and the changes in the mud bed can be inferred from changes in suspended sediment concentration, measured using one of the techniques described above. The shear force driving the circulation can be measured directly as the force needed to rotate the lid. Laboratory annular flumes generally rotate the flume and the lid in opposite directions, to minimize secondary radial circulation of the water, and to obtain a more uniform distribution of shear on the mud bed. An increase in suspended sediment corresponds to a decrease in the mass of sediment on the bed *m* and a decrease in suspended sediment to an increase in the mass of the bed. It is then possible to extrapolate these measured shear stresses to τ_c and τ_s, at which erosion or deposition ceases; and to interpolate the Parthenaides coefficient M_p and fall velocity *w* (Example Problem III-5-1).

(10) Calibration techniques. Ideally, one should be able to measure all the hydrodynamic properties of a cohesive sediment, and proceed directly to a model of the shore. The model, however, will always need calibration and verification against measurements of erosion and deposition, to fine tune measurements and confirm that the model represents the shore. With some of the hydrodynamic properties unknown, the model can be used to choose between high and low values of the unknown properties, using "design of experiments" techniques (Willis and Crookshank 1994). This still requires intelligent estimates of high and low values of the unknown properties and good field measurements of erosion, deposition, or transport.

III-5-7. Erosion Processes

a. Shear stress.

(1) The formulae for predicting the movement of cohesive sediment predict rates of erosion and deposition, not transport. Try putting a typical cohesive floc size into a noncohesive sediment transport formula (Part III-6) and you will predict virtually infinite transport rates, in which the predicted density of 'sediment in suspension' may exceed its real density on the bed. Noncohesive sediment formulae are generally based on transport limitations: assuming there is a sediment supply to match the transport potential. Cohesive sediment formulae are generally based on supply limitations, and assume the flow can transport all eroded sediment. They define the sediment exchange between the bed and the water column.

(2) The cohesive sediment formulae are also less theoretically based. They form a simple numerical framework for interpolating and extrapolating observed hydrodynamic behavior of cohesive sediment. Generally erosion or deposition is correlated with the excess shear stress.

b. Erodibility of consolidated sediments.

(1) There have been many studies of the erosion resistance of cohesive soils to flowing water. Very few of these investigations have considered the much more complex flow conditions encountered in the coastal zone. Nevertheless, a basic understanding, such as it is, of the complex process of erosion of consolidated cohesive soil provides a basis for assessing the erosion resistance of cohesive soils in the coastal environment.

(2) The erodibility of cohesive soils is controlled by the bonds between cohesive particles. Many tests of remolded cohesive sediments have found that the most important parameters in describing the erodibility of cohesive sediments include consolidation and physio-chemical conditions, both of which influence the degree of bonding between particles. Aside from the difficulty of obtaining intact cohesive samples, the reason for testing remolded and reconsolidated samples is to establish the erodibility of recompacted clays, used as construction materials. Extensive investigations into the relationships between the properties of these 'homogenous' cohesive sediments (e.g., consolidation and other geotechnical parameters), and the physio-chemical properties of the fluid (including temperature, pH, and salt or cation content) have found a direct relationship between these parameters and erodibility (Croad 1981; Arulanandan, Loganatham, and Krone 1975).

(3) More recent investigations of intact consolidated sediment samples have found that the natural structure of the material, including the presence of fissures, fractures, and seams of noncohesive materials such as silt and fine sand, is the most important factor in determining erodibility (Lefebvre, Rohan, and Douville 1985; Hutchinson 1986). The natural structure is uniquely defined by the environmental conditions during the original deposition and subsequent weathering of the sediment (including overconsolidation during glacial periods for some sediments). Conventional geotechnical parameters such as clay content and shear strength do not provide a direct measure of the influence of the natural structure of consolidated sediments as it relates to erodibility. Nevertheless, Kamphuis (1987) suggested that the presence of fissures in samples may be indirectly reflected in the undrained shear strength, consolidation pressure, and clay content.

(4) A technique for assessing the hydraulic erodibility of natural and engineered earth materials including both soil and rock is described by Annandale (1996). This empirical method, which provides a relationship between threshold stream power for erosion and an erodibility index, was developed from field observations of spillway performance downstream of dams. The erodibility index is determined as a scalar product of indices representing the following material properties: (1) mass strength, (2) block/particle size, (3) discontinuity/interparticle bond shear strength, and (4) shape of the material units and their orientation relative to the flow velocity.

(5) For clay materials as for mud, the erodibility index is primarily a function of the shear strength of the soil. Stream power is calculated as the product of near-bed velocity and shear stress. This approach was applied to the scour of weak rock in the presence of waves and currents to investigate scour potential around bridge piers (Anglin et al. 1996).

(6) In summary, in the absence of a reliable and standardized technique for assessing the natural structure of consolidated cohesive sediment, as it relates to erodibility in the coastal environment, more empirical approaches must be followed, such as establishing erodibility coefficients from laboratory tests or field data.

(7) One such example of a direct empirical technique for estimating erodibility is described by Kamphuis (1990) and Parson, Morang, and Nairn (1996). In these tests, intact (undisturbed) samples of consolidated sediment were placed in a drop section of the floor of a high-velocity unidirectional flow flume or tunnel with transparent walls or windows. The shear stress over the samples was determined indirectly by measuring the vertical profile of velocity just above the bed. The average erosion rate was then determined by measuring the volumetric erosion experienced on the surface of a sample within a test period. Rates were determined for velocities in the range of 0.5 to 3 m/sec. Further details on this test procedure are presented in Part III-5-6a. All results determined by using this technique for various types of consolidated sediment (including mudstone, till, and lacustrine clay) are summarized in Figure III-5-18 (from Parson, Morang, and Nairn (1996)). It was found that shear stresses in the range of 0 to 18 Pa resulted in erosion rates in the range of 0 to 8 mm/hr.

(8) In a further extension of this type of testing, Kamphuis (1990) found that the erosion rate increased dramatically when sand was added to the flow. The results of all of the erodibility tests using this technique with sand in flow are presented in Figure III-5-19 (from Parson, Morang, and Nairn (1996)). A comparison of the clear water erosion results of Figure III-5-18 and the sand in flow results of Figure III-5-19, indicates that for the same shear stress and sediment sample, the erosion rate is increased by a factor of 3 to 8 when

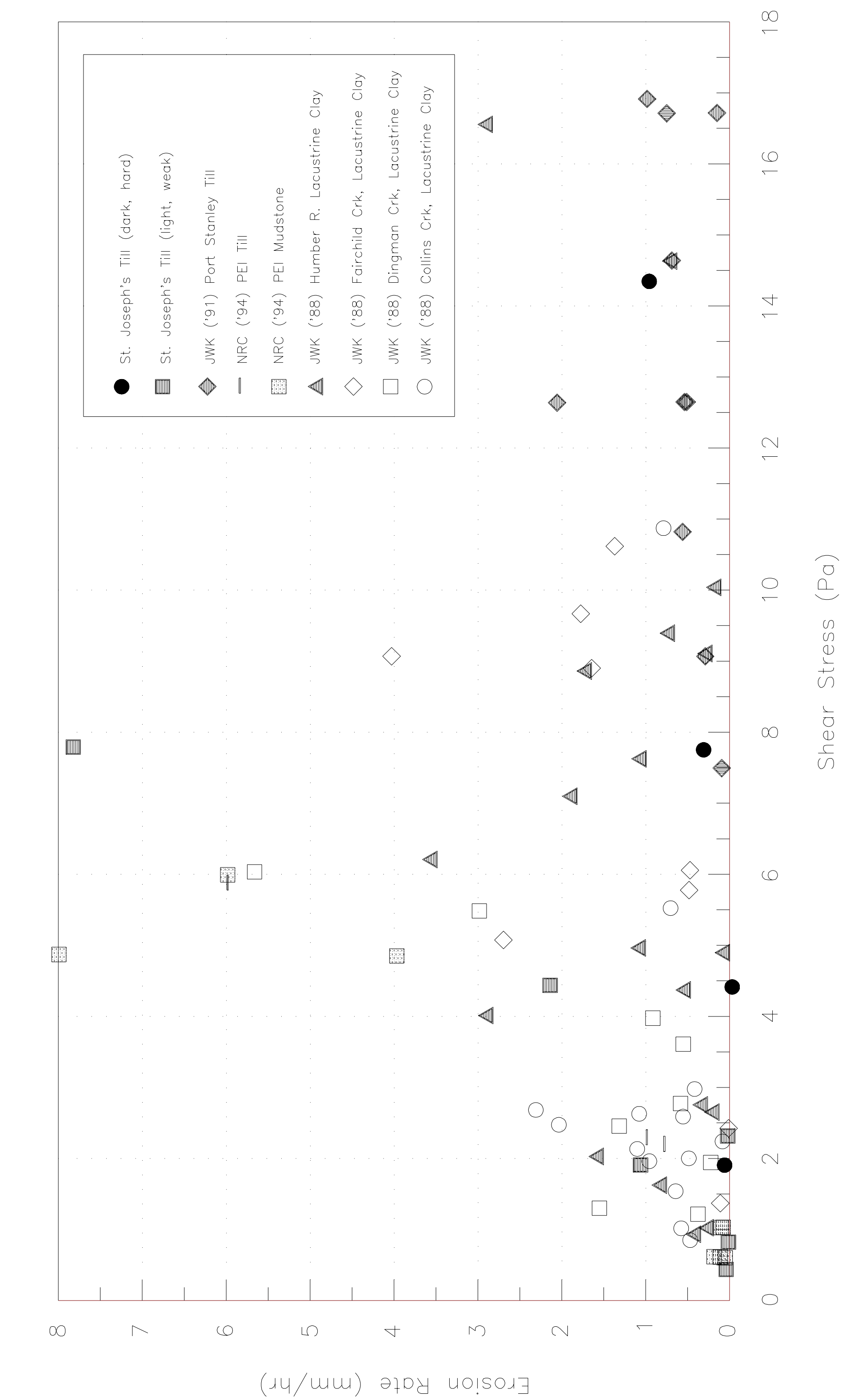

Figure III-5-18. Clear-water erosion rates from unidirectional flow flume and tunnel tests for various materials

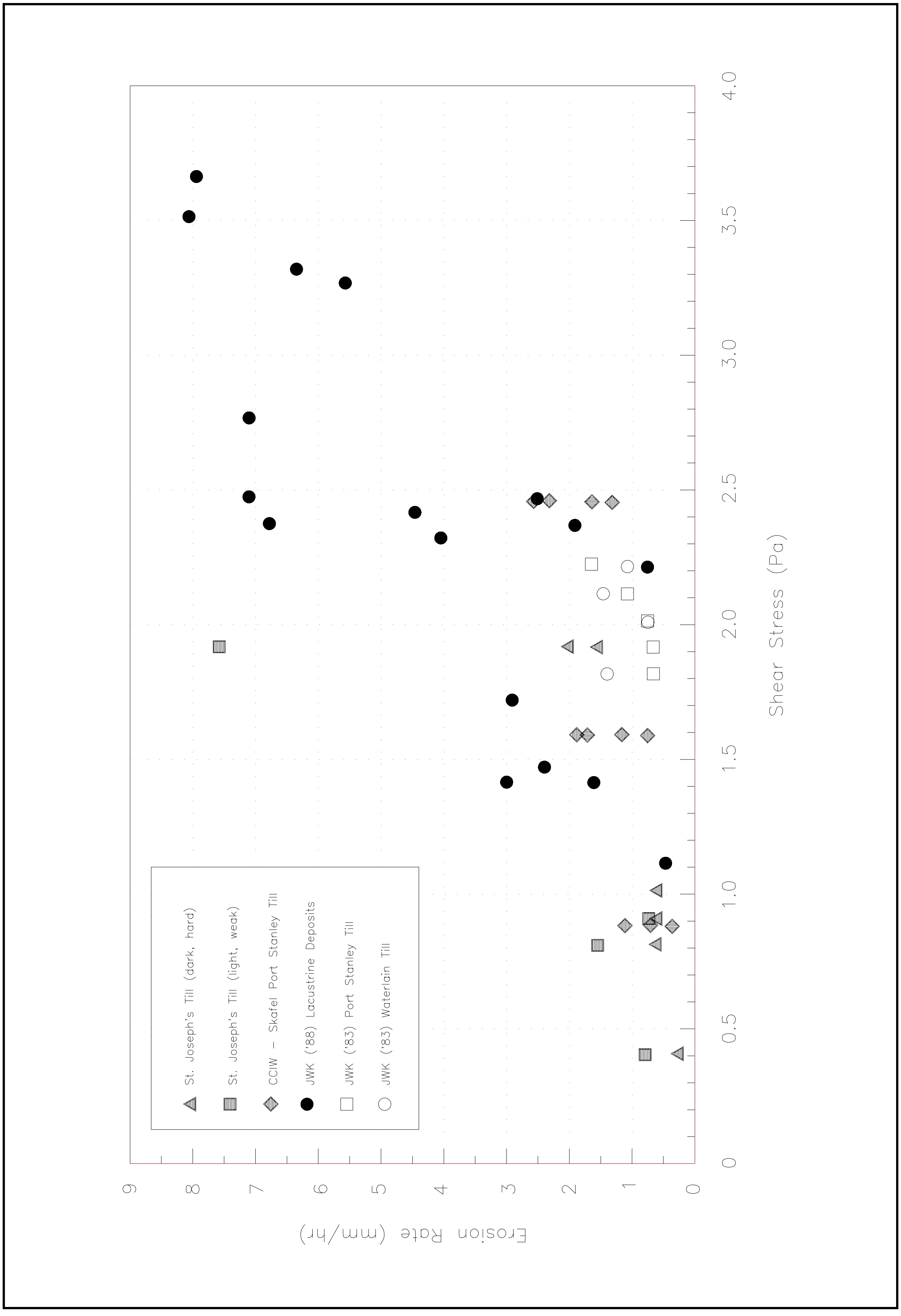

Figure III-5-19. Sand in flow erosion rates from unidirectional flow flume and tunnel tests for various materials

sand is introduced into the flow. It should be noted that the sand-grain size was selected so that sand transport occurred in a saltating (bed load) manner in the tests.

(9) A close review of the results presented in Figure III-5-18 reveals two populations of erosion response. The less erosion-resistant group consists of soft cohesive sediments as well as sediments characterized with a high degree of fracturing and/or the presence of silt seams. In contrast, the more erosion-resistant group consisted of firm, homogenous cohesive sediments. The shear strengths for the material tested ranged from 50 to 200 kPa.

(10) Some of the most recent research into the erodibility of cohesive sediments in the coastal environment has focused on the potential for softening of the exposed surface layer. Davidson-Arnott and Langham (1995) report that at the western end of Lake Ontario, previously overconsolidated till featured shear strengths in the range of 10 to 100 kPa. The softer sediment (10- to 20-kPa shear strength range) was associated with deeper water areas where the till was always exposed (i.e., the noncohesive sediment cover was insignificant) and where the frequency of erosion events was low. In contrast, in shallower areas (depths less than 3 m or 10 ft), the shear strengths were found to be in the range of 30 to 80 kPa. In the shallower depths, the lake bed is exposed to erosion events more frequently and sand cover thicknesses were greater. Using an adapted micro shear vane for a location in a water depth of 3.5 m, they found that the shear strength was lower in the upper 0.1 m of the lake bed and much lower in a 10-mm-thick surface layer. In fresh water, the softening process may be a result of cyclic loading by wave action, whereas in a seawater environment, factors such as salinity and biological activity (such as burrowing organisms) may also be important (Hutchinson 1986).

(11) Finally, an indirect approach to determining the erodibility of consolidated sediment at a specific site is through the use of numerical models to 'back-calculate' the erodibility coefficient through an analysis of environmental conditions (wave action and water levels) and the observed shoreline or lake bed erosion over a given period of time. This approach is discussed in more detail in Part III-5-12.

c. Subaqueous erosion of consolidated sediments.

(1) In this section, the underwater erosion process is described. In the previous section, we explained that the erosion of hard cohesive soils consists of the destruction of bonds between clay particles and the natural structure or framework created through consolidation of the soil matrix. Erosion of consolidated sediment is irreversible. Once the sediment, which often consists of 80 to 90 percent fines, is eroded, it cannot be reconstituted in its consolidated form in the littoral zone. The eroded fines (silt and clay) are winnowed, carried offshore, and deposited in deep water in contrast to the small fraction of sand and gravel, which remains in the littoral zone. Therefore, the erosion of cohesive sediment is fundamentally different from the erosion of noncohesive sediment. In the latter case, for every volume of eroded sand, a large portion of that material will be deposited somewhere in the littoral zone (in some specific instances, onshore or offshore losses of sand from beaches can occur). Therefore, on sandy shores, the process of erosion is reversible.

(2) An extensive study of nearshore profiles on the north central shore of Lake Erie described by Philpott (1984) revealed that the profile shape remained relatively constant over an 80-year interval despite dramatic shore recession. This led Philpott (1984) to conclude that the controlling process in bluff or cliff recession on cohesive shores is not restricted to wave action at the toe (as proposed by Sunamura (1992) for eroding rocky coasts) but by the erosion of the nearshore profile by waves. Boyd (1992) cites many earlier references that also suggest that the nearshore has a controlling influence on shoreline recession. The shoreward shift of the dynamic equilibrium profile implies that erosion or downcutting is proportional to the gradient of the nearshore profile and is, thus, greatest close to shore. Davidson-Arnott (1986) describes field measurements of downcutting for a till profile (through the deployment of micro-erosion meters across a

transect) at a site near Grimsby on Lake Ontario. The results confirm the hypothesis on downcutting, i.e., the rates increase towards the shore in a manner related to the local bed slope, thus allowing for the preservation of the profile shape as it shifts shoreward with time. The downcutting hypothesis has now been confirmed by many other field investigations including 9 years of profile retreat data at Maumee Bay State Park in Ohio (Fuller 1995). Hutchinson (1986) and Sunamura (1992) also note that the rate of lowering or downwasting of the intertidal platform on erodible rocky coasts probably determines the long-term rate of cliff retreat in most instances.

(3) In general, it has also been shown that the underlying cohesive profile, for cases where the properties of the cohesive sediment are uniform along the profile, follows an equilibrium profile shape as defined by Dean (1977). Kamphuis (1990) went on to show that the specific exponential shape of the equilibrium profile at a cohesive shore site was related to the grain size of the overlying noncohesive sediment. In other words, the profile shape could be determined using the grain size of the sand veneer to define the sediment scale parameter A (see Equation 3-15). The shape of the profile can also be influenced by other factors such as variable stratigraphy and the presence of lag deposits as discussed in Part III-5-5. In these cases, a smooth equilibrium profile with an exponential shape will not exist, at least not over the full profile. Riggs, Cleary, and Snyder (1995) note that in most instances along the North Carolina coasts, the complexity of the underlying stratigraphy is such that the profile rarely resembles the equilibrium form found on truly sandy shores.

(4) The downcutting process is illustrated in Figure III-5-20 for a cohesive shore site located east of Toronto along the Scarborough Bluffs. The bluff face has retreated approximately 30 m (100 ft) in a 37-year period. The underlying cohesive profile shape in 1952 is very similar to that in 1989; it has simply shifted shoreward by 30 m. Therefore, the long-term bluff or cliff retreat rate is equivalent to the profile retreat rate. This figure also shows that there can be a significant quantity of sand covering an underlying cohesive profile. The position of the underlying cohesive profile shown in Figure III-5-20 was estimated based on observations that the cohesive sediment is usually exposed or very thinly covered in the troughs between the bars. Also, it is known that the till is exposed at the toe of the bluff (i.e., at the back of the beach).

(5) Figure III-5-20 demonstrates that there is not a cross-shore balance of erosion and deposition. All of the eroded material from the cohesive profile and the bluff is either winnowed offshore (clay and silt fractions) or transported alongshore (sand and gravel fractions).

(6) The profile retreat model for cohesive shores implies that: the amount that the driving forces for erosion exceed the resisting forces is inversely proportional to the water depth. In other words, the most active subaqueous erosion occurs at the shoreline. In general, it may be assumed that the erosion resistance of the cohesive sediment is consistent across the profile (if anything, the sediment may be less erosion-resistant in deeper water due to the increased role of softening). Therefore, the driving force for erosion must increase in the shoreward direction. These observations provide important evidence on the nature of the driving forces for cohesive profile erosion.

(7) Coakley, Rukavina, and Zeman (1986) proposed that outside the surf zone, the downcutting process is driven by shear stresses generated by the orbital motion under waves. Outside the surf zone, this driving force is inversely proportional to water depth. However, considering that wave heights and the related orbital velocity decreases in the surf zone, this mechanism cannot explain the inverse relationship between depth and driving force in this zone. Nairn, Pinchin, and Philpott (1986) proposed that the complex combination of driving forces in the surf zone may be represented by the rate of energy dissipation (described by the rate of wave height decay). Using a model of wave energy dissipation for random waves, it was shown that the rate of energy dissipation was directly proportional to the rate of downcutting in the surf zone. In the surf

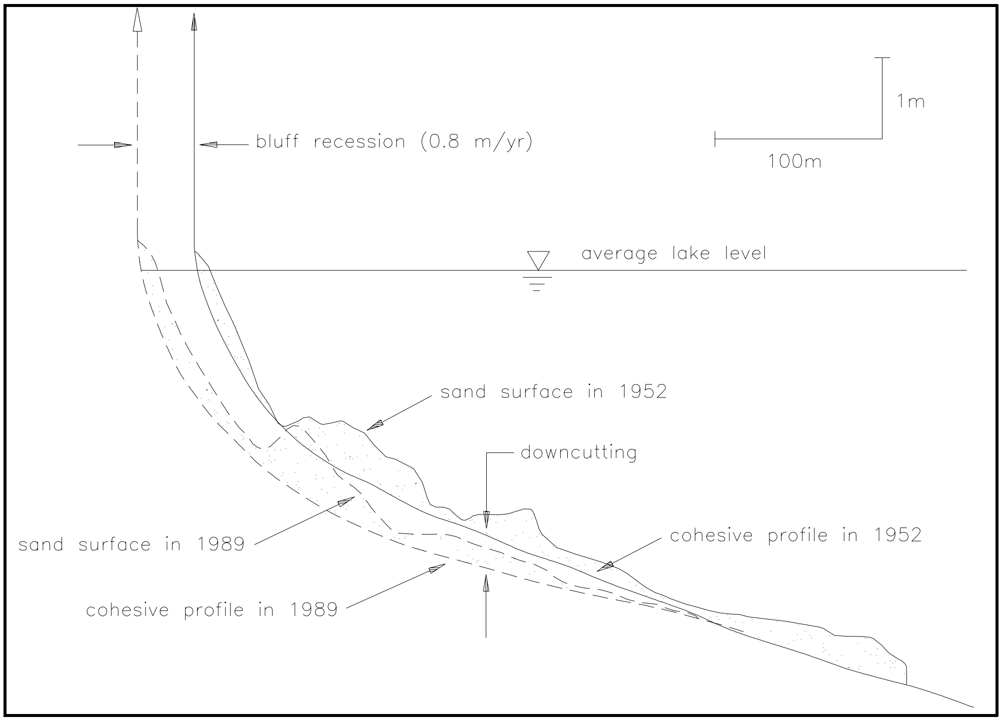

Figure III-5-20. Bluff retreat and profile downcutting over a 37-year period at Scarborough Bluffs, located east of Toronto on Lake Ontario

zone, the rate of energy dissipation provides a good indicator of the possible driving forces such as intensity of flows (e.g., undertow and alongshore currents) as well as the intensity of near-bed turbulence (i.e., plunging breakers, which generate high near-bed turbulence, are associated with rapid wave decay). The erosive nature of plunging breakers (and the associated near-bed turbulence) on cohesive profile erosion was demonstrated through laboratory experiments performed in a wave flume with intact cohesive samples (Skafel 1995).

(8) Sand acts as an abrasive agent in the erosion of consolidated shores. Referring to Figure III-5-20, it is also likely that the presence of alongshore bars can have the opposite effect of protecting the underlying cohesive sediment from exposure and subsequent downcutting. As these bars migrate with changing wave and water level conditions, different areas of the underlying cohesive profile become exposed in the troughs between the bars. The influence of the quantity and mobility of sand cover over a cohesive profile is explored in greater detail in Part III-5-2.

(9) At the boundary between the action of subaqueous and subaerial erosion processes, the bluff or cliff toe can in some instances experience notching. This notching typically occurs in more competent materials such as rock, frozen sediments, and harder cohesive sediments, (i.e., in materials that are capable of withstanding failure when undercut).

(10) In summary, the fundamental principles of consolidated cohesive shore erosion are:

(a) The erosion of consolidated cohesive sediment is irreversible.

(b) The long-term rate of shoreline retreat is directly related to the rate of nearshore downcutting and the associated profile retreat.

(c) The local rate of downcutting is proportional to the gradient of the nearshore slope at any location across the profile.

(d) In addition to acting as an abrasive agent that accelerates consolidated cohesive sediment erosion, sand can also serve to protect an underlying cohesive profile from erosion.

d. Subaqueous erosion of mud. The Parthenaides Equation for mud erosion (Parthenaides 1962, Mehta et al. 1989) is an excess shear equation describing the erosion rate of a cohesive sediment:

$$\frac{dm}{dt} = M_p \frac{(\tau_c - \tau)}{\tau_c} \qquad \text{(III-5-1)}$$

where

m = mass of sediment on the bed, kg/m^2 (lb/ft^2)

t = time, sec

τ = bed shear, Pa (lbf/ft^2)

τ_c = critical shear for erosion, Pa (psf) (lbf/ft^2)

M_p = 'Parthenaides Coefficient,' erosion rate at twice τ_c, kg/m^2/sec (lb/ft^2/sec)

e. Fluid mud.

(1) This assumes a well-defined interface between bed and water column: sediment on the bed remains at rest; while that in the water column moves with the water. Often, cohesive sediment forms an intermediate layer of fluid mud (denser than water; less dense than the bed; still capable of motion, but slower than the ambient flow). Fluid mud layers are frequently found near the shoreline, where wave activity can 'pump up' excess pore pressures within the fluid mud mass, slowing down drainage and consolidation. Hydrographers generally define fluid mud as having a density of less than 1,100 to 1,200 kg/m^3 (70 to 75 lb/ft^3).

(2) Erosion processes in fluid mud are similar to mixing processes at a salt-wedge interface. A layer of lighter fluid (water with suspended sediment) flowing above the denser fluid mud eventually induces waves in the interface when a critical Richardson Number or densimetric Froude Number is exceeded. Wind-wave activity in the upper layer increases the interfacial waves, diverting energy from the surface waves. As the difference in flow rates increases, the interfacial wave energy increases until breaking occurs, putting some fluid mud back in suspension and entraining clearer water in the fluid mud.

(3) The densimetric Froude Number is defined as:

$$F_r = \sqrt{\frac{(V_w - V_{fm})^2 \rho_{fm}}{g h_{fm} (\rho_{fm} - \rho_w)}} \qquad \text{(III-5-2)}$$

where

V_w = mean velocity in the water layer, m/sec (fps), above

V_{fm} = mean velocity in the fluid mud layer, m/sec (fps)

ρ_w = density of the water layer, kg/m^3 (lb/ft^3)

ρ_{fm} = density of the fluid mud layer, kg/m^3 (lb/ft^3)

g = acceleration due to gravity, m/sec^2 (ft/sec^2)

h_{fm} = thickness (depth) of fluid mud layer, m (ft)

(4) This reduces to the more familiar Froude Number at the water surface, where the upper layer is air of relatively negligible density and the lower fluid mud layer is the water. It defines the ratio of the velocity differential at the interface (V_w - V_{fm}) to the celerity of a small gravity wave in the interface. When that ratio is unity, the equivalent of a hydraulic jump (breaking wave) may be expected in the interface, with resultant entrainment of fluid mud in the water (erosion of the fluid mud) balanced by entrainment of water in the fluid mud.

(5) At the interface between the fluid mud and the bed, the fluid mud protects the bed from erosion. Shear that develops between the mud and the bed is generally too low to entrain stationary particles. The processes at this interface are entirely deposition, as water drains from the fluid mud.

f. Subaerial erosion processes.

(1) Subaerial erosion processes on cohesive shores do not necessarily have anything to do with the air or wind, although strengthening of mud flats has been noted by Amos et al. (1992) due to evaporation at low water in the macrotidal Bay of Fundy. On consolidated shores, the primary subaerial erosion process is slumping of oversteep bluffs or cliffs.

(2) As stated earlier, the long-term bluff or cliff retreat rate is determined by the rate of profile downcutting. In a review of shoreline erosion data from the Lake Erie shoreline, Kamphuis (1987) points out that cliff height does not exert much influence on the process (in fact, a distinct lack of correlation was noted) because erosional debris from a shore cliff is quickly swept away, winnowed offshore, and deposited in deep water. Exceptions to this generalization include locations where the debris is not easily removed from the toe of the cliff (e.g., in the case of eroded rock cliffs or blocks of frozen sediment along Arctic shores). The primary reason for slope failures along a cohesive shore is the oversteepened nature of the slope owing to the ongoing profile and toe erosion. Nevertheless, even though subaerial processes do not determine the long-term rate of shoreline recession on cohesive shores (i.e., the frequency of slope failures), these processes are critical in determining when and where a failure will occur.

(3) Slope stability is a function of the balance between the downward force of gravity and the strength of the geologic materials in a bluff or cliff. The strength of the geologic materials depends on the cohesion of particles and the presence or absence of groundwater. The stratigraphy of a bluff or cliff can have a significant influence on slope stability. Weak clay layers can provide slip planes for slope failures or serve to confine groundwater flows, which may appear as springs at the bluff face. Where groundwater exits the bluff face, seepage erosion can occur. Also, depending on the sequence of layering, groundwater flows can act to increase pressures within the slope and contribute to instability. This will often occur when seepage pathways at the bluff face are blocked by talus from a slide further up the slope. In some instances, the