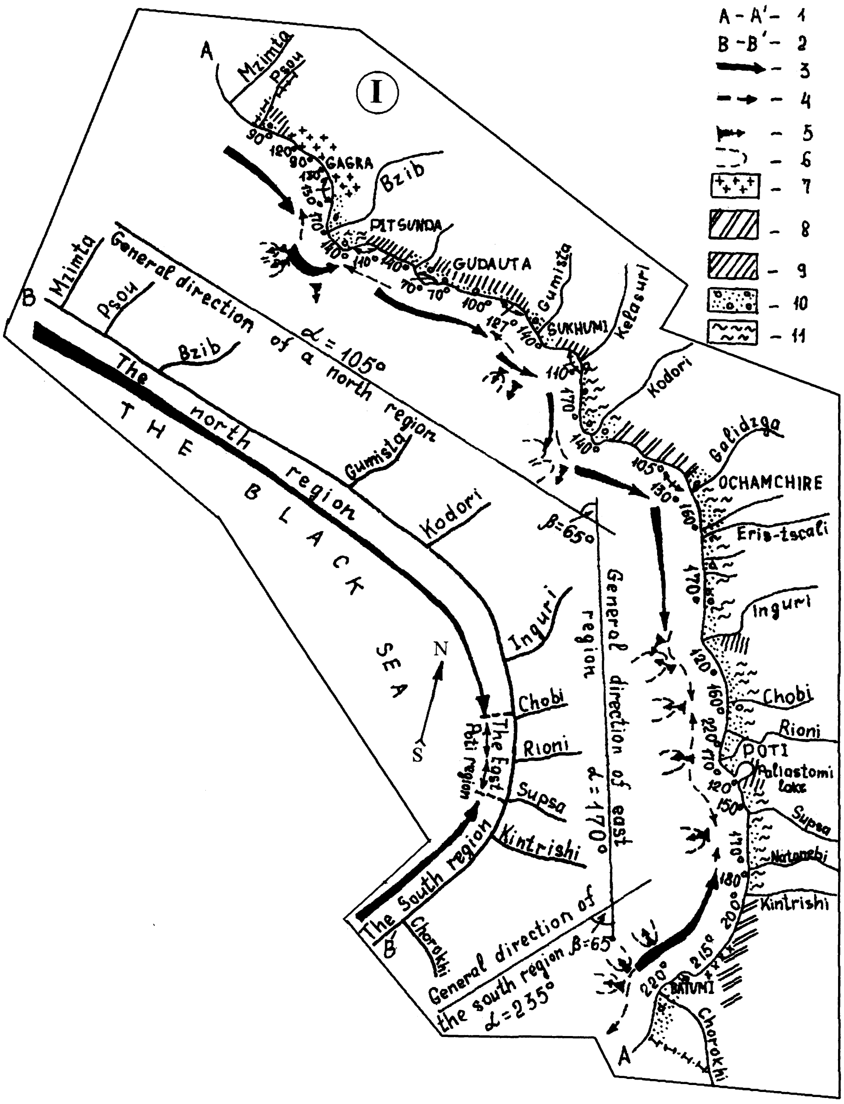

Figure B56 Schematic map of morpho- and litho-dynamics of the Black Sea coast in Georgia (after Kos'yan, 1993). 1, modern coastline; 2, coastline during the period of the drop in sea-level rise 6–5 ka; 3, longshore sediment streams, their direction and relative actual capacity; 4, direction of migration and transport of finer sediment; 5, partial loss of sediments at considerable depths; 6, canyon heads and steep falls; 7, cliffed rocks with erosional relief; 8, semi-cliffed rocks (conglomerates, marl, schists, etc.) with erosional relief; 9, related rocks (marine and lagoonal clays) with plain relief; 10, loose deposits (pebbles, gravel, sands of terraces, dunes and beaches); 11, bog and lacustrine deposits.

the river mouths and consist of the Holocene alluvium (Figure B56). These promontories protrude far offshore and overlie a significant portion of the continental shelf. No other large accumulative landforms are present on the Caucasian coast. The beaches are associated either with numerous river mouths or places of longshore sediment drift discharge. They are mainly composed of gravel and pebbles.

As shown above, accumulative forms are subjected to active erosion. Its intensification is caused by natural reasons: sea-level rise at a rate of 1–2 mm/year and decrease in river discharge due to regulation of rivers and removal of beach sediments. Heads of submarine canyons contribute to coastal dynamics since part of the material transported by alongshore drift is accumulated there. For example, the Akula submarine canyon near Pitsunda accumulates about 80 thousand m^3 of sediments per year (Kiknadze, 1977; Kos'yan, 1993).

Within the Georgian coastal zone alongshore drift is directed to the southeast (Figure B56). Each sediment stream represents a dynamic system with its own source of sediment supply and areas of sediment loss (submarine canyons and steep slopes) or final discharge. The capacity of alongshore sediment streams ranges from 3–15 to 150–220 thousand m^3/year. A small alongshore sediment stream is directed to the north from the Chorokh river mouth to the Colchis Lowland.

High erosional shores are typical of the mountainous coasts of the Crimea. They are subjected to active erosion since they are affected by severe winds (and waves) blowing from the southwest and southeast. Shore destruction is accelerated by landslides occurring in clays. Sometimes the landslides have an area of hundreds of square meters. For instance, the town of Alupka is located on six large landslides and its stability is conditioned by several factors. Of these are influence of underground and surface waters, abrasion, load of buildings and other construction.

Many shores of the Southern Crimea are formed by the slopes of ancient volcanoes (Karadag region) and tectonic faults. Outcrops of volcanic rocks and limestones form capes separated by shores represented by soft shales, clays, and sandstones. Ria-coasts occur near Sevastopol and Balaklava.

Beaches of the Southern Crimea are formed of pebbles, because finer sediments (more than 0.03) are transported down the steep submarine slopes. Removal of pebbles for building purposes caused the disappearance of beaches. However, some of them have been recently restored.

Many shores of the Southern Crimea are artificially protected. Dynamic interaction between different regions is weak due to the absence of large rivers supplying sufficient amounts of alluvial sediments to the coastal zone. Thus, local shore protection is successful and has no negative influence on adjacent coasts. Different, usually complex, engineering structures are used that protect coasts from both landslides and abrasion. Of these are embankments with seawalls, traverses, breakwaters, groins with artificial sediment filling between them, etc. (Zenkovich, 1982; Kos'yan, 1993).

Steep coasts of the Southern Black Sea are formed by densely forested northern slopes of the high Eastern and Western Pontus mountains stretching subparallel to the coastline. The mountains gradually lower westward and near the Bosphorus Strait their height does not exceed 300 m. Erosional and denudation coasts with steep rocky cliffs are widespread in Turkey. Only in separated small bays do the sandy-pebbly "pocket" beaches occur. Areas of sediment accumulation are associated with mouths of such large rivers as the Kizil-Irmak, Sakarja, and Eshil-Irmak. These rivers form rather large deltas prograding far offshore and nearly reaching the edge of the narrow continental shelf. Violent storms produced by severe northwesterly winds deflect the pathways of alluvium to the east thus forming flanked barriers (Kos'yan, 1993).

The largest curves of the Turkish coast correspond to the lowland peninsulas of Bafri and Djiva, related to river deltas and the mountainous Injeburun Peninsula.

Western and northwestern coasts of the Black Sea are rather low with hilly plains of different origin (alluvial, marine, and alluvial–marine) facing the sea. The delta of the Danube, the largest river of Western Europe, is located here. It has a complicated structure. Besides common channel bars there are a series of cheniers (local name "grindu") marking the stages of delta progradation. The river mainly discharges through its northern Kiliiskii channel. Thus, the southern part of the delta is smaller and is being slightly eroded. Active utilization of the Danube water for irrigation by five countries reinforces erosion of the southern part of the delta. There are numerous water reservoirs in the delta: lakes-limans (northern part), complexes of lakes and lagoons (southern part), lakes (inner part). From the north the delta is bounded by the Budzhak plateau, and from the southwest by the lake–lagoon complex of Rozelm-Synop. Abundance of warm water and high fertility of soils favor plant and animal life.

Coasts to the northeast from the delta are represented by plains and low plateaus. The only exception is the anticline of the Tarkhankut Peninsula. Its steep slopes are mainly composed of easily eroded loesses and clays. The rate of erosion ranges from 7 to 20 m/year (Shuiskii, 1974).

These erosional coasts alternate with lagoons and limans. Limans represent the lower parts of river valleys that have been flooded during the Holocene transgression. Most of them are separated from the sea by sandy–shelly accumulative forms (spits or baymouth barriers). Specific environmental conditions exist in limans since their waters are warmer and less salty. As a result, productivity of waters is higher.

A considerable part of the coast is subjected to landslides. Both landslides and coastal abrasion destroy valuable territories of the Ukrainian steppes. At present, the accumulative forms in the mouths of limans are eroding. They are composed of sand layers overlying lagoonal clays, thus giving evidence for migration of the accumulative forms toward the lagoons (Shuiskii and Schwartz, 1988).

Two opposite longshore drifts exist in the region stretching to the northeast from the Danube River to Odessa (Figure B57).

Jagged coasts are characteristic of the region to the east from Odessa including the western Crimea. Adjacent lowlands experience relative submergence at a maximal rate of 30 cm/ 100 years as recorded in the inner part of the Karkinit Bay. Large accumulative forms are the most interesting elements of the coastal relief, that is, the Kinburn spit and the system of the Tendra-Dyarylgach spits related to it.

The Kinburn spit and Odessa shoal (to the west of it) originated in the place of the Dnieper and South Bug deltas junction. Under the

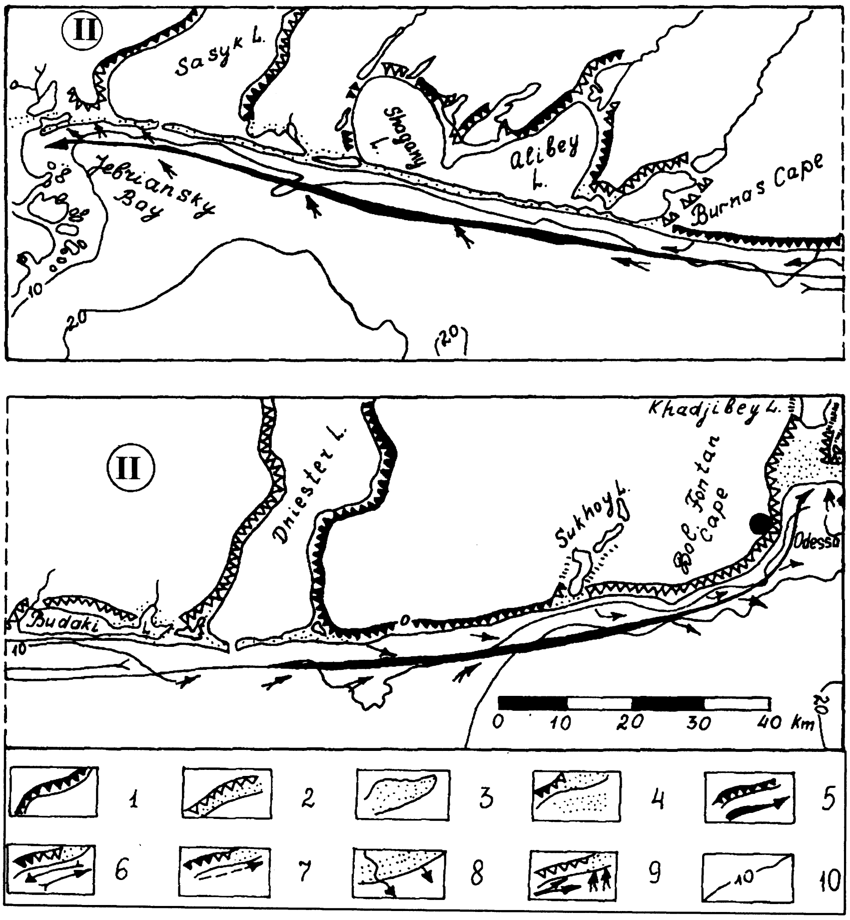

Figure B57 Geomorphological map of the northwestern Black Sea coast between the Danube River delta and Odessa (after Zenkovich, 1958). 1, active erosional scarps; 2, passive erosional scarps; 3, emerged coastal accretion bodies and coastal ridges; 4, emerged coastal accretion bodies and coastal ridges; 5, longshore sediment streams (thickness of arrows proportional to the capacity of a stream); 6, longshore sediment drift; 7, prevailing sediment drift; 8, offshore sediment drift; 9, onshore sediment drift; 10, depths in meters.

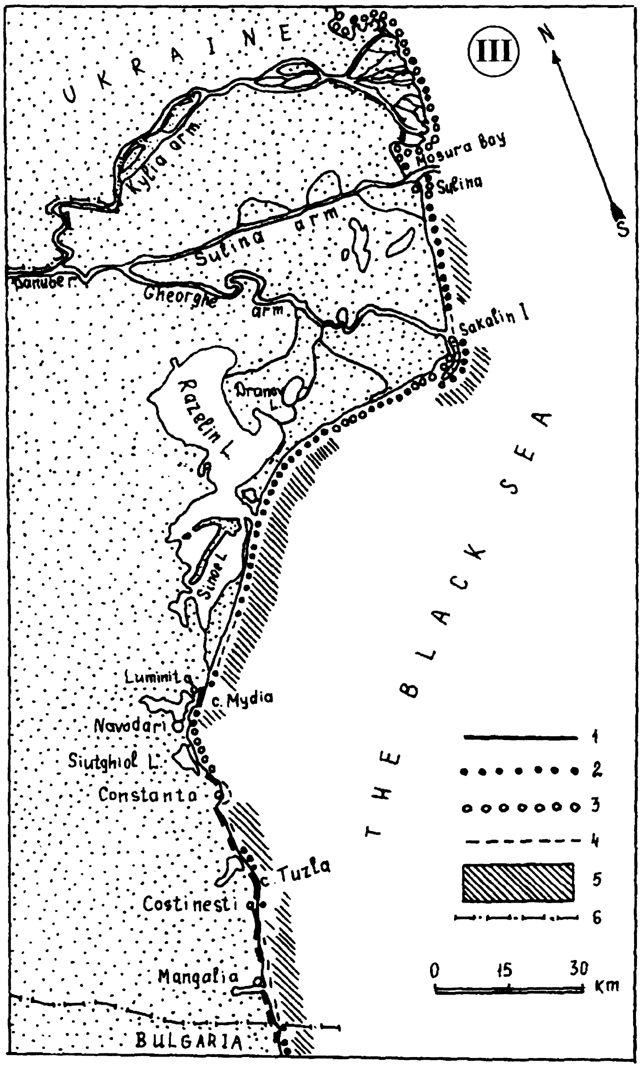

Figure B58 Morphology of the Romanian coast of the Black Sea (after Kos'yan, 1993). 1, active cliffs; 2, retreating coastline of accumulative coasts; 3, prograding coastline of accumulative coasts; 4, stable coastlines; 5, coastal sections with predominance of erosional processes in nearshore zone; 6, state frontier.

Holocene sea-level rise deltaic sediments were reworked and a system of subaerial and submarine sand bars was formed. Dunes and salt lakes located on the Kinburn spit are parallel to the coastline.

The Tendra and Yarylgach spits represent a joined accumulative form that continues to grow. However, landward migration also takes place. As a result, the central part of the accumulative form became attached to the continental shore, and its distal end formed two separate spits.

The eastern part of the Crimean Peninsula, together with the Taman Peninsula, form a single coastal region divided into two parts by the Kerch' Strait. The territory is covered by limans and lagoons associated with the ancient and modern delta of the Kuban' River. However, erosional shores are dominant. The coastline represents a series of arcs where clays of Maikopian age intercalate with solid rocks of Neogene age that form headlands.

Along the Kerch'–Taman' coast relics of the ancient accumulative forms and lagoonal silts were reported that allowed for reconstructing the Holocene history of the coastal area.

The western coast of the Black Sea lies in Romania and Bulgaria. The Romanian coast is subdivided into two parts. Its western part corresponds to the Danube delta that equals 78% of the delta surface. As mentioned above this part of the delta is now eroding at the rate of up to 7 m/year (Kos'yan, 1993).

Southward from the delta the coast is graded. It is composed of loesses, clays, and limestones of Neogene and Pleistocene age. A considerable part of the southern Romanian coast (51 km of the total 101 km) is abraded and consists of active cliffs 2–40 m high. The rate of abrasion averages 1–2 m/year, sometimes reaching 7 m/year. Maximal rates are characteristic for cliffs composed of loesses and clays. Capes are usually formed of Neogene limestones. Submarine benches are typical of the Romanian erosional coasts. Their width sometimes reach 1,100 m (Figure B58).

Coastal accumulative forms are represented by sandy beaches resting against cliffs and, near the river mouths, and by barrier forms separating lakes and lagoons. Sintghiol is the largest sandy barrier. Sedimentary material is supplied to the sea by the Danube delta and active cliffs.

The Romanian coast is actively used for recreation. To protect the coast from destruction certain efforts have been taken: construction of seawalls, cobble filling, etc.

The coasts of Bulgaria are mainly erosional. In southern Bulgaria erosional forms are restricted to the small bays of the zone of Alpine orogenesis. Cliffs of eight different types are distinguished in this area: from 15 to 20 m high cliffs with even surfaces composed of uniform loess and clayey deposits to high cliffs (up to 60–90 m) with uneven surfaces and a series of landslide steps on the slopes. Such cliffs are widespread in the region between Kavarna and Balchik and Kranevo-Zlaty Pyaski. Earthquakes facilitate landslides thus considerably accelerating retreat of the coast. The average rates of erosion vary from 0.005 to 1 m/year. Maximal rates reach 30 m/year. The estimated amount of material released due to abrasion of cliffs is 1,344,100 m^3/year.

Material produced by coastal abrasion and alluvium forms the beaches that occupy 28% of the Bulgarian coast (Simeonova, 1976). Some of the beaches, like that at Varna, prograde at a rate of 0.75 m/year. Similar process operate near the mouth of the Kamchiya River and in the region of the popular resorts of Albena and Zlaty Pyaski. However, most of the accumulative coasts retreat. For instance, between Cherny Nos Peninsula and the Albena resort the rate of retreat is 0.12–0.63 m/year. Generally the rate of retreat grows in the northward direction. Erosion often results from the negative influence of human activities and underestimation of the role of coastal processes. To protect coasts from erosion Bulgarian engineers fill up tetrapods with stones and construct dams separating the bays from the sea and straighten the coastline.

The coasts are also protected by groins (often short and without any filling between them), seawalls and other less effective structures. The most effective means, such as creation of artificial beaches, are not implemented in Bulgaria.

Sea-level oscillations played an important role in the recent evolution of the Black Sea coasts. From the available data it follows that during the 20th century sea level was steadily rising at the average rate of 2.1 ± 1.3 mm/year (Nikonov *et al.*, 1997). This estimation is based on the data collected at 70 points on the Russian, Ukrainian, Georgian, and Bulgarian coasts. The values exceed the average rate of the global sea-level rise, probably due to tectonic submergence of the Black Sea depression. Different estimations of the sea-level oscillations could be definitely attributed to different tectonic movements. The highest rates of sea-level rise were recorded in the Colchis Lowland, while the lowest ones were in the northeastern Black Sea.

Thus, it might be concluded that submergence of the coasts has been the main trend in their recent evolution. This process leads to abrasion of the cliffed coasts and erosion of the accumulative ones. However, under predicted conditions of more rapid sea-level rise, erosion will be intensified and many of the unique accumulative forms will be destroyed. First of all this affects the accumulative forms of the limans in the northwestern coastal area (Tendra, Binburn, Yagyrlach spits). Their destruction may have a severe impact on the ecology of limans that are the zones of extremely high bioproductivity. Sea-level rise will accelerate destruction of the Holocene accumulative forms on the Georgian and Turkish coasts. In this connection, all countries of the region must plan enhancing protective activities and carry out a long-term policy of coastal management.

Coastal zone of the Caspian Sea

The coasts of the Caspian Sea display a great variety of natural environments being located in different landscape zones. Recently, the problems associated with the rapid rise of its sea level have generated particular interest, especially in the context of the expected accelerated rise of the global sea level (Dolotov and Kaplin, 1996).

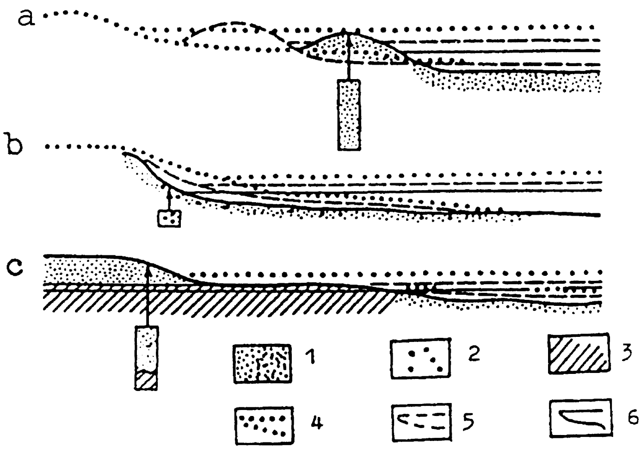

Figure B59 Processes of relief formation and sediment accumulation in shallow-water neashore zone under sea-level fall (after Dolotov, 1996). a, continuous sediment accumulation; b, erosion of external edges of accumulative coasts; c, preservation of sandy accumulative body; 1, sand; 2, pebble; 3, bedrock; 4, 5 and 6, successive positions of sea level.

The history of investigations on the Caspian Sea coasts was discussed in detail in the monograph of Leont'ev and Khalilov (1965). Further generalization was given in the monograph written by Leont'ev with co-authors (Leont'ev *et al.*, 1977). Present environmental problems of this area were outlined in the monographs of Kuksa (1994) and Zonn (1999).

Environmental problems of the Caspian Sea coastal zone arise from active economic development of not only the Caspian Sea itself, but its drainage area and adjacent territories (Kaplin and Ignatov, 1997). This region is distinguished by repeated sea-level changes, both seasonal and multi-annual. That is why the Caspian Sea is a natural laboratory for studying evolution of coasts under different sea-level oscillations.

In the modern historical period a rapid sea-level fall (of nearly 1.7 m) occurred from 1929 until the early 1940s. Dynamic changes in the coastal zone mainly depended upon the rate of sea-level fall (or decrease in depth in the nearshore zone) and the amount of sediments in the coastal area (Dolotov, 1961).

On the shallow-water sand coasts that are typical of the Caspian Sea the relief-forming processes are controlled by sediment budget, gradient of the coast, and configuration of the coastline under sea-level fall going on at different rates. Three types of the coasts with different patterns of relief changes have been identified.

Continuous accumulation of sediments and progradation of coasts (Figure B59(a)) takes place in case sufficient amounts of sediments are supplied to the coastal zone from adjacent land and shores (positive sediment budget over prolonged time period). If a positive sediment budget is replaced by a negative one progradation of the coast changes to landward retreat of the coastline (Figure B59(b)) (irregular sediment supply). Under insufficient sediment supply erosion is replaced by preservation of sandy accumulative bodies that occurs when direct wave attacks over the former coastline have ceased (Figure B59(c)).

Under sea-level fall the evolution of the Caspian Sea coasts went on in the following manner: continuous accumulation of sediments, emergence of the seafloor, and continuous seaward advance of the coastline. The area of coasts enlarged, and economic activity occupied new territories where settlements, roads, oil- and gas-pipelines, and resorts were constructed (Dolotov, 1996). Taking into consideration the predicted future sea-level fall the Soviet Government decided to create a vast recreation zone in the coastal regions of Dagestan and Azerbaidzhan (Molchanova, 1989). The Caspian coast offers several advantages over the Caucasian coast of the Black Sea. Sandy beaches are wider and longer here, solar radiation is more active and the number of sunny days in summer is higher (Veliev *et al.*, 1987). Part of these constructions has already been built.

In 1978, an unexpected and sharp sea-level rise occurred. The average rate of sea-level rise was 14–15 cm/year, but in some years it was as high as 30 cm/year and even more. The direct influence of the sea-level rise was flooding of coastal lowlands and acceleration of coastal erosion. The indirect impact included the rise of groundwater, swamping of the coastal area, salinization of soils, and expansion of surge areas. Environmental conditions of both the nearshore shallow zone and adjacent land sharply changed. Sea-level rise favored erosional processes and general landward migration of the coastline.

At the same time, contrary to the general trend of sea-level rise, in some patches of the western coast accumulation of sediments went on and the coastline migrated seaward. This happens, when sediment input exceeds the amount of unconsolidated sediments flooded by the advancing sea. These coastal regions are of special interest since they give a unique opportunity for future economic development (primarily recreation) even under the ongoing sea-level rise.

Generally, the character of relief changes under sea-level rise depends upon the rate of this rise, relief-forming environmental processes (hydrodynamics), and sediment balance. Coastal morphology determines the character and rate of the natural catastrophic processes together with their impact on economy and population (Dolotov, 1996).

Coastal geomorphology

Based on differences in relief, the Caspian Sea coast is subdivided into four regions (Leont'ev *et al.*, 1977).

The western coast receives about 50% of the total solid river runoff, while the material produced by abrasion is considerably less abundant. As a result of a positive sediment balance coastal erosion is suppressed. Under dominant northwesterly and southeasterly winds longshore drift streams are formed that smooth the coast and create accumulative forms.

River runoff is practically absent in the eastern regions. Active erosion in the recent past has not produced sufficient amounts of sedimentary material, since the cliffs are mainly composed of clays and limestones. Biogenic and chemogenic sedimentation went on slowly. As a whole, the coast is more embayed than in the west. In places where the coastline remains relatively straight over long distances, prevailing waves that are transversal to the coast form large accumulative barriers.

The northern coast is distinguished by extreme shallowness. Southeasterly and northerly winds predominate here. Abundant fine-grained alluvial material supplied by rivers is transported in the form of suspension. Winds and on- and offshore currents form the coasts characterized by frequent and significant displacement of the coastline.

Southern coasts are represented (Voropaev *et al.*, 1998) by coastal lowlands ranging in width from 1 (in the central part) to 60 km (near the large deltas of the Sefidrud and Gyurgyan Rivers). More than 40 small rivers discharge into the Caspian Sea in this region. During the past 50 years, solid river runoff was considerably reduced due to construction of reservoirs on all the large rivers. Granulometric composition of the beaches changes from gravel–pebble (western Mazenderan) to sand (Gilidzhan, central Mazenderan), and silt (eastern Mazenderan).

Several types of accumulative coasts occur in the Caspian Sea: lagoonal shores, coasts with terraces and other accumulative forms, erosional and deltaic shores, mudflats, coastal lowlands formed by onshore winds and waves (Figure B60).

Deltaic coasts occupy considerable parts of the coast—these are the deltas of the Volga and the Ural in the north, Terek and Sulak in the northwest, and Kura in the southwest. Even, quite recently, these rivers discharged considerable amounts of sediments into the sea thus supporting progradation of deltas. This process intensified during sea-level falls. In the northern Caspian Sea, where the Volga and Ural deltas are located, coasts have retreated seaward by dozens and hundreds of kilometers since 1929. In the late 1950s and 1960s large-scale hydroengineering projects were launched. This caused dramatic decrease in river discharge of such rivers as the Kura, Terek, and Sulak. The rise of the Caspian Sea level caused erosion of deltas. Coastal lowlands of the Northern Caspian region became inundated, and the previously accumulated coastal landforms were destroyed.

Large portions of the Caspian Sea coasts are flat. They were formed in course of regression that took place from the 1930s to the 1970s. Mudflats occur in the northern Caspian region, around Kirov Bay in southern Azerbaidzhan, in the region surrounding Krasnovodsk Bay and to the south from it. During regression aggradation of the northwestern coasts of the sea proceeded at a rate of 60–100 m/year, in the Kizlyar Bay it was faster and reached 150–200 m/year, and northward of it-even 700–800 m/year. In the Kirov Bay wind-induced mudflats were up to 1.5 km wide (Kaplin, 1997). The sea-level fall resulted in considerable advance of land in the eastern coastal regions. For instance, southward from the Cheleken Peninsula the average rate of land advance during the period from 1929 to 1957 was 34–36 m/year (Leont'ev *et al.*, 1977). Naturally, shallow bays of the northern coast were dried up.

The change from regressive to transgressive regime has dramatically affected the drained shores with mudflats. Gentle gradients (close to 0.0001) of submarine slopes caused passive flooding of the coasts.

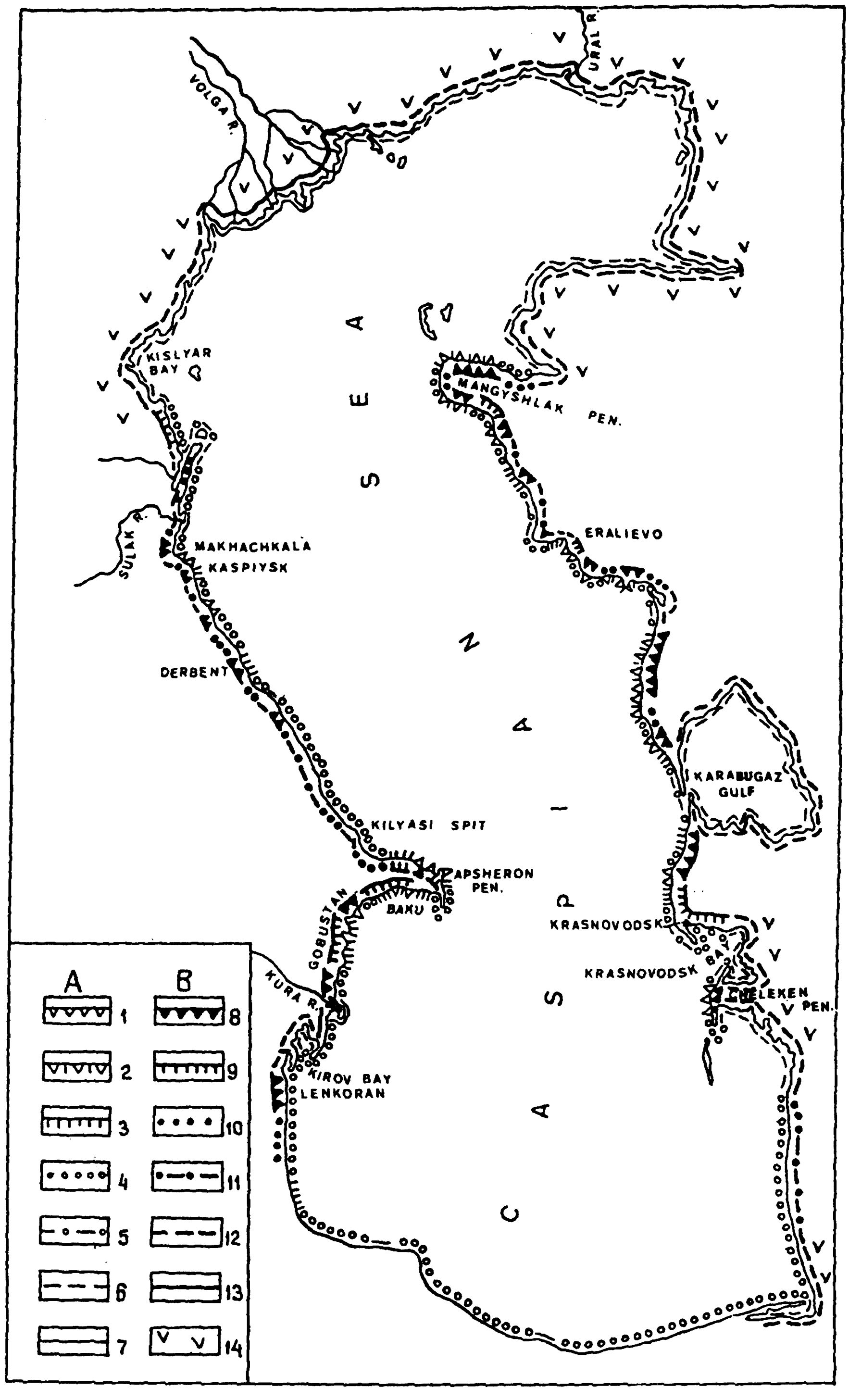

Figure B60 Types of the Caspian Sea coasts (after Ignatov *et al.*, 1993).
A, regressive stage (prior to 1977). 1, erosional shore; 2, erosional shore with passive cliff; 3, erosional–accumulative shore; 4, progradingbeach; 5, accumulative lagoonal shore; 6, mudflats formed by wind-induced surges; 7, deltaic coast. B, transgressive stage (after 1978). 8, erosional shore; 9, erosional–accumulative shore; 10, prograding beach; 11, accumulative lagoonal shore; 12, mud flats; 13, deltaic coast; 14, areas affected by transgressive flooding.

Sea-level rise resulted not only in submergence of land, but rise of the groundwater table and, hence, salinization of groundwater and swamping of the adjacent lowlands. The Kirov Bay was filled with water again. In the vicinity of the Kilyazinskaya spit (northern Azerbaidzhan), a flat coastal terrace that formed in 1940 due to sea-level fall has been partly flooded and swamped. Considerable parts of coastal lowlands in the northern near Caspian region on both sides of the Volga River have been flooded. Accumulation of sediments went on in such bays as the Komsomolets, Kaidak, Mertvyi Kultuk, Bol'shoi Sor in the Bugaz Peninsula, and in depressions between Baer knolls that are widespread in the northern and northeastern Caspian lowlands. Wind-induced surges up to 1.5–2 m high caused passive flooding of this vast area. These phenomena resulted from a longshore drift and caused episodic abrasion of the coasts and erosion of the seafloor, producing furrows and ridge-and-runnel erosional forms. Wind-induced currents re-suspended sands and silts previously accumulated on the seafloor. Suspended material was removed by currents and precipitates near the shoreline to form saltings (mudflats).

Sea-level rise changes accumulative coasts as well. As mentioned above, accumulation that took place during the regressive stage affected nearly the entire coastline, but the pattern was somewhat different. As sea level was falling, cliffs became passive, longshore drift ceased, and spits were eroded, especially at their base. We can exemplify the Kura and Agrakhan spits on the western shore, the accumulative form of Cape Rakushechnyi, Kenderli, and Krasnovodsk spits on the eastern one. With the drop in sea level, the middle and lower parts of submarine slope became eroded. The produced clastic material is gradually transported upward the slope and accumulated near the shore. Intensive landward transportation of sediments inhibited the longshore drift.

Due to the sea-level rise former cliffs became active again and abrasion intensified. Transgression changed the profile of submarine slopes, especially their upper parts. These changes are accompanied by erosion of the frontal part of accumulative forms or creation of bars near the water edge that later turn into beach barriers separating the sea from lagoons. Many lagoonal shores were formed in Dagestan and northern Azerbaidzhan, where gradients of the submarine slope equal 0.005. The present sea-level rise is responsible for accumulation of beach barriers separating lagoons. The beach barriers are overlapping lagoons thus giving an impression that the coastline retreats. However, lagoons keep expanding despite the landward migration of beach barriers since flooding of drained territories and the rise of the groundwater table favor further expansion and deepening of lagoons.

The coasts of Dagestan and Northern Azerbaidzhan, with steeper gradients (up to 0.01), have shore-attached bars formed of coquina. These bars have an asymmetrical profile giving evidence of their landward migration towards young terraces behind them. No lagoons are formed because the coasts are steep and lie above sea level.

Finally, slopes with gradients exceeding 0.01 are subjected to active erosion of both Holocene and recent accumulative forms. This trend leads to significant landward retreat of the coastline (Kaplin, 1997). Evolution of such coasts follows the well known “Bruun’s rule” (Bruun, 1962).

The transgression has also affected erosional coasts. The latter are typical of the Eastern (Mangyshlak Peninsula, regions northward from the Kara-Bogaz-Gol Bay, Cheleken Peninsula) and, partially, Western (Dagestan, Lenkoran’, Apsheron Peninsula) Caspian Sea. The share of erosional shores on the Dagestan coast increased from 10% to 40%, on the Azerbaidzhan coast—from 20% to 55%, Kazakhstan—from 8% to 13%, and Turkmenistan—from 7% to 22%. Until recently, some places on the Dagestan coast have been protected from erosion by offshore submarine ridges composed of limestone. Benches and ridges have been flooded by the transgression, during storms (especially surges) waves reach the cliffs. If the cliffs used to be protected by accumulative terraces, clastic material has been actively reworked; part of it being transported into the longshore drift and the rest removed down the submarine slope. In general, the above-water parts of the slopes were more actively abraded by sea waves, and the coastline rapidly migrated landward. In some parts of the Dagestan and Lenkoran’, the rate of the process reached 20–25 m/year.

Therefore, the recent rise of the Caspian Sea level has significantly modified the dynamics of all identified coastal types. Evolution of the coasts subjected to relative submergence depends upon the gradient of the submarine slope (Kaplin, 1997). Different ways in evolution of accumulative coasts under the sea-level rise (Figure B61) have been discussed in several publications on the Caspian Sea (Kaplin, 1989, 1990; Ignatov *et al*., 1992, 1993). The coastal dynamics of the Caspian Sea display a certain discrepancy between transgressive and regressive regimes on the one hand, and cycles of the coastal evolution on the other hand. A regressive regime usually corresponds to the cycle of accumulation. However, erosion of coastal accumulative forms started in the 1960s when sea level was still falling. Certainly, one of the reasons why erosion became active is economic activity, such as construction of barrages and irrigation networks that reduced the solid river runoff and led to the deficiency of sediments in the coastal zone. On the other hand, it is natural that an erosional cycle succeeds the accumulative one. The reason is that both a drop in the sea level and its rise under transgression are associated with reformation of the submarine slope and with its erosional cutback. Yet, during regression the zone of the shore slope erosion shifts seaward, not landward, involving parts of the outer slope where benthonic material is of a finer grain size. With time the deposits of fractions that can be transported up the slope and that can built the accretion forms are depleted. Although the rates of the Caspian coastal processes are much higher than of those in the world ocean, these processes are essentially similar. Therefore, research into the Caspian Sea coastal dynamics has importance beyond its regional significance. It may be of great use for simulating the formative laws applying to the coastal dynamics of the world ocean, particularly as the ocean level is rising and is likely to keep rising in the future as a result of global warming (Kaplin, 1997).

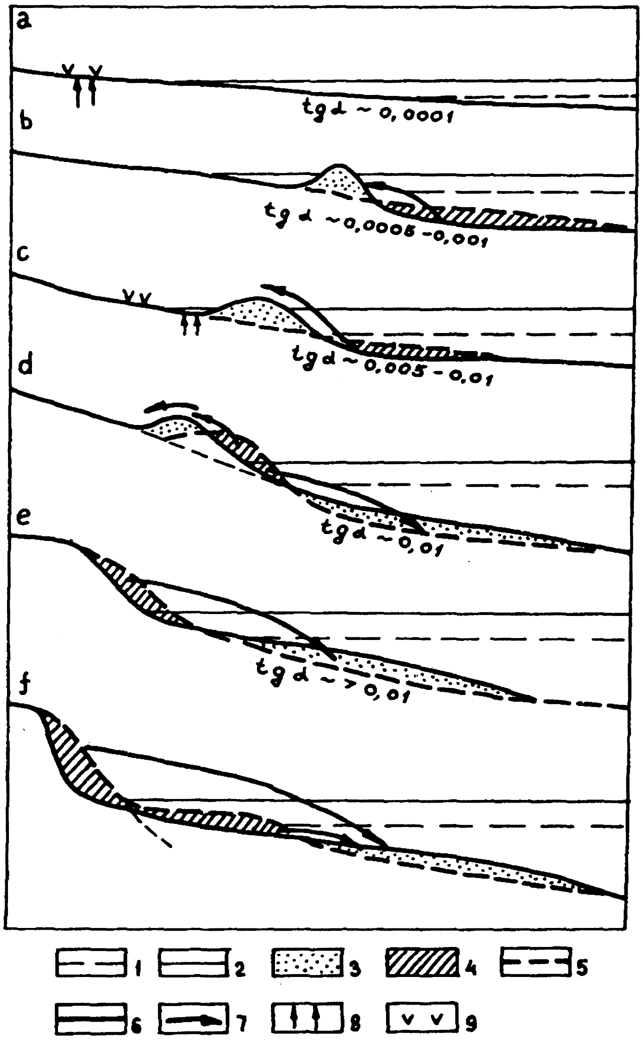

Figure B61 Schematic representation of the Caspian Sea transgressive coasts in function of the offshore gradient (after Kaplin, 1989).
1, regressive sea level; 2, transgressive sea level; 3, sediment accretions; 4, erosion lens; 5, former profile of coastal zone; 6, present-day profile; 7, sediment drift; 8, groundwater rises; 9, bogs.

Environmental problems of the coasts

The present environmental conditions of the Caspian Sea are determined by the effect of rising sea level and increasing anthropogenic impact (Kaplin, 1997).

Considerable sea-level rise has already adversely affected the economy of the coastal states of the Caspian region including the Russian Federation that occupies its northwestern part—the Dagestan, Kalmykia, and Astrakhan’ region. Many industrial and habitable buildings, recreational structures and other objects are now in the zone of flooding or in the zone of subsoil water penetration (Dolotov, 1996).

The rising sea level poses threats as follows: flooding and underflooding of coastal areas earlier occupied by communication facilities, livestock farms, grazing lands, fish hatcheries, piers, fish spawning grounds, wildlife and nesting bird habitats. It also prevents cattle from accessing fertile pastures. Incursions of seawaters into areas occupied by human settlements or farms generally lacking treatment facilities resulted in capture and retransportation of technogenic products, municipal effluents and wastes toward the Caspian Sea increasing the supply of pollutants. Flooding causes malfunctioning of irrigation channels and transverse drains in the irrigated areas. Intensive shore erosion has caused losses of considerable land areas.

Special environmental problem is the underflooding of running and suspended wells in the oil and gas fields and subsequent propagation of oil products. Since oil production has or had been carried out over many decades it is evident that significant amounts of oil have been accumulating in the soil, finding their way into, and heavily polluting the sea. The oil film formed over large areas, dramatically reducing water evaporation and affecting the Caspian water balance (Kaplin, 1997).

Starting up of the first turn of the Astrakhan' gas-condensate complex together with exploitation of oil and gas-deposits present a potential threat to existence of unique ecosystem of the river mouths in the Northern Caspian region (Kuksa, 1994). Development of shelf oil and gas fields by the Caspian states strongly affects natural processes by pollution of water and bottom sediments, destruction of plant cover, and considerable reduction of fish resources (Kurbatova, 1994).

In course of the Caspian sea-level rise some of the productive oil and gas fields on the coast will be flooded. The flooded area will also include prospective sites for deep exploratory well-drilling.

The major sources of pollution of the Caspian Sea water consist of: river (surface) runoff; untreated effluents discharged from enterprises, farms, or human settlements in coastal areas or in the river mouths; navigation accidents and technical processes in industries operating directly in the Caspian water area; surface washout during surges; and flooding of producing oil and gas fields, industrial sites, agricultural lands, and human settlements.

River runoff is the largest source of pollutants to the Caspian Sea accounting for 90% or even more of the total influx of pollutant. It is attributed to the fact that the Volga, Ural, Kura, and Terek rivers receive polluted effluents from various industrial facilities and farms along their entire courses. Concentration of various pollutants in river water at river mouths exceeds the maximum permissible levels, frequently by a very wide margin (up to 10-fold or more).

It should be noted that pollutants transport with river runoff is a rather constant process, varying only slightly in different years. The most important consequences are as follows: large amounts of pollutants carried to the sea by rivers under the influence of hydrometeorological factors (wind, currents, waves) penetrate the entire water column and bottom sediments, that later act as the sources of secondary pollution of seawater. Self-purification processes are unable to neutralize the water continuously impacted by chemical inputs. Environmental conditions in river deltas are seriously threatened, mainly in deltas and river mouth beaches. The latter represent the most valuable natural water complexes that act as the principal fish spawning and foraging grounds as well as waterfowl nesting places and rest areas.

The second important source of pollutants transported to the sea are effluents discharged from enterprises, farms, or human settlements situated directly on the coast. An extremely adverse effect on the marine environment is produced by sudden discharge of pollutants resulting from accidents at enterprises of treatment facilities as well as various failures of sewage systems (Kaplin, 1997). All maritime cities, first of all, Astrakhan', Baku, Makhachkala, Turkmenbashi, are sources of pollution that drop to the sea sewage waters (Zonn, 1999).

The third major source of seawater contamination are accidents (oil and oil products spills) occurring during navigation and exploitation of offshore oil- and gas-fields, as well as, owing to the sea-level rise, flooded coastal oil fields. Accidental spills of oil and oil products are a source of significant damage to marine ecosystems, because concentration of pollutants may be extremely high exceeding the permissible level by hundreds or by even thousands of times.

Biological resources and fish reserves are affected as early as at the stage of seismic exploration of oil and gas fields. Special damage was caused by blasting operations that were responsible for the large-scale mortality of sturgeons. During drilling operations on the continental shelf, a special threat is posed by sustained discharges of liquid and solid wastes that are associated with the drilling process. Environmental consequences in the areas of offshore oil- and gas-field development are observable 5–12 km away from the drilling site and are manifested as high levels of oil pollution of water, bottom sediments, aquatic and benthic fauna and flora, and reduced species diversity of benthic communities and degradation of their structure (Kaplin, 1997).

The sea receives water wastes from many sources. From 2 to 5 tons of heavy metals, 60,000–200,000 tons of petroleum products, and more than 5 million tons of organic pollutants are dumped into deltas every year. The sediments are, therefore, contaminated by heavy metals, especially lead and cadmium, and their concentrations exceed many times those of the natural background. The composition of organic matter within the limits of deltas and their flood-plains causes deterioration of the oxygen regime and leads to hydrogen sulfide pollution of the beaches.

The flooding and associated contamination of land initiates specific biogeochemical processes whereby anaerobic gas generation is stimulated. The waters then become contaminated by metal compounds, heavy hydrocarbons, carbon dioxide, and nitrogen compounds, as well as bituminous substances and aromatic (benzene) polycyclic hydrocarbons. Generation of hydrogen sulfide poses the greatest danger (Kaplin, 1995).

Transformations of ecosystems in all rivers of the Caspian Sea basin occurred due to hydraulic engineering and hydro-energetic projects, exploration of oil- and gas-fields, oil-chemical production, irrigation of nearshore territories, and increase of industrial and domestic water supply (Zonn, 1999).

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Bibliography

Berlyant, A.M., Mamaev, V.O., and Musina, O.P. (eds.), 1999. GIS. The Black Sea. Moscow, Izd. Moskovskogo universiteta (in Russian).

Bruun, P., 1962. Sea-level rise as a cause of shore erosion. *Journal of Waterways and Harbors Division*, **88**(WWI): 117–130.

Dolotov, Yu.S., 1961. Formation and classification of marine depositional terraces on the coasts subjected to the affect of tectonic emergence. *Transactions of Institute of Oceanology of the Soviet Academy Science*, **48**: 172–192 (in Russian).

Dolotov, Yu.S., 1996. On coastal relief change conditioned by sea level oscillations and its influence on economic activity (Caspian Sea as an example). In *Coastal Change*. Proceedings of the International Conference BORDOMER and IOC-UNESCO, **1**: 496–506.

Dolotov, Yu.S. and Kaplin, P.A., 1996. Present shore zone of the Caspian Sea—natural model of coastal evolution under conditions of the global sea-level rise. *Oceanology*, **36**(6): 926–930 (in Russian).

Ignatov, E.I., Kaplin, P.A., Luk'yanova, S.A., and Solov'eva, G.D., 1992. The effect of the recent transgression of the Caspian Sea on its coastal dynamics. *Geomorphology*, **1**: 12–21 (in Russian).

Ignatov, E.I., Kaplin, P.A., Luk'yanova, S.A., and Solov'eva, G.D., 1993. Evolution of the Caspian Sea coasts under conditions of sea-level rise: model for coastal change under increasing "greenhouse effect". *Journal Coastal Research*, **9**(1): 104–111.

Kaplin, P.A., 1989. Shoreline evolution during the 20th century. In Ayala-Costanares, W., Wooster, A. and Yanes-Arencibia, A. (eds.), *Oceanography-1988*. Mexico D.F.: UNAM Press, pp. 59–64.

Kaplin, P.A., 1990. Practical problems for coastal submergences in the light of secular trends. In Paepe, P. *et al*. (eds.), *Greenhouse Effect, Sea Level and Drought*. Ddorecht: Kluwer, pp. 385–393.

Kaplin, P.A., 1995. The Caspian: its past, present and future. In Mandych, A.F. (ed.), *Enclosed Seas and Large Lakes of Eastern Europe and Middle Asia*. Amsterdam: SPB Acad. Publ., pp. 71–117.

Kaplin, P.A., 1997. Implications of climatic change and water-level rise in the Caspian Sea Region. Regional review—1995. UN Environment Program, Geneva, UNEP.

Kaplin, P.A., and Ignatov, E.I. (eds.), 1997. Geoecology of the Caspian region. Issue 1: Geoecological changes under conditions of the Caspian Sea level oscillations. Moscow, Geogr. Faculty of the Moscow State University (in Russian).

Kaplin, P.A., Leont'ev, O.K., Luk'yanova, S.A. and Nikiforov, L.G., 1991. The shores. Moscow, Mysl' (in Russian).

Kaplin, P.A., Porotov, A.V., Selivanov, A.O. and Esin, N.N., 1993. The North Black Sea and the Sea of Azov coasts under a possible greenhouse-induced global sea-level rise. In Kos'yan, R. (ed.), *Coastlines of the Black Sea*. New York: American Society of Civil Engineers Publication, pp. 316–354.

Kiknadze, A.G., 1977. The dynamic system and sediment budget of the Georgian coast. In *Man and Natural Environment*. Sukhumi: Alashara Publication, pp. 5–11 (in Russian).

Kos'yan, R. (ed.), 1993. *Coastlines of the Black Sea*. New York: American Society of Civil Engineers Publications.

Kuksa, V.I., 1994. *Southern Seas* (Aral Sea, Caspian Sea, Sea of Azov and Black Sea) *Under Conditions of Anthropogenic Stress*. St. Petersburg: Gidrometeoizdat.

Kurbatova, I.E., 1994. Study of anthropogenic disturbance of natural environment in the north-eastern coastal areas of the Caspian Sea using space images. *Water Resources*, **21**(4): 487–491 (in Russian).

Leont'ev, O.K., and Khalilov, A.I., 1965. *Natural Conditions of Coastal Formation in the Caspian Sea*. Baku: Academy of Science Azerbaijan SSR (in Russian).

Leont'ev, O.K., Maev, E.G., and Rychagov, G.I., 1977. The geomorphology of the shores and floor of the Caspian Sea. Moscow, Izd. Moskovskogo Universiteta (in Russian).
Molchanova, L.V., 1989. Recreational potential of the Caspian Sea coasts. In Korzun, V.I. (ed.) *Recreational Resources of the Caspian Sea: The Problems of Use and Defense*. Moscow: Nauka, pp. 4–13 (in Russian).
Nevesskii, E.N., 1967. *Depositional Processes in the Nearshore Zone*. Moscow: Nauka, (in Russian).
Nikonov, A.A., Ekman, S.V., and Mishin, A.V., 1997. Modern vertical movements of the Earth's crust in the coasts of the Black Sea and the Sea of Azov (according to the data on sea level change). *Doklady RAS*, **357**(6): 818–822 (in Russian).
Sapozhnikov, V.V., 1992. Ecological state of the Black Sea nearshore zone. In Sapozhnikov, V.V. (ed.) *Ecology of the Black Sea Nearshore Zone*. Moscow: VNIRO, pp. 5–17 (in Russian).
Shuiskii, Yu.D., 1974. Erosional processes and their rates on the Ukrainian shores of the Black Sea and the Sea of Azov. Izvestiya AN SSSR, ser. *Geografiya*, **6**: 10–16 (in Russian).
Shuiskii, Yu.D., and Schwartz, M.L., 1988. Human impact and rates of shore retreat along the Black sea coast. *Journal of Coastal Research*, **4**(3): 405–416.
Simeonova, G.O., 1976. Onshore processes of the Black Sea within Bulgaria. In XXIII International Geographical Congress, Dynamics of sea shores, Tbilisi, pp. 32–35 (in Russian).
Veliev, Kh.A., Mamedov, S.B., and Aliev, Yu.G., 1987. Perspectives of recreational assimilation of the Azerbaidzhan coast of the Caspian Sea. In *Geographical and Economic Problems of investigations and Assimilation of the Southern Seas of the USSR*. Leningrad: USSR Geographical Society (in Russian).
Voropaev, G.V., Krasnozhen, G.F., and Lachojani, H.K., 1998. River drainage and steadiness of the Iranian coast of the Caspian sea. *Water Resources*, **25**(6): 747–758 (in Russian).
Zenkovich, V.P., 1958. *Dynamics and Morphology of the USSR Shores of the Black Sea*. Moscow: USSR Academic Science Publication, vol. 1 (in Russian).
Zenkovich, V.P., 1959. *Dynamics and Morphology of the USSR Shores of the Black Sea*. Moscow: USSR Academic Science Publication, vol. 2 (in Russian).
Zenkovich, V.P., 1982. The Black Sea and the Sea of Azov. In Aseev, A.A., and Korzhnev, S.S. (eds.). *The Far East Shores and Other Shores of the USSR*. Moscow: Nauka, pp. 199–205 (in Russian).
Zenkovich, V.P., and Schwartz, M.L. 1987. Protecting the Black Sea-Georgian SSR gravel coast. *Journal of Coastal Research*, **3**(2): 201–209.
Zonn, I.S., 1999. *The Caspian: Myths and Reality*. Moscow: TOO "Korkis" (in Russian).

Cross-references

Barrier
Beach Processes
Deltas
Marine Debris—Onshore, Offshore, Seafloor Litter
Oil Spills
Sea-Level Rise, Effect
Shore Protection Structures
Water Quality

BLUFFS—See CLIFFED COASTS

BOGS

Terminology

The term bog is used to describe certain forms of wet terrestrial vegetation. Unfortunately, in common with the words employed for many other categories of wetland, there are variations and inconsistencies in usage, regionally (particularly within Europe) as well as globally. Bog has been broadly defined so as to encompass all types of peat forming vegetation (see entry on *Peat*) or narrowly defined to denote only plant communities which are dependent upon precipitation and dust for supplies of water and nutrients. The term peatland is more appropriate for the former. The latter "ombrotrophic" condition may be an absolute state, however, this is by no means always the case and it is perhaps not surprisingly, therefore, that such communities have floristic affinities with other wetland vegetation types. Consequently, recent authors (Wheeler and Proctor, 2000) have preferred to use the term bog to describe a type of vegetation, that is, one which is usually dominated by either (or a combination of) sphagna (mosses), ericoids (dwarf shrubs), or Cyperaceae (sedges). Bogs are characteristically base-poor (with a pH < 5.0) and generally, though not exclusively, occur over a substratum of peat. Rather than peatland, an additional term, mire is preferred here to describe all forms of wet terrestrial vegetation. Bogs frequently grade into base-rich mires, which in Europe are referred to as fens. Fens may be herbaceous or wooded (fen carr in Europe), for which in the United States the terms marsh and swamp are, respectively, commonly used.

Classification and distribution

Mire vegetation is strongly influenced by hydrology and topography and hydro-topographical relationships have been widely used in mire classification schemes. Fundamental is the division made between the ombrotrophic (atmospheric) and minerotrophic (via surface runoff or percolating groundwater) supply of water and nutrients. Waterlogged peat can accumulate above the groundwater level (to a maximum depth of about 10 m) leading to the formation of raised mires. Such mires are entirely ombrotrophic and therefore base-poor, supporting only bog vegetation. Other mires are to some extent influenced by a mixture of ombrotrophic and minerotrophic sources, local scale variations in which are expressed through hydrochemical and floristic gradients. Such mires are by no means always base-rich and bog vegetation will occur either where the soils of the catchment are poor in soluble minerals (minerotrophic bogs) or where precipitation/evaporation ratios are particularly high. The latter situation is not uncommon along the Atlantic seaboards of North America and northwestern Europe. Here extensive, primarily ombrogenous, "blanket" bogs can cover the landscape spreading over relatively steep slopes and descending down to sea level.

On a regional and continental scale, the geographic distribution of mire types reflects the climatic regime. The limit of ombrogenous bog development in southeastern Labrador, for example, coincides with the 1,100 mm precipitation isopleth (Foster and Glaser, 1986). Globally mires are more widely distributed at high and low latitudes and at high altitude. Nevertheless, extensive lowland mires, such as the Everglades, occur even in the subtropics and tropics where in coastal districts they merge into brackish marshes and mangrove swamps. Regional and continental level reviews of bog and related vegetation types can be found in Gore (1983).

A variety of mire types are found in coastal situations (see entry on *Wetlands*). Some North American and northwestern European sites show complete spatial gradations from salt marsh, through fen and minerotrophic bog to raised bog. Similar temporal gradations can be reconstructed from peats deposited within coastal sedimentary sequences. Species composition in coastal mires is likely to be additionally influenced by the input of sodium and chloride via salt-spray, brackish groundwater, or as a result of flooding episodes.

Origins and development

Mire communities can develop from waterbodies (a process referred to as hydroseral succession) or over formerly dry surfaces. Both circumstances apply in coastal situations where vegetation changes in stratigraphic sequences are frequently used to infer waterlevel movements resulting from fluctuations in relative sea level. However, it should be noted that any environmental process influencing waterlevel elevation or nutrient status is capable of producing vegetation change. Such change can also result from internal (termed autogenic) processes, most obviously through the accumulation of sediment. Therefore, while the growth of mire over a dry surface indicates rising waterlevels, mires can develop over marine/brackish sediments as a result of falling, stable, or even slowly rising waterlevels (if exceeded by the rate of sediment accumulation).

Vegetation sequences and the processes influencing the development of coastal mires are reviewed in Waller *et al*. (1999) with particular reference to stratigraphic information collected from the Romney Marsh depositional complex in southeastern England. *Alnus glutinosa* (alder) dominated fen carr (swamp) vegetation developed, above salt marsh clays, and prevailed at sites close to the upland edge and in neighboring river valleys, from ca. 6000 to 2400 yr BP. This community appears to have been sustained both by inflowing base-rich water and rising relative sea-level (preventing vertical isolation from groundwater). At sites immediately behind a coastal barrier peat formation began later. Here salt marsh clays are followed by a sequence of herbaceous fen, minerogenic bog, and ombrotrophic bog. Bog development appears to have required both vertical and spatial isolation from base-rich water sources. The former was induced by a decline in the rate of relative sea-level rise and by climate change. An additional factor appears to be the mobility of the peat matrix. Peat formed from herbaceous vegetation is more mobile than woody peat and acidiphilous vegetation developing on such surfaces is therefore less likely to be flooded with base-rich water. Spatial isolation seems to have been achieved by the extensive landward accumulation of peat and the presence of the barrier. The importance of the latter is demonstrated by the widespread occurrence of bog in back-barrier environments in the Low Countries during the Holocene epoch. Having achieved independence from groundwater the ombrotrophic vegetation of Romney Marsh was able to continue growing for a further 1000 C14 years after other mire types within the depositional complex were subject to renewed marine inundation.

Paleoenvironmental reconstructions using bog sediments

Sediments derived from bog, in common with other organic deposits, comprise an important paleoecological archive. Preserved plant material (seeds, pollen, and vegetative remains) and faunal remains such as Rhizopods (testate amoebae) and Coleoptera (beetles) can be used to elucidate *in situ* environmental changes and in some cases changes occurring in adjacent habitats. Analysis of the pollen preserved in organic sediments has proved a particularly powerful tool for understanding long-term vegetation trends. Mires may also contain archaeological artifacts (see entry on *Archaeology*) and in northwestern Europe a number of exceptionally well-preserved human remains, referred to as Bog bodies, have been recovered. Ombrotrophic bogs, being dependent upon precipitation for their growth, have additionally been an important source of information on climate change during the Holocene epoch. In particular, stratigraphic changes from darker more decomposed peat to lighter fresh *Sphagnum* peat have been taken to indicate periods of faster peat growth and therefore wetter climatic conditions. Changes in bog stratigraphy at many locations across northwestern Europe (including coastal locations) indicate such a climate shift occurred around 2650 yr BP (van Geel *et al*., 1996). Unfortunately, changes in bog stratigraphy are not always synchronous between, or even within, bogs. Growth rates vary not only geographically but also in response to local hydrological features.

Organic sediments derived from coastal situations are commonly employed in reconstructions of former sea level as they can be radiocarbon dated and certain plant communities can be related to a specific ("reference") waterlevel range (see entries on *Peat* and *Sea-Level Indicators, Biological in Depositional Sequences*). The latter is clearly not the case with sediments derived from ombrotrophic bog. Given the difficulties distinguishing between ombrotrophic and minerotrophic bogs on the basis of floristic composition, and the gradations possible between these conditions, organic sediments derived from acidophilious vegetation should be avoided when collecting material for this purpose.

Human exploitation

Mires are exploited as a source of peat for fuel and horticulture and, following drainage, for cultivation. Ombrotrophic sediments are best suited for the former purpose, minerogenic for the latter. The extensive exploitation of peat for fuel can be traced back to the medieval period in northwestern Europe. For example, large quantities were removed from a series of valleys on the edge of the coast of Norfolk in eastern England. Subsequent flooding created a series of shallow lakes referred to as Broads. Peat continues to be a major energy resource in a number of countries (Russia, Ireland). Reclamation of the lowland mire complexes in Europe occurred from the 17th century onwards through the construction of effective drainage channels subsequently aided by pumping (see entry on *Reclamation*). Such activities result in the lowering of the land surface both as a result of sediment compaction (the loss of interstitial fluid) and erosion (as the surface organic sediments decompose). At Holme Fen, in Eastern England, where *Sphagnum* is an important peat constituent, the ground surface fell by 3.87 m between 1848 and 1978 (Hutchinson, 1980). The large-scale exploitation of bogs has increasingly led to calls for their conservation. Along with other forms of wetland they are included within the RAMSAR Convention (see entry on *Wetlands*).

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Bibliography

Foster, D.R., and Glaser, P.H., 1986. The raised bogs of south-eastern Labrador, Canada: classification, distribution, vegetation and recent dynamics. *Journal of Ecology*, **74**: 47–72.

Gore, A.J.P., 1983. *Ecosystems of the World, 4B: Mires, Swamp, Bog, Fen and Moor. Regional Studies*. Amsterdam: Elsevier.

Hutchinson, J.N., 1980. The record of peat wastage in the East Anglian Fenlands at Holme Post, 1848–1978 AD. *Journal of Ecology*, **68**: 229–249.

van Geel, B., Buurman, J., and Waterbolk, H.T., 1996. Archaeological and palaeoecological indications of an adrupt climate change in The Netherlands, and evidence for climatological teleconnections around 2650 BP. *Journal of Quaternary Science*, **11**: 451–460.

Waller, M.P., Long, A.J., Long, D., and Innes, J.B., 1999. Patterns and processes in the development of coastal mire vegetation: Multi-site investigations from Walland Marsh, Southeast England. *Quaternary Science Reviews*, **18**: 1419–1444.

Wheeler, B.D., and Proctor, M.C.F., 2000. Ecological gradients, subdivisions and terminology of north-west European mires. *Journal of Ecology*, **88**: 187–203.

Cross-references

Archaeology
Coastal Climate
Hydrology of Coastal Zone
Peat
Reclamation
Salt Marsh
Sea-Level Indicators, Biological in Depositional Sequences
Wetlands
Wetlands Restoration

BOULDER BARRICADES

Definition, distribution and historical development

Boulder barricades are elongate rows of boulders that flank the coastline, separated from the shore by an intertidal flat (Figure B62). They are the result of ice transport and therefore are found only in Arctic and sub-Arctic regions. They are formed by the grounding of boulder-laden ice rafts in nearshore zones during spring ice break-up.

In North America, boulder barricades have been reported in Labrador (Daly, 1902; Rosen, 1979, 1980); Hudson Strait, east Foxe Basin, Baffin Island (Bird, 1964); and the St. Lawrence River (Brochu, 1961; Dionne, 1972). In other areas they have been reported in the Baltic Sea and Fennoscandia (Lyell, 1854; Tanner, 1939). Lɸken (1962) utilized uplifted boulder barricades in northern Labrador as an accurate sea-level indicator to delineate the Holocene regression. While Tanner (1939) observed a decrease in barricade development corresponding to a reduction in tide range from 1.3 to 0 m in Labrador, the features do occur in other nontidal (i.e., Baltic Sea) areas.

Lyell (1854) first recognized boulder barricades as an ice-deposited landform. Daly (1902) introduced the term, but believed that they were an accumulation of boulders at the seaward limit of wave backwash, with ice playing a secondary role. Tanner (1939) concluded that the features were the result of boulder-laden ice cakes piling up against a fixed shore icefoot. Conversely, Brochu (1961) hypothesized that intertidal ice-cakes were moved seaward during ice breakup, pushing boulders to the low water line.

Boulder barricade formation

Monitoring of coastal ice in central Labrador during both winter and spring breakup are the basis for a model for the entrainment of boulders into ice and the transportation during spring breakup. In tidal regions, such as Labrador, intertidal ice freezes downward with increased freezing at each high tide. Boulders are frozen in the ice and become lifted from the intertidal bottom. Observations on a broad intertidal flat indicate that more boulders are lifted from the upper intertidal zone, so apparently the less-frequent lifting of the ice during spring tides was more effective at encasing boulders than the diurnal lifting from the lower intertidal zone. High melting rates occur from the ice surface in the late winter, so the continued freeze-down and surface-melt result in the transportation of boulders up through the ice.

Figure B62 Boulder barricade in Makkovik Bay, Labrador.

Figure B63 A boulder adrift on an ice cake, Makkovik Bay, Labrador.

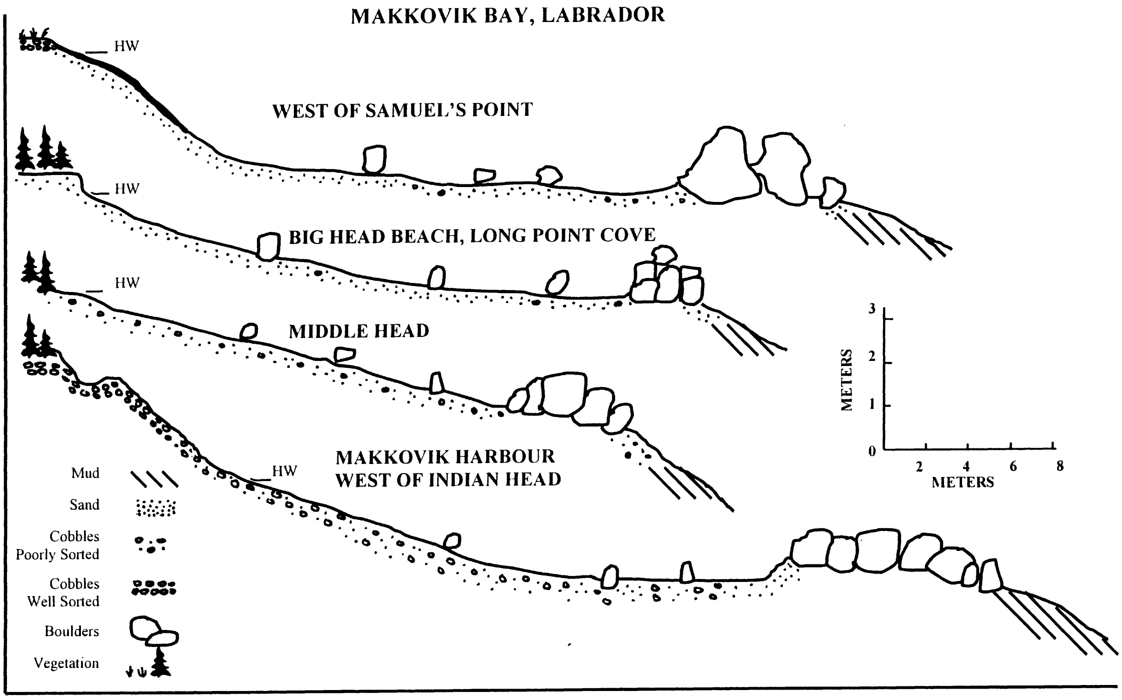

Figure B64 Nearshore profiles at selected sites in Makkovik Bay, Labrador. Random boulders are common in the intertidal zone, while most accumulate as a boulder barricade landward of a steep drop off to deeper water (from Rosen, 1982, with permission of Kluwer Academic Publishers).

In the Baltic Sea off Tallinn, Estonia, which is nearly non-tidal, the infrequent lifting of ice for freeze-down and encasement, and floating for spring transportation may be due to meteorological tides that are a major cause of sea-level fluctuations in the region (Maurice Schwartz, personal communication).

In spring when the snow cover has melted, boulders have been observed sitting on intertidal ice pans (Figure B63). The intertidal zone breaks up before offshore areas because of numerous tidal cracks and the decreased albedo of the mud-laden nearshore ice. Shore leads up to 1 km wide serve as thoroughfares for these wind-transported ice rafts. The boulders may be randomly deposited in the nearshore as *boulder flats*, as commonly occurs on the deltaic flats at the heads of embayments. However, in central Labrador many of the intertidal zones consist of uplifted marine clays with the top surface planed-off by contemporary wave and ice processes. This results in a slope-break near the low water line (Figure B64). Since the ice thickness is comparable to the tide range, there is a high probability for ice-rafts to ground at this position. Accumulation of boulders over successive seasons results in an intermittent barricade, which further serves to trap ice rafts during breakup. Landward of the slope break/boulder barricade position, random boulders, or boulder flats are also common.

At Tallinn, Estonia, the boulder barricades form in a similar setting. In this area, the nearshore is a rock-cut bench and the barricades accumulate at the seaward limit of this bench. (M. Schwartz, personal communication). Conversely, in the St. Lawrence Estuary, there is a range of nearshore boulder forms, including boulder flats, mounds, ridges, and pavements (Dionne, 1972), which corresponds with no evidence of a nearshore slope break.

Summary

Boulder barricades are the result of grounding of boulder-laden ice rafts in the nearshore zone during spring ice breakup. Wind and tides are the major transport mechanisms. The requisite conditions for the formation of boulder barricades are: a rocky coastal setting, sufficient winter ice, and water-level fluctuations to entrain boulders in ice rafts; a distinct slope break in the nearshore zone. Without the third condition, boulders will be desposited as boulder flats.

Boulder barricades are distinctly different from *boulder ramparts* and *ice-push ridges*, which are common on coastal and lake shorelines in that they are located seaward of the strandline, rather along the water line, due to the conditions discussed above.

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Bibliography

Bird, J.B., 1964. *The Physiography of Arctic Canada*. Baltimore, MD: Johns Hopkins Press.
Brochu, M., 1961. Déplacements de Blocs par la Glace le long du Saint Laurent. Étude Géographique No. 30, Ministere des Mines et des Relevés techniques, Ottawa, Canada.
Daly, R., 1902. Geology of the northeast coast of Labrador: Bull. Harvard Mus. Comp. Zool., no. 38, pp. 270–302.
Dionne, J.C., 1972. Charactéristiques des blocs erratiques des rives de l'estuaire du Saint Laurent. *Rev. Géogr.*, Montreal, vol. XXVI, 2, pp. 125–152.
Lϕken, O., 1962. The late-glacial and post glacial emergence and the deglaciation of northern most Labrador. *Geography Bulletin*, **17**, p. 23–56.
Lyell, C., 1854. *Principles of Geology*, New York: D. Appleton and Co.
Rosen, P.S., 1979. Boulder Barricades in Central Labrador. *Journal of Sedimentary Petrology*, **49**: 1113–1124.
Rosen, P.S., and McCann, S.B., 1980. Coastal environments of the Makkovik region Labrador. *Geologic Survey of Canada*, **80–10**: 267–280.
Rosen, P.S., 1982. Boulder Barricades. In Schwartz, M.L. (ed.), Encyclopedia of Beaches and Coastal Environments, Stroudsburg: Hutchinson Ross, pp. 176–178.
Tanner, V., 1939. Om de blockrika strandgördlarna (Boulder barricades) vid subarktiska oceankustar, förekomstsätt og uplomst. *Terra*, **51**: 157–165.

Cross-references

Arctic, Coastal Geomorphology
Ice-Bordered Coasts
Paraglacial Coasts

BOULDER BEACHES

Strictly speaking a boulder beach is one where the mean clast size meets the formal definition of "boulder" in the terms of the Wentworth grade scale, that is with a mean particle dia-meter of >256 mm (−8 *phi*) (Wentworth, 1922). However, the term is also sometimes used in a general sense to describe beaches where the sediment is a mixture of boulders and large cobbles.

Boulder beaches are found in high wave-energy environments where clasts of these large dimensions are released directly by erosion of bedrock, or where material is delivered to the shore zone by slope movements such as rockfall. In both cases sediment size is a function of joint spacing. Preformed boulders may also be supplied by erosion of Quaternary deposits such as glacial till, and by infrequent high-magnitude river floods.

While a considerable body of literature exists on gravel beaches, (usually concerned with pebbles and small cobbles), beaches of large cobbles and boulders have been neglected. This is partly because morphological response times are too long for consideration in the normal time-frame of academic field programs, and also because very large clasts are difficult to characterize. Some publications purporting to be general overviews of coarse sediments completely disregard boulders. Others set arbitrary upper limits to their area of interest in that, even where boulders form part of the beach sediment, sampling, experimental work, and subsequent analyses are confined to, at maximum, the cobble sizes. Most workers are reasonably precise about the lower size limit of the sediment studied, but the upper limit often remains vague: this may be partly due to the prevalent use of the *phi* scale, which becomes increasingly generalized in the coarser sizes. Terminology is often imprecise, for example, the description "gravel" is commonplace, although it covers all clasts coarser than 2 mm diameter (-1 *phi*), and so does not distinguish among pebbles, cobbles, and boulders.

Sedimentology

The few studies that have been carried out on boulder beaches have tended to look at details rather than broad patterns of sedimentation and morphology, for example, Bartrum (1947), Shelley (1968), and Hills (1970). The neglect of large clast beaches has led to attempts to apply sedimentation models derived from studies of pebble and cobble beaches to the boulder beach environment. Doubts concerning the validity of such extrapolations were fully confirmed by a comprehensive study of boulder beaches by H.L. Oak along the coast of New South Wales in Australia (Oak, 1984). Oak proposed that boulder beaches demonstrate certain unique sedimentary characteristics that distinguish them as fundamentally different from pebble and cobble beaches. Hence, relationships established in the many studies of gravel beaches are often inapplicable.

The dominant characteristics of boulder beaches listed by Oak are:

1. A high wave-energy environment, competent to move large clasts.
2. Upbeach fining of sediment.
3. Abundant breakage of sediment.
4. Positively skewed size distributions.
5. Upbeach decrease in roundness.
6. No shape zonation.
7. No sphericity grading.
8. Low foreshore slopes, decreasing as particle size increases.

Of these characteristics numbers 2–4 and 6–8 contrast strongly with the known sedimentary characteristics of pebble and cobble beaches.

Sediment size

Mean clast size is the most significant parameter determining the sedimentary character and behavior of a boulder beach. The pattern of general upbeach coarsening typical of smaller grade gravels is a function of two interlinked processes: (1) storm swash can carry most of the range of available clast sizes upslope, and (2) backwash is competent to carry a major part of this range at least some distance seawards. Neither of these is true of a boulder beach, and so clast size is the primary determinant of movement. Clast size decreases upbeach because the dominant boulders are so large that they can only be moved by traction as bedload, and even then perhaps only in very infrequent intense storms. Only the sub-population derived from breakage of the larger clasts can be suspended. As wave uprush moves up the beach face, permeability reduces its volume, and gravity effects and turbulence reduce its velocity so quickly, that only increasingly finer material can be transported. Backwash effects are negligible, so the smaller clasts, including breakage products, remain where swash deposits them. Waves can move the larger boulders to the trim line at the base of the beach, but cannot move them any distance upslope.

As high-energy marine processes act on predominantly boulder-sized sediment winnowing of the fines produces a sediment assemblage dominated by a relatively small number of well-sorted large clasts, with a very minor subordinate population of smaller fragments, most of which have survived in the high-energy area only because of entrapment. The distinctive positively skewed size distribution of a boulder beach is attributed to the presence among the beach sediments of this tail of fines derived as breakage products of the dominant boulder population. Since large clasts resist continuous movement, spasmodic breakage during storms is the dominant size-reducing process (Bluck, 1969; Matthews, 1983). Abrasion is limited to the effects of passive sandblasting or the small movements of *in situ* abrasion, both of which are largely confined to a limited area at the base of the beach. In contrast, on pebble/cobble beaches breakage is minimal and most size reduction is achieved by attrition. The very fine products of this process will be removed in suspension, unlike pebble-sized breakage products, which may be retained on the beach.

Sediment shape

Shape sorting (and the related characteristic of sphericity sorting) is poorly developed on a boulder beach. This forms a contrast with the characteristic shape zoning of a pebble beach, which results from selective clast transportation in which backwash plays an important role. On a boulder beach the sedimentation process is fundamentally different because the morphology is purely swash-formed. Selective shape sorting becomes increasingly ineffective where clasts are large and where wave conditions are turbulent. On a high-energy boulder beach shape sorting is insignificant because shape is only a dominant influence when entrainment forces are at critical thresholds for selective transport. When forces are not marginal, mass rather than shape, is the dominant control on net up- and down-beach transport potential.

Shape-controlled sorting processes are inoperative because, (1) even when storm waves are competent the prevailing bedload transport mechanism is basically insensitive to clast shape, and size will remain the dominant factor, (2) the large clast sizes and the high porosity of the beach means that backwash does not have the energy potential to create shape sorting, and (3) the rugosity of the beach surface militates against the gravity-induced downslope movements of pebble grades, which are instead trapped in the voids between boulders. Only storm swash leaves its fingerprint and as a result the only primary structure imprinted on a boulder beach is swash-controlled upbeach fining. Size, therefore, exercises not only an initial control on upbeach sorting, but it is also the terminal control. As mean clast size decreases the size control typical of boulder sedimentation gradually gives way to the shape control associated with pebble and cobble sedimentation.

Sediment roundness

Beach boulders are typically smoothed and rounded. On high-energy coasts clast transport, given free movement conditions, is very rapid. This casts considerable doubt on whether the angular, rough-textured blocks produced by wave quarrying and rockfall could possibly acquire such a degree of rounding and smoothing on a short (both spatially and temporally) unimpeded journey from source to boulder beach. Active and passive abrasion and rounding processes continue to take place in the beach environment, but their effects are largely confined to the base of the beach. Clasts at higher elevations on the beach face may have been emplaced by one high-magnitude storm. Marine influences will rarely reach these elevations, and only slow weathering processes can contribute to further rounding. It is doubtful whether the sum of these beach-face processes can entirely account for the evolution of clast shapes from an initially angular form controlled by geology, to the rounded, marine form characteristic of boulder beaches. It seems more likely that the majority of boulder beach clasts have spent some time in the intensely turbulent and abrasive hydrodynamic environment represented by traps such as potholes, gullies, and channels. Eventually a storm liberates the clasts to continue their journey to the beach.

Roundness is most pronounced toward the base of a boulder beach because the large clasts in this area experience marine action over longer time periods, and also because angular breakage products will be transported upbeach by wave action. Rounding cannot be taken as evidence of clast movements within the beach deposit, as rounded profiles can be attributed to pre-emplacement history (see above), and can also be created and maintained by *in situ* processes such as breakage, mutual attrition, and "water load abrasion". The large well-sorted clasts found in the high-energy zone near the seaward margin of the beach characteristically demonstrate rounded, flattened ellipsoidal profiles. The remarkable stability of the lower part of a boulder beach in the face of high-energy wave action is probably due to a combination of large clast size, imbrication caused by strong unidirectional flows, and a variety of other fitting and interlocking processes acting on the beach fabric (Shelley, 1968; Hills, 1970; Bishop and Hughes, 1989).

With distance upbeach angularity tends to increase, especially in the finer grades with more compact and platy shapes. This is a general comment, as the size- rather than shape-controlled swash transport mechanism will occasionally carry clasts exhibiting the full range of roundness well up the beach face. The major reason for the upbeach increase in angularity is the influence of breakage during infrequent storms. The products of breakage will remain as a component of the beach sediments because at higher positions on the beach face wave action that might winnow small-grade material is infrequent, backwash is ineffective, and the only agency acting to increase roundness is the relatively slow process of spheroidal weathering. On the boulders near the landward margin of the larger beaches, surface soundness deteriorates as weathering processes produce a rougher texture, and lichen colonization is common. In a general sense it is probably valid to consider the development of weathering rinds and lichen cover as indices of decreasing marine influence and movement. However, recent rockfall blocks on any part of the beach face carry a weathering/lichen signature from a subaerial environment, not a beach environment.

Sediment orientation

On gravel beaches pebbles transported as bedload generally tend to be oriented with the long axis parallel to the shore, that is; transverse to the direction of swash movement. Such preferred orientation patterns are weakly developed on boulder beaches because a high velocity turbulent flow on a coarse bed leads to decreased regularity in orientation patterns. Clast collisions can change orientations, disturbing or perhaps even completely obscuring the pattern imprinted by the transport process. Clasts can also be oriented by the prevalent waves without undergoing net transport. Thus, while a bedload transport mechanism does generate preferred orientations, the generally weak development of this characteristic on boulder beaches is probably due to the interplay of high-energy wave action with the particularly rough surface of the beach deposit.

Beach profile

On all types of beaches relationships involving slope are regarded as particularly significant because slope is usually considered the primary index of morphological response to wave action. There have been so few published studies on boulder beaches that discussion on their profiles must be tentative.

The most notable feature of the typical boulder beach profile is a lack of variation over time. Even during relatively severe storms changes involve only individual clasts, with the beach slope itself remaining unchanged. Some boulder beaches exhibit obvious concave upwards profiles. In detail, many have one basically rectilinear main facet extending down to mean high-water mark or below, with overall concavity produced by narrow and rudimentary low angle facets in the intertidal area. The beach profile can be conceptualized as providing a "fingerprint" of the resultant of earlier swash/backwash interaction. On boulder beaches backwash is minimal so beach material is pushed shorewards to rest at an angle controlled by the balance of gravity and the dominant swash forces. For this reason, coarse beaches are more likely to be concave upwards than fine beaches.

On boulder beaches mean beach slope decreases as mean size increases. This characteristic appears to be in direct conflict with one of the basic tenets of coastal studies, that is, that coarser sediments produce steeper slopes, partly because the angle of rest is higher, but mainly because high percolation reduces backwash, which would tend to draw down material and lessen the slope. Shepard (1963) published a table in which the predicted average beach face slope for clasts in the 64–256 mm size range (-6 to -8 *phi*) was 24°. However, the slopes of boulder beaches are considerably below this predicted value, an illustration of the dangers inherent in extrapolating from work on finer grade material. Most studies of boulder beaches record slopes in the range 6°–14° with the mean lying around 12°.

Oak (1984) formulated an explanation for the finding that the slopes of boulder beaches are gentler than predicted. On all beaches of whatever mean sediment size, storms produce an equilibrium profile that is flatter than the pre-storm profile. The accepted principle that a beach must adjust to wave energy by flattening its profile holds true then, even for boulder-sized clasts. Beach angle does indeed increase with clast size, but only if waves can move all sediment. In practice only high-energy storm waves are competent to move large clasts, so the profile of a boulder beach is in fact a "lag" storm profile, adjusted to and formed by storm waves. On sand and pebble beaches lag times are short, and in the days after a storm infill will steepen the beach face. However, this does not happen on a boulder beach, as normal wave action cannot bring about a steep fairweather profile, so the beach typically exhibits a relatively low-angle storm profile. Therefore, the relatively gentle slope of a boulder beach, with its concave upwards profile, can be considered indicative of high swash velocities and minimum sediment storage, that is, a storm profile. The persistence of the profile simply reflects the fact that competent storms are infrequent.

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Bibliography

Bartrum, J.A., 1947. The rate of rounding of beach boulders. *Journal of Geology*, **55**: 514–515.

Bishop, P., and Hughes, M., 1989. Imbricate and fitted fabrics in coastal boulder deposits on the Australian east coast. *Geology*, **17**: 544–547.

Bluck, B.J., 1967. Sedimentation of beach gravels, examples from south Wales. *Journal of Sedimentary Petrology*, **37**(1): 128–156.

Bluck B.J., 1969. Particle rounding in beach gravels. *Geological Magazine*, **106**(1): 1–14.

Hills, E.S., 1970. Fitting, fretting and imprisoned boulders. *Nature*, **226**: 345–347.

Matthews, E.R., 1983. Measurements of beach pebble attrition in Palliser Bay, southern North Island, New Zealand. *Sedimentology*, **30**(6): 787–799.

Oak, H.L., 1984. The boulder beach: a fundamentally distinct sedimentary assemblage. *Annals Association American Geographers*, **74**(1): 71–82.

Shelley, D., 1968. Fitting boulders: the result of an important shore process. *Nature*, **220**: 1020–1021.

Shepard, F.P., 1973. *Submarine Geology*. New York: Harper and Row.

Wentworth, C.K., 1922. A scale of grade and class terms for clastic sediments. *Journal of Geology*, **30**: 377–392.

Cross-references

Beach Sediment Characteristics
Boulder Barricades
Boulder Pavements
Cliffed Coasts
Cliffs, Lithology versus Erosion Rates
Gravel Barriers
Gravel Beaches
Rock Coast Processes
Shore Platforms

BOULDER PAVEMENTS

Striated boulder pavements can form either on intertidal surfaces in areas affected by floating ice (Martini, 1981; Hansom, 1983, 1986) or at the base of glaciers or on grounded ice sheets (A.G.I., 1974; Boulton, 1978; Visser and Hall, 1984). Pavements have also been described from fluvial environments (Mackay and Mackay, 1977). Their distinctive nature also allows them to be used in the sedimentary record to assist in the reconstruction of past ice-affected environments (Eyles, 1988). Pavements deposited subglacially are argued to be the result of accretion of boulders around an obstacle and to carry striations that are largely unidirectional. Although there are no detailed descriptions of such pavements forming in present glacial environments, they have also been described from the top surface of Quaternary deposits as well as buried within such deposits (Hansom, 1983; Eyles, 1988). Pavements formed on present cold-climate intertidal surfaces are thought to be the result of abrasion and bulldozing of boulder-lag surfaces by floating ice and small icebergs (Martini, 1981; Hansom, 1983, 1986, Gilbert *et al*., 1984; Forbes and Taylor, 1994). The striations that the boulder surfaces carry are then controlled by the direction of movement of blocks of floating ice together with the rotational striations imparted when such blocks become stranded. Prerequisites for the development of intertidal boulder pavements are held by Hansom (1983) to be: (1) a boulder source; (2) frequent onshore movement of floating ice; and (3) a low-gradient intertidal zone. Given such conditions, the degree of development of the pavement seems to be controlled by the frequency of onshore ice movement, because the best formed pavements occur in areas subject to the highest frequencies of freely moving ice rather than areas that remain frozen for substantial parts of the year.

Figure B65 A well-developed boulder pavement in the South Shetland Islands, Antarctica. The polygonal depressions are caused by tidally grounded ice blocks which smooth and striate the boulders.

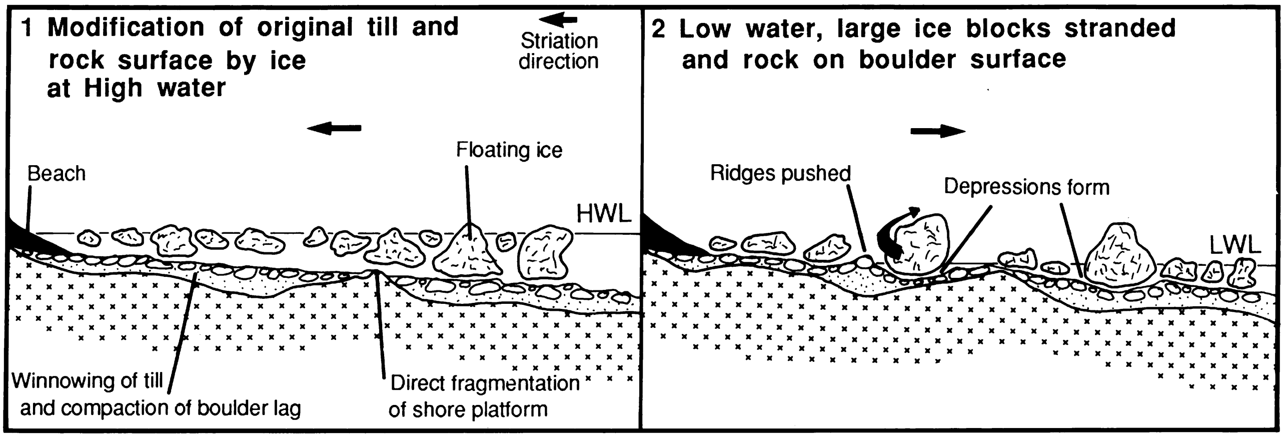

Figure B66 A model showing the development of a boulder pavement at the extremes of the tidal cycle. Progressive stranding of ice blocks causes compaction, polishing, and striation of the boulders as well as forming depressions in the pavement surface.

Marine boulder pavements are composed of smoothed boulders, often of up to 1 m in diameter, that are tightly packed together in the intertidal zone, the pavement surface appearing as a smooth, highly polished, and striated mosaic. The pavement surface is often interrupted by outcropping bedrock together with shore-normal furrows and polygonal depressions that can be up to 5 m across (Figure B65). In the South Shetland Islands (see *Atlantic Ocean Islands*) they have been described as comprising a single layer of boulders underlain by a layer of clay containing locally derived lithologies, whereas in South Georgia, pavements are underlain by glacial till into which the boulders have been packed (Hansom, 1983). The main processes involved in pavement development are summarized in Figure B66. Floating ice blocks coming ashore onto a low gradient boulder-strewn shore bulldoze and pack loose boulders in the zone of grounding, initially in the upper intertidal but increasingly at the seaward edge. Some boulders may come from direct fragmentation of rock outcrops of any underlying shore platform that may exist and some may come from ice-rafted exotics. Polishing of the boulder surface is achieved by rock-shod floating ice abrading and striating the surface of the boulders (Hansom, 1983). The orientations of the striations also inform the development processes of the surface polygonal depressions since the spread of striations on boulder ridges parallel to the shore can only be achieved by partially stranded ice blocks rotating on the pavement surface (Figure B66).

In the Antarctic, boulder pavements are found in varying degrees of development across 10° of latitude from South Georgia to the Antarctic Peninsula and in Victoria Land, pavements of tightly packed and smoothed boulders locally veneer the shallow subtidal shore platforms of the Adare and Hallett Peninsulas (Hansom and Kirk, 1989). The distribution and development of the Antarctic pavements show clear relationships between the frequency of floating ice grounding and wave processes. Where the frequency of ice grounding is high then the pavements are well developed. In the South Shetlands, the probability of floating ice and the percentage of ice concentration are both high. This limits the wave processes that destroy the pavement surface while ensuring frequent ice smoothing and packing. The result is a morphogenetic environment with a mix of ice/wave processes that is optimal for pavement development. Moving south away from this optimal ice/wave zone the incidence of grounding ice is reduced and so in the Antarctic Peninsula, wave processes are negligible, the incidence of fast ice is high and the frequency of ice grounding is low. Pavements here are poorly developed and embryonic. On the open coasts of South Georgia, well to the north of the optimal ice/wave zone, wave processes dominate, the frequency of grounding ice is low and so pavements are again poorly developed. However, within the sheltered inner fjords of South Georgia, wave processes are restricted, the frequency of floating ice is higher on account of glaciers calving into tidewater and, as a result, the boulder pavements are better developed (Hansom and Kirk, 1989). In the fjords of Vestfirdir in Iceland, similar forms also exist but are poorly developed as a result of the juxtaposition of a very limited ice-climate and a very low energy wave environment (Hansom, 1986).

Martini (1981) suggests that the incidence of boulder pavements can be taken as a reliable indicator of intertidal ice action. Thus, the occurrence of emerged pavements at 5, 9, 12.5, and 17 m above sea level in the South Shetland Islands is convincing evidence of unchanged morphogenetic conditions in the area at least since the uppermost of the pavements was formed some 9000 years BP (Hansom, 1983). Eyles (1988) uses boulder pavements in a similar way to reconstruct fluctuations in ice environments in the Gulf of Alaska during the early Pleistocene.

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Bibliography

A.G.I., 1974. *American Geophysical Union Glossary of Geology*. Washington, D.C.: American Geophysical Union.

Boulton, G.S., 1978. Boulder shapes and grain size distribution of debris as indicators of transport path through a glacier and till genesis. *Sedimentology*, **25**: 773–779.

Eyles, C.H., 1988, A model for striated boulder pavement formation on glaciated shallow-marine shelves: an example from the Yakataga Formation, Alaska. *Journal of Sedimentary Petrology*, **58**(1): 62–71.

Forbes, D.L., and Taylor, R.B., 1994. Ice in the shore zone and the geomorphology of cold coasts. *Progress in Physical Geography*. **18**(1): 59–89.

Gilbert, R., Aitken, A., and McLaughlan, B., 1984. A survey of the coastal environments in the vicinity of Nain, Labrador. *Maritime Sediments*, **20**: 143–155.

Hansom, J.D., 1983. Ice-formed intertidal boulder pavements in the sub-Antarctic. *Journal Sedimentary Petrology*, **53**(1): 1035–1045.

Hansom, J.D., 1986. Intertidal forms produced by floating ice in Vestfirdir, Iceland. *Marine Geology*, **71**: 289–298.

Hansom, J.D., and Kirk, R.M., 1989. Ice in the intertidal zone: examples form Antarctica. *Essener Geographische Arbeiten*, **18**: 211–236.

Mackay, J.R., and Mackay, D.R., 1977. The stability of ice-push features, Mackenzie River, Canada. *Canadian Journal Earth Science* **14**: 2213–2225.

Martini, I.P., 1981. Ice effect on erosion and sedimentation on the Ontario shores of James Bay, Canada. *Zeitschrift fur Geomorphologie*. N. F. **25**(1): 1–7.

Visser, J.N.J., and Hall, K.J., 1984. Boulder beds in the glaciogenic Permo-Carboniferous Dwyka Formation in South Africa. *Sedimentology*, **32**: 281–294.

Cross-references

Antarctica, Coastal Ecology and Geomorphology
Arctic, Coastal Geomorphology
Boulder Barricades
Boulder Beaches
Glaciated Coasts
Ice-Bordered Coasts
Shore Platforms

BYPASSING AT LITTORAL DRIFT BARRIERS

Definition

A littoral drift barrier is an obstacle against the littoral drift or migration of material along the shore. Such barriers may be natural, for example, major headlands on the shore, or man-made such as jetties, breakwaters, or dredged channels, which established a hindrance for the normal drift of material along the shore.

Natural barriers may be responsible for major changes in the natural uninterrupted shore. The California saw-toothed headland shore is a large example of that. Bypassing is transportation of materials across the barrier, breaking the barrier-effect.

Bypassing by nature

Bypassing is the way that material, after a short interruption caused by an inlet, channel, jetty, or other kind of littoral barrier, is given back to the normal littoral drift zone a distance downdrift from the littoral barriers. If nature did not bypass sand across inlets, passes, and channels on seashores, many *marine forelands* including barriers, spits, and entire peninsulas would not exist. A typical example of this is Florida, which was built of sand washed down by rivers and streams from the Appalachian highland, and carried southward, for final deposition in the huge barrier and ridge systems.

Bar bypassing—limited tidal action

Figure B67 shows a barrier with an inlet. Littoral drift material passes along the barrier. At the downdrift end, it continues on its way across the inlet on a submerged bar, the extent and depth of which depends on the amount and character of the material which bypasses and the intensity of wave and current action. By increasing amounts of littoral material, the bar area increases and depth decreases.

In most cases, migration of tidal channels takes place in the direction of the littoral drift. Sand is transported over the bar under the influence

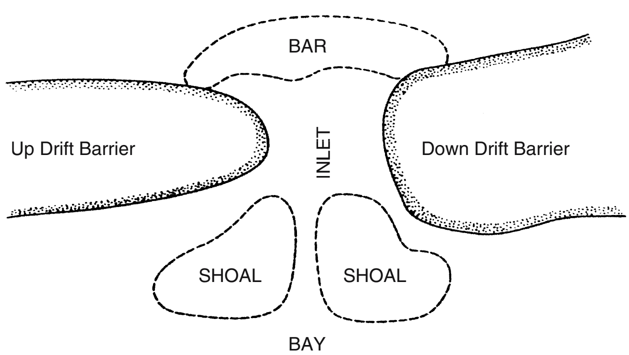

Figure B67 Coastal inlet with predominant bar bypassing (from Bruun, 1990). (Reproduced by permission. From Port Engineering V2 4E copyright (c) 1990, Gulf Publishing Company, Houston, Texas, 800-231-6275. All rights reserved.)

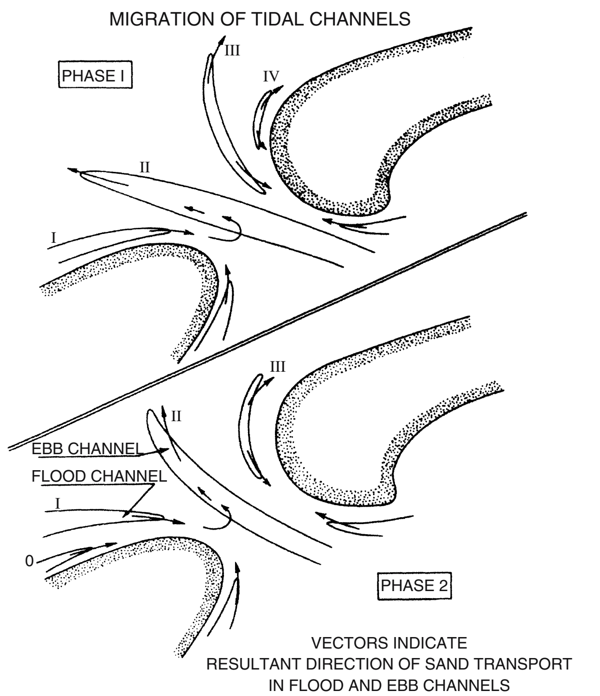

Figure B68 Migration of tidal channels (from Bruun, 1961). (Reproduced by permission of the publisher, ASCE.)

of waves and deposited on the updrift bank of the channels, thus forcing the shifting.

In the vicinity of tidal inlets, the generally strong tidal currents in the inlet change the littoral drift pattern entirely. Along the uninterrupted coastline, wave action is generally the predominant cause for the transportation of material. In the vicinity of tidal inlets, however, transport of material takes place under the combined effect of waves and tidal currents.

In tidal rivers, estuaries and inlets, tidal channels can usually be identified as either *flood* or *ebb* channels. Flood channels carry predominantly flood flow, causing a resultant sand transport in a bayward direction; they usually have a shoal at the end. Ebb channels carry predominantly ebb flow and have resultant material transport seaward and a bar or shoal at the end (Figure B67).

Principles involved in bypassing by tidal flow action—unimproved inlets

In general, sand transfer by tidal flow takes place in two different ways, namely by migration of channels and bars and by transport of sand by tidal flow in the channel. Tidal channels in inlets, particularly those running between the gorge and the ocean, are subject to migration. This means that they change location continuously, moving from one side of the inlet to the other. In Figure B68, this principle is demonstrated by Phases 1 and 2 of a tidal channel system. Channels in Phase1 are numbered, I, II, III, and IV. In Phase 2 the locations of these channels have changed compared with Phase 1, and a new channel, 0, has developed. In this example, the channels move from left to right, and bars or shoals between the channels move in the same direction with the result that a bar occasionally joins the downdrift coast.

One may distinguish between inlets that are mainly bar-bypassers (Figure B67) and inlets that are rather tidal flow bypassers by considering the ratio between tidal prism during spring tide (Ωm^3) and the total amount of material carried to the inlet entrance by the littoral drift (M_{tot} in cubic meters per year). A great many cases were analyzed and showed that inlets with $\Omega/M_{tot} < 50$ were mainly bar-bypassers, while inlets with $\Omega/M_{tot} > 150$ were mainly tidal flow bypassers. Inlets with $50 < M_{tot} < 150$ combined the two modes. Inlets or harbors without or with only little tidal prism have only one bypassing style-man-made bypassing or dredging.

Man-made littoral drift barriers

Human intervention of coastal processes started when they erected shore-perpendicular or parallel breakwaters for protecting ports against waves and sediments and groins for coastal protection on open littoral drift shores. This type of construction began in the 19th century in the Mediterranean and on the shores of the British Isles (Bruun, 1990). The problem of man-induced erosion was magnified when the Dutch invented dredging in an effort to provide greater channel depths for navigation. It was by hard and very expensive experience that they learned that when they put something out in the sea, "something is going to happen." Commonly, shoaling occurs on one side and erosion on the other side of an obstruction. In most instances, this probably came as a surprise and often initiated "desperate efforts" in order to maintain depths at an entrance (e.g., by extending updrift breakwaters or jetties or by dredging operations with available equipment or by both). This provided only a temporary relief for navigation and usually the greater the efforts to maintain depths, the more severe the erosion on the downdrift side.

The first technical counter-measures were the construction of groins and/or seawalls. While both mitigated the nearshore or onshore erosion problem, they also aggravated the downdrift erosion. Not until the late 1930s was it realized that the only practical solution to the problem was the elimination of the barrier effect. This was done by establishing sand bypassing whereby material is pumped or trucked across the barrier to the downdrift beaches.

The need for bypassing was supported by legislation such as the Florida law (1987), which reads as follows (Section 161.142, Declaration of Public Policy Relating to Improved Navigation inlets):

"(1) All construction and maintenance dredging of beach-quality sand should be placed on the downdrift beaches; or, if placed elsewhere, an equivalent quality and quantity of sand from an alternate location should be placed on the downdrift beaches.

(2) On an average annual basis, a quantity of sand should be placed on the downdrift beaches equal to the natural net annual longshore sediment transport."

Quantitative considerations

The quantitative aspect of longshore drift blocking by barriers is very simple. If the barrier causes the loss of a certain quantity of material

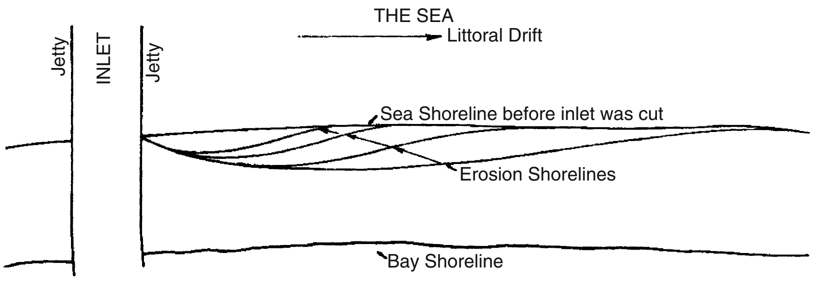

Figure B69 Shoreline development downdrift of the Fort Pierce Inlet, Florida, schematics (from Bruun, 1990). (Reproduced by permission. From Port Engineering V2 4E copyright (c) 1990, Gulf Publishing Company, Houston, Texas, 800-231-6275. All rights reserved.)

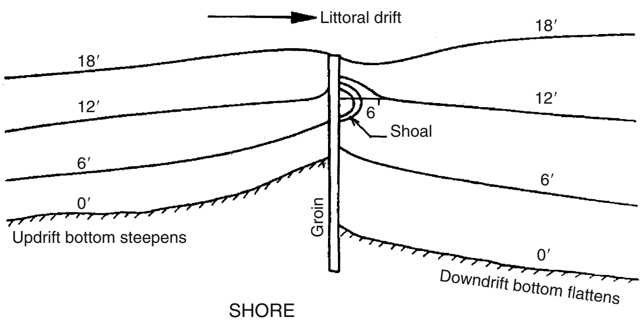

Figure B70 The development of bottom configuration downdrift of a littoral drift barrier (from Bruun, 1990). (Reproduced by permission. From Port Engineering V2 4E copyright (c) 1990, Gulf Pubhlishing Company, Houston, Texas, 800-231-6275. All rights reserved.)

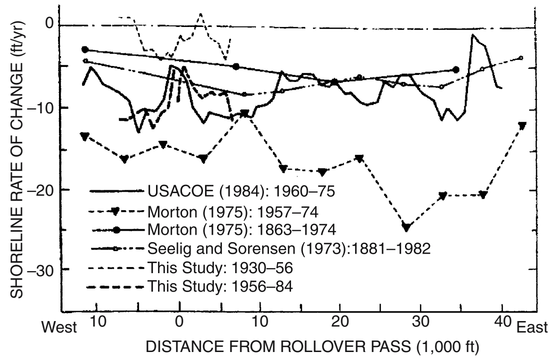

Figure B71 Comparison of shoreline rates of change near Rollover Pass, Texas (from Bruun, 1995, reprinted by permission of the Journal of Coastal Research).

which was "locked up" by the barrier, this quantity is unavailable to downdrift beaches, which consequently will suffer erosion of that magnitude. The more difficult question is: how is erosion, due to loss of sand, distributed downdrift as a function of time.

Coastal geomorphological considerations

Three parameters are important in this context: the length of the adversely affected shore, the cross-sectional retreat of the erosion cut and the rate of expansion of erosion, and its dis-tribution downdrift as functions of time. Length and cross-sectional evolution of the erosion cut give the geometric development as a function of time. The corresponding development in the offshore bottom follows the same general pattern, but there is usually a material change in the configuration of the offshore profiles, which tends to flatten in the downdrift areas (Bruun, 1990, chapters 7 and 8). Figure B69 shows a typical longshore shoreline development trend, Figure B70 (Bruun, 1990) shows the offshore development as well.

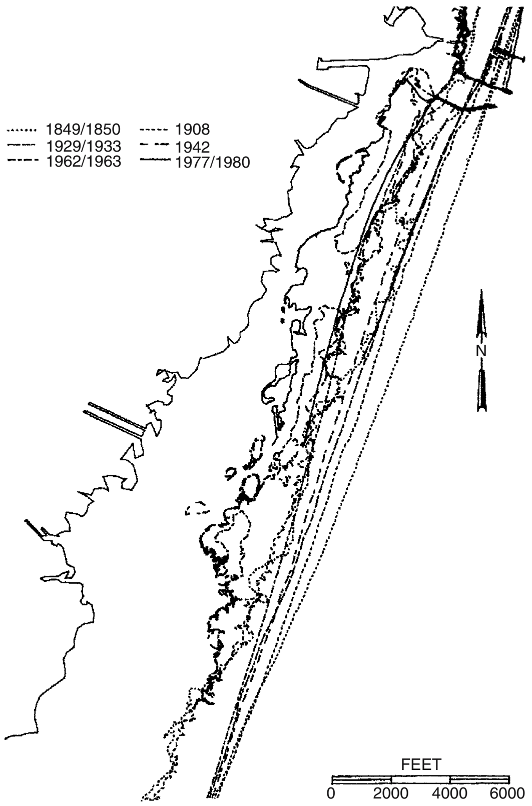

Figure B72 Barrier Island migration showing landward displacement of both Ocean and Bay high tide shorelines of Assateague Island (1850–1980). The Barrier Island maintained its width, 120–210 m, with this rapid translocation. The mainland bayshore has remained essentially stable (from Bruun, 1995, reprinted by permission of the Journal of Coastal Reasearch).

The dominant sediment bypassing mechanism at a tidal inlet affects the extent and magnitude of the downdrift high-tide shoreline response. The updrift coastline response to the introduction of a jetty is fairly localized and little dependent upon the sediment bypassing mechanisms active at the tidal inlet. Tidal inlets which are predominantly tidal-flow bypassers have more severe, downdrift effects on high-tide shoreline response than tidal inlets which bypass sediment through bar bypassing. Tidal inlets which are combined tidal flow and bar bypassers have relatively constant downdrift effects (magnitude and rate of change) through time. Significantly deepening the channel through the bar can alter the dominant sediment bypassing mechanism.

Bodge (1992, 1999), Rosati and Ebersole (1996), and Bruun (1995) made efforts to quantify the response of adjacent shores by tidal inlets. Rosati and Krauss (1999) have continued these efforts. Bodges (1999) paper is universally applicable. Bruun (1995) gives about 20 examples of which two are mentioned below.

The literature only mentions few examples where the downdrift long distance development was recorded as function of time to obtain a rate. Examples in the literature include Fields *et al*. (1989) and Bodge (1993, 1999). Theoretical approaches are available, but they concentrate on immediate downdrift reactions. Although admittedly, the effects continue to expand downdrift "infinitely" as indicated by Pelnard-Considere (1956).

Obviously, the migration rate of the downdrift erosion depends upon the quantitative magnitude of the *barrier effect*, for example, the loss of material to inlet shoals. A large loss will expand faster than a small loss.

"Beach/Inlet Processes and Management", Special Issue No. 18, of the *Journal of Coastal Research* (1993, A.J. Mehta, Editor) has a great number of examples on the influence of coastal inlets on the littoral drift system. The short distance effect of the inlet on the downdrift erosion is shown in several figures, but the development downdrift is cut short by only examining the development for a limited distance downdrift.

Rollover Pass, Texas

Conditions at Rollover Pass on the Texas Gulf, 31 km northeast of the Galveston Inlet, are described by Bales and Holley (1989) and by Bruun (1995). The pass is a man-made artificially stabilized inlet on the Bolivar Peninsula. Improvements of the pass were completed in 1959. Figure B71 compares shoreline rates of change near Rollover Pass. Referring to the period 1957–74 in the figure, it may be seen that the downdrift effect extended at least 40,000 ft. or 13 km and probably more. Based on 1957–74, one arrives at a migration rate of erosion of 13/17 = 0.8 km/year. The rate is 0.9 km/year if 1959, the completion year for the improvements, is used.

Ocean City Inlet, Maryland

Shoreline evolution on either side of the inlet is dealt with by Leatherman (1984). Historical shoreline changes on the downdrift side of the inlet on Assateague Island are shown in Figure B72. Construction of the Ocean City Inlet jetties, in combination with a net southerly longshore littoral drift, has resulted in severe erosion along north Assateague Island. It appears that none of the 120,000 m^3 of sand that annually flows south along this coastal sector reaches Assateague Island. Since the jetties, built 1934–35 are filled to capacity (1984), the material is largely moving offshore to build a huge ebb tidal delta that is detectable from space through analysis of Landsat imagery. A comparison of the 1942 and 1962 coastlines clearly shows the trend since the

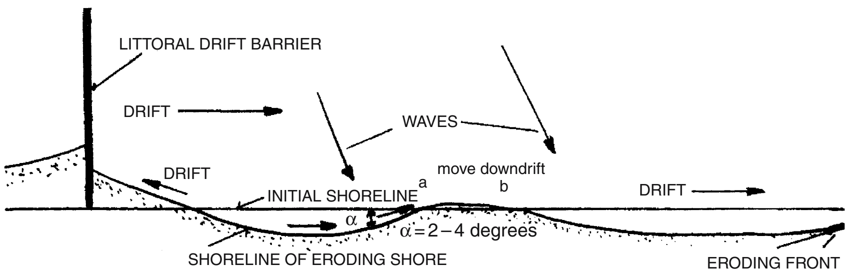

Figure B73 Schematics. The development of downdrift erosion at littoral drift barrier (from Bruun 1995, reprinted by permission of the Journal of Coastal Research).

jetty construction. The arc of erosion south of the inlet is clearly evident when considering historical changes 1950–80 in Figure B72. The historical high-tide shoreline's high-tide changes tend to converge further downdrift. This artificially induced erosion continues to impinge further downdrift through time.

Figure B72 does not extend far enough downdrift to indicate the front of the jetty-induced erosion. By a slight extrapolation, it was found that it is most likely that in 1962 it had reached 10 km south. That is, a front movement of 10 km/20 years— 0.5 km/year.

This was further confirmed by Rosati and Ebersole's (1996) quantitative research, which demonstrated that the downdrift erosion extended at least 14 km downshore (Bruun, 1995, 2000).

The zero or slow down area

The peculiar "zero-area" which sometimes appears downdrift at a rather short distance from the barrier (Figure B73) is a coastal geomorphological feature which *does not necessarily indicate the extreme limit of leeside erosion*. Obviously, the high-tide shoreline has to resume its initial direction following its change of direction on the downdrift side of the barrier. This can only be accomplished by an S-curve, which in turn develops a kind of a "corner" at point "a" as seen in Figure B73. This makes the local following section of the shore resemble "a groin" with some (minor) stabilizing effects updrift, but at the same time aggravating the large-scale erosion downdrift caused by the littoral drift barrier. A consequence of that is that the bump is most developed when the drift is very predominant and less visible for a more neutral situation.

Conclusion

The downdrift high-tide shoreline development at a littoral drift barrier may in some cases, but not always, be described by a short (local) as well as a long distance effect which both move downdrift at various rates; the long distance movement being two to three times faster than the short distance, or about ~0.5 km/ year versus ~1–1.5 km/year. These figures may be subject to considerable variances depending upon wave intensities, barrier morphologies and littoral drift magnitudes as well as upon the relative predominance of the drift. The short distance effect is a coastal geomorphological feature, the long distance a materials deficit feature. Quantitative research is making progress (Bodge, 1999; Rosati and Kraus, 1999).

Bypassing technologies

Figure B74 (Bruun, 1990) is a review of bypassing plants and principles distinguishing between non-scouring and scouring conditions (tidal entrances). Bypassing may be undertaken by bypassing plants or by bypassing arrangements.

The most simple examples of bypassing are found at breakwaters, single or double, perpendicular or parallel to shore provided with dredged entrance channels for navigation (Tables 9–16 of Bruun, 1990, is a comprehensive overview of bypassing plants and arrangements). Table B8 summarizes the situation as of 1990 (Bruun, 1990).

The main difference between bypassing at harbors and at tidal entrances lies in the action of the tidal currents. It may therefore be said that while in the case of harbors, bypassing arrangements may be

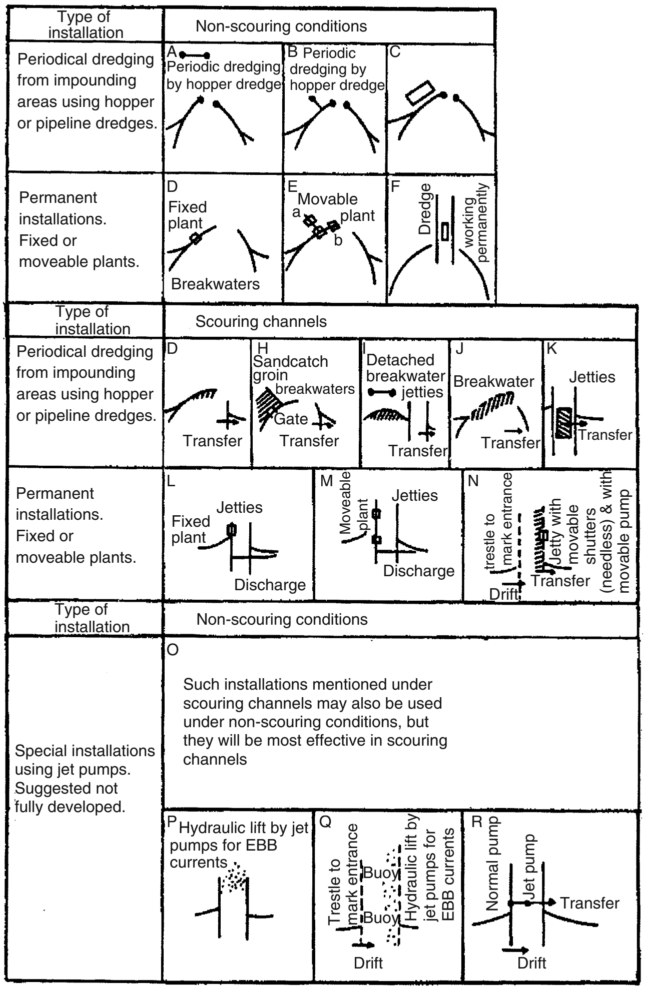

Figure B74 Various principles of bypassing (from Bruun, 1990). (Reproduced by permission. From Port Engineering V2 4E copyright (c) 1990, Gulf Publishing Company, Houston, Texas, 800-231-6275. All rights reserved.)

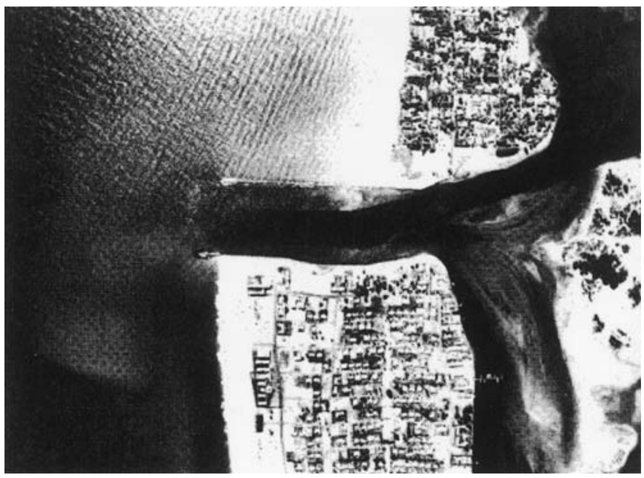

Figure B75 The Palm Beach Inlet, Florida (from Bruun, 1990). (Reproduced by permission. From Port Engineering V2 4E copyright (c) 1990, Gulf Publishing Company, Houston, Texas, 800-231-6275. All rights reserved.)

Table B8 Number of bypassing arrangements established or under Construction (Bruun, 1990, with updates)

	Fixed plants (including jet pumps)	Movable plants	Detached breakwaters	Weir jetties	Sand catch or dredged trap updrift in channel or bay
Built	7	3	3	6	20
Suggested	2	—	1	—	—

This table gives the right order of magnitude of current projects.

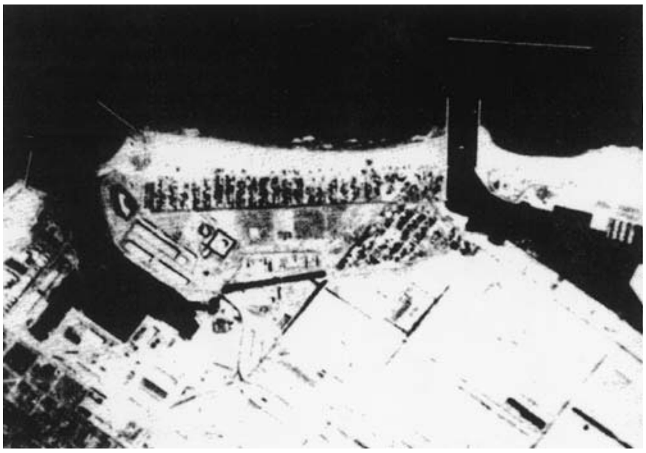

Figure B76 Detached Breakwater, Ventura Harbor, California (from Bruun, 1990). (Reproduced by permission. From Port Engineering V2 4E copyright (c) 1990, Gulf Publishing Company, Houston, Texas, 800-231-6275. All rights reserved).

designed solely on wave mechanics principles, the design at tidal inlets also includes current mechanics.

At Paradeep, State of Orissa, Bay of Bengal, India, a large movable plant which included a 750 hp pump producing 500 t sand/h was installed on a 370-m steel trestle running perpendicular to the updrift breakwater, in the middle sixties (Figure B74(E)). The specifications required that the dredge pump combined with a booster pump should be capable of handling this quantity of slurry through an 46 cm reducing to 41 cm pipeline about 2,200 m long. The plant was supposed to work fairly regularly throughout the year in most weather conditions. The trap capacity, however, proved to be too small to handle the strong deposits during the monsoon and sand bypasses the trap when it is filled, some of it in suspension even before it is filled. The result is that it has become necessary also to operate a hydraulic pipeline dredge in the entrance to remove the sand, which escaped the trap. It is probably rather doubtful that more fixed plants, although proposed or under discussion, will be built as the most recent experiences are not very promising. The 191,000 m^3/yr plant at Palm Beach Inlet in Florida, Figure B75, has not been satisfactory either and has seldom operated at the planned full capacity. It is going to be replaced by a more effective arrangement.

Future development of bypassing

The most reliable or effective trap arrangement is undoubtedly the detached breakwater built offshore on the updrift side (Figures B74(A), (I) and Figure B76). But it is an expensive solution requiring a large offshore area, a usually rather expensive breakwater and an effective suction dredge of a seagoing, therefore also expensive, type plus a rather long and therefore also costly pipeline, possibly with one or more booster stations to push the material all the way to the downdrift side beaches.

Generally, it may be said that developments that are taking place favor the most flexible arrangement of traps to be dredged by floating equipment, which bypass the material across the littoral drift barrier. The success of such arrangements, however, depends partly upon the correct placement of the trap from a sedimentation as well as a practical viewpoint in regard to transfer of material and partly upon the equipment available for transfer and the economics involved.

Trap arrangements at or on the updrift side of a sand catch breakwater like Figures B74(B), (C), (H), however, leave the dredging equipment somewhat exposed to wave action. The submerged weir (Figure B74(J)) was introduced to alleviate this drawback but it has not been fully satisfactory in all cases. While the Hillsboro Inlet, Florida, the "old timer" in the group, must be classified as a success, the arrangement at the Masonboro Inlet, North Carolina and the arrangement at East Pass, Florida, have experienced some difficulties due to weir operation.

The Hillsboro Inlet

The Hillsboro Inlet in southeast Florida, is a natural inlet to the Atlantic Ocean, connecting the Atlantic Intracoastal Waterway to the ocean. It provides free access for commercial and recreational boats, and storm water drainage for a large interior land area.

There was about a 1.2 m depth over a rock bar at the entrance in its natural condition. The inlet channel was improved in the 1960s by an excellent design developed at the University of Florida, confirmed by a hydraulic model study. The channel was cut to a depth of 3 m. A 61 m jetty on the north side for the predominant littoral drift and a 122 m jetty on the south side, were constructed.

The north side jetty has a natural weir (low section) for material transfer across the jetty, and a sand deposit basin (trap) inside for storage of material. The stored material is later transferred by a hydraulic dredge to beaches south of the inlet.

A worn-out dredge was replaced in 1983 by a 41 cm dredge reduced to 30 cm. The year-round dredging operation is very successful due to the weir design. The channel is able to be kept at an operating depth of 2.4 m 92,000 $\pm$ 15,000 m^3 of beach quality material is bypassed each year, essentially all the littoral drift sand deposited in the inlet basin and channel.

When sand is dredged promptly from the channel after a storm, less material is lost to the ocean by ebb tides or deposited in the interior channel by flood tides. The beach south of the inlet has accreted yearly and has not had to be renourished since 1983. It appears to be reaching equilibrium. A planned project will increase the outer channel depth to 6 m, improve the geo-metry of the entrance, increase the material captured for bypassing, and improve navigation safety and drainage.

Alternative bypassing systems using jet-pumps

Jet pumps submerged in the entrance for transfer by normal pumping power (Figure B74(R)) may also prove a useful procedure but it only covers a rather local area, although its influence may be expanded for some distance to either side by several pumps and pipelines (Boyce and Polvi, 1972).

The application of jet pumps to stir up material, Figures B74(P), (Q), (Bruun, 1990) using ebb currents as the main flushing or carrying agent, may prove to be a very practical arrangement, but it only helps to carry the material away from a certain local area like agitation dredging and does not transfer the material.

A jet pump system was built at the Nerang River entrance in Queensland, Australia, as described by Bruun (1990). It is based on an updrift array of jet pumps. The pumps have a large capacity, but there are problems with clogging of the pumps by debris.

Use of submerged pumps combined with fluidization

Weisman *et al.* (1982) describe improvement of channel and bypassing stabilities by perforated hydraulic pressure pipes placed below the bottom. A few examples are mentioned here.

Case one, inlet with a dredged, otherwise unprotected channel

It may be improved by lift pipes placed across the bar, at the same time improving bypassing by combined wave and (ebb) current action. A trap may also be placed in the channel to accumulate materials carried to the trap by ebb as well as by flood currents. This trap has a "lift system" in the middle that may be emptied whenever needed, for example, by a fluidization pump.

Case two, inlet entrance improved by special geometry jetties for channel stability and bypassing

Lift pipes are used to obtain optimum stability of the channel across an entrance bar or shoal (almost standard). This also improves bypassing by combined ebb currents and wave action. Channel stability is further improved by a trap in the channel operated continually for lift during ebb flows, so that the channel always stays clear. The trap may be emptied intermittently for transfer, by fluidization. Outside the updrift jetty a large trap is established for continuous transfer of material carried to the trap by littoral currents and onshore bottom creep due to wave action. This transfer may also be undertaken by fluidization using the same pump as for the bar lift.

Advantages of using hydraulic lift for channel stability

The advantages of using hydraulic lifts to increase flushing abilities are well demonstrated in nature by the influence of wave action in "opening up" a cross section. It may also be observed at places where nature delivers—free of charge—the hydraulic pressure. Some natural tidal inlets placed are accordingly all over the world. The lift may be operated according to needs and particularly during and after heavy storms. The lift is able to direct the sediment transport oceanward. Such a lift system may be used in connection with a submerged pump like the Punaise which is an underwater pump of Dutch origin. It has the shape of a thumb-tack. The pointed part digs itself down in the bottom by hydraulic pressure pumps. The fluidized material is carried from the cone-shaped pit through pipeline to the discharge area. The Punaise has been tested in the Netherlands and is being tested in the United States by the US Army Corps of Engineers.

The Punaise

The shape of the dredge gave it the name "The Punaise" (Dutch word for thumb-tack). The first "PinPoint" dredge, Punaise PN250, was commissioned in 1990, the second dredge Punaise PN400 was commissioned in late 1993.

The Punaise works on a very simple principle. A pump and suction pipe are connected to ballast tanks and then the entire structure is submerged to come to rest in the bottom where dredging is to be carried out.

The link to a small shore-based control unit and energy supply is provided by cables and hoses while a pipeline is used to transport the dredged material. The submerged pump excavates by hydraulic erosion. It creates an unstable slope upon which sediment flows to the suction intake. The unique support system makes use of the suction pipe that is embedded into the bottom to a level below the dredging depth (1–3 m) thus providing both horizontal and vertical stability.

Conclusion

1. Fixed bypassing plants will be replaced by more flexible plants. For major projects large floating plants like the trailing hopper dredger which discharges downdrift is common.
2. For medium size projects bypassing most likely will be by proper size, but smaller, hopper, or pipeline dredgers.
3. For projects of more modest size bypassing will be by shallow water hopper dredgers in some cases combined with underwater pumps and fluidized on arrangements (Visser and Bruun, 1997).

Per Bruun

Bibliography

Bales, D.J., and Holley, Ed.R., 1989. Sand transport in Texas tidal inlet, In *Proceedings ASCE Waterway, Port, Coastal and Ocean Engineering*, **115**(4) 427–443.

Bodge, K.R., 1992. *Port Canaveral Inlet Management Plan*. Port Canaveral: Olsen Associates, Inc.

Bodge, K.R., 1993. Gross transport effects and sand management strategy at inlets. *Journal of Coastal Research*, **18**(Special Issue,) 111–124.

Bodge, K.R., 1999. Inlet Impacts and Family of Solutions, *Coastal Sediments 1999*, Reston: American Society of Civil Engineers.

Boyce, A.R., and Polvi, R.D., 1972. Jet pump application in estuary maintenance. Corvallis: *Civ. Eng. Dept., Oregon State University*.

Bruun, P., 1961. Natural by-passing of sand at coastal inlets. *ASCE Translations*, **126**(4): 823–854.

Bruun, P., 1990. *Port Engineering*, Houston: *The Gulf Publishing Co*.

Bruun, P., 1995. The development of downdrift erosion. *Journal Coastal Research* **11**(4): 1242–1257.

Bruun, P., and Gerritsen, F., 1959. Natural bypassing of sand at coastal inlets. *Transactions of the American Society Covil Engineers*, Paper #3293, 823–854.

Fields, J.L., Marino, J.N., and Weishar, L.L., 1989. Effect of Florida tidal inlets on adjacent shorelines, *Florida Shore and Beach Preservation Association Proceedings*, pp. 383–391.

Leatherman, S.P., 1984, Shoreline evolution of North Assateague Island, Maryland, *Shore and Beach*, **52**(3): 3–10.

Mehta, A.J. (ed.), 1993, Beach inlet processes and management. *Journal of Coastal Research*, Special issue No. 18.

Pelnard-Considere, R., 1956. Prediction of Beach Planforms with Essai de Theorie de l'Evolution des Formes de Rivage an Plage de Sable et de Galets, *4th Journees de l'Hydraulique: Les Energies de la Mer*, Question III, No. 1, 239–300.

Rosati, J., and Ebersole, B.A., 1996. Littoral impact of Ocean City Inlet, Maryland, USA, *Chapter 23 of Proceeding of the ASCE Conference*, pp. 2779–2791.

Rosati, J.D., and Kraus, N.C., 1999, Advances in coastal sediment budget methodology. *Shore and Beach*, **67**(2–3): 56–65.

Visser, K., and Bruun, P., 1997, The Punaise underwater dredger. *Journal of Coastal Research* **13**(4): 1329–1333.

Weisman, R.N., Collins, C.C., and Pars, J.M., 1982. Maintaining tidal inlets channels by fluidization. *Journal of Waterway, Port Coastal Ocean Division*, **108**(4): 526–538.

Cross-references

Barrier
Dredging of Coastal Environments
Littoral Cells
Longshore Sediment Transport
Navigation Structures
Tidal Prism