DUKE GEOLOGICAL LABORATORY

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TRIPS ON THE ROCKS

Guide 18: Beach at Robert Moses State Park, Fire Island, Long Island, New York

Trip 24: 26 September 1992

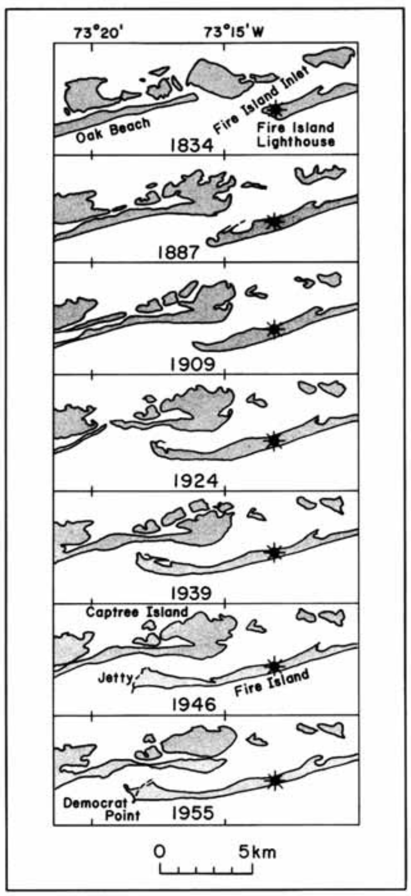

Figure 1. Migration of Fire Island inlet by westward growth of Fire Island barrier beach. Fire Island Lighthouse, built on the shore of former inlet in 1834 now stands 8 km away from the inlet. (From Friedman and Sanders, 1978, Fig. 11-11B, p. 316.)

Field Trip Notes by:

Charles Merguerian and John E. Sanders

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Trip 24: 26 September 1992

Logistics:
Departure from NYAS: 0830
1000: Arrive Robert Moses State Park, Parking Field 3; short break for restrooms, cafeteria, etc.; Assemble outside pavilion for walk to narrow beach and eroded scarp. SST and beach strata.
1230: Lunch at picnic pavilion.
1300: Board vans to drive to W end of Parking Field 2: walk on a wide beach; dig 2 SST's.
1600: Board vans for pit stop at pavilion; trip back to NYAS.

INTRODUCTION

Fire Island is widely known for many reasons. Among geologists, it is world famous as an example of rapid lateral growth of a barrier island as a result of a persistently shifting inlet. This growth is proved by a series of dated maps (Figure 1, on cover) showing the position of Fire Island Inlet with respect to the Fire Island Lighthouse. This lighthouse was built alongside the inlet in 1834. Today, the lighthouse is 8 km (5 miles) away from the inlet. Between 1834 and 1940 (when the Corps of Engineers built the Federal Jetty at Democrat Point), the average rate of westward growth of Fire Island was 1 meter per week! Figure 2 shows an aerial view of the west end of Fire Island after the great hurricane of 1938 and prior to the construction of the Federal Jetty.

In addition to this historical evidence of rapid westward growth of Fire Island, the open-ocean beaches have been studied by the Corps of Engineers in connection with the persistent movement of the sand by waves and by several geologists. The results of the Corps of Engineers studies are contained in various House (of Representatives) Documents and a few papers that were published in engineering journals. The Corps of Engineers accepted the general conclusion that the source of the sand for the open-ocean beaches on Fire Island was erosion of the cliffs at Monauk Point (Taney, 1961a, b). Maurice Rosalsky, a longtime member of the geology faculty at City College, published three papers on his observations of the behavior of the beach at Fire Island (1949 NYAS minor beach features; 1950 beach drifting; 1964 swash and backwash). Although these papers by Rosalsky did not make much impact on the ongoing rush of studies about beaches, JES has found them to be accurate and valuable contributions.

JES has been studying shore processes at Robert Moses State Park since 1968, when Naresh Kumar began to investigate the geologic record left beneath the western end of Fire Island by the shifting inlet (Kumar, 1973; Kumar and Sanders, 1972, 1974, 1975). As part of Kumar's research project, supported by a grant to JES from the National Science Foundation, we persuaded the Corps of Engineers to let us borrow their collection of vibracores that had been drilled seaward of Fire Island as part of the Sand Inventory Program. These cores contained many graded/laminated sediments that we interpreted as the effects of storms (Kumar and Sanders, 1976, 1978).

Figure 2. Oblique aerial view eastward from above Fire Island Inlet showing curved spits at west end of Fire Island. Atlantic Ocean is at upper right; east-west segment of Fire Island Inlet, at left; and south shore of Jones Beach barrier, at upper left. (Photo by Fairchild Aerial Surveys, date not known, but must have been in the mid-1930s, prior to the great 1938 hurricane and also prior to construction of the Federal Jetty at Democrat Point in 1939-40; published in A. K. Lobeck, 1939, p. 334.).

Two byproducts of Kumar's project were that JES began to realize the important connection between the behavior of the open-ocean beach at Fire Island and the behavior of spits at Democrat Point and the late Imre Baumgaertner began to study the mechanisms by which beaches are cut back during storms. Baumgaertner (1975, 1977 ms.) proposed the concept that beach erosion takes place chiefly by the process of grazing-swash undercutting (Sanders and Baumgaertner, 1977 ms.).

Although JES began the inlet study firmly convinced that everything that needed to be known about beaches had already been established by previous workers, especially in light of the many research projects that had been supported by the Corps of Engineers through their Coastal Engineering Research Center (CERC), many emphasizing the importance of "wave energy" and the behavior of waves. After a few years, JES decided that important points were being overlooked because the basic definition of "beach" was too restrictive and that in determining what happened to a beach, "wave energy" was not the most-important variable. The definition of a beach used by the coastal engineers was the one proposed in 1933 by the organization ancestral to the Coastal Engineering Research Center, namely the U. S. Army Corps of Engineers Beach Erosion Board (BEB). The BEB defined the outer limit of a beach at mean low water. As JES began to see more and more effects brought about by the flow of coast-parallel currents of water, he began to think in terms of an expanded definition of beach to include all the coastal strip subjected to the effects of breaking waves. The importance of the connection between the beach and the part of the nearshore bottom, known as the shoreface, where the waves undergo shoaling transformations began to be emphasized by several workers. In 1978, JES presented a paper in which he defined 3 dynamic zones on a beach based on tidal-reference levels. Later JES (1980) argued that the nature of the post-storm recovery of the beach at Westhampton proved that sand was being transported landward from the shoreface and onto the beach through the breaker zone.

Audrey Massa (1980, 1981, 1982, 1988 ms.) began to study the build up of sand on the east side of the Federal Jetty on spits. (For details of her study of the spits at Democrat Point, see the guidebook for On-The-Rocks Trip 07, Sedimentology of Robert Moses State Park, 17 September 1989.)

The studies of Fire Island and the offshore cores yielded evidence related to the origin of barrier islands (Sanders and Kumar, 1975a) and on another mechanism for origin of elongate coastal sand bodies known as "shoestring sands" that became famous among petroleum geologists as a result of the large numbers of oil pools found in the Pennsylvanian-age strata in central Kansas (Sanders and Kumar 1975b).

Further particulars on the subsurface relationships beneath the Jones Beach barrier were found by Rampino (1978 ms.). The relationships found in the boreholes made in connection with the construction of sewage-treatment plants and outfall pipes leading from these plants into the open ocean indicated two units of outwash sand separated by a marginal-marine deposit named the Wantagh Formation (equivalent to the "20-foot clay" of the U. S. Geological Survey informal usage; Rampino and Sanders, 1980, 1981a, b).

What we present in this guidebook includes driving directions from the Academy to Robert Moses State Park, a brief summary of the geologic features of the territory that we shall drive across en route, and detailed descriptions of the beach that we shall investigate today.

DRIVING DIRECTIONS

Turn L on 5th Avenue, drive S to 34th Street, turn L and use Midtown Tunnel and Long Island Expressway. Leave LIE at exit for Northern State Parkway and Meadowbrook Parkway, in direction of Jones Beach. After entrance gate to Jones Beach State Park, follow Ocean Parkway E to Robert Moses Causeway. Enter Causeway S bound, cross the bridge and turn R at the traffic circle, following sign for Parking Field 3. Drive into Parking Field 3 and park as close as possible to the east end of the parking lot. After lunch, we shall re-board the vans and drive to Parking Field 2. To do this, we may have to return to the traffic circle and go all the way around to make a U turn. In any case, to get to Parking Field 2, it is necessary to proceed all the way to end of the divided highway and there make a U turn. After the U turn, drive E for about 1/4 mile, turn R into first entranceway into Parking Field 2. At the circle, turn R and drive as far west as is possible; find a parking space. Leave vans and walk to beach via path from walkway to clubhouse for miniature golf course.

GEOLOGIC RELATIONSHIPS ON TRIP ROUTE (MANHATTAN, LONG ISLAND, AND MODERN OPEN-OCEAN COAST ON SOUTH SHORE OF LONG ISLAND)

In driving from Manhattan to Robert Moses State Park, we shall cross over (but rarely see) many geologic formations (Table 1 is a summary of the geologic time scale with major subdivisions listed) that we shall generalize into three "layers" (Table 2). From oldest to youngest: Layers I and II together, that we shall consider as forming one, the "basement" complex of ancient metamorphic rocks (ages greater than 350 million years); Layer VI, coastal-plain strata (sediments deposited on the passive eastern margin of the North American plate, which formed when the Atlantic Ocean opened, late in the Jurassic Period, 180 million years ago and still exists today and continues to subside while the Atlantic Ocean continues to widen); and Layer VII, Quaternary sediments that we shall subdivide into Layer VIIB, upper Pleistocene deposits (draped over the eroded edge of the elevated and tilted coastal-plain strata; their ages are not precisely known, but the tried-and-true geologic method of "guessing" gives numbers in the range of 70 thousand to 40 thousand years); and Layer VIIA, the modern (= Holocene; some would say, "obscene") sediments being deposited along the margins of the Atlantic Ocean; these are no older than about 2 thousand years. Although we shall be seeing only sediments of Layer VIIA today, it will be helpful to outline the general relationships among the "layers" that we will drive over and to relate them to the geologic conditions associated with the extraction of ground water on Long Island.

Layers I and II: the "Basement" Complex

The solid bedrock, which is exposed in many parts of Manhattan, underlies the tall buildings, and is encountered in the subway tunnels, ranges in age from about 350 million years to 1,100 million years. Collectively, it forms a "basement" complex on which the coastal-plain and younger sediments accumulated. This complex includes many kinds of metamorphic rocks and has provided the garnets and other dark-colored "heavy" minerals (specific gravity greater than 2.80) that are locally abundant in the beach sands on Fire Island. The "basement" complex extends regionally beneath the coastal-plain cover of Long Island, underlies Long Island Sound, and is exposed at the surface the Western and Eastern Highlands of Connecticut (Figure 3). We have described the rocks of this complex in several of our On-The-Rocks guidebooks (notably Trip 16, Manhattan and The Bronx; Trip 21, Cameron's Line and The Bronx Parks) and thus will not give them more than passing notice here.

The kinds of rocks composing the "basement" complex form only at temperatures and pressures that prevail deep within the Earth (possibly 20 to 30 kilometers). Accordingly, the exposure of such rocks at the Earth's surface means not only that the territory in which they are found sank to great depths for long enough for the metamorphic reactions to take place, but also that it was afterward re-elevated and its former cover eroded (Figure 4). The age of 350 million years marks the last time the rocks of the complex were heated to temperatures great enough to cause their minerals to recrystallize and to drive out the gaseous radioactive-decay products so that their isotopic "clocks" were re-set to zero.

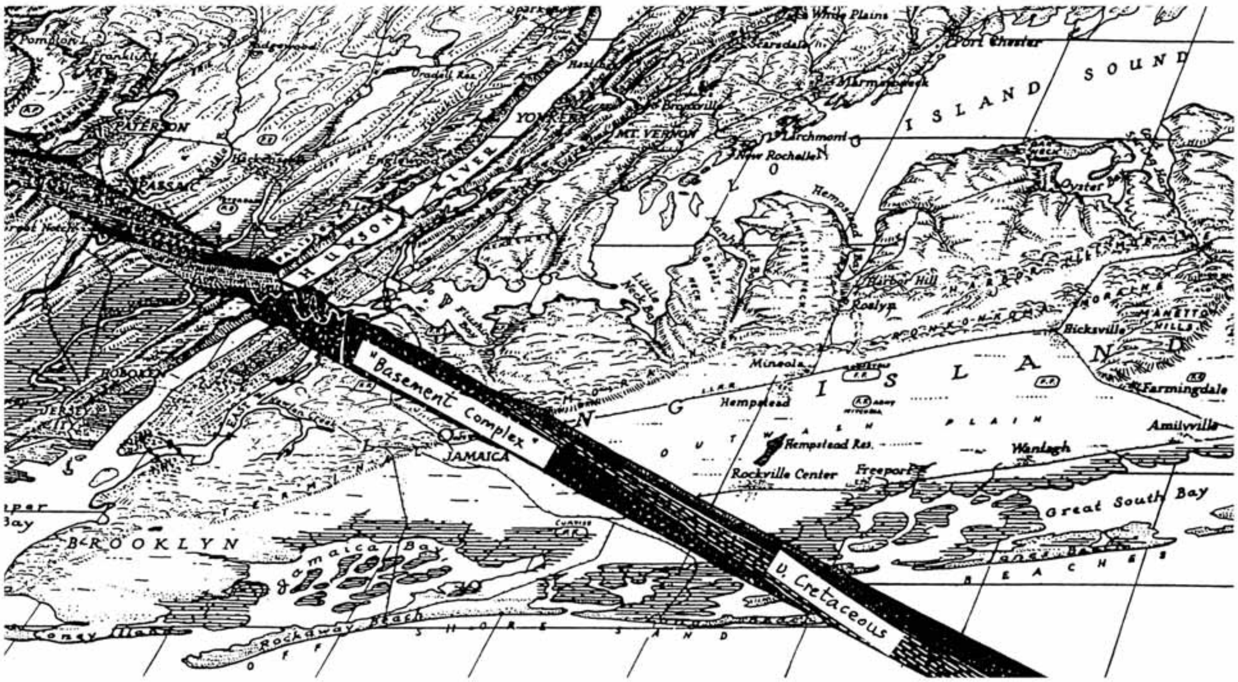


Figure 3. Physiographic block diagram with cutaway face to show geologic structure. (Modified from E. Raisz by showing U. Cretaceous, not Tertiary, beneath the south shore of Long Island.)

The history of ups and downs of the complex (See Figure 4.) involves its first re-appearance at the Earth's surface about 220 million years ago (at the beginning of the episode of subsidence of the Newark Basin); another sinking of part of it equal to the thickness of the Newark strata (possibly as much as 10 kilometers) plus closely coupled zones that were elevated to provide the coarse sediment forming the fill of the Newark Basin; and a post-Newark, pre-coastal plain elevation amounting to whatever was the amount of Newark subsidence plus another amount great enough to cause the formerly horizontal Newark strata to acquire their present-day dips.

After the Newark strata had been deformed and eroded, the tectonic setting of eastern North America changed profoundly. JES thinks that this change was associated with the opening of the Atlantic Ocean and the formation of a subsiding, passive continental margin, the setting that prevails today. By contrast, many geologists believe that the Newark episode coincides with the opening of the Atlantic Ocean. JES thinks that this belief has resulted from repeated assertions in articles published in geologic journals that these two events were connected and written by the "high priests" of plate tectonics whose knowledge of the Newark rocks had not progressed beyond what they learned in geologic "kindergarten." (JES subscribes to many of the points made by the author of that delightful book: "We learned it all in kindergarten;" but as for geology of the Newark strata, "kindergarten" is not good enough.)

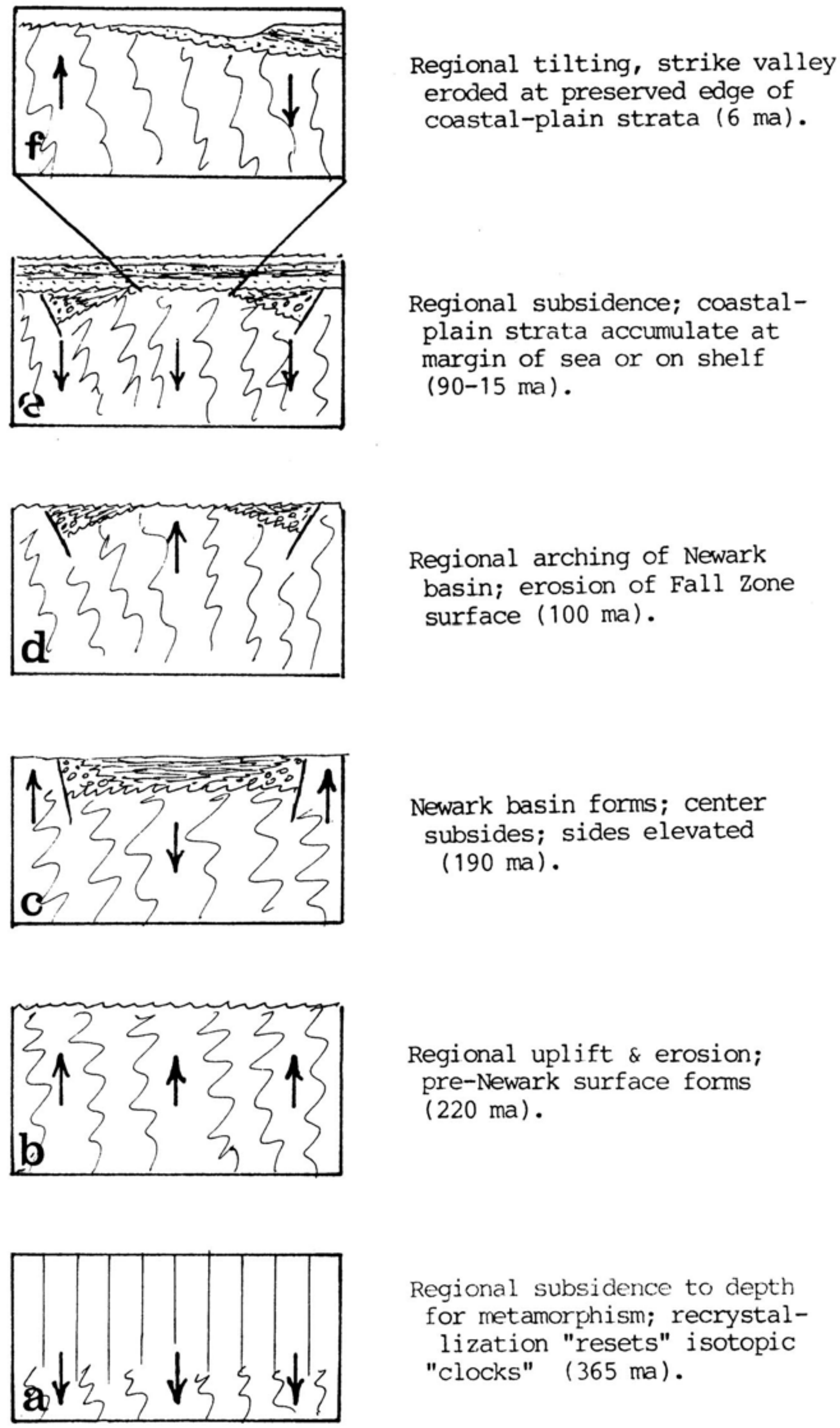


Figure 4. Stages in development of Long Island and vicinity, beginning with the subsidence and metamorphism of mid-Devonian age (a) and ending (f) with regional southward tilting of Layer VI and erosion of inner lowland and inner cuesta of coastal plain (Pliocene?).

After all the post-Newark elevation and great erosion, an erosion surface having low relief (= a peneplain) formed. The surface that formed early in the Cretaceous Period is the Fall Zone surface. (See Flint, 1963, for a description of the Fall Zone surface in Connecticut.) No sooner had the Fall Zone surface formed but another period of subsidence began that enabled the coastal-plain strata to accumulate.

Do Any Parts Of Layer V (Newark Basin-Filling Strata) Exist Beneath Long Island?

An off-again, on-again debate that has aroused the curiosity of no more than an "illustrious few" geologists is whether or not any remnants of Newark basin-filling strata exist beneath Long Island. This idea was first proposed by W. O. Crosby, a Professor of Geology at MIT who served as a consultant to the New York City Board of Water Supply. Crosby (1910 ms.) described the cuttings from a deep water well drilled at Duck Island on the north shore of Long Island (Figure 5). He claimed that the cuttings indicated red Newark-type sandstone beneath the usual Upper Cretaceous coastal-plain sediments (our Layer VI).

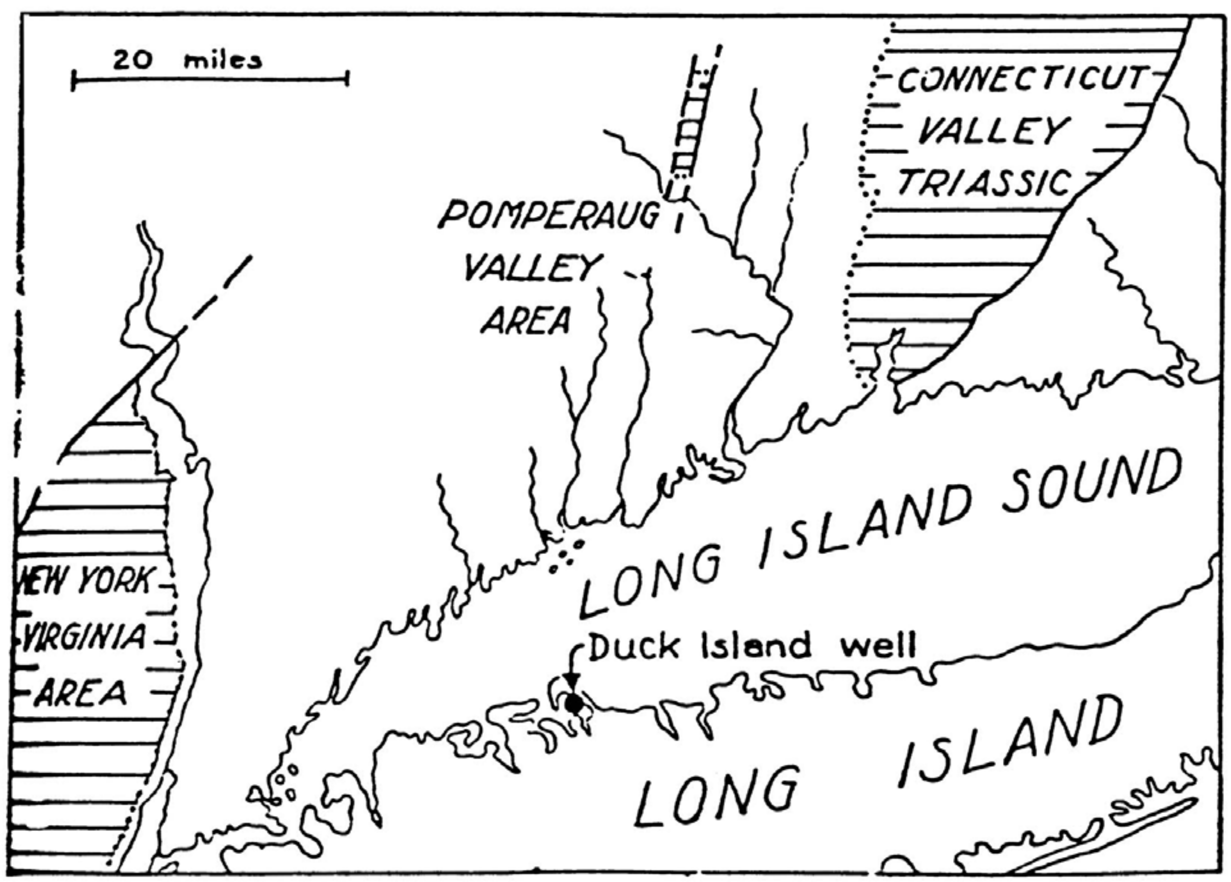


Figure 5. Regional map showing locations of Newark-type basin filling strata (parallel horizontal lines) with respect to location of Duck Island well, northern Long Island. (Girard Wheeler, 1938, p. 141.)

Girard Wheeler (1938) accepted Crosby's conclusions about the Duck Island well and introduced them into the geologic literature in a short paper in which he discussed the implications of Crosby's report on the debate over the former extent of the Newark-type basins (in modern language, the Newark basin of Pennsylvania-New Jersey and the Hartford basin of Connecticut-Massachusetts).

By contrast, the general opinion among the geologists of the U. S. Geological Survey's Long Island office of ground water is one of skepticism about Crosby's report. The ground-water experts doubt that Newark-type strata underlie the Cretaceous on Long Island for two reasons: (1) the coastal-plain Cretaceous contains red sandstones that resemble the typical Newark redbeds (for example, the red sandstones from the Cretaceous found as pebbles on the beach at Garvies Point during On-The-Rocks Trip 17, Glacial geology of Long Island, 17 and 18 November 1990); and (2) hundreds of other wells that penetrate the bottom of the Cretaceous coastal-plain strata encounter rocks of the basement complex. The Geological Survey's geologists scoff at the Crosby report; they think the driller of the Duck Island well was interested in keeping the job going and knew it would stop if he reported basement rocks.

Enough interest was aroused by Wheeler's paper to prompt Oliver and Drake (1951) to locate two seismic lines across Long Island Sound in places that would serve as tests of the idea that the Hartford basin extended southward from New Haven harbor. Subsequently, other continuous seismic profiles made in Long Island Sound have indicated that no Newark-type strata underlie the Upper Cretaceous (Grim, Drake, and Heirtzler, 1970; Lewis and Needell, 1987).

When JES was getting started in his studies of the Newark-age strata filling the Hartford basin in southern Connecticut, he became impressed with the significance of folds having axes normal to the basin-marginal faults. Many of these folds could be seen to involve the underlyling basement rocks. Accordingly, JES proposed the concept that the axis of the transverse Danbury anticline crosses Long Island Sound diagonally in such a way as to end the outcrop of the Hartford basin in New Haven harbor, but could also bring the Newark strata back down again beneath the Duck Island well (Figure 6). Woollard's (1943) regional gravity map shows a belt of negative anomalies that matches the JES projection of Newark strata (Figure 7). In connection with attempts to assess the offshore petroleum situations off the Atlantic coast, the U. S. Geological Survey carried out many geophysical investigations, including continuous seismic-reflection profiles that show Newark-type basins beneath the Upper Cretaceous in New York Bight south of Long Island (Hutchinson and Klitgord, 1988; Hutchinson, Klitgord, and Detrick, 1985; Klitgord and Hutchinson, 1985; Klitgord and Behrendt, 1979; Figure 8). These new geophysical results indicate that several belts of Newark-type basins exist south of Long Island; none of these matches the belt projected across Long Island in the Sanders 1960 paper.

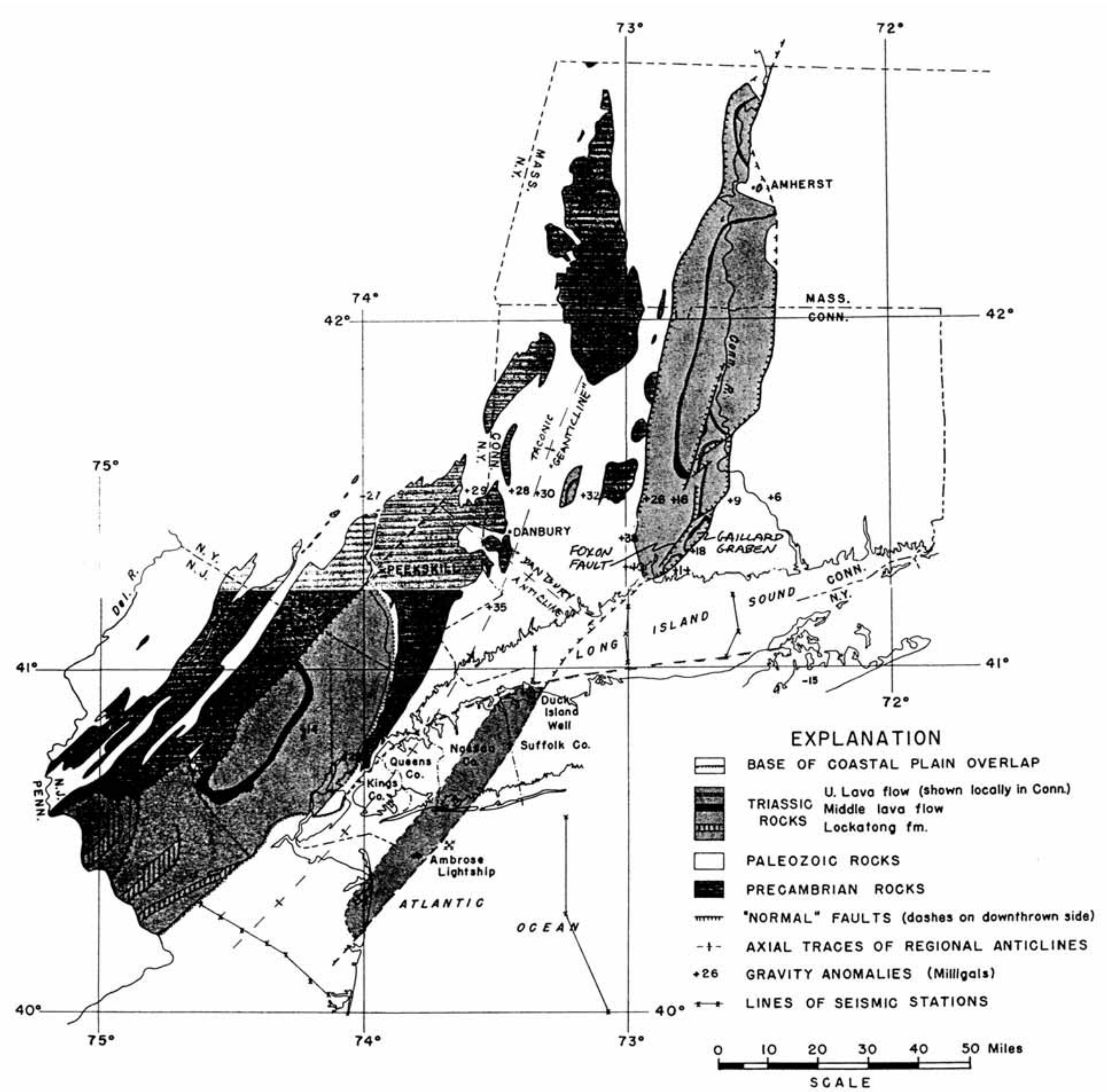


Figure 6. Schematic tectonic map of Upper Triassic-Lower Jurassic basin-filling rocks of Newark basin and Hartford basin, with speculative projection of buried basin beneath western Long Island. Compilation done at time when the entire "New York City Group" was assigned to the Precambrian; hence, the peculiar rendering of southern Westchester County, NY. Projected Newark area beneath western Long Island is separated from the Newark basin by the regional Taconic "geanticline" and from the Hartford basin by the Danbury transverse anticline. (J. E. Sanders, 1960, fig. 1, p. 120-121.)

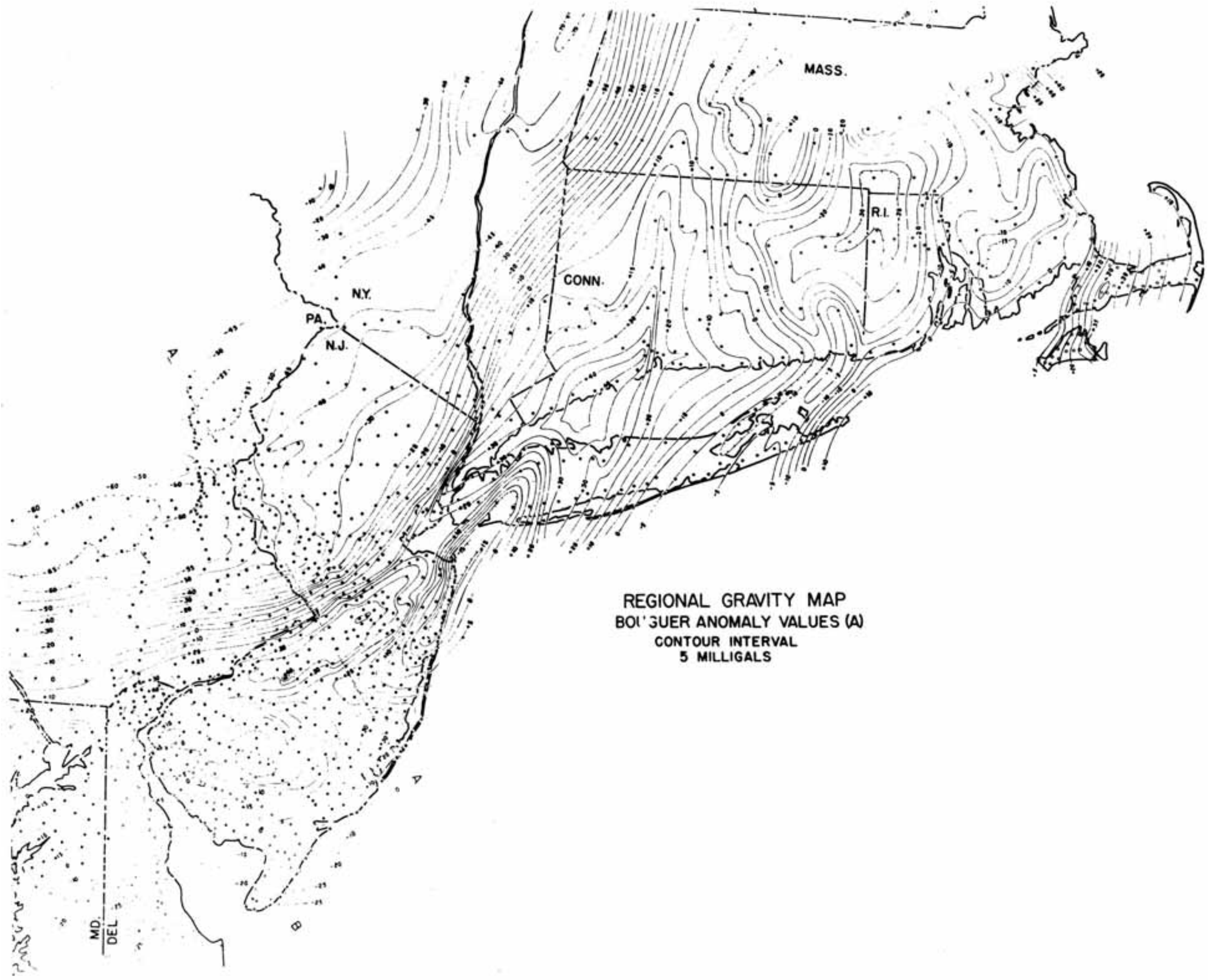


Figure 7. Regional gravity-anomaly map, New Jersey to Massachusetts shows elongate belt of negative anomalies extending beneath western Long Island and ending in the vicinity of Sandy Hook, NJ. Compare with projected buried Newark area shown in Figure 6 (which JES drew before he ever saw this gravity map). (G. P. Woollard, 1943.)

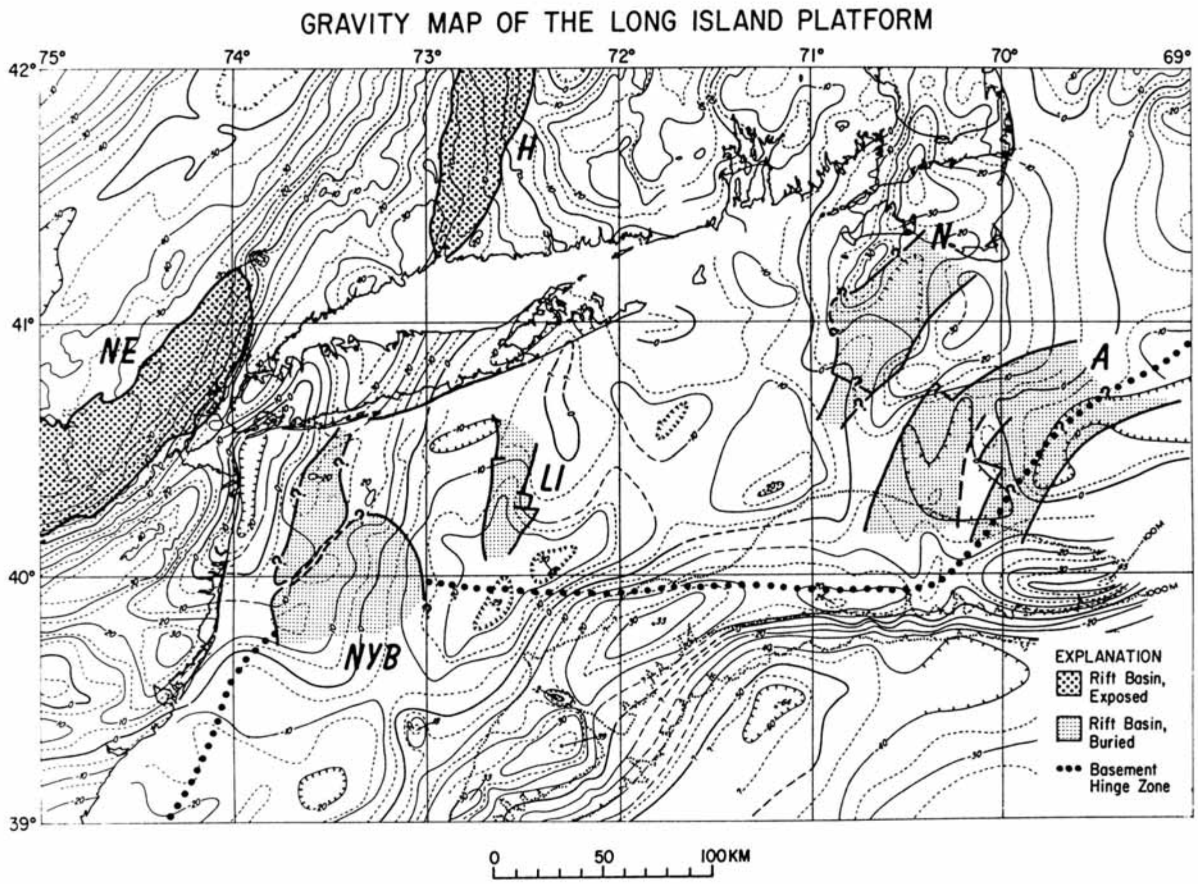


Figure 8. Regional gravity-anomaly map and locations of Newark type basins (2 exposed, NE for Newark basin; H, for Hartford basin; and 4 buried, NYB for New York Bight basin, LI for Long Island basin; N, for Nantucket basin; and A, for Atlantis basin. (D. R. Hutchinson, K. D. Klitgord, and R. S. Detrick, 1986, fig.14, p. 697; compare with Woollard's 1943 map shown in Figure 7.)

Layer VI: Coastal-Plain Strata; Mineralogic Maturity and the Lloyd Aquifer

Whatever is your geologic "religion" about the Newark and the opening of the Atlantic Ocean, no argument surrounds the conclusion that the coastal-plain strata are products of a subsiding, passive continental margin on the trailing edge of the North American plate as it moved away from the spreading center of the Atlantic Ocean, the mid-Atlantic ridge.

A mineralogic characteristic of the local coastal-plain strata is the presence of sand-size minerals such as quartz and zircon that resist chemical decomposition during weathering and a general absence of feldspars and other minerals that are easily decomposed during weathering. Such accumulations of the resistant survival products of weathering are said to be mineralogically mature. In its simplest form, mineralogic maturity or -immaturity is judged on the basis of feldspar in the sand fraction. Mineralogic maturity means a lack of feldspar (where feldspar is present in the bedrock being eroded). The presence of feldspar indicates immaturity. In addition, the clay minerals of the coastal-plain strata are typified by kaolinite, a product of intense chemical weathering of the kind that today is found mostly in the tropical zones.

The coastal-plain strata beneath Long Island consist of a lower nonmarine part and an upper, marine part. A basal sand, the Lloyd Sand, contains large supplies of potable water (= ground water). The recharge area of this sand is Long Island itself. No long-distance subsurface route exists, as it does, for example, at Atlantic City. The deep wells at Atlantic City tap ground water from a formation that is exposed at the land surface in the New Jersey pine barrens. As shown in Figure 9, the Cretaceous strata generally stop along the south side of Long Island Sound. A few remnants are known out in the Sound (example: Stratford Shoal; see continuous seismic-reflection profiles of Tagg and Uchupi, 1967; Grim, Drake, and Heirtzler, 1970), and some Cretaceous strata may be preserved as the fillings of deep valleys (as in the West Haven valley described by JES in 1965; see also Haeni and Sanders, 1974; and Sanders, 1989 ms.; and for eastern Long Island Sound, Lewis and Needell, 1987). The geologic relationships shown by the continuous seismic-reflection profiles absolutely preclude the possibility that any of Long Island's ground water from the Lloyd aquifer comes from a distant source to the north (say, from Vermont, as JES has heard some students in his geology classes, residents of Long Island, assert). As was first pointed out by a French geologist, Jacques Avias, the fact that the ground water in the Lloyd aquifer comes from rain which falls on Long Island means that the overlying Upper Cretaceous clays are not as impermeable as many geologists suppose. Although these clays may be impermeable on the scale of laboratory specimens, on the much-larger scale prevailing in nature, they must be cut by cracks that enable the water to penetrate from the surface to the base of the Cretaceous formations.

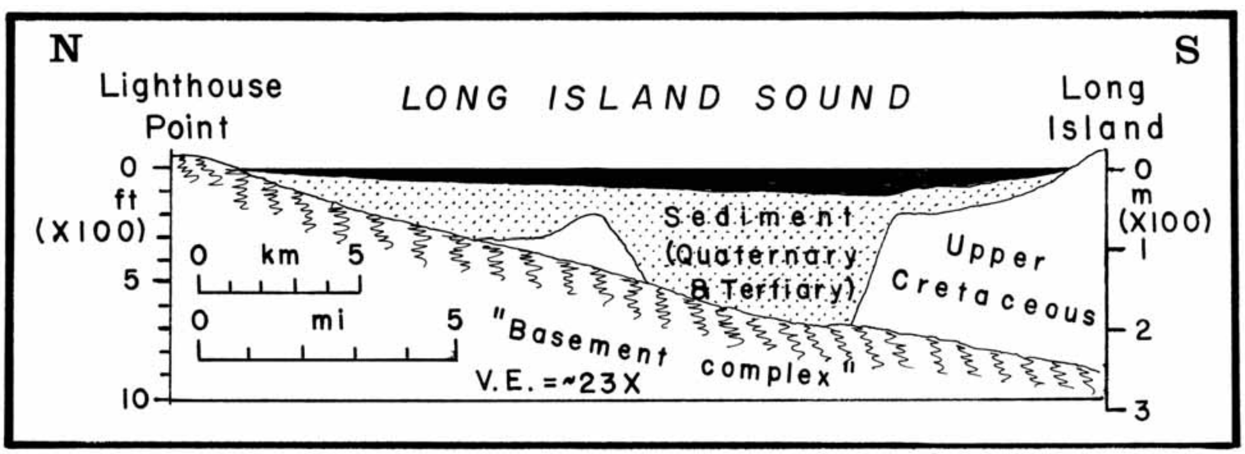


Figure 9. North-south profile-section from Connecticut, across Long Island Sound, Long Island, along longitude 72° 53' West. Water depths from NOS chart. Subbottom relationships from Grim, Drake, and Heirtzler, 1970, Figure 12, p. 661. Water of Long Island Sound shown in black.

The deep wells that supply water to Jones Beach and Robert Moses State Parks (stored in the tall towers in the traffic circles at each park) tap the Lloyd Sand. Immediately beneath the

Pleistocene formations in the park wells are Upper Cretaceous marine formations (Perlmutter and Todd, 1965). No Tertiary formations are present (as incorrectly shown on the original version of the Raisz block diagram).

The episode of coastal-plain sediment accumulation that began late in the Cretaceous Period (about 95 million years ago) continued until the Miocene Epoch (about 25 million years ago). From the record in New Jersey, we can infer that a series of fans built southeastward from the rising Appalachians. These fans pushed the shoreline southeastward and ended the episode of accumulation of marine sediments in what is now New Jersey. (However, marine sediments continued to accumulated in areas now submerged.) In the Pliocene Epoch (starting about 6 million years ago), regional uplift, possibly combined with the first of many episodes of eustatic lowering of sea level, enabled deep valleys to be eroded. One such valley, named a strike valley because it is parallel to the strike of the gently tilted coastal-plain strata, formed along the eroded updip edge of the preserved Cretaceous strata. (This feature has also been named the inner-cuesta lowland.) The depression that has been filled with sea water and forms Long Island Sound began its career as such a stream-eroded strike-valley lowland.

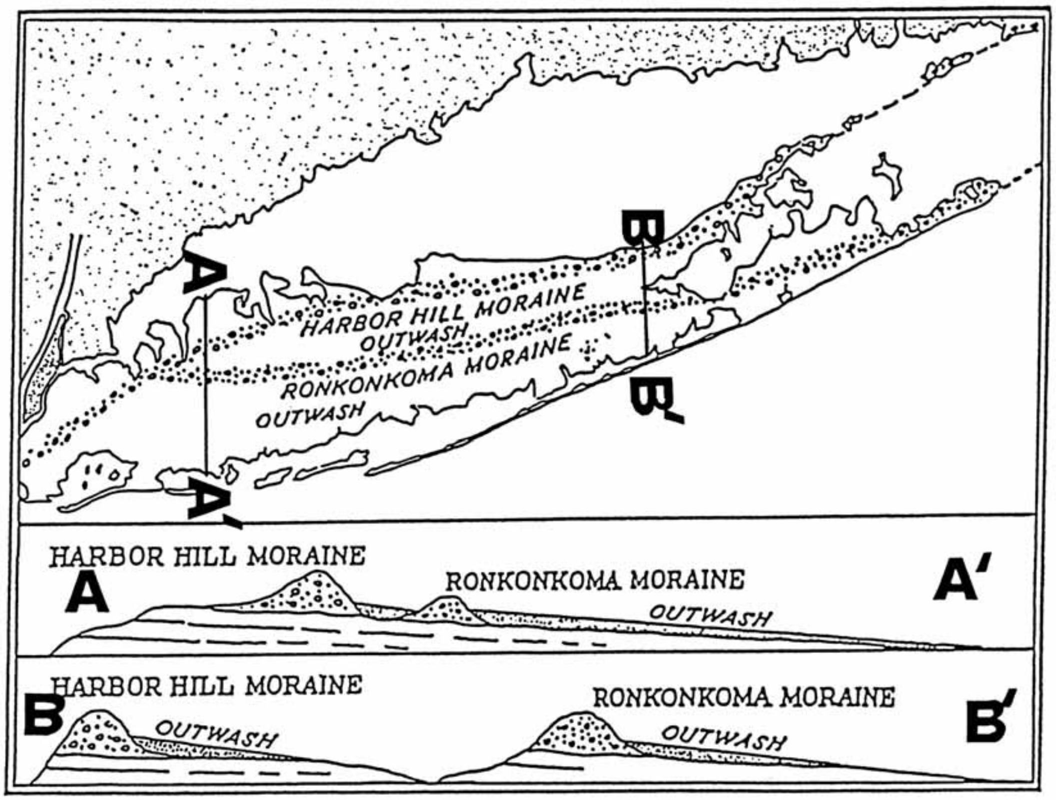
Layer VII: Quaternary Sediments

The Quaternary includes the Pleistocene and the Holocene that may be closely related in time, but include sediments deposited under utterly contrasting conditions. Accordingly, we shall subdivide Layer VII into two parts, Layer VIIB for the Pleistocene glacial deposits and Layer VIIA for the Holocene sediments deposited by the rising sea and along the modern coast.

Layer VIIB: Pleistocene Deposits: Terminal Moraines and Outwash; the "Glacial" Aquifer

Draping over the eroded edge of the Cretaceous strata are varying thickness of Pleistocene sediments deposited close to the margin of a continental glacier. At the margins of a continental glacier, the usual pattern of sediments includes two contrasting kinds. (1) The material deposited directly by the flowing ice of the glacier is till, a heterogeneous mixture of debris ranging in size from boulders to clay. Till usually lacks internal layers (strata) and if it contains abundant clay-size particles in its matrix, is impermeable (thus is not only not an aquifer, but may drain water so slowly that low places in its surface are characterized by swamps). (2) The sediment that was deposited by water from the melting glacier is known as outwash. Much outwash is deposited by braided streams that build fans adjacent to the terminus of the glacier. The top surface of these fans is sometimes referred to as an outwash plain.

The most-prominent features of the Long Island landscape are the terminal-moraine ridges that extend nearly the entire length of the island and form the "backbones" of the two "forks" of the eastern end of Long Island. The ridge forming the "backbone" of the South Fork is the Ronkonkama Moraine and that of the North Fork is the Harbor Hill Moraine (Figure 10). At Lake Success, the Harbor Hill Moraine truncates the Ronkonkama Moraine and continues westward through Queens and Brooklyn; it crosses New York Harbor at the Narrows and extends along the shore of Staten Island. Each terminal-moraine ridge is associated with an adjacent lowland on the south that is underlain by outwash. Notice that in the two profile-sections, the the symbol used for the moraine differs from that of the outwash. The implication is that the moraine ridges consist of till, in contrast to the meltwater-deposited outwash. According to JES observations, much of the material in the moraine ridges closely resembles outwash. On Long Island, till is rare. It is present at the extreme western and eastern ends if the island, but is hard to find in between.



MORAINES AND OUTWASH PLAINS ON LONG ISLAND

Figure 10. Map of Long Island showing two prominent moraines and associated outwash plains, with schematic N-S profile-sections along lines shown. (After A. K. Lobeck; lines of profile-sections added by J. E. Sanders, 1985.)

As we explained in our On-The-Rocks Guidebook for Trip 15 on the glacial geology of Long Island, the prevailing opinion among Pleistocene geologists is that Long Island's two terminal-moraine ridges were deposited by the latest glacier, now known by the name of Woodfordian, which retreated from the region about 13,000 years ago. [According to the JES-CM reinterpretation of the local Pleistocene stratigraphy, these two moraine ridges were not built by the very latest Wisconsinan continental glacier (the Woodfordian), as is generally supposed, but by two older glaciers, each of which flowed regionally from NNW to SSE. JES and CM support the older work of J. B. Woodworth (1901), who found that in western Queens, the Harbor Hill moraine consists of red-brown till resting on striated bedrock showing grooves oriented NW-SE. JES and CM contend that the terminal moraine deposited by the Woodfordian glacier, which flowed from the NNE to the SSW, is not present on much of Long Island, but extends along the south coast of Connecticut (Flint and Gebert, 1974, 1976).]

Another noteworthy feature of many Upper Pleistocene outwash sediments is their compositional maturity (lack of feldspar and other minerals that decompose readily during weathering). Because the chief activity of a continental glacier is to grind up whatever materials it passes over and not to decompose them, mineralogically mature outwash must mean that the advancing continental glacier was traveling over materials which generally lack feldspar, such as the Upper Cretaceous sediments. Where the continental glacier flowed directly over the pre-Cretaceous "basement" complex, its sediments are mineralogically immature.

JES thinks that both the foregoing characteristics of the Upper Pleistocene sediments underlying much of Long Island can be explained by inferring that most of the time, the terminus of the glacier was located in the lowland that is now Long Island Sound and that the ice flowed over the Lloyd Sand as shown in Figure 11. Eventually, the glacier must have stripped away all of the Lloyd Sand so that it began to grind up the abundantly feldspathic rocks of the "basement" complex and thus to deposit mineralogically immature sediments. (If this JES idea is correct, then the first appearance of mineralogically immature Pleistocene sediments should be a useful marker.) The two moraine ridges may have been built when the ice surged forward and pushed the adjacent outwash into linear ridges.

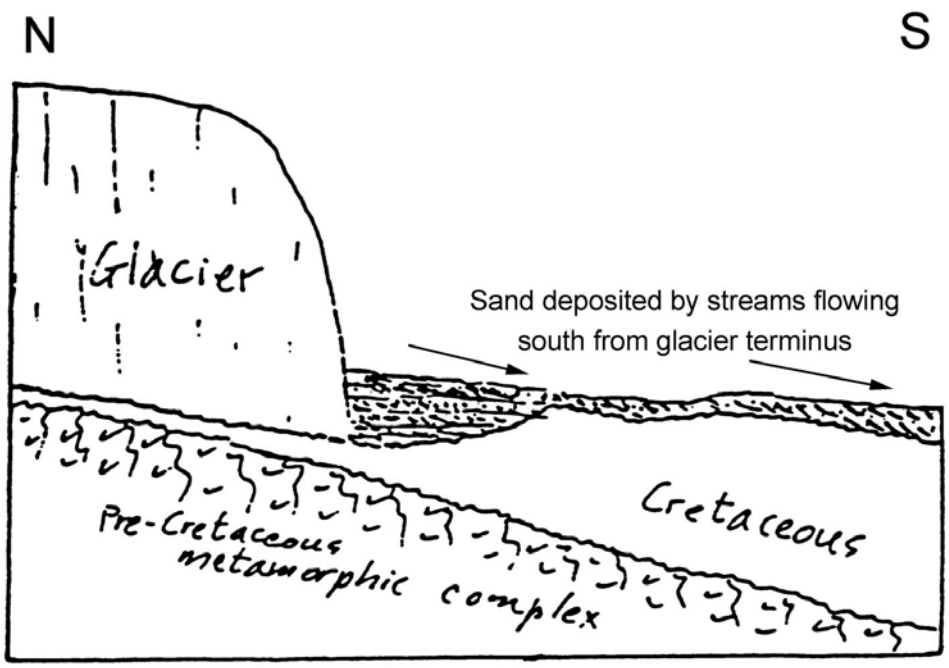


Figure 11. Restored profile-section from Connecticut to Long Island showing terminus of continental glacier standing in what is now Long Island Sound and spreading mineralogically mature outwash sand and gravel southward to bury the Upper Cretaceous strata of Long Island. Extension of Cretaceous beneath glacier is schematic, but is based on the lack of feldspar in much of the Long Island outwash. (Drawn by J. E. Sanders in 1985 using same regional relationships shown in Figure 4, from W. deLaguna.)

That much of Long Island is underlain by outwash is confirmed by the almost-universal presence of the so-called "glacial" aquifer, the name given by the geologists of the Ground-Water Branch of the U. S. Geological Survey to the shallow blanket of sand from which most domestic wells on Long Island derive their water.

Rampino's subsurface results along the sewer-outfall pipes across the Jones Beach barrier (Figure 12) indicate that two outwash sheets are separated by the Wantagh Formation (the 20-foot clay of the U. S. Geological Survey). If each of these two units of outwash sand correlate with the two major Long Island moraines, then they add another argument to the proposition that these moraines were not deposited by a single glacier and that the older of the two antedates a high stand of sea level. Given the information presented by Howard Ricketts (1986) that the Gardiners Clay is a deposit of the Yarmouth Interglacial Epoch ca. 200ka (as first correlated by M. L. Fuller in 1914) and not of the Sangamon Interglacial at ca. 100 ka (as asserted by MacClintock and Richards in 1936 and accepted by the multitude), then a probable correlation for the Wantagh Formation is Sangamonan. We have assembled the proposed new correlations of the Long Island Pleistocene units in Table 3.

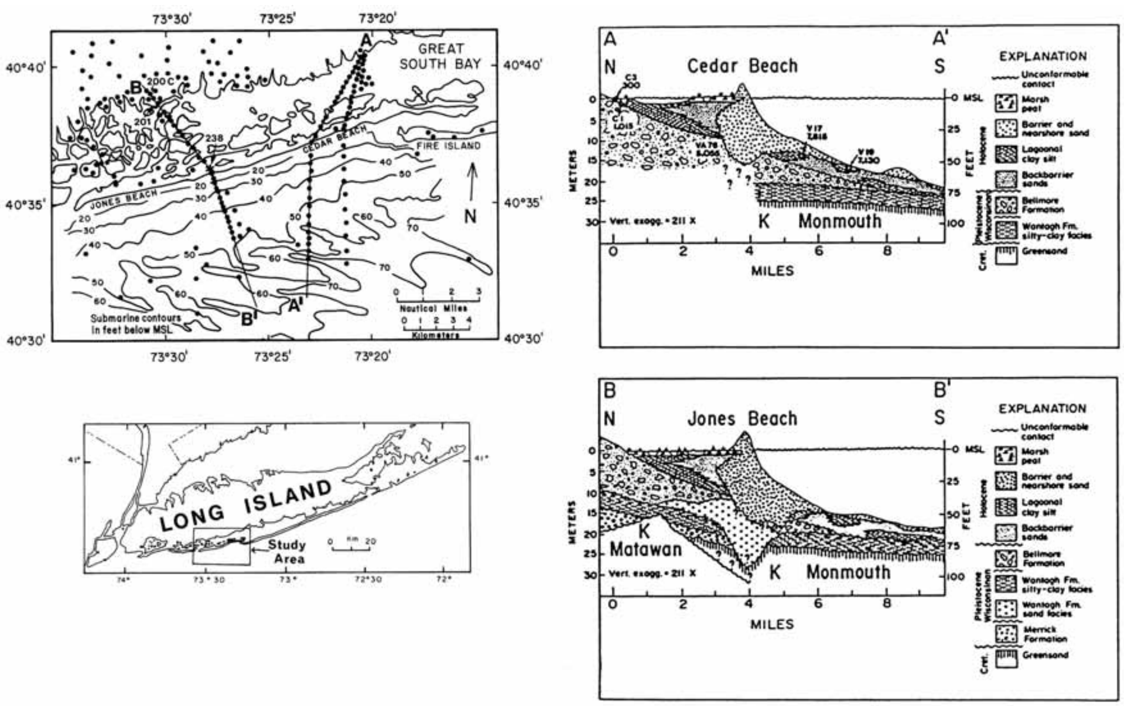


Figure 12. Index maps and profile-sections along effluent pipes from sewage-treatment plants across Jones Beach barrier at Jones Beach (Nassau Co. plant, BB') and Cedar Beach (Suffolk Co. plant, AA') showing stratigraphic relationships of Holocene sediments deposited during the postglacial submergence and the interbedding of the two units of outwash (Merrick and Bellmore formations) with marginal-marine Wantagh Formation on the Jones Beach profile-section (BB'). Notice also on AA' the downward-projecting base of the Cedar Beach barrier, a "fossil" inlet-filling sand body, and on BB', the existence of a buried Pleistocene barrier island with downward-projecting inlet-filling sand body beneath Jones Beach (M. R. Rampino and J. E. Sanders, 1980.)

Layer VIIA: Holocene Sediments: Barrier Islands and Intertidal Salt Marshes Fringing Great South Bay

The south coast of Long Island is a place where waves and tides of the transgressing sea are reworking the Upper Pleistocene sediments (cliffs composed of till near Montauk, outwash sand and gravel in most other localities). No rivers deliver new sediment eroded from inland. The sediment present is that which happens to be available from past geologic activities.

The waves have built narrow, linear sandy islands known as barrier islands. Between this string of islands and the main part of Long Island is a "lagoon," the largest of which is Great South Bay. Over much of Great South Bay, tidal action brings in silt, which builds up mudflats that are soon colonized by salt-tolerant grasses to build intertidal salt marshes whose top surface approximates the level of mean high water. Once the marsh grasses have become established, they are able to survive further submergence at rates of up to 4 millimeters per year. They do this by trapping silt among their stalks and by growing upward. As they grow upward, they do two important things: (1) they spread landward, and hence can overlie any pre-existing material, including granitic bedrock; and (2) they build an ever-thickening layer of marsh peat beneath themselves. Initially, salt-tolerant grasses overlie intertidal mudflats and the thickness of the marsh peat equals half the mean tidal range. Thus, on a coast where the mean tidal range is 2 meters, the thickness of marsh peat is 1 meter. Marsh peat thicker than half the mean tidal range or marsh peat overlying non-tidal sediments indicates submergence (Sanders and Ellis, 1961). The intertidal salt marshes of Great South Bay are young; no older than 2 thousand years (Rampino, 1978 ms.; Rampino and Sanders, 1981b). This age coincides with that determined for the adjacent barrier islands (Kumar, 1973; Sanders and Kumar, 1975a, b; Rampino, 1978 ms.; Rampino and Sanders, 1981a, c).

The tidal oscillations cause water to enter and to leave Great South Bay through narrow inlets within which swift currents flow. At the ends of the inlets are tidal deltas, composed of sand that has been transported back and forth in the deep parts of the inlets by the tidal currents but eventually is deposited where these currents spread out as thin sheets. A flood-tidal delta is in a lagoon; an ebb-tidal delta, in the open sea.

OBJECTIVES

1) To discuss the topics: "What is a beach?" and what are smaller "Sons of Beaches".

2) To compare a narrow beach subject to chronic erosion with a wide beach subject to deposition.

1) To become familiar with the three sediment populations found at RMSP:
   a. well-sorted white medium sand;
   b. well-sorted dark-colored, esp. dark reddish, medium sand; and,
   c. poorly sorted coarse brown sand, gravel, and shell debris.

4) To recognize the various parts of an ocean beach, including the shore-parallel ridges of sand (are they dunes?), berm, beach face (and/or beach scarp), and the three morphodynamic zones of an ocean beach: supratidal, intertidal, and subtidal.

5) To study the relationship between deposition of new layers of sediment and sediment surfaces, including both small-scale bed forms and large-scale depositional "slopes;" and to recognize plane-, parallel-, and cross strata.

6) The understand the general geologic relationships of Long Island and the occurrence of ground water in various geologic units.

7) To be duly impressed by the evidence for the rapid westward growth of the west end of Fire Island as a result of inlet migration (average rate of 1 meter per week in the interval 1834-1940).

7) To realize how the effects of the operation of the geologic cycle through time create a geologic record of sediments and of sedimentary rock.

PLAN FOR THE DAY: (To have a nice day at the beach!)

1. Study a narrow beach and adjacent eroded scarp near the picnic pavilion at Parking Field 3; sketch profile and dig SST ("Sanders Scientific Trench"). An SST differs from any old trench in that one side is carefully chosen so that the relationship between the sediment surface and the vertical trench wall will be in the best light for photography after the digging has been completed. ALL SAND IS SHOVELED TO THE OPPOSITE SIDE. NO EXCEPTIONS WILL BE TOLERATED!

We will post guards to keep curious onlookers from wandering up to look from the pristine side.

2. Study a wide beach near W end of Parking Field 2 and select a site near each berm crest for digging an SST.

One important geologic technique we need to use is a system for locating ourselves on a map. JES prefers to use the UTM metric grid system. (UTM means Universal Transverse Mercator; it refers to the kind of map projection used.) In using the UTM grid coordinates, one specifies first the east coordinate, and then the north. (Remember this by the phrase: "Read right up;" right, for going E; up, for going N.)

As this field trip was to a single locality, detailed descriptions are in the guidebook. [UTM grid coordinates of the beach located just beyond the picnic pavilion at Parking Field 3, the starting point of our westward traverse to the jetty, is 646.85E / 4498.05N, Bay Shore West quadrangle.]

JES has added the UTM grid lines to a Park Commission map of RMSP (Figure 13). We will use this map and the UTM coordinates to establish other locations throughout the day.

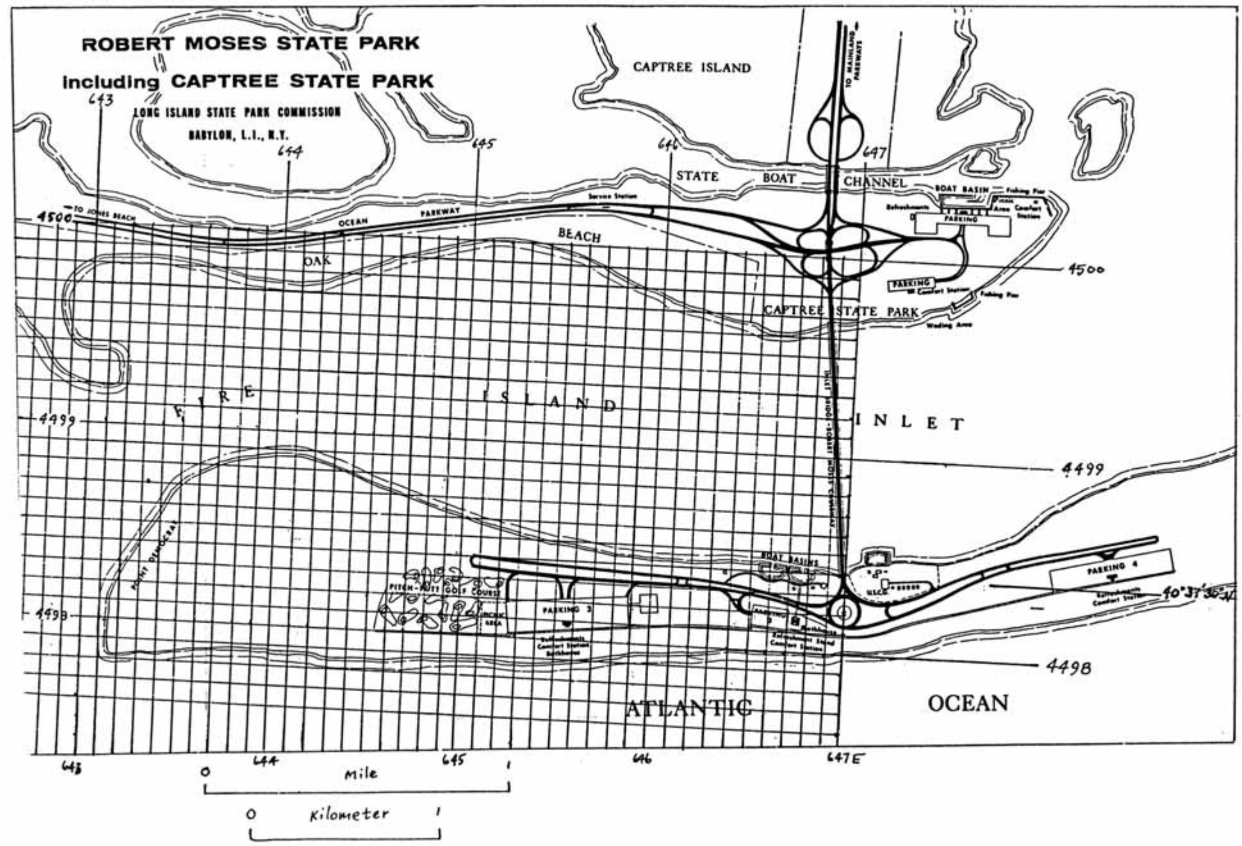


Figure 13. Map of Robert Moses State Park and vicinity. (Long Island State Park Commission map, with UTM metric grid coordinates added by JES, 1985.)

GENERAL REMARKS ABOUT BEACHES AND ASSOCIATED MORPHODYNAMIC ZONES

Most students of beaches take it for granted that everyone knows what a beach is and no formal definition is needed. As we shall see, several definitions have been proposed. In the following paragraphs, I review some of these definitions of a beach and some important beach-associated morphodynamic zones.

What Is A Beach?

As a first approximation toward a definition of beach, one might say "a body of sand washed by waves." This statement contains two elements: (a) kind of material (i. e., sand), and (b) dynamic process (i. e., washing by waves).

The "standard" definition of a beach used in the USA resulted from the work of the Beach Erosion Board of the U. S. Army Corps of Engineers in 1933. Shortly after the U. S. Congress had established the Beach Erosion Board (abbreviated BEB; subsequently the BEB became the Coastal Engineering Research Center, abbreviated CERC) as a part of the Army Corps of Engineers and had charged them with the responsibility for defending the nation's coasts against erosion, the engineers set about defining their "turf," so to speak. After all, if they were to function as the Beach Erosion Board, they had to spell out what they considered to be a beach. JES thinks that the BEB decided it did not want its engineers to get their feet wet; for whatever reasons, BEB defined a beach as including certain bodies of coastal sediment that extend seaward only to the line of mean low water.

The BEB's definition is probably the most widely cited \one, as exemplified by Wiegel's (1953) glossary of waves, tides, currents, and beaches. In this widely used book, Wiegel gave two definitions of beach:

> (1) "Commonly, the zone (SHORE) of unconsolidated material which (sic) extends inland from the waterline to the place where there is a marked change in material or physiographic form, or to the line of permanent vegetation (usually the effective limit of normal storm waves). The seaward limit of the beach, unless otherwise specified is the LOW WATER (sic) DATUM. A beach includes FORESHORE AND BACKSHORE." (This definition concludes with a reference to a figure showing the parts of a beach.)
>
> (2) "Sometimes, the material which is in more or less active transit alongshore or on-and-off shore, rather than the zone involved."

The BEB definition served a very useful purpose for the Corps of Engineers: it set the outer limit of a beach at the low-water datum. (This definition's "hidden agenda" amounted to a recognition that what's in the water falls under the jurisdiction of the U. S. Navy.) Most writers of geology textbooks generally follow the BEB definition. For example, Plummer and McGeary (1991, p. 308) define a beach as: "a strip of sediment (usually sand or gravel) that extends from the low-water line inland to a cliff or a zone of permanent vegetation."

Totally ignored in the BEB's engineering definition of beach was any consideration of the classic work on the topographic features of lake shores written by one of the greatest American geologists of all time, G. K. Gilbert. Gilbert's treatise on shore features (1885) begins with a discussion of wave erosion and then takes up "littoral transportation." Gilbert ended his section entitled "littoral transportation" by discussing the sediment sorting caused by waves. He emphasized the fundamental principle of separation of the fine particles from the coarse particles. Because of its large settling speed, a coarse particle "cannot be carried beyond the zone of agitation, and remains a part of the shore." Only these coarse particles are "the subject of littoral transportation," a process he called shore drift (Gilbert, 1885, p. 86).

According to Gilbert (1885, p. 87):

> "The zone occupied by the shore drift in transit
> is the beach. Its lower margin is beneath the water,
> a little beyond the line where the great storm waves
> break. Its upper margin is usually a few feet above
> the level of still water"

Another American geologist, W. O. Thompson (1937) defined a beach as a "deposit that rests on the shore;" by shore he meant the "zone over which the water line, the line of contact between land and sea, migrates" (Thompson, 1937, p. 725). Thompson's definition incorporates the concept that a beach is a deposit, but places rather vague limits on its width ("the zone across which the waterline migrates"). One can infer that Thompson's definition includes both the littoral zone (between the tide levels) and the zone washed by the waves.

Although he adopted in a general way the usage of Wiegel (1953), Ingle (1966) commented on the lack of standardization of terms for the parts of beaches. Accordingly, Ingle proposed his own subdivisions, as follows:

> "The three principal dynamic zones of the beach
> are the swash zone, the surf zone, and the breaker
> zone" (Ingle, 1966, p. 11).

Ingle admitted a possible fourth zone, a place of transition where the backwash collides with water from the inner part of the surf zone, which he defined as "the area between the effective seaward limit of backwash (swash zone) and the breaker zone" (p. 11).

By including the breaker zone as part of the beach, Ingle implied that the seaward limit of the beach is the seaward limit of the breaker zone. In this sense, Ingle sided with G. K. Gilbert (whom he, too, did not mention) and departed from the BEB-type definition that Ingle otherwise generally accepted.

After sifting through dozens of definitions (details omitted here), JES concludes that most agree that the landward limit of a beach should be placed at any point where the kind of material changes significantly (from well-sorted, cohesionless sediment to something else), or, in the absence of such a pronounced change of material, at the upper limit of storm-wave action. On the seaward (or lakeward) side, no agreement exists. Proposals range from the low-tide datum (of the BEB) to the depth at which waves stir the bottom sediments (which could extend to the deep-sea floor, depending on the effects of tsunami).

JES prefers to restrict beaches to the zone affected by breaking waves. This includes direct breaking; river-like currents of water circulating as a result of influxes of water from breaking waves (as longshore currents in the surf zone, rip currents, swash and backwash, flow of thin sheets of water that spill over the crest of the berm, currents created by the nourishment and draining of berm-top spillover pools, and through-flowing washovers that breach the shore-parallel ridges); and the shifting of the breaker zone as a result of tides. Accordingly, JES prefers the following definition of a beach as:

> A continuous body of well-sorted, coarse-textured (>0.0625 mm), cohesionless sediment subject to the effects of breaking waves, and extending along the shore of a lake or the sea between the outermost line of breakers (if the appropriate sediment is continuous under the water surface; otherwise to the outer limit of the specified sediment if such a change takes place inshore of the outer line of breakers) and the landward limit of wave action on the continuous body of cohesionless sediments or to the contact of wave-washed cohesionless sediments and other materials (if such contact lies within the zone subjected to the effect of breaking waves).

As thus defined, a beach includes all the breaker bars and troughs between them; the surf zone; the rip-current channels; the zone of swash and backwash; the zone between tidal limits; the berm and associated spillover pools (of Rosalsky, 1949), plus spillover deltas deposited in these pools as well as channel deposits associated with flow of water between and out of these pools; and deposits made in breakthrough channels and other gaps in the shore-parallel ridges as well as any associated washover fans (Price, 1947).

According to this definition, a muddy shore is not a beach, nor is a rocky shore only partly covered by cohesionless sediments.

One of the most-important implications of acceptance of the definition of beach advocated by JES is a drastic expansion of ideas about what constitutes "beach sediment." The geologic literature abounds with studies of "beaches" wherein the only samples collected were carefully selected from mid-tide level on the beachface (in the zone of swash and backwash). Ignored in such studies are all the channels, the trough-cross-stratified sediments, and the other irregular features made by thin, fast-moving sheets of water within the various various morphodynamic zones next outlined.

Beach-Associated Morphodynamic Zones

However they define a beach, all students of coastal processes recognize that the coastal sediments and the water interact in several distinct zones. Because these zones are characterized not only by distinctive water motions, but also by interactions between the moving water and underlying sediment, that create characteristic morphologic features composed of sediment, the zones are named morphodynamic zones. With respect to tide levels, three critical morphodynamic zones are: (1) subtidal (always submerged); (2) intertidal (alternately wet and dry, twice a day hereabouts); and (3) supratidal (mostly dry, but wet during higher-than-normal tides or when particularly large swells are breaking). Some of the distinctive water motions within these morphodynamic zones change as tidal levels oscillate up and down. The following paragraphs summarize what happens in these three zones.

SUBTIDAL ZONE: WAVE BEHAVIOR AND WATER CIRCULATION

In the subtidal zone, three subzones are important: (1a) the subzone within which the waves undergo shoaling transformations; (1b) the subzone in which the waves break; and (1c) the subzone where breaking waves reform and/or are converted to sheets of water (surf) that move toward and/or along shore.

The following paragraphs sketch the changes that waves undergo as they approach the shore, break, and reform and/or are transformed into surf. (Summarized from Friedman and Sanders, 1978; and Freidman, Sanders, and Kopaska-Merkel, 1992.)

Subzone 1(a): Zone of Shoaling Transformations

Between the zone where the waves do not affect the bottom (and thus are named deep-water waves) and the breaker zone, waves interact with the bottom in various ways. These interactions are collectively designated as shoaling transformations.

Initial Wave/bottom Interactions: Wave Base and Shallow-water Waves

In deep water, defined as water deeper than half the wavelength of the waves (L), the speeds (or celerities = C) of the waves are determined solely by their wavelengths; water depth is not a factor. The longer waves travel faster than shorter waves. Wave orbits are circles having diameters at the surface that are equal to wave height (H). These diameters decrease rapidly with depth. At a depth of half the wavelength (L/2), the diameters are only 1/23 their value at the surface (i. e., H/23). The depth L/2 at which the bottom and waves first begin to interact is known as wave base. As waves interact with the bottom, the factor of depth begins to affect wave celerities.

Waves traveling in water shallower than wave base (L/2) are known as shallow-water waves. Farther landward, some shallow-water waves break; others become very-shallow-water waves, as is explained farther along.

Between depth L/2 and L/8, changes in the waves are relatively insignificant. The orbits are still circular, wave profiles closely resemble those of deep-water waves, and wave celerities are about the same as they are in deep water.

Zone of Significant Changes; Transformation Base and Outer Zone of Wave Refraction

In water shallower than L/8, however, waves and water motions change significantly. Circular orbits become ellipses, wave profiles change (crests become steeper and troughs wider and flatter than in deep water), and wave celerities decrease. Such transformed waves display the typical properties that are used to define shallow-water waves.

Because L/8 is the depth at which shoaling transformations become noticeable, JES named this depth as transformation base. At the bottom, the back-and-forth oscillations of the water as it moves around in "orbits" that are much-flattened ellipses become strong enough to form small-scale ripples where the water/sediment interface is underlain by fine sand.

Zone of Very-shallow-water Waves: Inner Zone of Wave Refraction

Before they break, some waves (all the low swells but almost none of the steep sea waves) enter a zone shallower than 1/20th of their wavelengths (L/20), at which point they become very-shallow-water waves. A fundamental principle that applies to all very-shallow-water waves is that their speeds are no longer controlled by their wavelengths but only by the depth of water. The mathematical function is that wave celerity equals the square root of the product of the Earth's gravitational acceleration times the depth of water. The following table illustrates some important changes:

Period of waves	**Speed of wave in deep water**	**Depth of water at wave base (L/2)**	**Depth where waves become very shallow-water waves (L/20)**	**Speed of very-shallow water waves at depth of conversion**
(sec)	**(m/sec)**	**(m)**	**(m)**	**(m/sec)**
8	12	50	5.0	7
5	9	19.5	2.0	5
4	6.6	12.5	1.5	4
3	3.4	7	1.0	3.1

Two implications of the numbers in this table are: (1) that at the depth where they are converted to very-shallow-water waves, the celerities of long-period swells are much reduced from their values in deep water (for the example 8-sec waves, 7 m/sec compared with 12 m/sec) and that as water depth decreases to the other values shown in the table, celerity of such waves decreases even further (now being independent of the sizes of the waves and controlled solely by water depth); and (2) short-period waves will break before they become very-shallow-water waves. As a result, they never enter the inner zone of wave refraction within which swells are refracted so prominently outside the breaker zone.

JES thinks that his recognition of two zones of wave refraction explains many of the contrasting results obtained by students of beaches on the Pacific coast of the USA compared with those obtained by students of beaches on the Atlantic coast of the USA. The fundamental paper on wave refraction by Munk and Traylor (1947) illustrates the behavior of long-period swells crossing a narrow shelf and refracting notably just outside the breaker zone as they become very-shallow-water waves and pass through the inner zone of wave refraction and then break. On the Atlantic coast, as they travel across the wide, shallow continental shelf, many swells are refracted in the outer zone of wave refraction. By the time such refracted swells become very-shallow-water waves and thus enter the inner zone of wave refraction, their crests have already become nearly parallel to the breaker zone.

Zone 1(b): Breaker Zone

The key feature within the subtidal zone that JES uses to define the outer limit of the beach is the low-tide breaker zone. The typical morphologic expression of the breaker zone is a ridge (known as a bar) composed of coarse sediment (sand or even gravel) built by the surges of water associated with each line of breakers (toward the land on the sea side and toward the sea on the land side).

Zone 1(c): Surf Zone

Just landward of each of these breaker-bar ridges is a trough. Generally, the depth of the water in these troughs is great enough to enable water from the breakers to reform as oscillatory waves that continue landward. The innermost trough associated with a breaker bar is the surf zone. Within this zone, waves move both landward and seaward. The landward-moving waves are those that have reformed from the collapse of the water in the breaker zone. The seaward-moving waves are those that form when the swash water "piles up" against the water in the surf zone.

What has been referred to as the surf zone does not fit neatly into the three morphodynamic zones being discussed. Strictly speaking, the surf zone belongs with the subtidal morphodynamic zone; it would not exist without the water. However, the part of a beach that is a surf zone at high tide can become dry at low tide. Therefore, this part of the beach belongs with the intertidal zone. Accordingly, important parts of the activities within the surf zone are discussed in a following section entitled intertidal zone. And, as is explained farther along, what is a surf zone at low tide can become part of the zone of wave transformations at high tide.

Water Circulation In The Subtidal Zone

Oblique approach of waves drives water into the subtidal morphodynamic zone at an angle. As a result, the water tends to "pile up;" it adjusts by flowing parallel to shore. The "piling up" is most pronounced within the troughs landward of the breaker bar(s). At intervals, this water flowing parallel to shore within these troughs, particularly within the surf zone, cuts gaps through the breaker bar(s) and returns seaward as narrow currents named rip currents.

The collective name for coast-parallel transport of sediment in any morphodynamic zone is longshore drift. The shore-parallel transport of sediment in the surf zone contributes significantly to the total longshore drift. Such longshore transport of sediment in the surf zone takes place independently of what may be happening in the other morphodynamic zones. In other words, if one uses a mathematical formula based on wave parameters to predict the amount of longshore sediment transport in the surf zone, then one has a tool that is valid only for predicting longshore sediment transport in the surf zone--period. This statement may sound ridiculous, but it has been made deliberately. Too many attempts have been made, for example, which purport to relate the longshore transport in the surf zone to erosion of the beach face. As we shall see, such erosion takes place in the intertidal zone or in the supratidal zone; it is not closely coupled to the transport in the subtidal zone.

Depending on the range of the tides and slope of the nearshore bottom, the kinds of water circulation just described may be active only at low tide. At high tide, this circulation pattern may not be active; instead, the subzone may become a part of the zone of shoaling waves. As such, at high tide, JES would exclude that part of this subzone from the beach.

INTERTIDAL ZONE

As a first approximation, one may visualize that the intertidal- and supratidal zones are affected by some of the same general kinds of water-circulation processes which operate within the landward part of the subtidal zone: sheets of water from breaking waves flow landward, possibly obliquely, over ridges of sediment and into lower-lying areas to landward; these form currents of water that flow parallel to shore; and at selected points, the water cuts channels through the ridge and within these channels, swift currents flow seaward.

Within the intertidal zone, the key points are how the surf behaves and how it relates to the sheets of water that flow up the beach face (steep part of the beach) (these sheets are known as swash) and those that flow back down the beach face (=backwash); and how the breaker zone migrates during each tidal cycle.

Water Circulation

The kind of circulation in the intertidal zone depends on the state of the tide. At high tide, the circulation is as just described for the subtidal zone. Longshore currents form inside the line of breakers and at intervals, rip currents cut through the breaker bars to return water seaward. Also at high tide, the inner edge of the intertidal circulation zone is the zone of swash and backwash on the beach face (the place where the sheets of water known as swash flow upward and those known as the backwash flow downward). As long as this remarkable zone of swash and backwash is operating, it serves as a buffer along the landward side of the longshore current in the high-tide surf zone. This flow of water parallel to shore in the high-tide surf zone tends to create a flat bottom. At low tide, this flat surface may be exposed. If so, then most workers would probably refer to it as a "low-tide terrace." (JES thinks that this name is slightly misleading because its distinctive slope is determined by what happens at high tide.)

At low tide, the breakers may shift seaward exposing the sand ridge that had been functioning as the innermost breaker bar. This ridge and the trough that accompanies it on its landward side (the floor of the high-tide surf zone), which become dry at low water, have been named ridge and runnel (for the trough). On a ridge-and-runnel beach at low tide, it is possible to see currents of water flowing parallel to shore as the tide goes out. These currents return seaward through exposed gaps in the ridge. (These are the same gaps through which the rip currents flow at high tide.) When the tide is out, the zone of swash and backwash may be transferred to the seaward side of the ridge, and the floor of the runnel may become dry. Now, the high-tide beach face also is completely dry and is not affected by the water. Along its base, ground water from within the berm seeps out, forming a line of miniature springs.

When the tide returns, the low-tide zone of swash and backwash migrates up to the crest of the now-submerging ridge. Water floods into the runnel. Water from large swashes may spill over the crest of the ridge and pour into the runnel. These sheets of water do not become backwash; they flow landward and may deposit their entrained sand as tiny spillover delta lobes in the quiet water they encounter in the runnel. As the tide rises higher and higher, a point is reached where the zone of swash and backwash abandons the now-drowned ridge and skips across the runnel to resume its activity on the high-tide beach face. JES thinks that this twice-daily shifting of the surf zone from its low-tide position on the seaward side of a ridge to the high-tide beach face is a small-scale model of what might happen to an entire barrier island on a larger scale during a rapid submergence (i. e., relative rise of sea level).

Formation of Berm Scarps

At certain times, and for reasons that are not altogether fully understood, the behavior of water in the high-tide surf zone as just described is supplanted by a totally different pattern. One of the keys to this change is what JES refers to as "reverse refraction." Within the deeper water filling the trough landward of the breaker bar, waves that have broken re-form into new waves. The crests of these newly re-formed waves swing around so as to become perpendicular to shore rather than parallel to shore. (In normal refraction, the trends of wave crests tend to become parallel to shore. In this "reverse refraction," just the opposite happens--the crests tend to become perpendicular to shore.) Under certain conditions, water motions in the surf zone cause the zone of swash and backwash to disappear. When this happens, the sheets of water undercut the beach-face sediment and cause it to collapse and thus transform it into a vertical scarp. Sheets of water flowing more parallel to the shore than toward it undermine the base of this scarp causing more and more sand to collapse along vertical fractures (Rosalsky, 1964). (This behavior of the sand is related to the pore pressure of the interstitial capillary water; the surface tension of the water acts as a "cement" to hold the sand particles together.) If the undercutting mode is established on a rising tide, then the entire berm (the more-or-less flat part of the beach where most visitors take sunbaths), can be eroded away in a few hours. If the tide rises higher than normal, the landward-retreating scarp can "eat" into any higher ridges of sand that may be present along the landward edge of the berm ("shore-parallel ridges" as a general, non-genetic term; "dunes" of most popular usage). JES thinks that such retreating berm scarps are the chief cause of erosion of the subaerial parts of beaches and adjacent sand bodies.

To summarize, the dynamic activities in the intertidal morphodynamic zone range from those of the surf zone at high tide to the results of wave spillover and associated return currents as the tide rises.

SUPRATIDAL ZONE

In the supratidal zone, water appears only rarely, but when it does, incredibly swift changes can take place. When the tide is higher than normal or when large swells are breaking, or both of these conditions prevail, sheets of swash water flow up to the crest of the beach face, building it upward. Within a few hours, the crest of the beach (berm crest) may stand well above the general level of the rest of the berm. How high depends on the tidal height and sizes of the waves; it could be as much as a meter or more. The swashes that wash completely across the berm crest continue to flow landward down the back slope that leads away from the crest. If the berm-top sand is dry, as it typically is during dry weather, then many of the first-arriving sheets of spilled-over swash disappear by sinking into the dry sand. Eventually, however, the berm-top sand may become saturated. After the sand has become saturated, the water from the spillover sheets begins to accumulate; it may create spillover pools of varying sizes. These spillover pools may continue to enlarge until their ever-rising water level finds low places to serve as outlets. Such outlets may be low swales along the crestline of the berm, on the seaward side, or gaps in the crest lines of the shore-parallel ridge(s) ["dune(s)"] on the landward side. At Robert Moses State Park, some of these berm- top spillover pools have attained impressive dimensions. Their widths perpendicular to shore have reached many tens of meters; their lengths parallel to shore, a half a kilometer or so; and their maximum water depths, 2 meters. Water draining seaward from such substantial spillover pools has eroded gaps in the beach crest that were 15 meters wide and 2 meters deep. JES has even seen an example in which the skipper of a small power boat in distress guided his vessel to safety from the waves by navigating through one of these return channels and beaching his craft on the landward side of the quiet water of a spillover pool. Two days later, he returned with a low-loader trailer and hauled his boat away overland.

Within these bermtop spillover pools, strong currents flow parallel to shore; in the outlet channels, powerful currents flow away from shore.

The amounts of sand that can be shifted by the swift currents flowing in the intertidal circulation zone at high tide and along the top of the berm in the supratidal circulation zone can be tremendous. As we shall see, most of the times when spits at Democrat Point have grown rapidly have coincided with episodes of coast-parallel flow in these two zones.

BEACH FACE

The beach face displays three contrasting appearances. These are: (1) a more-or-less plane surface dipping toward the sea at angles ranging from about 5° to about 8°, (2) a vertical scarp; and (3) a "corrugated" surface consisting of rounded projections toward the sea that are comparable to rocky "headlands" and intervening semicircular recessed areas comparable to "bays." Collectively, these corrugated features are named beach cusps.

A plane beach face is probably the most-typical appearance. It is the product of slightly oblique approach of the waves. Not much needs to be added about plane beach faces. The scarps have been mentioned in a previous section. The rest of this section is devoted to beach cusps.

Beach Cusps

The spacing from headland to headland may vary from a few tens of meters to hundreds of meters. Some cusps are symmetrical; others, asymmetrical.

Cusps, especially the symmetrical ones, are created by the interactions of several sets of waves, the chief ones being a set that approaches the beach almost head on. Presumably, a second set of waves traveling along the beach (edge waves) interferes with those coming straight in. Such interference is thought to be responsible for creating the even spacing of the cusps. However they get started, once cusps become established, they tend to be self perpetuating. Water from swash is concentrated in the cusp bays and tends to flow seaward in streams along the axes of the bays. When these returning streams of water collide with the bores traveling landward across the surf zone, they cause the bores to divide. Thus, the directions of the arriving bores are changed; the wave forms are refracted. Part of the bore flows to the left, and part to the right, both tending to flow parallel to shore. Thus, in front of each headland, a bore approaches from the adjacent bay. The two, coming from opposite directions, cross in front of the headland, and there tend to drop their sediment. Accordingly, thicker layers of sediment are deposited on the headlands than accumulate in the bays. A very distinctive pattern of strata results. Cusps build seaward by adding successive relatively thick layers that are composed mostly of light-colored minerals of quartz density (2.65). Simultaneously, sediment may also be added to the bays, but at a much-slower rate than on the headlands. Successive layers are relatively thin and contain abundant dark-colored "heavy" minerals (densities exceeding 2.8; local examples are garnet, magnetite, and ilmenite). In walls of special trenches dug here at Robert Moses State Park, cusp strata have been traced continuously from bays to headlands. In one bay, a layer was one-particle-diameter thick (0.25 mm). When traced to the adjacent headland, this layer thickened to 25 cm, an increase of 100 times.

The foregoing analysis of cusps implies that under certain wave conditions, a corrugated beach face is a stable condition. On such a beachface, downbeach longshore transport has been changed into a whole series of small cells within which the transport direction reverses. Instead of a continuous flow down the beach in a single direction, as along a plane beach face or along the toe of an eroding beach scarp, on a cusped beach, thousands of local flow cells become established. Water flows away from the bays and toward the headlands. And the reason it does so seems to be refraction. What JES concludes from this is that if one is intent on "managing" a beach, then one might do well to intervene in such a way so as to cause waves to refract where they are not otherwise doing so. For example, by placing a series of invisible shoals in the subtidal zone, one might be able to establish a permanent criss-crossing refraction pattern, so that no matter how the waves approached the beach, they would always be refracted so that the water would divide and thus create cusps. Instead of fighting fire with fire, this would be fighting water with water. The result would be to eliminate the long one-way stretches of longshore transport that characterize the non-cusp intertidal- and supratidal zones. If this could be done, it might impede erosion of the berms. (This scheme probably would be very difficult to implement; it means working in the zone of breaking waves and in the surf zone. The notion of having the waves work against themselves, however, seems worth further investigation.

RELATIONSHIPS OF STRATA TO SLOPES OF BEACH SURFACES

The most-important lesson about the strata in beach sediments to be learned from this trip to Fire Island is that the strata form where sediment is added to parts or all of the visible parts of the beach. Two important slopes on which sediment builds outward and/or upward are the foreshore (which dips steeply toward the ocean) and the backshore (which dips gently away from the ocean). The part of the foreshore on which we shall concentrate is the beach face. The backshore surface is synonymous with berm top. These two oppositely dipping surfaces are separated by the sharp crest of the berm (Figure 14).

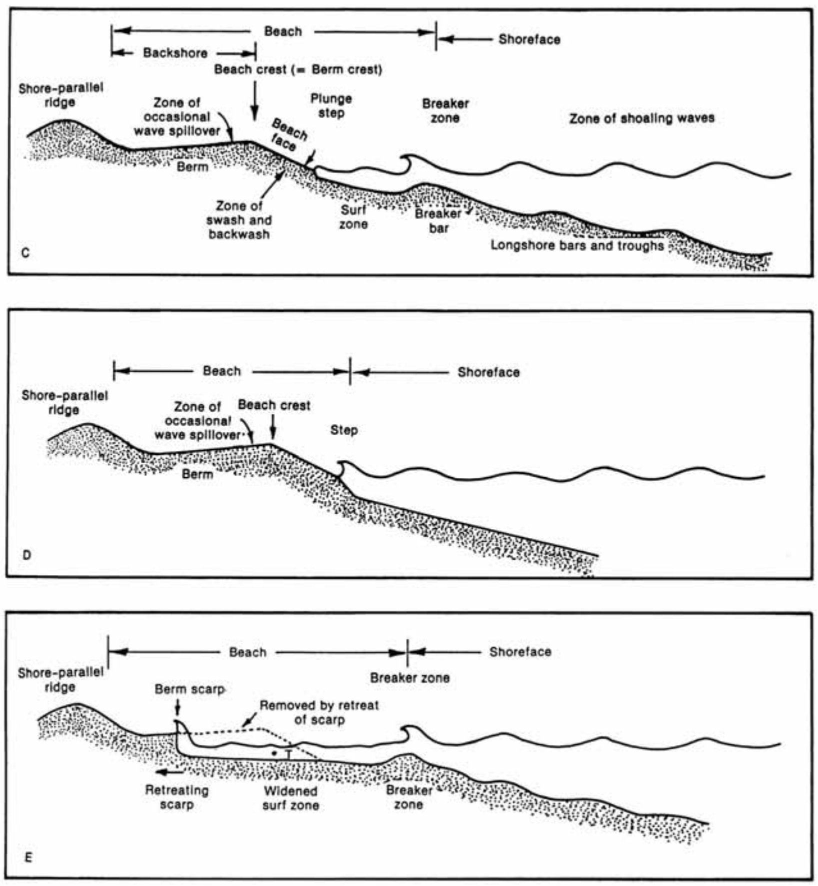


Figure 14. Schematic profiles at right angles to shores illustrating morphodynamic subdivisions of beach as used by JES.
C. Beach inshore from gently sloping bottom exhibits longshore bars on inner shoreface beneath zone of shoaling waves, breaker bar, surf zone and trough, small plunge step, plane beach face, and berm.
D. Stepped beach inshore from steeply sloping bottom lacks breaker bar; waves come almost to water's edge before collapsing over a prominent step.
E. Profile of beach as in C, but one in which berm is being rapidly removed. Powerful, wide longshore current in surf zone directed along shore (axis of current represented by dot with letter T to show flow out of plane of profile and toward viewer) has destroyed beach face and created vertical beach scarp. (JES in Friedman, Sanders, and Kopaska-Merkel, 1992, fig. 11-1, p. 405.)

Foreshore (=Beach-face) Strata

The diagnostic feature of beach-face strata is that they dip toward the ocean at angles in the range of 5° to 8°. The sand to build the beach face seaward is eroded from the zone of shoaling waves and to reach the beach face must pass through the breaker bar(s). This passing through the breaker bar(s) while the breaker bar(s) remain(s) more or less in place implies that the morphologic form of these bars can persist while at the same time sand is transiting through them. Usually beach-face strata consist of well-sorted, light-colored sand. A few such strata consist of dark-colored heavy minerals, such as garnet, magnetite, and ilmenite.

Backshore (Berm-top) Strata

Berm-top strata are formed only when conditions enable sheets of swash water to flow over the crest of the berm and to spread out on the top of the berm. Such conditions are associated with higher-than-normal spring tides and/or with the coming ashore of large swells.

Whenever numerous people visit the beach (mostly on weekends except during July and August when the beach is crowded daily), they tend to disturb the natural strata (by digging, driving beach buggies, or simply walking). Given the extent to which "people marks" are in the sand, it is rather surprising that the walls of most SST's display strata lacking such marks. One of the reasons for this is that when the first sheets of swash water flow over the crest of the berm, they smooth out the sand, thus eliminating many "people marks." After this initial smoothing, subsequent layers build up rapidly on a natural surface.

The natural smoothing by the thin sheets of swash water also eliminates what may be extensive evidence of wind action on the top of the berm.

Relationships Among Strata: Significance of Truncated Strata

If the beach face accretes seaward at the same time that the berm top builds upward, then a continuous layer will result. The beach-face part of this layer dips seaward, the berm-top part dips landward, and the place where the dip direction changes marks the position of a former berm crest. Given the dynamic setting of a beach, however, and the possibility for rapid removal of the seaward part (or even all of) the berm by undercutting, one should not be surprised to find that the seaward sides of berm-top strata are truncated (Figure 15). Usually this truncation takes place as a beach scarp (Figure 16) forms and migrates landward with a rising tide. But, such scarps do not last very long; they are reworked and replaced by a normal beach-face slope. As a result, beach-face strata truncate berm-top strata. Every such truncation implies that a berm crest formerly existed at some point seaward of the truncated surface.

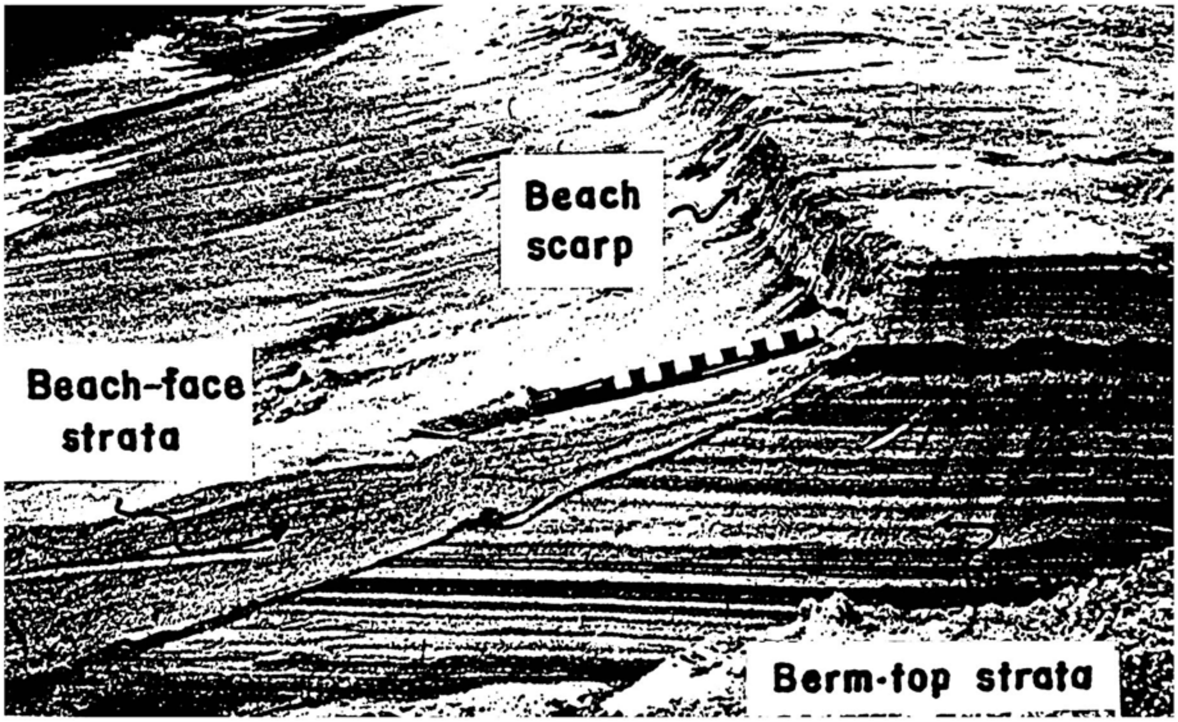


Figure 15. Light-colored beach-face (=foreshore) strata, dipping seaward and ending landward at low beach scarp truncate nearly horizontal berm-top (=backshore) strata consisting of distinct alternating layers of light- and dark-colored sand. Robert Moses State Park, Fire Island, Long Island, NY on 26 September 1983. (A. Massa.)

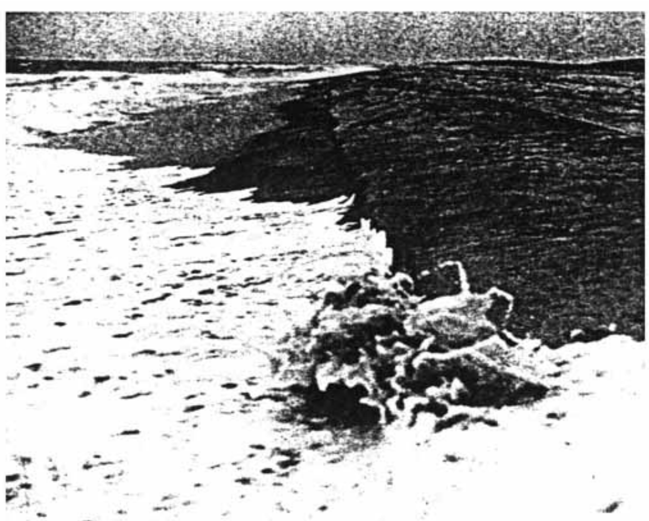


Figure 16. Initial stages in cutting of berm scarp. Onrushing very oblique swash moving toward beach face and away from observer has curled back after striking notch cut at base of small scarp (narrow dark area in center of view); at right, some swash water has flowed over the top of the small scarp. Robert Moses State Park, Fire Island, Long Island, NY, view W on 18 September 1975. (Imre Baumgaertner.)

Under certain conditions, a new beach-face, berm-crest, and berm-top combination may be deposited seaward of an older such combination. In that case, the gently dipping newer berm-top strata end landward by being cut off at the former beach-face slope (which is steeper and dips in the opposite direction, namely seaward).

Pay careful attention to the strata exposed in the walls of the SST. Label berm-top strata, beach-face strata, and any surfaces where the berm-top strata have been truncated. See if strata from more than one cycle of sediment accumulation on a beach-face, berm-crest, berm-top combination are present.

Relationship of New Layers of Sediment to Depositional Surfaces

A problem of never-ending interest for geologists is the relationship between aspects of the environment and strata deposited. At RMSP, Fire Island, it is possible to: (1) examine directly the relationships between the surface of the beach and individual strata or groups of strata deposited by sheets of water from breaking waves and (2) find examples of fixed sequences of strata that have been deposited as a result of lateral shifting of the depositional surfaces (such as an inlet channel, a spit, or the beach and shoreface).

Surfaces of the beach and individual strata

The best examples of the relationships of individual strata to parts of the beach are the contrasts among beachface (plane or cusped), berm top, and spillover delta lobes. RMSP beach sediment generally contains dark-colored heavy minerals in abundance; these form discrete, highly visible layers that contrast decidedly with the light-colored sand in which light minerals of quartz density predominate. The key point to remember is that the new individual layers generally follow the inclination of the surface on which they are deposited. Thus, the beachface strata dip rather steeply seaward, and the berm-top strata, gently landward. The dips of cusps and on spillover delta lobes display great directional spreads.

In order to show the relationships between strata and depositional surfaces, we shall dig a few SST's. We shall take suitable precautions to keep the sides from collapsing and to steer curious visitors from approaching the trench from the pristine side.

Beach-face strata

As noted previously, the beach face may be a plane surface or may be cusped. Consider first the strata deposited on a plane beach face. Such strata dip seaward at steep angles, usually ranging between 5° and 8°. (We will measure some beach-face dips to see for ourselves.) Because of the constant change from erosion to deposition and back again along the beach face, the beach-face strata rarely attain thicknesses of more than a few centimeters. They usually form thin wedges that taper out landward and thicken toward the sea.

Where a beach face is building seaward at the same time that the berm is accreting upward, then a continuous layer of sand may be deposited from the beach face, across the berm crest, and landward across the top the berm.

As mentioned in a previous section, the strata of a cusped beach face display a broadly corrugated aspect, with thicker layers of light-colored sand on the headlands, and thinner layers of dark-colored minerals in the bays (Figure 17).

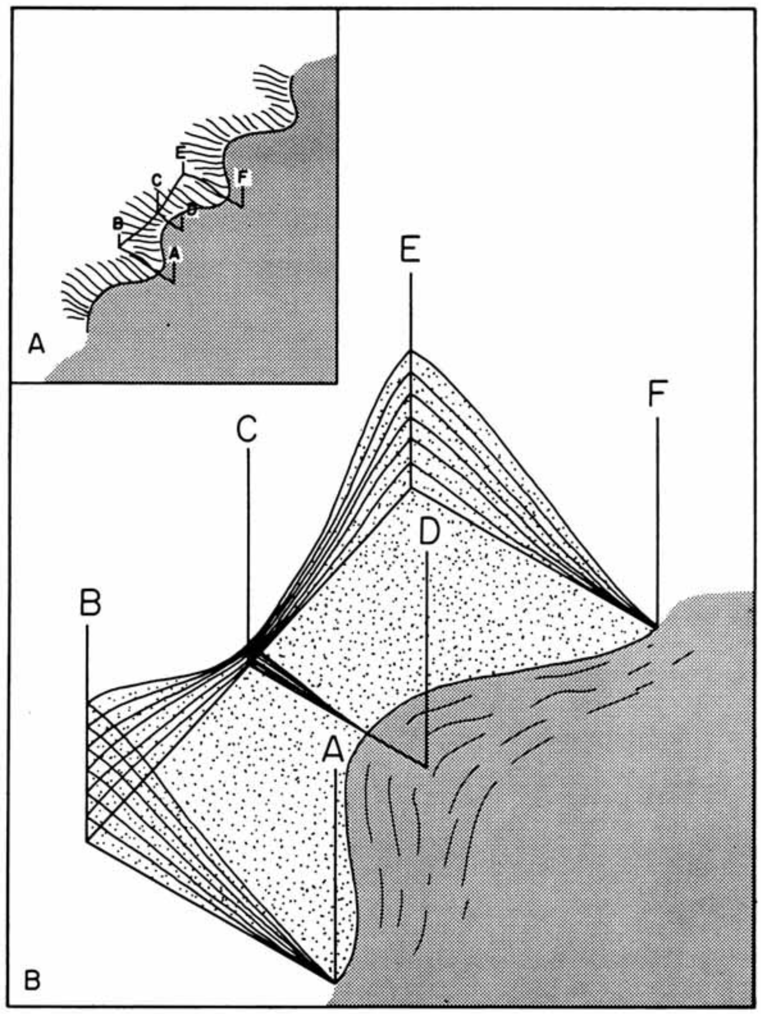


Figure 17. Beach cusps.
A. Regular, symmetrical beach cusps, sketched in oblique view from above. Lettered lines refer to fence diagram in B.
B. Fence diagram through sediments forming symmetrical beach cusps showing variation in the thicknesses of individual layers of sand as seen in cutaway views of three trench faces normal to shore [AB and EF through cusp "headlands" and CD through cusp "bay," and one trench parallel to shore (BCE). Thin layers of dark-colored heavy minerals at C (and extending toward D) and at A and F contrast with thicker layers composed of light-colored light minerals. (From Friedman, Sanders, and Kopaska-Merkel, 1992; A, fig. 11-3, p. 407, B, fig. 11-18, p. 421.)

Berm-top Strata

The sheet-like layers of berm-top strata dip gently landward, conforming to the landward dip of the top of the berm away from the berm crest. A remarkable feature of most of the berm-top strata JES has seen in SST's dug at RMSP is that they are uniform and planar. Only rarely is any expression preserved of the irregularities of footprints, tire tracks, and so forth, that always cover the top of the berm at the end of every weekend. As mentioned, this implies that the sheets of water which spill over the berm crest rapidly smooth out nearly all irregularities.

Berm-top strata cannot be deposited by themselves; they are always accompanied by a beach face and a berm crest. But, these two features, beach face and berm crest, may not be preserved. Because several factors can cause the seaward side of a berm to be eroded and thus to become narrower (seaward side shifts landward), truncated berm-top strata are common. By carefully tracing the various sets of berm-top strata and beach-face strata, it is possible to work out the history of erosion and deposition of sand on the berm.

On occasion, a beach scarp may be preserved. Scarps are the end products of an episode of berm erosion; no sediments layers are deposited parallel to the vertical scarps.

Strata of Spillover Delta Lobes

What are referred to as spillover delta lobes can be found in two places on beaches and in one place on a spit. On the beach, they are immediately landward of the crest of a ridge on a ridge-and-runnel beach, or landward of the berm crest. On a spit, they are landward of the crest of the spit berm. The strata of these lobes dip landward at the angle of repose of the sand, which is about 35°. These are the classical foreset beds of a so-called Gilbert delta (named for the American geologist, G. K. Gilbert, who described and named them from his observations of the sediments deposited around the margins of ancient Lake Bonneville. The steep dips of the foreset beds prove that the lobes were built into "standing" water (no such steep dip is found on washover fans, which are spread over a newly inundated former land surface by thin sheets of water). JES thinks spillover delta lobes illustrate an important aspect associated with deltas in which such steeply dipping foresets are present: a shallow channel supplied sand to a body of water deeper than the channel.

Topsets and bottomsets associated with these lobes are to be expected, but JES has never seen them in any of his many SST's.

Strata Associated with Bed Forms

Currents of water commonly fashion the sand over which they flow into a series of rhythmic relief features named bed forms. Included are such things as ripples, megaripples, sand waves, and dunes.

Strata are associated with bed forms as a result of the migration of the bed forms in response to the current. In the simplest case, the bed form migrates downcurrent by eroding sand from its upcurrent side (the gently dipping side) and depositing it on the downcurrent side. (The action resembles the deposition of foresets on a growing spillover-delta lobe.) The strata dip in the downcurrent direction and are named cross strata (or cross laminae, if they are very thin).

The only places where we are likely to see any bed forms that we can dig into and look to see if the cross strata are visible are the top of a berm where currents related to a berm-top spillover pool may have rippled the sand, or on surfaces exposed at low tide where currents flowed swiftly when water was still present. Such places include the floor of a runnel on a ridge-and-runnel beach and on the top of a spit platform.

Successions of Strata Deposited by Shifting Depositional Surfaces

Under this heading, we refer to larger-scale depositional surfaces that include several of the features discussed in the previous sections. Two examples are available at Fire Island: (1) beneath the triangular body of sand that built up on the E side of the Federal Jetty from 1940 to 1950 (Figure 18); and (2) beneath those parts of Fire Island through which the inlet migrated and thus deposited an inlet sequence (Figure 19).

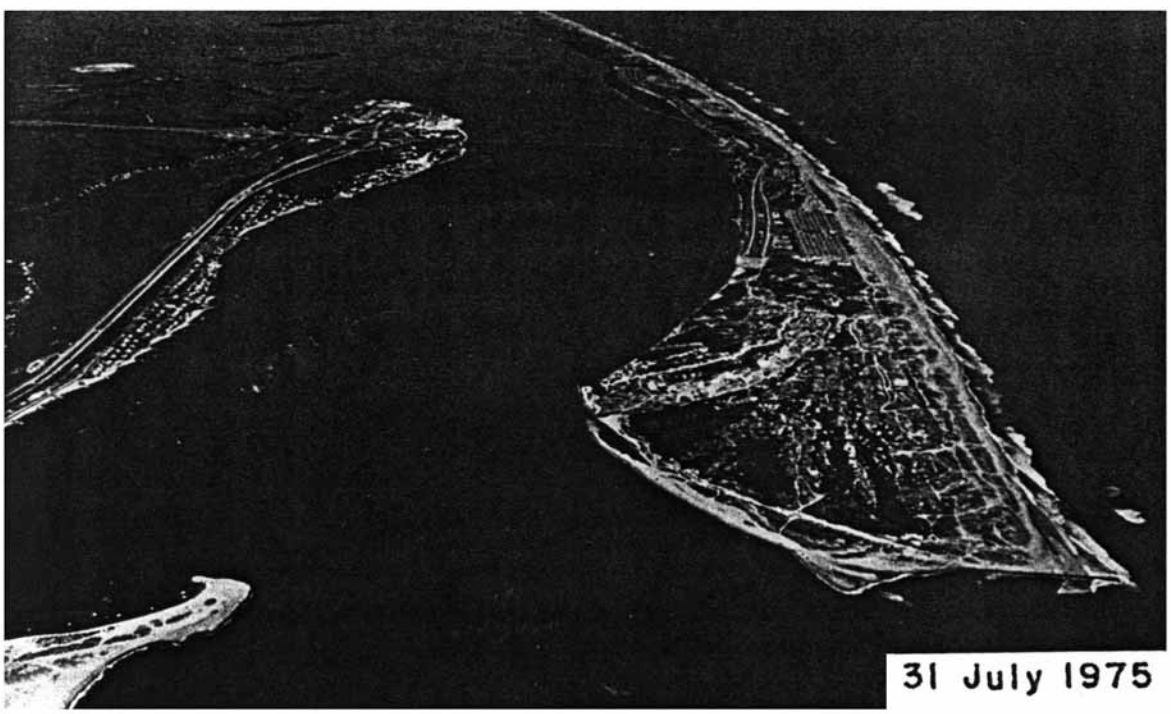


Figure 18. Small triangular strand plain at SW end of Fire Island barrier, Long Island, NY, that in the ten years following the construction of the Federal Jetty (1940 to 1950) prograded seaward on the E side of the jetty, viewed obliquely from low-flying airplane on 31 July 1975 after dredge had removed most of the sand west of the jetty. Atlantic Ocean at right; Great South Bay and Oak Beach at left. (Bruce Caplan.)

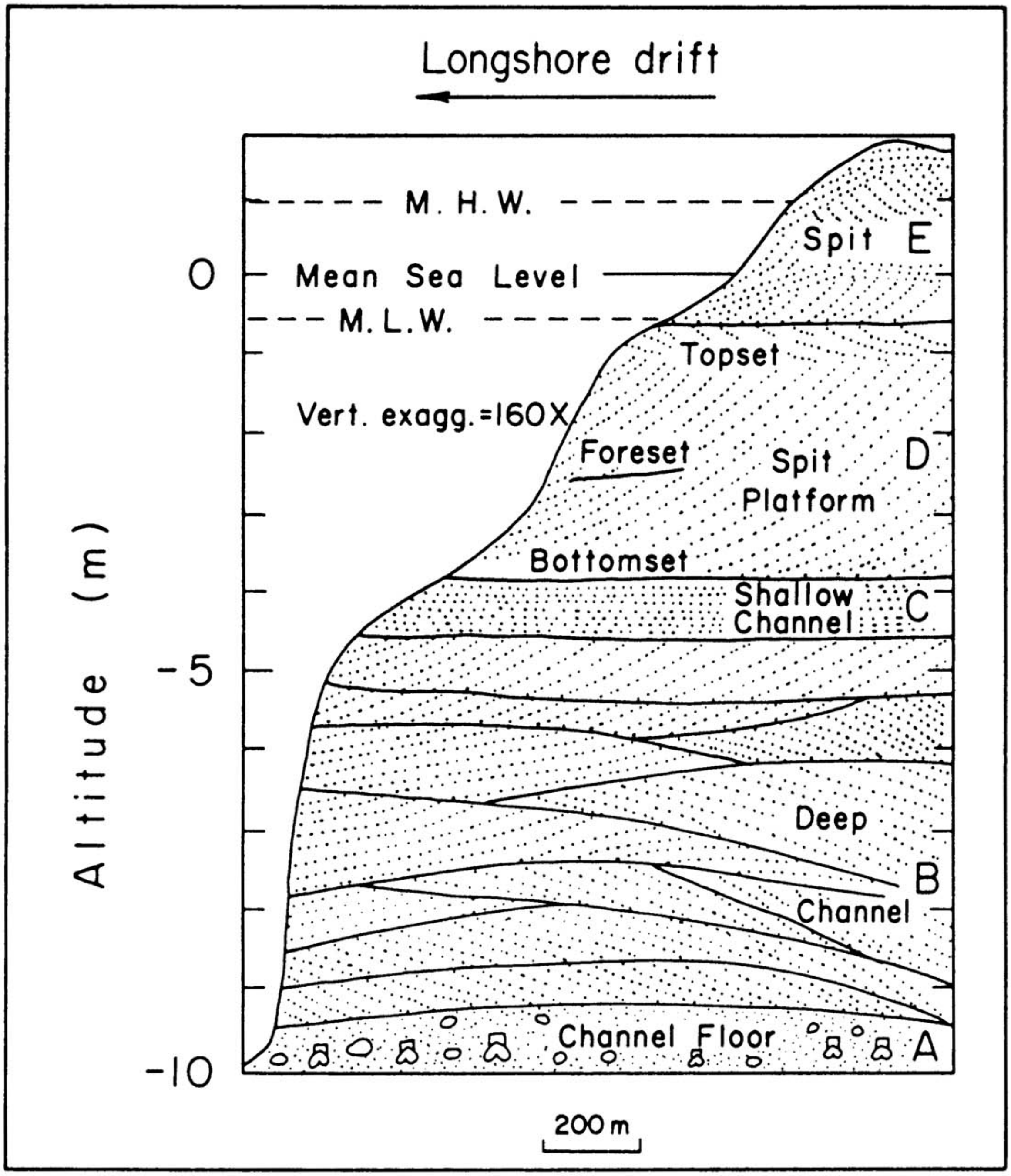


Figure 19. Schematic profile and section through east bank of Fire Island Inlet at Democrat Point spit, showing names and thicknesses of strata composing the succession deposited by the lateral migration of an inlet 10 m deep. The directions of dip
of the cross strata in the sketch (notably in the deep channel, topsets of the spit platform, and spit) are not related to the plane of this sketch. They have been drawn to show their distinctive aspects, not their true dip directions. (G. M. Friedman and J. E. Sanders, 1978, Principles of Sedimentology, Figure 11-13, p. 317; based on Naresh Kumar and J. E. Sanders, 1974, Fig. 6, p. 506.)

The principle involved is that along the various parts of a sloping sediment surface, contrasting and distinctive kinds of sediment will be deposited. The contrasts may be in particle sizes or in the kinds of primary sedimentary structures. For example, in an inlet deeper than 5 meters, the coarsest sediment, those of the channel-floor lag, will always be overlain by the deep-channel medium sands with their distinctive cross strata; and these, in turn, by the finer sands of the shallow parts of the channel. An inlet succession is one in which the pattern is coarsest at the base, and finer upward. (Such successions are named upward-fining successions.)

Where any sediment surface progrades, it will leave behind a distinctive sequence. On a barrier island, such prograding can take place where an inlet migrates parallel to shore (thus prograding along the shore) or where the whole seaward side of the barrier builds outward (prograding normal to the shore). Prograding of the seaward side of a barrier yields a barrier-island sequence that grades upward from shoreface sediments related to shoaling waves to various beach sediments that may be capped by wind-blown sand. The sediments of a barrier-island sequence are finer at the base and become coarser upward (=an upward-coarsening succession).

A First Look At The Beach: Composition And Subdivisions

A third population of sand that we can expect to encounter is black or deep red in color. The black is from an abundance of the iron-oxide minerals: magnetite (magnetic) or ilmenite (nonmagnetic iron-titanium oxide). The red is from garnets. The specific gravities of these dark-colored minerals are much greater than that of the light-colored minerals quartz (2.65) and feldspars (range of 2.56 to 2.76). The specific gravity of magnetite is 5.18; and that of garnet in the range of 4 (depending on the composition).

The dark-colored "heavy" minerals can be scattered throughout a population composed of light-colored "light" minerals, or the heavy minerals may have become segregated, usually as a kind of lag where the light-colored "light" minerals have been swept away. This segregation can be done by wind or water, and the heavy minerals may have been left essentially in place as a lag concentrate (as they typically are when the wind blows the light- colored minerals away or when the waves undercut the seaward side of a berm and create a berm scarp, leaving an accumulation of heavy minerals at the base of the scarp) or be transported and deposited to build up an area (as on the top of a berm).

The precise relationships among the aforementioned three populations of particles at the beach at RMSP are not well known. Other populations should perhaps be added to the list. Some questions needing answers are: (1) Where do these populations of particles come from? (2) Under what conditions are they brought ashore and added to the berm? (3) What becomes of them when they are not being brought ashore?

After we have made our profiles and dug an SST and Parking Field 3, we shall have lunch. After lunch, reboard the vans and drive to the far W end of Parking Field. From there, we shall leave the vans again and walk on a wider beach near the mini-golf club house.

A Walk On The Wide Beach: Morphology and Sedimentary Strata

We shall walk to the wide beach via a path near the clubhouse at the miniature golf course. We will measure a beach profile and dig two SST's, one each into the berm crests to display the relationships between the sedimentary layers and the surface of the beach.

THE GEOLOGIC CYCLE AND THE GEOLOGIC RECORD

Modern geology began about 200 years ago, when an unusually observant Scotsman, James Hutton, delivered two lectures to the Royal Society of Edinburgh, the first on 07 March 1785 and the second, on 04 April 1785. The title of his lectures was: "Concerning the Systems of the Earth, its Duration and Stability." In 1788, the paper was published in volume 1 of the Transactions of the Society under the title of: "Investigation of the Laws observable in the Composition, Dissolution and Restoration of Land upon the Globe." In 1795, Hutton published a two-volume book entitled "Theory of the Earth with Proofs and Illustrations." Chapter one was the paper from the Transactions. Hutton finished, but never published, the manuscript for a third volume. He died on 26 March 1797 (White,1956, p. v-xv).

In Hutton's time, geologic science had progressed to the point where it was generally agreed that beneath a surface layer of loose soils or "dirt," one encounters solid bedrock and that much of this bedrock consists of layers (named strata) that form parallel series, among which the thicknesses of some remain the same through long distances. In many places, the strata are horizontal, the same position in which they were formed initially, but in mountain chains, the strata are inclined at various angles, including vertical. Hutton's great contribution was the realization that the strata consist of "materials furnished from the ruins of former continents." In other words, the "system of universal decay and degradation" that he saw taking place all around him today was providing sediment that would be deposited in the sea, would be converted to strata, and later be elevated to form new land, only to be eroded and the debris carried into another sea, and so on, back and forth, up and down, and so on into the future. But equally important, this cycle of activities had been going on in the past. And, by tracing groups of strata and applying the concept that the lower strata are older than those covering them, Hutton was able to work his way backward in the geologic record. His most-startling conclusion was that the oldest rocks he could find in his searches within the geologic record consist of strata that resembled all the other strata. In other words, he was not able to find any rocks that had been built up under circumstances differing appreciably from those prevailing today. Accordingly, Hutton was confident that his generalization about the geologic cycle provided the key link in connecting the past, the present, and the future.

John Playfair, Professor of Mathematics at the University of Edinburgh, and a friend and pupil of Hutton's, published a memorial biography of Hutton in the Transactions of the Royal Society of Edinburgh in 1805. According to Playfair:

> "It might have been expected, when a work of so much originality as this Theory of the Earth, was given to the world, a theory which professed to be

> the result of such an ample and accurate induction, and which opened up so many views, interesting not to mineralogy alone, but to philosophy in general, that it would have produced a sudden and visible effect, and that men of science would have been every where eager to decide concerning its real value. Yet the truth is, that it drew their attention very slowly, so that several years elapsed before any one shewed himself publicly concerned about it, either as an enemy or a friend.
>
> "Several causes probably contributed to produce this indifference. The world was tired out with unsuccessful attempts to form geological theories, by men often but ill informed of the phenomena which they proposed to explain, and who proceeded also on the supposition that they could give an account of the origin of things, or the first establishment of that system which is now the order of nature...

To this list JES thinks should be added the turbulent state of affairs in western Europe, particularly in France, during the decade starting in 1789. Hutton had studied in Paris and was well acquainted with French-language publications by both French and Swiss investigators (for example, Dolomieu, Buffon, LaPlace, La Grange, and Saussure; and he read J. G. Lehman's 1756 book in its translation into French dated 1759). According to Donald B. MacIntyre (Heezen Memorial Lecture to New York Academy of Sciences, Section on Geological Sciences, in October 1988), during a decade of peace between France and England (1783-1793), Hutton's close friend and "scout," John Clerk, was available full time as an illustrator and collaborator with Hutton. In 1793, England and France were at war again, a condition that prevailed more or less continuously until 1815, when Wellington defeated Napoleon at the battle of Waterloo. During these wars, Clerk served with distinction in the Royal Navy. Despite the turmoil, however, French publications such as the Journal de Physique having dates in the 1790s are cited in the footnotes of Playfair's book (1802).

Two circumstances contributed to the eventual spread of Hutton's ideas. (1) Playfair book (1802) entitled: "Illustrations of the Huttonian Theory of the Earth" became more widely known than Hutton's original books. Moreover, Playfair's clear, elegant prose provided an easier access to Hutton's ideas than could be gained from Hutton's books. (2) Charles Lyell's textbook of geology (1830) was built on the Huttonian framework. Lyell's book founded modern geology.

APPENDIX A: WAVES: ORIGIN, CHARACTERISTICS, SHOALING, BREAKING, AND EFFECTS ON SEDIMENTS

Many kinds of waves cross the sea surface. They are caused by boat wakes, by winds blowing locally, by winds from distant storms, and by displacments of the sea floor. We shall skip over boat wakes, concentrate on wind-generated waves, and mention briefly in passing the kinds of waves generated by displacements of the sea floor. As we shall see, a close relationship exists between the kinds of waves and their sizes. The most-useful expession of wave sizes is the wave period, which is the time it takes for one wavelength to pass a given point. The periods of waves we shall discuss range from seconds to minutes.

Where wind stresses are applied directly to the water surface, they generate irregular, steep, choppy waves known as sea waves. The periods of sea waves range from fractions of a second for tiny ripples to about 10 seconds for large waves in an intense storm. The crest lines of sea waves do not persist laterally; a characteristic giving rise to their choppy aspect. Considering a large storm having a diameter of say 600 kilometers crossing an open ocean, one finds that the winds (which blow counterclockwise around a low-pressure storm in the Northern Hemisphere) generate waves that can be propagated in all directions away from the center of the storm (Figure 20). After they have traveled a few tens of kilometers away from the storm, the waves, which are now no longer subject to wind stresses, begin to reorganize and to form themselves into longer, lower, more-regular waves known as swell waves (or simply as the swell or swells). The periods of swells range from about 6 seconds up to 15 seconds. The periods of a few very large swells may reach 22 seconds or so. Off Fire Island, the periods of swells usually falls between 6 and 8 seconds. The ratio of wave height to wavelength (H/L) is used to express wave steepness. A ratio of 0.025 is used as the boundary between steep waves (usually sea waves) and low waves (swells; Figure 21).

One of the first observations to make about nearshore waves is the relationship between swells and sea waves. Notice their periods and directions of travel. Periods can be determined by counting the number of waves that break in 1 minute (60 sec). The period in seconds is the number of breaking waves counted divided into 60. Thus, if you counted 10 breaking waves in a minute, their periods are 60/10 = 6 sec.

If waves are crossing water that is deeper than twice their wavelengths, they are said to be deep-water waves. The speeds of deep-water waves are determined by their wavelengths. The longer-period waves travel faster than do shorter-period waves.

Water-surface waves that are generated by displacements of the sea floor are so long and low that in the open sea, it is not pos- sible to notice them. Their wavelengths measure in hundreds of kilometers and their heights, in a few meters. Their periods are about 15 minutes or so. The technical term for such waves is tsunami (Japanese for "harbor waves" and spelled the same in both singular and plural, as in the English word "sheep." JES used to make a big deal out of this until he learned a tiny bit of Japanese grammar, namely that all nouns are spelled the same in both singular- and plural forms!) The word "tsunamis" appears widely in the English-language technical literature and JES hereby abandons his previous DQ crusade to change it, but will continue his own usage of tsunami in both singular and plural. Because of their great wavelengths, tsunami are a special kind of waves known as very-shallow-water waves everywhere, even in the deepest parts of the oceans. Very-shallow-water waves are defined as water-surface waves in which the depth is 1/20 or less of the wavelength. (Using y for depth and L for wavelength, this is expressed as $y/L = 0.05$). A fundamental principle of very-shallow-water waves is that their speeds are controlled only by the depth of water. (The mathematical statement of this relationship is the speed equals the square root of the product of the Earth's gravitational acceleration times the depth of water.) Knowing the speeds of tsunami, it is possible to use this relationship to compute the depth of water over which they have traveled. On this basis, the first estimates of the depth of the Pacific Ocean were made. The great destructiveness of tsunami results from their piling up to great heights as they approach certain coasts, Hawaii, for example.

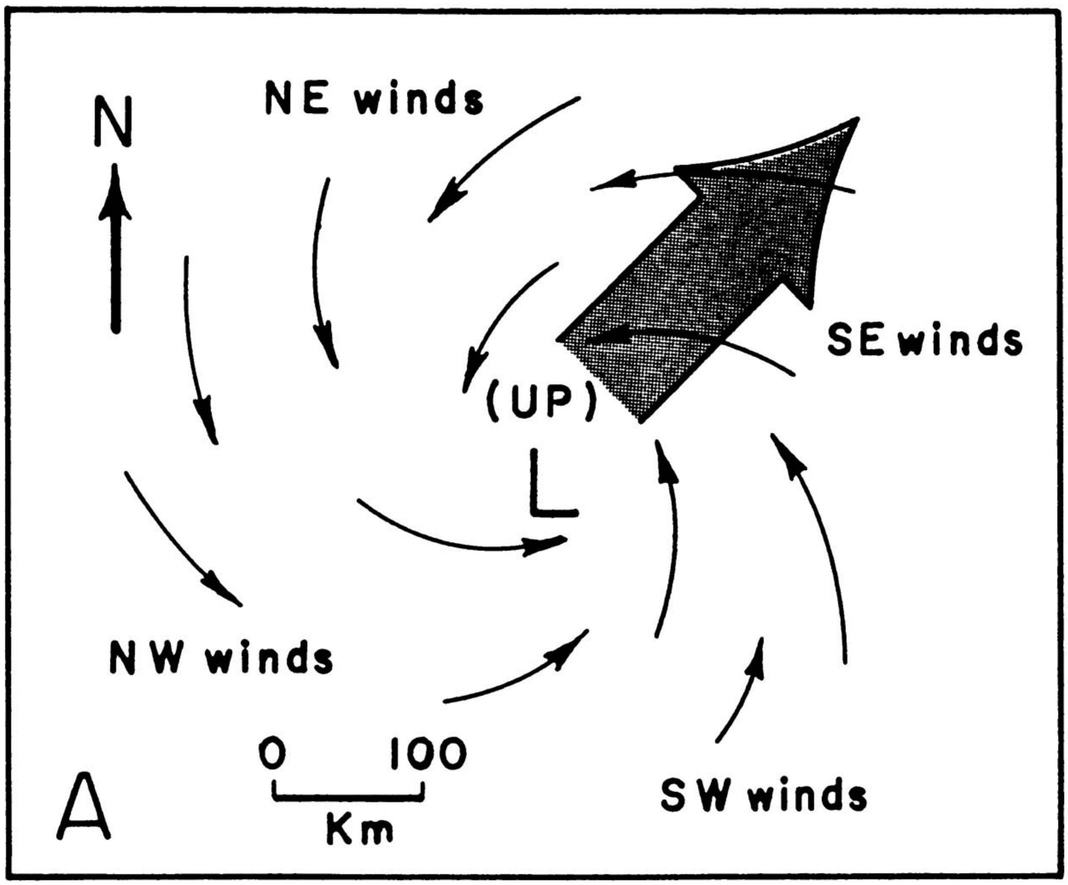


Figure 20. Schematic view, from above, of spiral low-pressure cell of a storm crossing an ocean in the Northern Hemisphere. Storm center is moving NE (large arrow; small arrows show winds at the Earth's surface. In SE quadrant, the forward speed of the storm adds to the speed of the SW circumferential winds. Here, surface winds reach maximum values. In NW quadrant, the forward speed of the storm is opposite to and thus subtracts from the the circumferential NE winds. Here, surface circumferential winds reach their minimum values. (G. M. Friedman and J. E. Sanders, 1978, Principles of sedimentology: New York, John Wiley & Sons, Inc., Figure A-2, A, p. 465.)

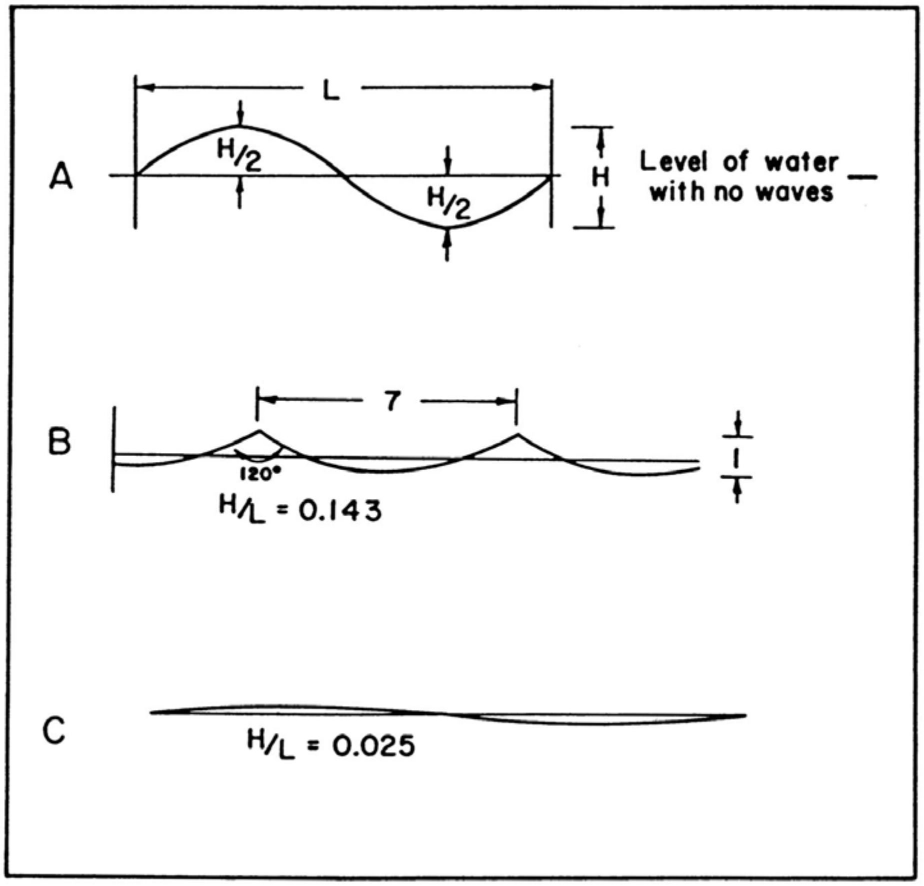


Figure 21. Profiles of waves sketched in a vertical plane parallel to their direction of travel.
A. Idealized, symmetrical wave; the same as the graph of a sine function. Much vertical exaggeration.
B. Symmetrical waves having critical limiting steepness. Waves break before they become any steeper than 1/7 (H/L = 0.143).
C. Symmetrical wave showing dimensions of waves used as the
boundary between steep and low; no vertical exaggeration. (Friedman and Sanders, 1978, Fig. A-1, p. 465.)

Approach of Waves to the Shore

As waves approach shore, they first react with the bottom and, eventually, break. The depth at which waves first begin to interact with the bottom is half their wavelength, or y/L = 0.5. Some investigators have applied the name wave base to this depth (Figure 22). At a depth of about 1/8 the wavelength (y/L = 0.125), the effects of the bottom become noticeable in that the waves begin to slow down and their profiles, to change. Crests become shorter and steeper, and troughs wider and flatter. The circular paths of the water "particles" typical of deep-water waves give way to elliptical paths, with the long axes of the ellipses parallel with the bottom (Figure 23). The waves are now known as shallow-water waves. The speeds of shallow-water waves are determined by a function that includes both the wave- lengths and the water depth.

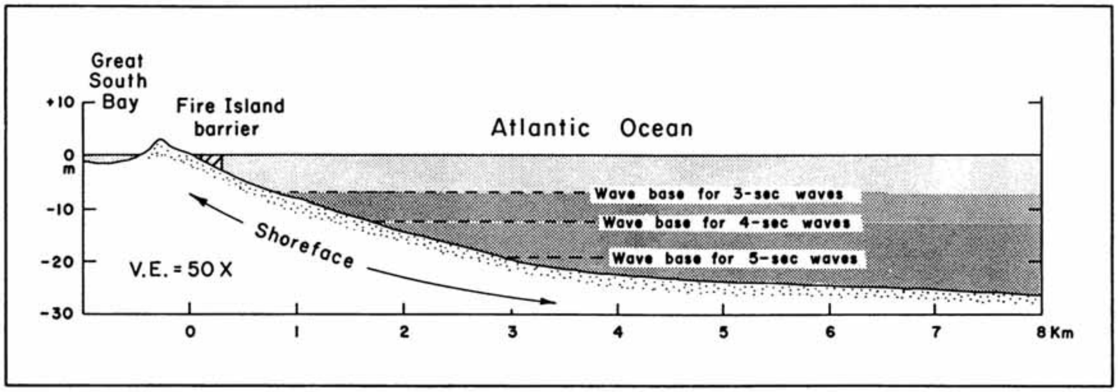


Figure 22. Profile at right angles to shore off Fire Island barrier island, Long Island, New York, showing depths to wave base for several sizes of waves (based on periods). Diagonal lines near the shoreline mark the zone where swells having 6-sec periods become very-shallow-water waves. Because of the wide, shallow shelf, most swells are refracted so that they approach the shore straight on in most localities. Thus, aerial photographs do not show pronounced refraction of the swells near shore, as on the Pacific coast. (Friedman and Sanders, 1978, Fig. A-15, p. 474.)

At depths of y/L of 0.05, shallow-water waves become very-shal- low water waves. (See discussion of tsunami.) A brief examination of the relationships of very-shallow-water waves will show their importance in explaining how waves behave near shore. Not all waves approaching shore become very-shallow-water waves; among those that do so, they do not experience this transformation at the same depth; as mentioned, it depends on their wavelengths, hence also on their periods. However, once a wave has become a very-shallow-water wave, its speed is fixed by the depth, and the wave slows down markedly as depth decreases. For example, at the limiting depth of 1/20 of wavelength (=L/20), the speed of a wave decreases to about 60 per cent of its former speed. By the time the wave encounters water about 2 meters deep, its speed may have been reduced to about one quarter of its speed in deep water.

In the opinion of JES, the fundamental importance of this change of many shoaling waves from shallow-water waves to very shallow-water waves has been generally overlooked. In some analyses, the nearshore waves are treated as if they were solitary waves. This may be justified on the basis of mathematical convenience, but it in no way coincides with reality.

JES thinks that this remarkable change to very-shallow-water waves explains some waves, notably swells arriving on the California coast undergo such obvious nearshore wave refraction (defined as the change of direction of waves as a result of decreasing depth or of an encounter with a current traveling against the waves).

In the zone of shoaling transformations, the shoaling waves impart an oscillatory motion to the bottom, which may give rise to wave-generated ripple marks (Figure 24). Two noteworthy geologic characteristics of such ripple marks are their transverse profiles and the directions of the trends of their crest lines.

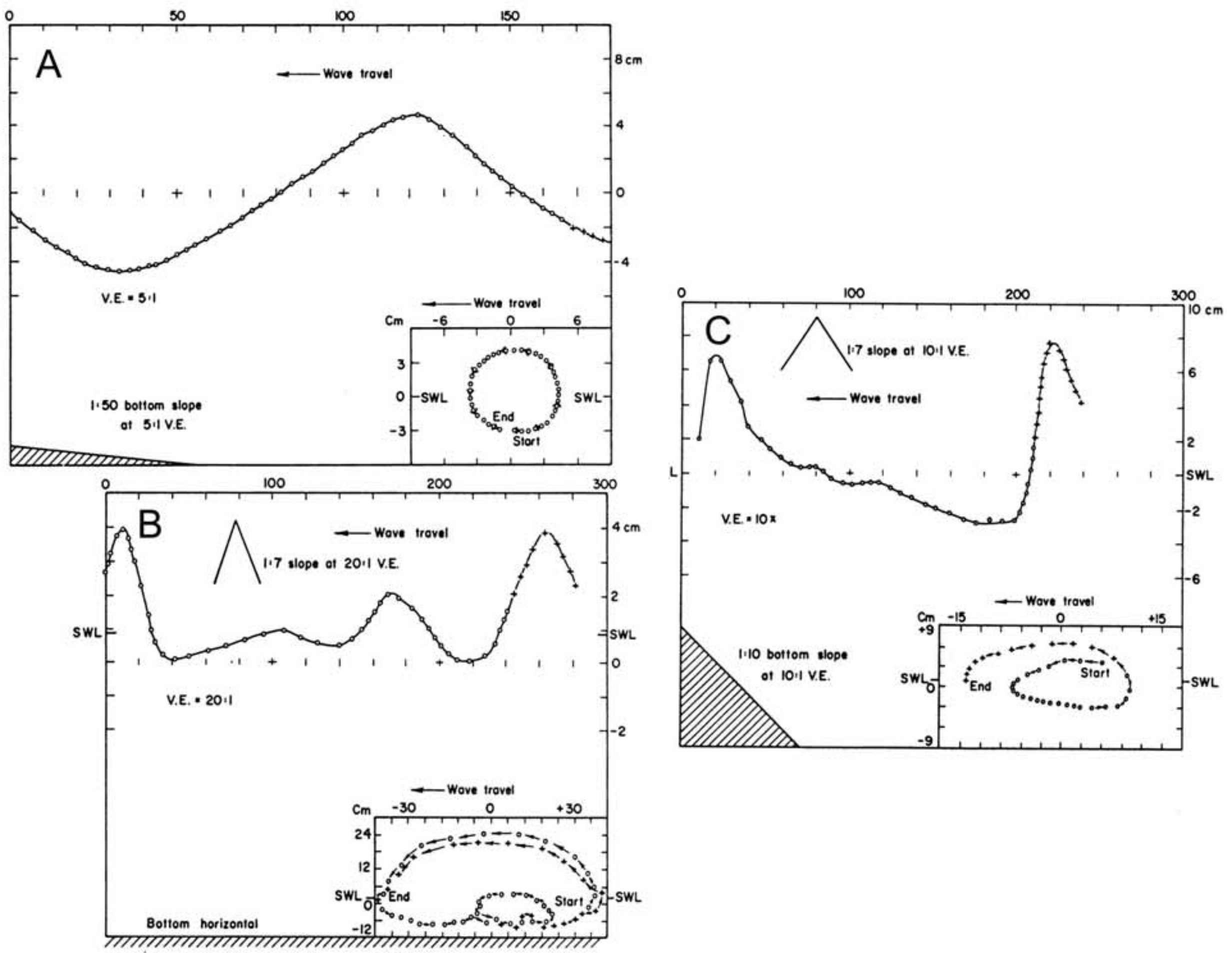


Figure 23. Shoaling transformations shown by single-frame pro- jection of movie film taken of neutrally buoyant particles through glass side of a large wave tank (dimensions 0.3 x 1 x 20 m).
A. Shoaling waves (H=9 cm; L=158 cm; period=1.06 sec; water 36 cm deep with bottom sloping 1:50; H/L=0.057; y/L=0.228) that are still nearly symmetrical and with water-particle orbits circular (lower right).
B. Much-transformed waves close to breaking limit (H=3.8 cm; L=254 cm; period=2.67 sec; water 8.9 cm deep; bottom horizontal; H/L=0.015; y/L=0.035. Secondary wave in the flattened trough has generated a smaller elliptical path at the base oe oe oe of the main ellipse.
C. Steep waves just prior to breaking (H of breakers, 11.3 cm; H/L of waves outside breaker zone=0.0206; period of breaking waves 1.51 sec; depth of water at breakers=9.14 cm; bottom slope=1:10. Orbits are elliptical; mass transport is in direction of wave travel. (Friedman and Sanders, 1978, Fig. A-11, p. 470-471.)

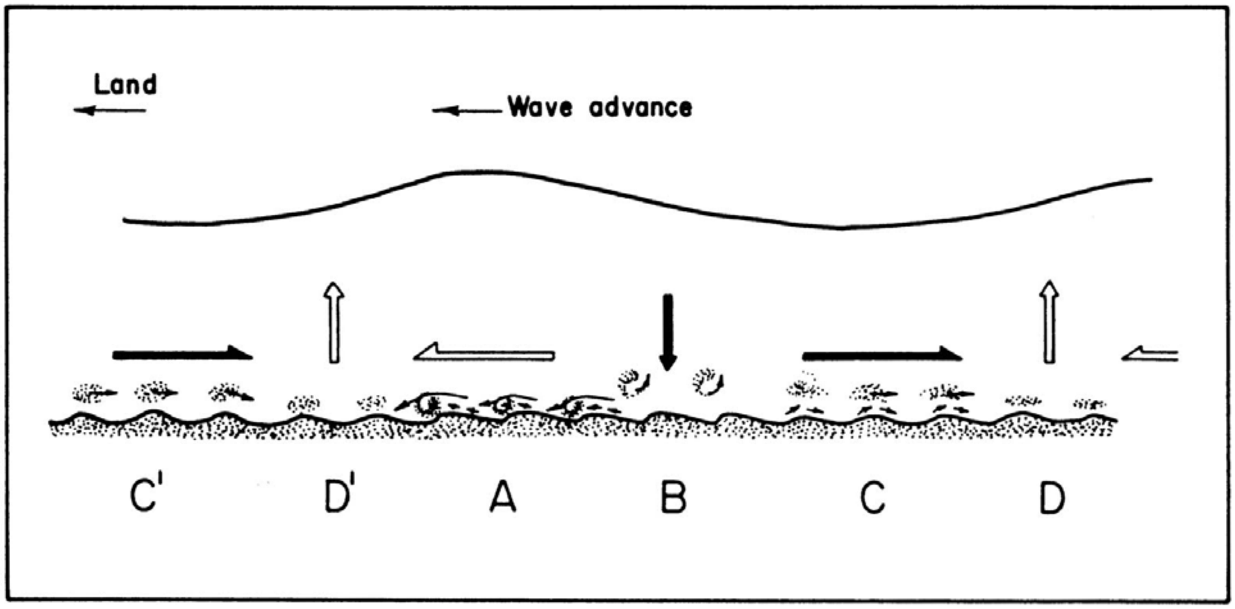


Figure 24. Schematic profile in vertical plane parallel to direction of travel of shallow-water waves showing relationships responsible for wave-generated ripple marks in bottom sand; sizes of bottom ripple marks not related to size of water-surface waves.

Under wave crest, at A, wave-induced bottom-water oscillation surges landward (open arrow with one barb), thus imparting shearing stresses from water to bottom, which creates asymmetrical ripple marks having steeper sides landward. The tiny eddies on the downshear sides of the sand ripples tend to rise through the water, to enlarge, and to dissipate.

After the water-wave crest has passed, at B, the water surges vertically downward (large vertical black arrow), carrying sediment back toward the bottom.

Beneath wave troughs, at C and C', bottom-water oscillates in a direction opposite to wave advance (large horizontal black arrows with one barb), imparting shearing stress to the bottom sediment that is directed away from shore and creating sediment- laden eddies on both sides of sediment ripples, whose profiles now become symmetrical or nearly symmetrical.

After a water-wave trough has passed, as at D and D', the bottom water surges vertically upward, lifting sediment-laden water, and thus tending to facilitate sediment transport by the landward surge accompanying the passage of the next wave crest. This is just the opposite of what happens beneath a water-wave trough, where a vertical downward surge of water precedes the seaward surge, a situation tending to inhibit sediment transport. (Friedman, Sanders, and Kopaska-Merkel, 1992, fig. 10-7, p. 363.)

The transverse profiles of wave-generated ripple marks differ according to location on the bottom. In the outer zone of shoaling waves, ripple profiles are symmetrical; their crests are pointed and their troughs are rounded and concave up. In the inner zone of shoaling waves, ripple profiles are asymmetrical; their crests are rounded and their steeper sides are toward shore (Figure 25). This configuration results from the changes in relative strengths of the back-and-forth oscillations of the bottom water as shoaling proceeds. In the outer part of the zone of shoaling waves, the oscillatory motion of the bottom water under water-wave crests equals that under water-wave troughs. In the inner part of the zone of shoaling waves, the shoreward surge beneath the wave crests exceeds that of the offshore surge beneath the wave troughs. (Compare Figures 23A and 23C.)

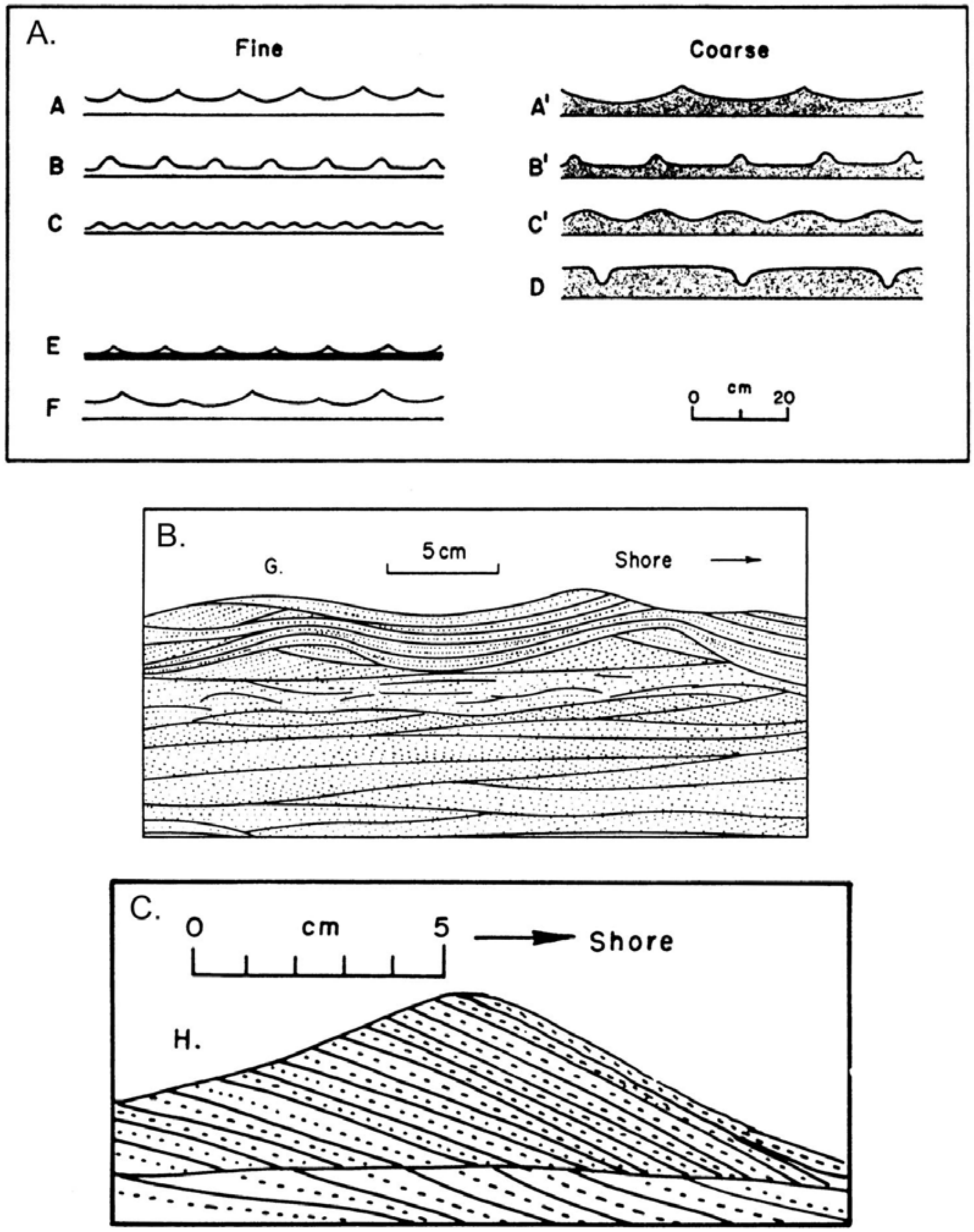


Figure 25. Varieties of wave-generated ripple marks shown by transverse profiles in vertical planes at right angles to ripple trends.
A. Symmetrical ripples that are smaller in fine sand than in coarse sand.
B and C. Sketches of internal laminae shown by relief peels made from box cores collected off the German coast of the Baltic Sea. B, slightly asymmetric ripples in which most cross laminae dip landward, but a few dip seaward. C, asymmetric ripple in which all cross laminae dip toward shore. (Friedman andSanders, 1978, Fig. A-17, p. 477; B and C from R. S. Newton, 1968.)

The second characteristic of geologic value is that the trends of the crests of wave-generated ripple marks are perpendicular to the direction of travel of the shoaling water waves. Where the shoaling waves are travelling straight in toward shore or if they arrive obliquely but have been significantly refracted, their crests parallel the shore. Correspondingly, the trends of ripple marks made by such shoaling waves are parallel to shore (Figure 26, B, Loc. 4 on 26 Aug).

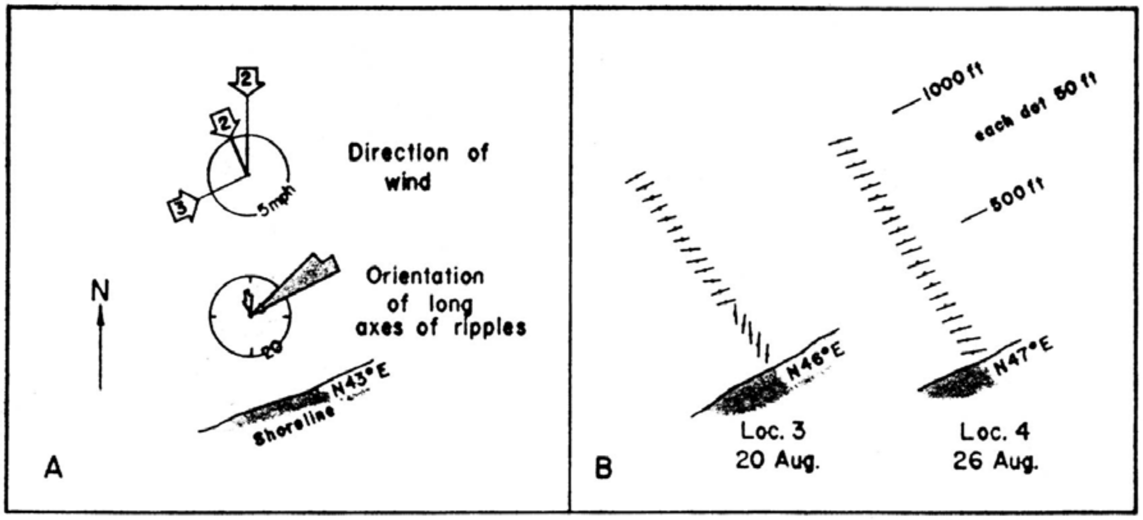


Figure 26. Trends of wave-generated ripple marks in nearshore sand, SW Lake Michigan, Berrien Co., Michigan, observed by R. A. Davis, Jr., in August 1963, compared with directions of winds and trend of shoreline.
A. Summary of all observations. Lines in upper rose diagram show wind speeds (lengths proportion to miles per hour; circle gives reference length for 5 mph) and wind directions. Numerals inside open arrows give number of days the wind blew from that direction. Lower rose diagram summarizes long axes of all wave-generated ripple marks measured (plotted in 5-degree sectors). Lengths of pie-shaped sectors are proportional to total number of ripple axes measured within each 5-degree class. Circle gives reference length for 20 readings.
B. Maps of ripple orientations at two closely spaced stations on different days. Observation points (dots) were spaced 50 feet apart on line perpendicular to shore. At Loc. 3 on 20 Aug 1963, inner ripples trend oblique to shore but are practically perpendicular to wind from the WSW; the trends of the outer ripple set are nearly parallel to shore. At Loc. 4 on 26 Aug 1963, the trends of all ripples nearly parallel the shore and are perpendicular to waves from the NNW. (Friedman and Sanders, 1978, fig 9-16, p. 251.)

Waves break because the crests, steepened by reactions with the bottom during the shoaling transformations, exceed the limiting crestal-peak angle of 120°. This can be expressed as a steepness ratio of deep-water wave height, H, divided by deep-water wavelength, L, as $H/L = 0.143$. (See Figure 21B.) Another relationship between deep-water wave height, H, and breaking depth is that breaking results at a depth of 1.28 H. If one recalls the limiting depth ($y/L = 0.05$) for conversion to very-shallow-water waves, then it is possible to use these relationships to calculate a limiting value of wave steepness (defined as H/L) for waves that will break before they become very-shallow-water waves.

We begin by setting the limiting breaker depth equal to the limiting depth for very-shallow-water waves:

Breaker depth = 1.28 H = 0.05 L (limiting depth for very-shallow-water waves)

Clearing and solving for H/L:

$$H/L = 0.05/1.28, \text{ which is } = 0.039.$$

This means that waves steeper than 0.039 will break before they can be converted into very-shallow-water waves. This means, further, that steep, short-period waves may break before they experience the rapid refraction related to the zone of very- shallow-water waves. Accordingly, such steep waves can approach the breaker zone from any angle. This relationship explains the observations of ripple axes trending at oblique angles to shore, as in Figure 25B, Loc. 3 on 20 Aug.

Off Long Island, the swells doubtless are refracted as they cross the wide, shallow continental shelf. Relationships comparable to those that have been shown for the Virginia shelf (Goldsmith, 1976) can be expected on the Long Island shelf. Correspondingly, sectors of Long Island beaches undergoing chronic erosion may be responding to convergence of wave energy as a result of refraction during their travel over the Long Island shelf.

Zone of Shoaling Waves; Shoreface

The seaward side of a barrier (or open coast) extending from the outer edge of a beach to a distinct change in slope is the shoreface. (See Figure 22.) The shoreface is underlain by sediments affected chiefly by shoaling waves. Under typical conditions, the shoaling waves establish a nearly symmetrical back-and-forth action within the lower flow regime. But, conditions on a shoreface range widely between two end-member extremes. At one end of this range is zero wave action. At the other end is intense bottom-water shearing within the upper flow regime. During fair weather, the shoaling waves erode the shoreface and transport sand to the beaches. During storms, the sediment dispersed into the water column may be transported offshore.

In fair weather, when the water surface is being crossed by only small sea waves and small swells, sediments underlying the outer part of the shoreface are not shifted by wave action. At such times, the chief processes affecting the bottom sediments result from the activities of bottom-dwelling organisms. All organisms that live on the bottom of the sea are collectively designated as the benthos (=benthic organisms). Depending on their life styles, benthic invertebrate animals are subdivided into the epifauna, invertebrates that live upon the sea bottom, and the infauna, invertebrates that live in burrows or in other shelters within the bottom materials. The moving epifaunal organisms leave tracks and trails on the sediment/water interface. The infauna (and a few kinds of epifauna that probe or dig into the bottom in their efforts to feet upon the infauna) make various kinds of holes, burrows, and tunnels within the sediment and in so doing, may create numerous mounds and semicircular depressions on the bottom. Their activities result in burrow-mottled sediment.

The bottom lying just seaward of the shoreface is rippled by the effects of shoaling waves of moderate storms and is jostled by the shoaling waves from severe storms. Otherwise, unless it is subject to strong tidal currents or wind-driven currents, it is quiet. During storms, the sediments underlying the entire shoreface are agitated, perhaps violently. No one really knows what happens on the shoreface during a storm. However, several pieces of indirect evidence contain important clues. These include reports from captains of sailing vessels; from one extraordinary accumulation of large clams on New Jersey beaches; from cores of shoreface sediments from off Long lsland, New York; and from distinctive curved strata deposited in mounds or hummocks.

Several published reports relate stories told by sea captains of finding sand on deck after storm waves had broken over the bows of their ships, which had been caught in storms near the shore where the charts showed water depths of 20 m and the bottom to be composed of sand. If these reports are true, they imply that the storm waves dispersed substantial quantities of sand from the bottom and perhaps even placed large amounts of sand into suspension. The feasibility of these stories of sand on the decks of ships during storms can be judged by comparing them with the following true-life adventure story of some New Jersey clams.

A remarkable accumulation on the beach at Ocean City, New Jersey of countless thousands of still-living pelecypods (Mercenaria mercenaria) took place on 27 and 28 February 1961 (Figure 27). Such organisms normally live in burrows in the sediments underlying the outer part of the shoreface. How could they possibly have been taken from their burrows and tossed on shore? Ordinarily, clams do not care for this sort of adventure. Quite the contrary; in order to remain as part of the infauna, burrowing clams are capable of deepening their burrows very rapidly.

Figure 27. Bulldozer clearing thick layer of large shells of Mercenaria mercenaria that inundated the beach at Ocean City, NJ on 27 and 28 February 1961. (Wide World Photos.)

Thus, if the bottom starts to be lowered, the normal thing is for the clams immediately to dig in deeper. However, during the winter, the temperature of the bottom water off New Jersey drops into the range of 1° to 4° C. This is not cold enough to kill the infauna, but it is cold enough to chill them to the point where they cannot react very quickly. Thus, if large shoaling waves apply shearing stresses to the bottom sediments while cold bottom water has immobilized the infauna, the waves may lower the level of the bottom and the animals are helpless to do anything about it. Result: the burrowing organisms are washed out of their burrows and become playthings for the waves.

If the foregoing explanation is correct, then on the days concerned, the waves off the New Jersey coast must have lowered the sandy bottom by at least 30 cm (depth of normal Mercenaria burrow). We can infer, therefore, that the shearing stresses applied to the bottom by the shoaling waves dispersed the bottom sediment and released countless thousands of clams from their burrows. Only the chilled clams were deposited on the heach. The result was a gigantic accumulation of "clams-in-the wholeshell." On exposure to the atmosphere, the clams died and the beach soon became one vast stinking mess. What became of the sand? We do not know for sure, but can guess that as the large shoaling waves died down, most of the dispersed and/or suspended sand probably was deposited back on the shoreface once again. The sand definitely was not deposited on the beach along with the clams.

We do not know of any further information about this New Jersey episode, but we do know that long cores of modern shoreface sediments off Long Island, New York, consist of thick (up to 2 m) sediment couplets. The basal parts of these couplets contain structureless gravel, up to several tens of centimeters thick and containing well-rounded pebbles of rock fragments up to 4 cm in diameter, and a few large broken shells. This coarse basal zone is overlain by slightly micaceous, very well-laminated fine sands up to 2 m thick. No skeletal remains of any kind have been found in these well-laminated shoreface sands.

We infer that these sediment couplets resulted from the effects of large shoaling waves, possibly storm waves. During the most-intense wave action, all the sand now forming the laminated part of these specimens was dispersed within the water (possibly kept in suspension), leaving the gravel as a widespread lag pavement on the shoreface. All large organisms and/or skeletal debris of large clams, for example, seem to have been separated from the dispersed (and/or suspended) terrigenous sediment. (Possibly the large skeletal materials were deposited on the beach, as were the New Jersey clams mentioned above.) At certain times on the beach at Fire Island, large disarticulated shells are present, but no live clams. Conclusion: waves dispersed the bottom sand, the living clams burrowed in deeper, but the skeletons of dead individuals could not do so, and thus were exhumed and tossed ashore. Other skeletal debris may have been broken into pieces too small to identify. As the intensity of the wave action diminished, the dispersed (and/or suspended) sand was redeposited on the bottom. The plane-parallel laminae imply that while the sand was being deposited, the stable bed configuration was a plane surface. This suggests conditions analogous to those in the transition from the lower flow regime to the upper flow regime of alluvial channels.

If shoaling waves could disperse and/or suspend enough sand to deposit 2 m of laminated sand after a storm, then the water column could very well have contained sand all the way to the surface. Using 30 percent porosity and the density of quartz, a prism of bottom sediment 10 cm on a side and 2 m long contains a volume of 20 liters within which are 37,100 grams of quartz, or 1855 grams per liter. If this same amount of sediment were uniformly dispersed through a water column 10 cm square and 20 m deep (volume = 200 l) the concentration would be 9.275 g/l. Perhaps after all, those old "sea dogs" gave accurate reports when they said that storm waves left sand on the decks of their ships.

If only deep-water waves are present, then burrowing organisms would recolonize the inactive bottom. In either case, the top 30 cm or so of the shoreface sediments would be reworked. Ripple marks and wave-ripple laminae appear in the sediments influenced by small shoaling waves; burrow mottles riddle the sediments not subject to wave action.

On the modern barrier off Long lsland, the distinctive modern examples of shoreface sediments just described are confined to depths ranging from 5 to 21 m. We regard these sediments as being extremely diagnostic and of potentially great value in reconstructing ancient marine environments where the shelf seaward of the shoreface of is underlain by sand.

In other localities, a sandy shoreface gives way seaward to silt or clay. In the absence of large waves, the depth of the sand-silt transition may be a reflection of the time-averaged wave base. If this sand-silt boundary moves back and forth, the result may be the interbedding of layers of sand with layers of silt. The mechanism by which shoreface sand moves outward over silt is not known. This is only one of the many areas of ignorance about the sediments of modern shorefaces. Doubtless, many surprising things will be learned as research on this topic progresses.

ACKNOWLEDGEMENTS

We are indebted to Matt Katz and Marcie Brenner of the New York Academy of Sciences for handling the registration logistics of our On-The-Rock field-trip series. Thanks also to the anonymous lifeguard at Robert Moses State Park who told JES about the "cliffs" exposed near Parking Field 3.

TABLES

Table 01 - GEOLOGIC TIME CHART

(with selected major geologic events from southeastern New York and vicinity)

ERA Periods (Epochs)	Years (Ma)	Selected Major Events
CENOZOIC		
(Holocene)	0.1	Rising sea forms Hudson Estuary, Long Island Sound, and other bays. Barrier islands form and migrate.
(Pleistocene)	1.6	Melting of last glaciers forms large lakes. Drainage from Great Lakes overflows into Hudson Valley. Dam at The Narrows suddenly breached and flood waters erode Hudson shelf valley. Repeated continental glaciation with five? glaciers flowing from NW and NE form moraine ridges on Long Island.
(Pliocene)	6.2	Regional uplift, tilting and erosion of coastal-plain strata; sea level drops. Depression eroded that later becomes Long Island Sound.
(Miocene)	26.2	Fans spread E and SE from Appalachians and push back sea. Last widespread marine unit in coastal-plain strata.
MESOZOIC	66.5	
(Cretaceous)		
	96	Passive eastern margin of North American plate subsides and sediments (the coastal-plain strata) accumulate.
	131	(Passive-margin sequence II).
(Jurassic)		Baltimore Canyon Trough forms and fills with 8,000 feet of pre-Cretaceous sediments.
		Atlantic Ocean starts to open. Newark basins deformed, arched, eroded.
	190	Continued filling of subsiding Newark basins and mafic igneous activity both extrusive and intrusive.
(Triassic)		Newark basins form and fill with non-marine sediments.

PALEOZOIC	245	
(Permian)		Pre-Newark erosion surface formed.
	260	**Appalachian orogeny.** (Terminal stage.) Folding, overthrusting, and metamorphism of Rhode Island coal basins; granites intruded.
(Carboniferous)		Faulting, folding, and metamorphism in New York City area. Southeastern New York undergoes continued uplift and erosion.
(Devonian)	365	**Acadian orogeny.** Deep burial of sedimentary strata. Faulting, folding, and metamorphism in New York City area. Peekskill Granite and Acadian granites intruded.
(Silurian)		
	440	**Taconic orogeny.** Intense deformation and metamorphism.
	450	Cortlandt Complex and related rocks intrude Taconian suture zone. (Cameron's Line). Arc-continent collision. Great overthrusting from ocean toward continent. Taconic
(Ordovician)		deep-water strata thrust above shallow-water strata. Ultramafic rocks (oceanic lithosphere) sliced off and transported above deposits of continental shelf.
		Shallow-water clastics and carbonates accumulate in west of basin (= Sauk Sequence; protoliths of the Lowerre Quartzite, Inwood Marble, part of Manhattan Schist Fm.). Deep-water terrigenous silts form to east. (= Taconic Sequence;
(Cambrian)		protoliths of Hartland Formation, parts of Manhattan Schist Fm.). (Passive-margin sequence I).
PROTEROZOIC		
	570	Period of uplift and erosion followed by subsidence of margin.
(Z)	600	Rifting with rift sediments, volcanism, and intrusive activity. (Ned Mountain, Pound Ridge, and Yonkers gneiss protoliths).
(Y)	1100	**Grenville orogeny.** Sediments and volcanics deposited, compressive deformation, intrusive activity, and granulite facies metamorphism. (Fordham Gneiss, Hudson Highlands and related rocks).
ARCHEOZOIC		
	2600	No record in New York.
	4600	Solar system (including Earth) forms.

Table 02 - Generalized Descriptions of Major Geologic "Layers", SE New York State and Vicinity

This geologic table is a tangible result of the On-The-Rocks Field Trip Program conducted by Drs. John E. Sanders and Charles Merguerian between 1988 and 1998. In Stenoan and Huttonian delight, we here present the seven layer cake model that has proved so effective in simplifying the complex geology of the region. Under continual scrutiny and improvement, we provide this updated web-based information as a public service to all students and educators of geology. We encourage any comments, additions, or corrections. References cited can be sought by following this link.

LAYER VII - QUATERNARY SEDIMENTS

A blanket of irregular thickness [up to 50 m or more] overlying and more or less covering all older bedrock units. Includes four or five tills of several ages each of which was deposited by a continental glacier that flowed across the region from one of two contrasting directions: (1) from N10°E to S10°W (direction from Labrador center and down the Hudson Valley), or (2) from N20°W to S20°E (direction from Keewatin center in Hudson's Bay region of Canada and across the Hudson Valley). The inferred relationship of the five tills is as follows from youngest [I] to oldest [V]. [I] - Yellow-brown to gray till from NNE to SSW, [II] - red-brown till from NW to SE, [III] - red-brown till from NW to SE, and [IV] - yellow-brown to gray till from NNE to SSW, and [V] - red-brown till from NW to SE containing decayed stones (Sanders and Merguerian, 1991a,b, 1992, 1994a, b; Sanders, Merguerian, and Mills, 1993; Sanders and others, 1997; Merguerian and Sanders, 1996). Quaternary sediments consist chiefly of till and outwash. On Long Island, outwash (sand and gravel) and glacial lake sediment predominates and till is minor and local. By contrast, on Staten Island, tills and interstratified lake sediments predominate and sandy outwash appears only locally, near Great Kills beach.

[Pliocene episode of extensive and rapid epeirogenic uplift of New England and deep erosion of major river valleys, including the excavation of the prominent inner lowland alongside the coastal-plain cuesta; a part of the modern landscape in New Jersey, but submerged in part to form Long Island Sound].

~~~~~~~~~~~~~~~~~~~~Surface of unconformity~~~~~~~~~~~~~~~~~~~~

LAYER VI - COASTAL-PLAIN STRATA (L. Cretaceous to U. Miocene; products of Passive Continental Margin II - Atlantic).

Marine- and nonmarine sands and clays, present beneath the Quaternary sediments on Long Island (but exposed locally in NW Long Island and on SW Staten Island) and forming a wide outcrop belt in NE New Jersey. These strata underlie the submerged continental terrace. The basal unit (L. Cretaceous from Maryland southward, but U. Cretaceous in vicinity of New York City) overlaps deformed- and eroded Newark strata and older formations. Also includes thick (2000 m) L. Cretaceous sands and shales filling the offshore Baltimore Canyon Trough. At the top are Miocene marine- and coastal units that are coarser than lower strata and in many localities SW of New Jersey, overstep farther inland than older coastal-plain strata. Capping unit is a thin (<50 m) sheet of yellow gravel (U. Miocene or L. Pliocene?) that was prograded as SE-directed fans from the Appalachians pushed back the sea. Eroded Newark debris is present in L. Cretaceous sands, but in U. Cretaceous through Miocene units, Newark-age redbed debris is conspicuously absent. This relationship is considered to be proof that the coastal-plain formations previously buried the Newark basins so that no Newark-age debris was available until after the Pliocene period of great regional uplift and erosion. The presence of resistant heavy minerals derived from the Proterozoic highlands part of the Appalachians within all coastal-plain sands indicates that the coastal-plain strata did not cover the central highlands of the Appalachians.

[Mid-Jurassic to Late Jurassic episode of regional arching of Newark basin-filling strata and end of sediment accumulation in Newark basin; multiple episodes of deformation including oroclinal "bending" of entire Appalachian chain in NE Pennsylvania (Carey, 1955), and one or more episodes of intrusion of mafic igneous rocks, of folding, of normal faulting, and of strike-slip faulting (Merguerian and Sanders, 1994b). Great uplift and erosion, ending with formation of Fall-Zone planation surface].

~~~~~~~~~~~~~~~~~~~~Surface of unconformity~~~~~~~~~~~~~~~~~~~~

LAYER V - NEWARK BASIN-FILLING STRATA (Upper Triassic and Lower Jurassic)

Newark-age strata unconformably overlie folded- and metamorphosed Paleozoic strata of Layer II and some of the Proterozoic formations of Layer I; are in fault contact with other Proterozoic formations of the Highlands complex. Cobbles and boulders in basin-marginal rudites near Ramapo Fault include mostly rocks from Layers III, IIB, and IIA(W), which formerly blanketed the Proterozoic now at the surface on the much-elevated Ramapo Mountains block. The thick (possibly 8 or 9 km) strata filling the Newark basin are nonmarine.

In addition to the basin-marginal rudites, the sediments include fluvial- and varied deposits of large lakes whose levels shifted cyclically in response to climate cycles evidently related to astronomic forcing. A notable lake deposit includes the Lockatong Formation, with its analcime-rich black argillites, which attains a maximum thickness of about 450 m in the Delaware River valley area. Interbedded with the Jurassic part of the Newark strata are three extrusive complexes, each 100 to 300 m thick, whose resistant tilted edges now underlie the curvilinear ridges of the Watchung Mountains in north-central New Jersey. Boulders of vesicular basalt in basin-marginal rudites prove that locally, the lava flows extended northwestward across one or more of the basin-marginal faults and onto a block that was later elevated and eroded. The thick (ca. 300 m) Palisades intrusive sheet is concordant in its central parts, where it intrudes the Lockatong at a level about 400 m above the base of the Newark strata. To the NE and SW, however, the sheet is discordant and cuts higher strata (Merguerian and Sanders, 1995a). Contact relationships and the discovery of clastic dikes at the base of the Palisades in Fort Lee, New Jersey, suggest that the mafic magma responsible for the Palisades was originally intruded at relatively shallow depths (roughly 3 to 4 km) according to Merguerian and Sanders (1995b).

Xenoliths and screens of both Stockton Arkose and Lockatong Argillite are present near the base of the sill. Locally, marginal zones of some xenoliths were melted to form granitic rocks (examples: the trondhjemite formed from the Lockatong Argillite at the Graniteville quarry, Staten Island, described by Benimoff and Sclar, 1984; and a "re-composed" augite granite associated with pieces of Stockton Arkose at Weehawken and Jersey City, described by J. V. Lewis, 1908, p. 135-137).

[**Appalachian terminal orogeny**; large-scale overthrusts of strata over strata (as in the bedding thrusts of the "Little Mountains east of the Catskills" and in the strata underlying the NW side of the Appalachian Great Valley), of basement over strata (in the outliers NW of the Hudson Highlands, and possibly also in many parts of the Highlands themselves), and presumably also of basement over basement (localities not yet identified). High-grade metamorphism of Coal Measures and intrusion of granites in Rhode Island dated at 270 Ma. Extensive uplift and erosion, ending with the formation of the pre-Newark peneplain].

~~~~~~~~~~~~~~~~~~~~Surface of unconformity~~~~~~~~~~~~~~~~~~~~

LAYER IV - COAL MEASURES AND RELATED STRATA (Carboniferous)

Mostly nonmarine coarse strata, about 6 km thick, including thick coals altered to anthracite grade, now preserved only in tight synclines in the Anthracite district, near Scranton, NE Pennsylvania; inferred to have formerly extended NE far enough to have buried the Catskills and vicinity in eastern New York State (Friedman and Sanders, 1982, 1983).

[**Acadian orogeny**; great thermal activity and folding, including metamorphism on a regional scale, ductile deformation, and intrusion of granites; dated at ~360 Ma].

LAYER III - MOSTLY MARINE STRATA OF APPALACHIAN BASIN AND CATSKILLS (Carbonates and terrigenous strata of Devonian and Silurian age)

(Western Facies)	**(Eastern Facies)**
Catskill Plateau, Delaware Valley monocline, and "Little Mountains" NW of Hudson-Great Valley lowland.	SE of Hudson-Great Valley lowland in Schunnemunk-Bellvale graben.
Kaaterskill redbeds and cgls.	Schunnemunk Cgl.
Ashokan Flags (large cross strata)	Bellvale Fm., upper unit
Mount Marion Fm. (graded layers,	Bellvale Fm., lower unit

marine)	(graded layers, marine)
Bakoven Black Shale	Cornwall Black Shale
Onondaga Limestone	
Schoharie buff siltstone	Pine Hill Formation
Esopus Formation	Esopus Formation
Glenerie Chert	
Connelly Conglomerate	Connelly Conglomerate
Central Valley Sandstone	
Carbonates of Helderberg Group	Carbonates of Helderberg Group
Manlius Limestone	
Rondout Formation	Rondout Formation
Decker Formation	
Binnewater Sandstone	Poxono Island Formation
High Falls Shale	Longwood Red Shale
Shawangunk Formation	Green Pond Conglomerate

[**Taconic orogeny**; 480 Ma deep-seated folding, dynamothermal metamorphism and mafic- to ultramafic (alkalic) igneous intrusive activity (dated in the range of 470 to 430 Ma) across suture zone (Cameron's Line-St. Nicholas thrust zones). Underthrusting of shallow-water western carbonates of Sauk Sequence below supracrustal deep-water eastern Taconic strata and imbrication of former Sauk-Tippecanoe margin. Long-distance transport of strata over strata has been demonstrated; less certain locally is proof of basement thrust over strata and of basement shifted over basement. In Newfoundland, a full ophiolite sequence, 10 km thick, has been thrust over shelf-type sedimentary strata].

~~~~~~~~~~~~~~~~~~~~~Surface of unconformity~~~~~~~~~~~~~~~~~~~~~

LAYER II - CAMBRO-ORDOVICIAN CONTINENTAL-MARGIN COVER (Products of Passive Continental Margin I - Iapetus). Subdivided into two sub layers, IIB and IIA. Layer IIA is further subdivided into western- and eastern facies.

LAYER IIB - TIPPECANOE SEQUENCE - Middle Ordovician flysch with basal limestone (Balmville, Jacksonburg limestones).

Not metamorphosed / Metamorphosed
Martinsburg Fm. / Manhattan Schist (Om - lower unit).
Normanskill Fm. / Annsville Phyllite

Subaerial exposure; karst features form on Sauk (Layer IIA[W]) platform.

~~~~~~~~~~~~~~~~~~~~~Surface of unconformity~~~~~~~~~~~~~~~~~~~~~

LAYER IIA[W] - SAUK SEQUENCE

Western shallow-water platform (L. Cambrian-M. Ordovician)

Copake Limestone
Rochdale Limestone
Halcyon Lake Fm.
Briarcliff Dolostone
Pine Plains Fm.
Stissing Dolostone
Poughquag Quartzite
Lowerre Quartzite [Base not known]

LAYER IIA[E] - TACONIC SEQUENCE

Eastern deep-water zone (L. Cambrian-M. Ordovician)

Stockbridge
or Inwood Marbles

(Ꞓ-Oh) Hartland Fm.
(Ꞓ-Om) Manhattan Fm.
(in part).

[**Pre-Iapetus Rifting Event**; extensional tectonics, volcanism, rift-facies sedimentation, and plutonic igneous activity precedes development of Iapetus [Layer II = passive continental margin I] ocean basin. Extensional interval yields protoliths of Pound Ridge Gneiss, Yonkers granitoid gneisses, and the Ned Mountain Formation (Brock, 1989, 1993). Followed by a period of uplift and erosion. In New Jersey, metamorphosed rift facies rocks are mapped as the Chestnut Hill Formation of A. A. Drake, Jr. (1984)].

~~~~~~~~~~~~~~~~~~~~Surface of unconformity~~~~~~~~~~~~~~~~~~~~

LAYER I - PROTEROZOIC BASEMENT ROCKS

Many individual lithologic units including Proterozoic Z and Y ortho- and paragneiss, granitoid rocks, metavolcanic- and metasedimentary rocks identified, but only a few attempts have been made to decipher the stratigraphic relationships; hence, the three-dimensional structural relationships remain obscure.

~~~~~~~~~~~~~~~~~~~~Surface of unconformity~~~~~~~~~~~~~~~~~~~~

[**Grenville orogeny**; deformation, metamorphism, and plutonism dated about 1,100 Ma. After the orogeny, an extensive period of uplift and erosion begins. Grenville-aged (Proterozoic Y) basement rocks include the Fordham Gneiss of Westchester County, the Bronx, and the subsurface of western Long Island (Queens and Brooklyn Sections, NYC Water Tunnel #3), the Hudson Highland-Reading Prong terrane, the Franklin Marble Belt and associated rocks, and the New Milford, Housatonic, Berkshire, and Green Mountain Massifs.]

~~~~~~~~~~~~~~~~~~~~Surface of unconformity~~~~~~~~~~~~~~~~~~~~

In New Jersey and Pennsylvania rocks older than the Franklin Marble Belt and associated rocks include the Losee Metamorphic Suite. Unconformably beneath the Losee, in Pennsylvania, Proterozoic X rocks of the Hexenkopf Complex crop out.

Table 03 – Proposed new classification of the Pleistocene deposits of New York City and vicinity (Sanders and Merguerian, 1998, Table 2)

Age	Till No.	Ice-flow Direction	Description; remarks
Late Wisconsinan ("Woodfordian"?)	I	NNE to SSW	Gray-brown till in Westchester Co., Staten Is., Brooklyn, & Queens (but not present on rest of Long Island); Hamden Till in CT with terminal moraine lying along the S coast of CT; gray lake sediments at Croton Point Park, Westchester Co.
Mid-Wisconsinan (?)			Paleosol on Till II, SW Staten Island.
Early Wisconsinan(?)	II	NW to SE	Harbor Hill Terminal Moraine and associated outwash (Bellmore Fm. in Jones Beach subsurface); Lake Chamberlain Till in southern CT.
Sangamonian(?)			Wantagh Fm. (in Jones Beach subsurface).
	IIIA	NW to SE	Ronkonkoma Terminal Moraine and associated outwash (Merrick Fm. in Jones Beach subsurface).
Illinoian(?)	IIIB		Manhasset Fm. of Fuller (with middle Montauk Till Member; in lower member, coarse delta foresets (including debris flows) deposited in Proglacial Lake Long Island dammed in on S by pre-Ronkonkoma terminal moraine.
	IIIC		
Yarmouthian			Jacob Sand, Gardiners Clay.
Kansan(?)	IV	NNE to SSW	Gray till with decayed stones at Teller's Point (Croton Point Park, Westchester Co.); gray till with green metavolcanic stones, Target Rock, LI.
Aftonian(?)			No deposits; deep chemical decay of Till V.
Nebraskan (?)	V	NW to SE	Reddish-brown decayed-stone till and -outwash at AKR Co., Staten Island, and at Garvies Point, Long Island; Jameco Gravel fills subsurface valley in SW Queens.
			Pre-glacial (?) Mannetto Gravel fills subsurface valleys.

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