Title

Coastal Sediment Transport Concepts and Mechanisms

Authors
Inman, Douglas L.
Masters, Patricia M.

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COASTAL SEDIMENT TRANSPORT CONCEPTS AND MECHANISMS

Douglas L. Inman
Coastal Morphology Group
Integrative Oceanography Division
Scripps Institution of Oceanography
dinman@ucsd.edu

and

Patricia M. Masters
Coastal Morphology Group
Integrative Oceanography Division
Scripps Institution of Oceanography
pmasters@ucsd.edu

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CONTENTS

LIST OF FIGURES

LIST OF TABLES

CHAPTER 5

COASTAL SEDIMENT TRANSPORT CONCEPTS AND MECHANISMS

5.1 OVERVIEW

This chapter identifies and develops the transport processes pertinent to the budget of sediment for a littoral cell. It is intended to provide a conceptual basis for the discussions which follow in Chapters 6 (Sources, Transport Modes, and Sinks of Sediment), 7 (Application of Shoreline Change Models) and 9 (Budget of Sediment and the Prediction of the Future State of the Coast).

The processes of primary interest here are the action of waves and coastal currents in eroding and terracing the land and in transporting sediment along the coast; the tractive forces of streams in eroding, transporting and depositing sediments; the effect of changing sealevel on beach erosion; and finally the influence of tides and streams in the maintenance and filling of coastal lagoons. For a more detailed discussion of these processes and their cumulative effects and interaction within the littoral cells of the San Diego Region, refer to CCSTWS 86-1 (Inman, et al., 1986).

5.2 LITTORAL PROCESSES

Historically, the principal sources of sediment for the California littoral cells were the coastal streams. Waves transported the sediment along the coast, and the main sinks for sediment were the submarine canyons.

Waves and the currents they generate are the single most important factors in the erosion, transportation and deposition of nearshore sediments. Waves mold beaches forming typical "summer" beach profiles in response to low waves and "winter" profiles in response to storm waves (Figure 5-1). Waves erode sea cliffs and cut terraces (Inman, 1983). When sediment is available, waves are effective in moving material along the bottom and in placing it in suspension for weaker currents to transport.

Parts of the Oceanside Littoral Cell are the most studied coastal sections in southern California. For this reason, many of the examples of coastal processes used in this chapter will be taken from this cell (e.g., Inman & Jenkins, 1983).

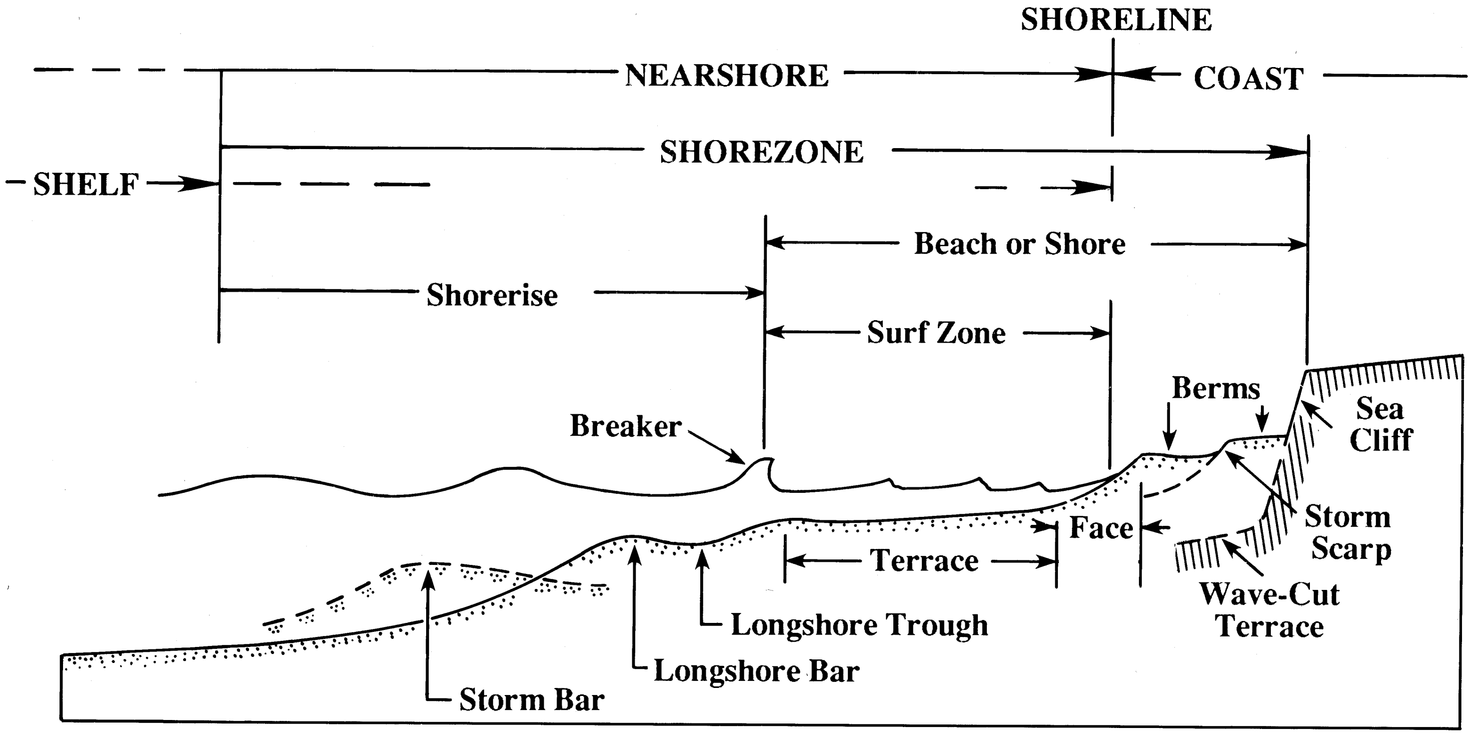

Figure 5-1. Nomenclature and schematic diagram for the summer profile of the shorezone of coasts with sea cliffs. Storms modify the beach profile as shown by the storm bar and storm scarp (after Inman, 1971).

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5.2.1 Wave-cut Terraces and Sea Cliffs

In the absence of beaches fronting sea cliffs, the direct force of the breaking waves erodes the cliffs and forms coastal terraces. The rising and lowering sealevels during the Pleistocene epoch caused the seas to transgress and regress across the land, both eroding and depositing material (Inman, 1983). Erosion is most pronounced during relative stillstands or pauses in the transgressive/regressive cycles. The signature for the sea's presence at a relative stillstand is usually in the form of a wave-cut terrace of gently sloping terrain, backed by sea cliffs when the near stillstand has been prolonged, as at present.

Active cliff erosion still occurs during severe winter wave conditions at many locations along the southern California coast. And, in the absence of beaches, the erosion products from sea cliffs supply sand to the littoral cell. Shepard & Grant (1947) found that wave erosion of the consolidated rocky coasts of southern California had been negligible during the preceding 50 years. On the other hand, they found a retreat of as much as a foot per year in unconsolidated formations. Based on a comparison of old subdivision maps, Kuhn & Shepard (1984) claim that the sea cliff at Encinitas retreated more than 180 meters (600 feet) between 1883 and 1891. Even though this is known to have been a very stormy period, the rate of sea cliff retreat of 23 meters per year (74 feet per year) appears to be an extreme example.

The wave-cut terrace associated with the sea cliffs at La Jolla is shown in cross-section in Figure 5–2. The decrease in the slope of the wave-cut terrace to one degree, beginning about 200 meters (650 ft) seaward of the sea cliff and at a terrace depth of 4 to 5 meters (13 to 16 ft) below mean sealevel probably represents the terracing that began about 6,000 yrs BP at the beginning of the slow (15 cm/century) rise to present sealevel (Inman, 1983).

Borings show that a wave-cut terrace also occurs at the base of the sea cliff and under the modern beach sand at many coastal locations (Figure 5–2). At Oceanside, the sea cliff is about 11 meters (35 ft) high and occurs just seaward of Pacific Street (Artim, 1981). Within the past two centuries, and during times of intense wave action and little sediment discharge from rivers, the beach was eroded back to the sea cliffs. Following periods of major flooding, the sandy deltas of the Santa Margarita and San Luis Rey Rivers built the beach seaward, forming a wide backshore area between the sea cliff and the beach berm. Photographs taken in 1916 show the sand delta of the San Luis Rey River extending out almost a pier length beyond the sea cliffs.

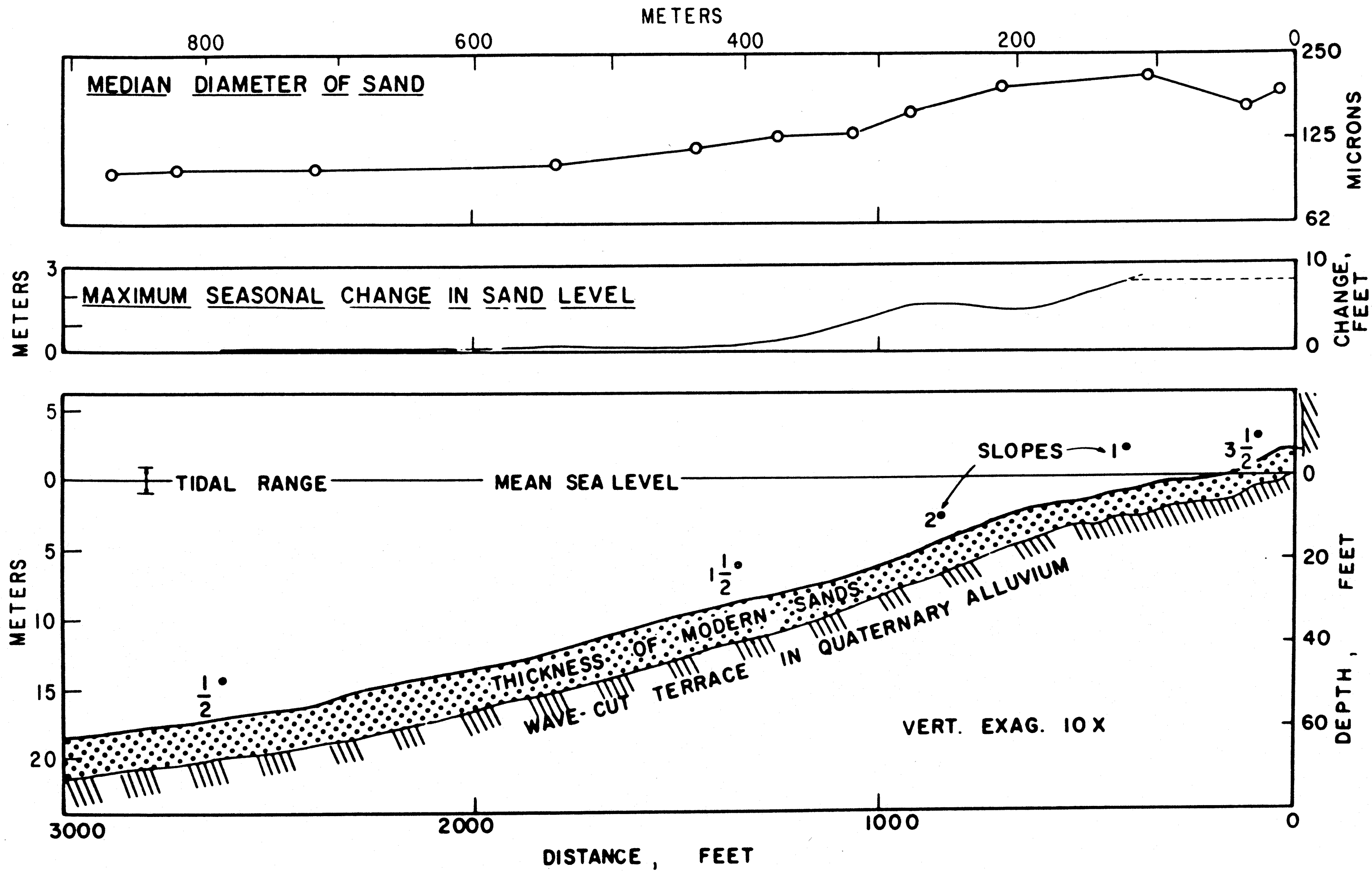

Figure 5-2. Profile of a fine sand beach at La Jolla, California, showing the sand size and the thickness of the modern nearshore sand over the wave-cut terrace (from Inman & Bagnold, 1963).

5.2.2 Formation of Beaches

Wherever there are waves and an adequate supply of sand or coarser sediment, beaches form. The initial and most characteristic event in the formation of a new beach from a heterogeneous sediment is the sorting out of the material, with coarse material remaining on the beach and fine material being washed away. Concurrent with the sorting action, the material is rearranged, some being piled high above the water level by the runup of the waves to form the beach berm, some moved back and forth by the swash to form the beach face, some carried back down the face to form the terrace that is characteristic of beach surf zones (Figure 5–1).

The action of waves on an inclined bed of sand eventually produces a beach profile that is in dynamic equilibrium with the energy dissipation associated with the oscillatory motion of the waves over the sand bottom. When a beach slope exceeds the natural equilibrium slope, an offshore transport of sand results and the beach slope flattens. Conversely, if a slope is less than the natural equilibrium slope, a shoreward transport of sand will result, and the beach slope will steepen. A dynamic equilibrium slope is attained when the up-slope and down-slope transports are everywhere equal. Such a beach profile is known as an equilibrium energy profile (Inman & Bagnold, 1963; Inman & Frautschy, 1965; Dean, 1977).

Many factors, such as rip currents and the presence of structures and promontories, affect local beach slopes. However, in general on long beaches composed of fine or medium-size sand, the following general description of the beach profile applies. The equilibrium beach slope steepens with the increasing onshore directed bottom stress that is associated with shoaling waves. The slope is usually gentle in deeper water over the shelf and steepens into the characteristic "shorerise" where the onshore stress is greatest just before the wave breaks. The slope decreases at the break point and is gentle over the terrace and longshore bar. The bore from the breaking wave traverses the gentle outer terrace, causing it to gradually steepen until it reaches the beach face where the remaining energy from the breaking wave is dissipated in the swash and backwash. The beach face is the steepest portion of the beach profile.

<u>Seasonal Beach Cycles</u>

Changes in the character and direction of approach of the waves cause a migration of sand between the beaches and deeper water. In general, the beaches build seaward during the low waves of summer and are cut back by higher, steeper winter storm waves (Figure 5–3). There are also shorter cycles of cut and fill associated with spring and neap tides and with nonseasonal waves and storms.

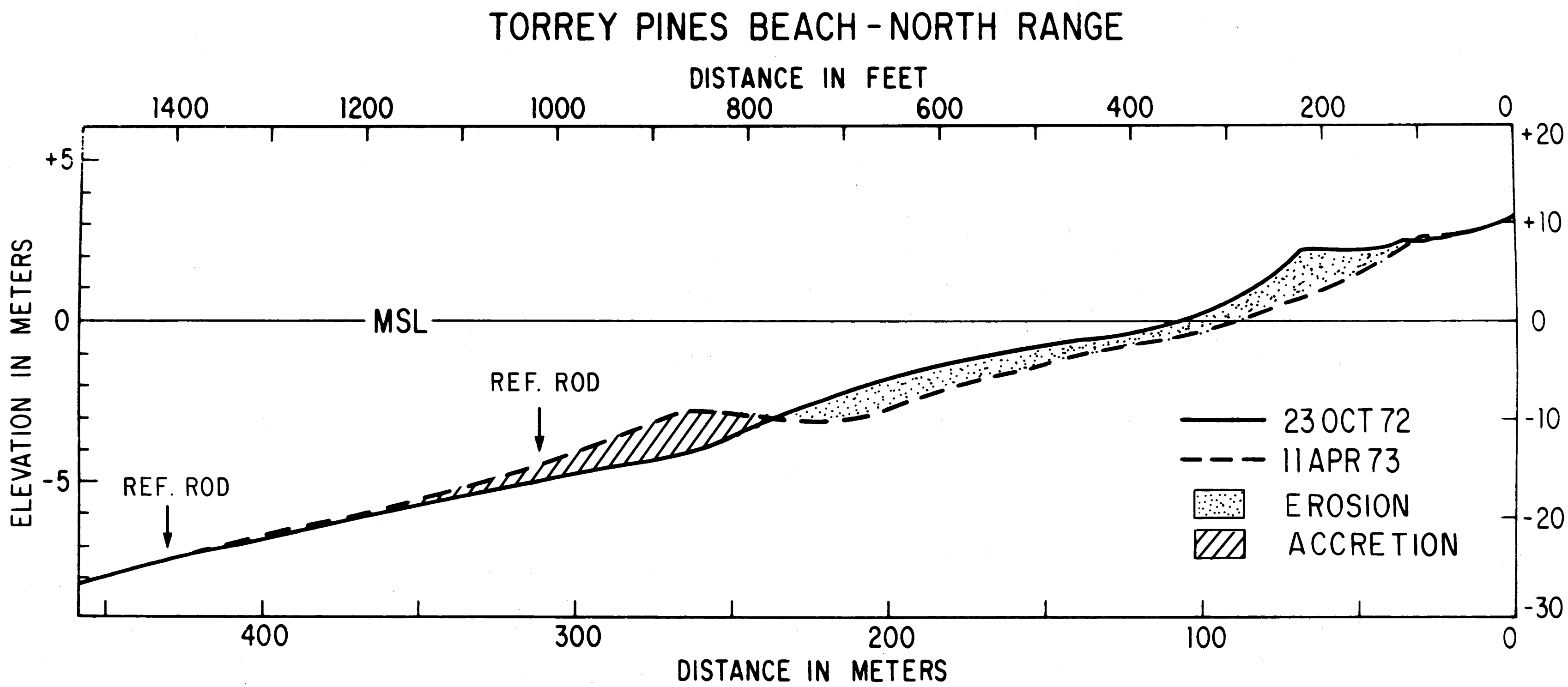

Figure 5-3. Comparison of end of summer (23 Oct 72) with end of winter (11 Apr 73) profiles at Torrey Pines Beach, California. Arrows indicate the positions of reference rod stations (from Nordstrom & Inman, 1975).

Bottom surveys indicate that most offshore-onshore interchange of sand occurs in depths less than about 13 meters (43 ft) but that some effects may extend to depths of 30 meters (100 ft) or more (Nordstrom & Inman, 1975).

More extensive measurements extending over periods of several years, using the technique of fathometer corrections derived from bottom reference rods, have been made off Torrey Pines Beach, California. Monthly measurements at Torrey Pines clearly show seasonal changes. The more gentle beach face slope that occurs for fully developed winter profiles is in contrast with the wider berm and steeper beach face of the summer profiles (Figure 5–3).

The seasonal change in beach width at Torrey Pines, from summer to winter berm, ranges from about 25 m to 70 m (80 to 230 ft) and averages about 38 m (125 ft). The volume changes associated with the seasonal beach changes range from about 60 to 130 m^3 per meter length of beach (72 to 156 yd^3/yd) and average about 92 m^3/m (110 yd^3/yd) (Nordstrom & Inman, 1975).

Long-Term Beach Changes

In addition to the seasonal beach changes, two types of longer term shoreline changes are important to the budget of sediment for a coastal segment. These are the shoreline changes due to local accretion/erosion and the changes associated with sealevel rise. With sealevel constant, local net accretion (or erosion) of the beach causes a corresponding progradation (or recession) of the shoreline, Δx (Figure 5–4a). Following Bruun's rule, both a shoreline recession and a redistribution of sand in the shorezone are assumed to be associated with a sealevel rise, Δz (Figure 5–4b) (Schwartz, 1967; Bruun, 1983). These changes are difficult to separate as they often occur together. However, since their causes are different, they are treated separately.

The analysis assumes that the shape of the shorezone portion of the profiles remains unchanged. For the case of constant sealevel, it is assumed that accretion or erosion causes the beach profile to be displaced a horizontal distance Δx (Figure 5–4a). For rising sealevel, it is assumed that the profile is raised vertically a distance Δz, then moved horizontally a distance Δx such that the volume eroded from the upper beach (hatched) is equal to the volume deposited further offshore along the shorerise (Figure 5–1).

Consider the case of constant sealevel and a horizontal recession of the shoreline Δx as in Figure 5–4a. The area between two simple offset translational profiles converging asymptotically at the depth of closure h can be approximated by the product of the horizontal displacement and the vertical distance Z over which the erosion occurs (e.g., Inman & Dolan, 1989). In this case, $Z=h+c$, where c is the height of the beach berm crest above MSL.

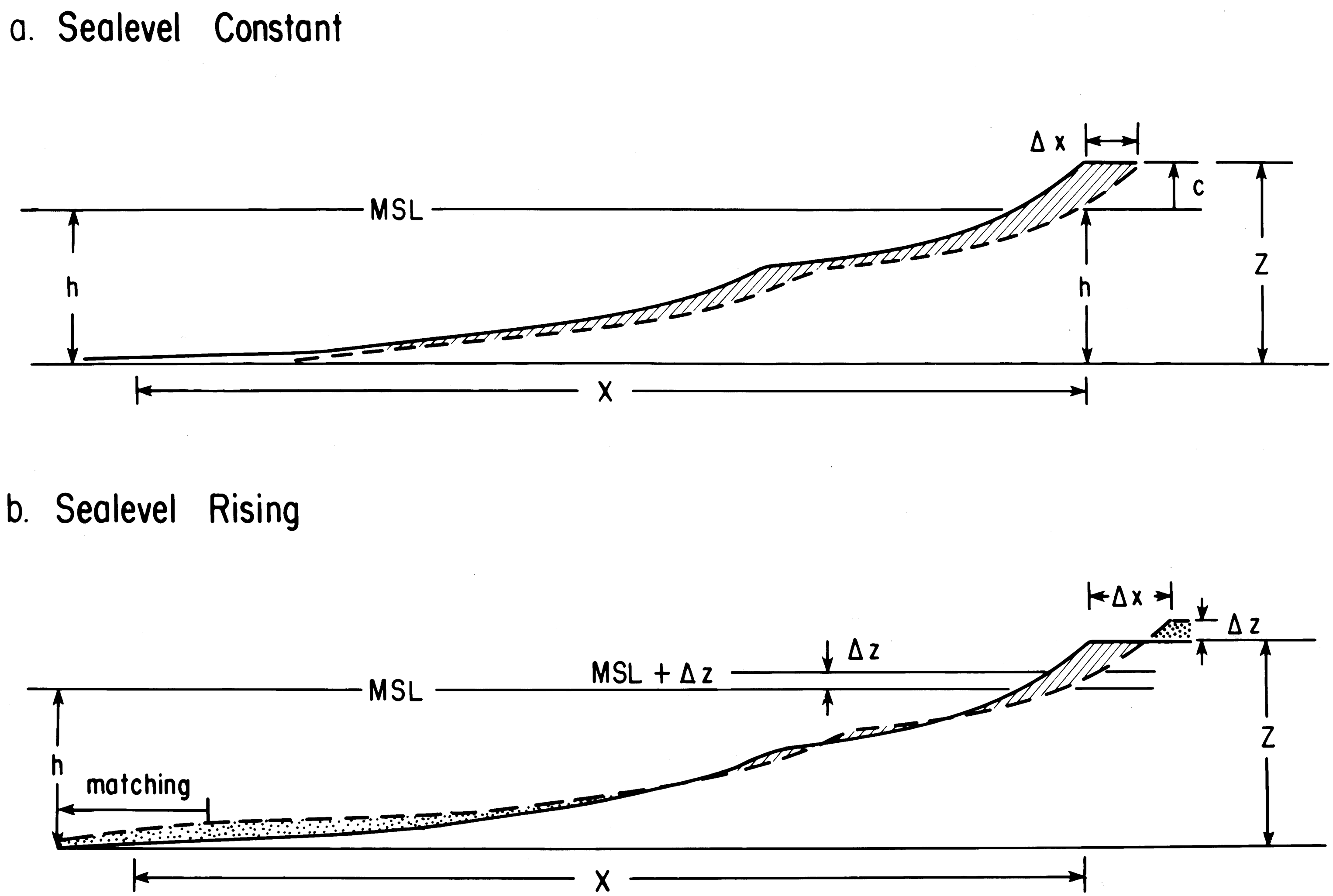

Figure 5-4. Schematic diagrams of beach profiles showing the volume of sand associated (a) with net recession of the shoreline, Δx, and (b) with sealevel rise, Δz. In both cases it is assumed that the equilibrium beach profile remains constant (from Inman & Dolan, 1989).

Therefore the volume change q_c per unit length of beach (i.e., m^3/m) will be

$$q_c \simeq \Delta x \tag{5-1}$$

From equation (5-1) it is apparent that the volume-equivalent factor for converting unit distance of shoreline change (x = 1 m) to volume change is

$$q'_c = q_c / \Delta x \equiv Z \tag{5-2}$$

where q'_c has units of length (i.e., m^3/m^2), and Z is the closure height.

When sealevel rises a vertical distance Δz, the beach profile is adjusted upward such that the area between the two profiles is given by the product ΔzX, where X is the horizontal distance over which the adjustment occurs. Bruun (1962; 1983) assumed that the beach profile would remain constant, and that a redistribution of material would occur such that the amount eroded from the upper beach would equal that deposited offshore. Thus the volume change of relation (5-1) was set equal to the volume ΔzX, and the resulting shoreline recession due to sealevel rise becomes

$$\Delta x_r \simeq \Delta zX/Z = \Delta z/\tan\beta \tag{5-3}$$

where $\tan\beta = Z/X$ is the extended beach slope. This relation is known as Bruun's rule (e.g., Schwartz, 1967; Hands, 1983). The volume of material associated with this redistribution of the profile is given approximately as

$$q_r \simeq \Delta xZ/2 = \Delta zX/2 \tag{5-4}$$

where a factor of about ½ enters because the erosion which occurs in the upper portion of the profile is balanced by the accretion which occurs in the lower portion (Figure 5-4b).

The parameters necessary for evaluating long-term beach changes are not well known, nor are they generally agreed upon. The one exception is the rate of sealevel rise, which for the San Diego region has been about 20 cm/century (0.66 ft/century) during the past 60 years (Chapter 4.2; see also Barnett, 1984; Flick, 1988; Hicks & Hickman, 1988). For illustrative purposes, estimates of long-term beach changes based on studies conducted at or near Torrey Pines Beach are tabulated in Table 5-1. The estimate of Z = 15 m (49 ft) may be a reasonable long-term average, but it will be too large for mild years and may be too small for the most severe years (refer to Chapter 9).

It is to be noted that the volume-equivalent factor in Table 5-1 of 15 m^3 per meter length of beach per meter of erosion(Δx=1 m)

TABLE 5-1
ESTIMATES OF VOLUME-EQUIVALENT FACTOR[a]

a. Sealevel Constant					b. Sealevel Rise $\Delta Z = 0.002$ m/yr	
h (m)	Z (m)	X (m)	$\tan\beta = Z/X$	q'_c (m^3/m^2)	Δx_r (m/yr)	q_r ($m^3/m \cdot yr$)
11	15	650	0.023	15	0.09	0.65
					$\Delta Z = 0.0066$ ft/yr	
(ft)	(ft)	(ft)	$\tan\beta$	(yd^3/yd^2)	(ft/yr)	($yd^3/yd \cdot yr$)
36	49	2130	0.023	16	0.29	0.79

[a] Volume-equivalent factor $q'_c \equiv Z$ (closure height) as defined in Figure 5-4 and equations 5-1 through 5-4.

Source: Nordstrom & Inman (1975) for Torrey Pines Beach.

($16\ yd^3/yd^2$), requires that there be a horizontal translation of a beach profile of identical shape. Based on the measurements of Nordstrom & Inman (1975), the seasonal beach changes which involve variations in beach slope are very large compared to any net long-term changes: factors of 38 times larger in beach width and over five times larger in volume. Therefore, the seasonal "noise" will generally obscure attempts to evaluate long-term beach changes, particularly when comparisons are developed from a short period of record. This is especially true along the cliffed coasts of the San Diego region, where waves directly attack the seacliffs during severe winters.

The present rate of sealevel rise is expected to result in a net horizontal recession of the beach of about $\Delta x_r = 0.09$ m/yr (0.3 ft/yr) and a very small annual volume redistribution of less than 1 m^3/m/yr (1 yd^3/yd/yr). Both effects will in general be obscured by seasonal beach changes, except over long periods like a century. The effect of sealevel rise may be more apparent in terracing and cliff erosion than in beach erosion, as suggested by Figure 5-2.

In summary, a study of beach profiles of the San Diego Region by Inman, et al. (1993) and Inman & Masters (n.d.) showed that, for seasonal beach changes associated with years of normal wave climate, the closure height Z = 14 m (45 ft). However, the study found that there was no meaningful value of Z for long-term beach changes that included exceptionally stormy years. During the cluster storms of 1982/83, the beach profiles were in a state of disequilibrium, and erosion of the shorerise extended beyond the depths of measurement (refer to Sections 9.4 through 9.6).

5.2.3 Momentum Flux

The fluxes of momentum associated with waves in and near the surf zone are responsible for a number of important phenomena. The onshore flux causes a setdown of water level that reaches a maximum at the breakpoint, and a setup of water level on the beach face. The longshore component of momentum flux drives the longshore current (e.g., Bowen, 1969) and is important in the prediction of the longshore transport of sand (refer to section 5.3.3).

Relations for the momentum flux tensor in shallow water have been detailed in a number of papers (e.g., Lundgren, 1963; Longuet-Higgins & Stewart, 1964). It is most commonly referred to as the radiation stress. The components of the radiation stress of interest here are:

$$S_{xx} = En\cos^2\alpha + E(n - \tfrac{1}{2}) \simeq 3E/2 \qquad (5\text{-}5)$$

$$S_{yx} = En\cos\alpha\sin\alpha \qquad \text{(conserved)} \qquad (5\text{-}6)$$

where S_{xx} is the onshore flux (x-direction) of onshore directed momentum (x), and S_{yx} is the onshore flux (x) of longshore directed momentum (y) with units of joules/m (ft-lb$_f$ / ft^2). Here, $E = (\rho g H^2)/8$ is the energy per unit surface area of waves of height H, n is the ratio of the group velocity Cn to the phase velocity C of the waves, and α is the angle the wave ray makes with an orthogonal to the shoreline. As indicated in equation (5-6), S_{yx} is conserved during shoaling and refraction.

5.3 BASIC TRANSPORT RELATIONS

Sediment transport mechanisms are critically dependent upon the size of the particles in sedimentary material. Sedimentary solids initially consist of individual, separable particles that are collectively referred to as clastics. From the standpoint of their transport, and their bulk deformation, the clastics can be conveniently divided into muds and granular material. Muds are composed of the small sized particles of silt and clay that form the suspended "wash" load of rivers. They are generally deposited in the relatively deep waters of the outer shelf and slope and in the calm waters of estuaries. In the aggregate, muds are soft, cohesive (sticky) and tend to deform plastically.

Granular material is the cohesionless sand and coarser material that is commonly found on beaches and on the shallow water portions of the shelf that are subject to the action of waves and currents. We are primarily concerned with granular material of sand size (62 to 2,000 μ; 0.062 to 2 mm), which constitutes the "mobile" portion of the sediment transported in nearshore waters.

Sediment transport begins with the onset of grain motion at the bed. The classical relation for the onset (threshold) of grain motion under unidirectional flow gives the onset as directly proportional to the grain diameter, D, and the fluid stress (e.g., Raudkivi, 1976). However the onset criterion for sand motion under shallow water waves is more complex. The experimental laboratory work of Bagnold (1946) and its theoretical analysis by G.I. Taylor (1946) show that the onset relation has the form

$$\theta_t = \text{const } St^{3/5}\, Re^{1/5} \qquad (5\text{-}7)$$

where $\theta_t = \rho u_m^2/(\rho_s - \rho)gD$ is the wave form of the Shields number, $St = d_o/2D = u_m/\sigma D$ is the Strouhal number, and $Re = u_m D/\nu$ is the grain Reynolds number. In these relations ρ and ρ_s are the fluid and grain densities, u_m is the amplitude of the maximum orbital velocity, $\sigma = 2\pi/T$ is the radian frequency of waves of period T, and d_o is the orbital diameter. The constant in relation (5-7)is evaluated as 0.10 for Bagnold's experiments on quartz sand with diameters of 160 μ, 360 μ, and 800 μ. The constant is 0.087 for the experiments of Dingler & Inman (1976) which were based on field and laboratory experiments on natural nearshore sands with median diameters ranging from 180 μ to 400 μ. Figure 5–5 is a graph of onset velocities from equation (5–7) using an average value of 0.094 for the constant.

5.3.1 Crosshore Transport

The seasonal beach cycles described previously are examples of crosshore transport associated with the adjustment of the beach profile to changes in the intensity of wave energy. It was shown that these adjustments in beach profile are associated with crosshore volume changes ranging from about 10 m^3/m (12 yd^3/yd) per day to 100 m^3/m (120 yd^3/yd) per summer/winter seasonal change (Inman, 1987).

Basic principles for crosshore transport are closely related to the concepts leading to the various types of "equilibrium" beach profile. The term has been applied indiscriminately to mean either a final long-term profile (geology) (e.g. Fenneman, 1902) or a shorter term profile that achieves equilibrium with "prevailing" wave conditions (e.g. Johnson, 1919). Models for shoreline change require an on- or offshore translation of a profile of constant form that has the shape of some average, long-term profile of the beach. Accordingly, we will refer to the long-term profile as a translational profile and the short-term profile as an equilibrium profile.

Short-Term Crosshore Transport

The first quantitative measurements of crosshore transport on an ocean beach were those of Shepard & LaFond (1940), made from the Scripps Institution pier at La Jolla, California. The same year, Evans (1940) published a study of beach changes along the shores of the Great Lakes. The measurements of Shepard and LaFond clearly showed the profile response to day-to-day changes in wave conditions as well as those associated with storms and seasonal wave climate. They measured changes in sand level as great as 1 m (3.2 ft) during a 24 hour period.

Since the original study of Shepard & La Fond (1940), profile changes have been documented at numerous locations in the San Diego region (e.g. Shepard, 1950, b; Nordstrom & Inman, 1975; Aubrey, et al., 1976). An extensive data set covering the entire region has been collected under the Coast of California Storm and Tidal Waves Study. The profiles were obtained between 1983 and 1988, generally on a semi-annual basis, at up to 57 coastal transect locations (see Chapter 3).

Several models have been developed to explain features of the equilibrium profile and the beach response to changing wave

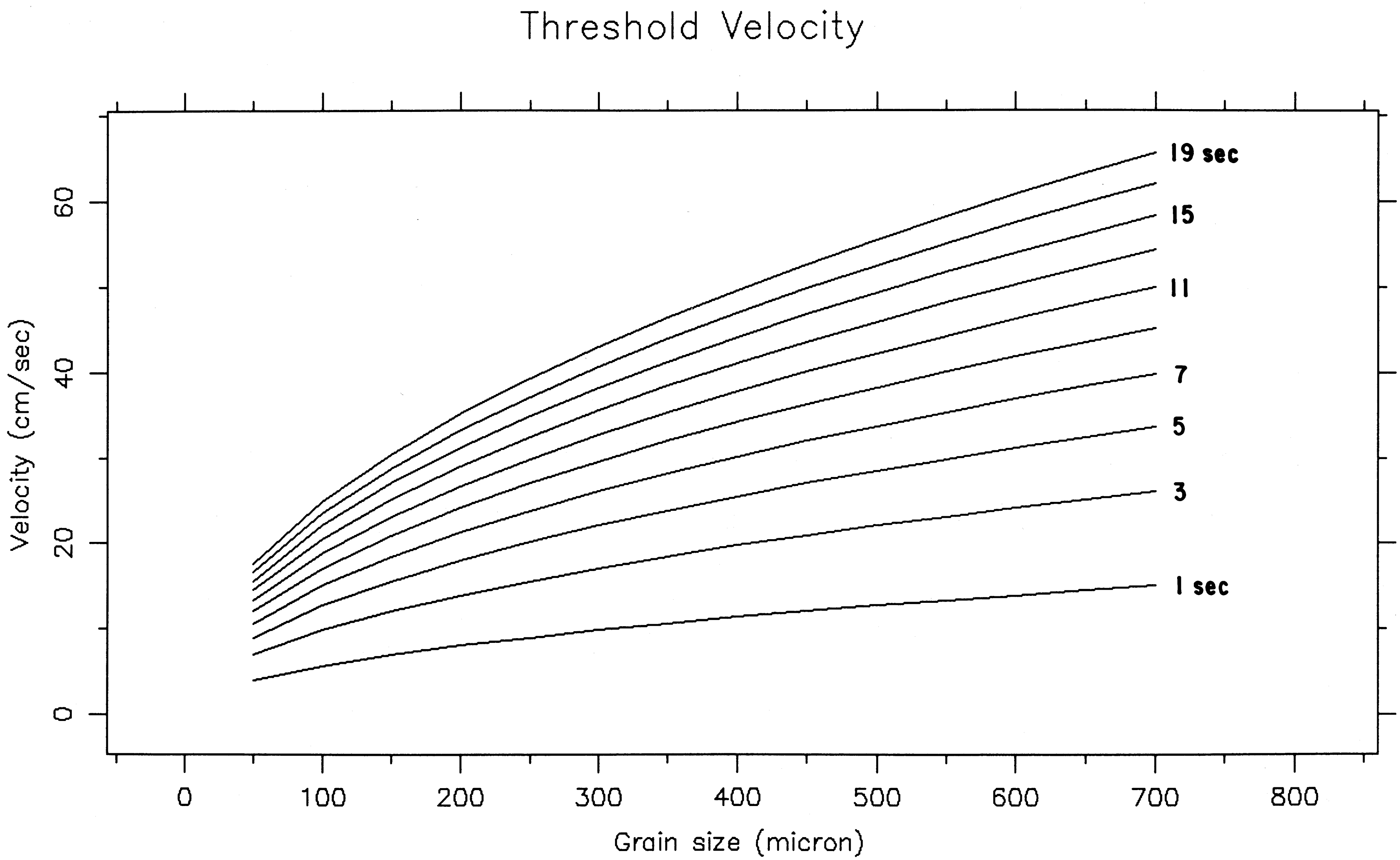

Figure 5-5. Threshold velocity u_{mt} for onset of motion of quartz grains under waves of various period. Equation (5-7) using data of Bagnold (1946) and Dingler & Inman (1976) with a constant equal to 0.094.

conditions. Grant (1943) points out that there is a differential between the on- and offshore velocities of shallow water waves, and a net shoreward transport of water is associated with their shoaling. Various combinations of the above concepts have been used to explain the complex features in the profile. These dynamic concepts were further developed and sometimes referred to as the "null" point hypothesis (e.g., Miller & Zeigler, 1958; Eagleson, et al., 1958; Johnson & Eagleson, 1966). However, as discussed by Bowen (1980), the null-point hypothesis for beach equilibrium "provides theoretical predictions seriously at odds with reality."

Long-Term Crosshore Transport

The long-term shapes of beach profiles have been studied extensively. Krumbein (1944) studied the relation of wave energy to beach slope and grain size at Half Moon Bay, California. Keuligan & Krumbein (1949) concluded that a shoaling solitary wave would result in a stable profile of the form

$$h = A\,x^m \qquad (5\text{-}8)$$

where h is depth, A is a proportionality constant, x is distance from shore, and the exponent m has a value of 2/5.

Bruun (1954) made an extensive study of beach profiles along the North Sea coast of Denmark and a somewhat limited study of profiles at Mission Beach, San Diego, California. He found that, on average, the above relation held for stable configurations with values of A and m of 0.20 and 2/3 for Denmark and 0.22 and 2/3 for Mission Beach. Dean (1977) found that the average beach profile along the Outer Banks of North Carolina followed the same form with values of A and m of 0.13 and 0.71, respectively. His profiles extended to depths of about 6 m (20 ft). However, considering the shorerise and the bar-berm portions of the beach as adjacent but separate curves gives values of m of about 0.4 for the beach profiles in the San Diego Region (see Chapter 9).

Net onshore transports of sand are known to have occurred on a world-wide basis where the nearshore bottom has gentle slopes. The coastal barriers of Holland were formed by onshore transport across the shallow bottom of the North Sea (e.g., Van Straaten, 1965). In this case the barriers appear to have built seaward until the increasing slopes terminated the onshore transport of sand. The barriers and extensive dune fields bordering Sebastian Vizcaino Bay, Pacific coast of Mexico, were formed from the onshore transport of sand across the shallow sea floor (Inman et al., 1966). Analysis of the budget of sediment led Bowen & Inman (1966) to conclude that the beaches north of Point Arguello, California, receive sand from the shelf. Clemens & Komar (1988) conclude, from an analysis of grain rounding and heavy mineral distribution, that some Oregon beaches are supplied by relict sand from the shelf.

However, there is controversy about the role that sediments on the nearshore portions of the shelf may have in the supply of sand to the beach. According to Moore & Curray (1974), this continuing controversy is "one of the most important problems in sedimentary geology." It is also fundamental to many aspects of coastal engineering (e.g., Dean, 1987).

Mechanisms for the offshore transport of sand have not been clearly demonstrated for broad-shelf coasts where there are no nearshore submarine canyons. However, it has been suggested that offshore transport of sand results from strong, downwelling bottom currents. The quantitative effects of a downwelling mechanism are unknown. Since the net bottom stresses due to wave motion are onshore, to be an effective offshore transport mechanism the downwelling effects must extend into water deeper than the shorerise which is formed by wave action. Thus it would appear that the potential effectiveness of a downwelling mechanism would be greatest along wide-shelf coasts that are subject to short period surface waves, a condition typical of the Atlantic and Gulf seaboards of the United States.

5.3.2 Longshore Transport

Traditionally, the amount of material trapped by coastal structures, such as jetties and breakwaters, divided by the time of trapping has been used to obtain the rate of longshore transport. In this case it must be recognized that there may be both up- and downcoast transports with rates indicated by Q_u and Q_d respectively. Their sum is the gross transport rate Q_g and their difference is the net transport rate Q_ℓ,

$$\text{gross} \qquad Q_g = Q_d + Q_u$$

$$\text{net} \qquad Q_\ell = Q_d - Q_u$$

These rates may change with season and with the time interval used to establish the rate. But in general at Santa Barbara Harbor, the west to east transport dominates to the extent that over periods of a year or more the net and gross transport rates are essentially equal (Johnson, 1953; Dean, 1989). At Oceanside, the annual downcoast rate is about twice the upcoast rate, and Q_g equals about $3Q_\ell$ (Inman & Jenkins, 1983).

When sediment trapping by coastal structures is used to estimate the rate of longshore transport, it must be borne in mind that some structures act as efficient sand traps, providing a good measure of the gross transport, whereas others may bypass significant portions of the littoral transport. In addition, the rate at which a structure retains sand usually differs for up- and downcoast transports, and this retention rate will change with time as the structure fills with sand. These factors have complicated the interpretation of littoral transport rates when trapping by structures occurs, leading to wide differences in opinion (e.g., compare Weggel & Clark, 1983, with Inman & Jenkins, 1983).

When waves approach at an angle to the shoreline, they transport sand along the beach. This longshore transport results from the combined effect of the breaking waves, which place sand in motion, and the presence of a longshore current in the surf zone, which aids in the movement of sand along the beach. Theory and field measurements of waves and the resulting longshore transport of sand show that the immersed-weight sand transport rate, I_ℓ, is proportional to the stress-flux factor, P_ℓ (Komar & Inman, 1970; Inman, et al., 1980; White & Inman, 1989),

$$\begin{aligned} I_\ell = K_\ell P_\ell &= K_\ell[P(\sin\alpha)(\cos\alpha)]_b \qquad (5\text{-}9) \\ &= K_\ell[C_b \cdot S_{yx}] \end{aligned}$$

where $K_\ell \simeq 0.8$ is a dimensionless factor ranging between 0.5 and 1.5, $P = ECn$ is the energy-flux of the waves (watts/m; ft-lb_f / sec-ft), E is the wave energy per unit area (joules/m^2; ft-lb_f / ft^2), C is the wave phase velocity (m/sec; ft/sec), Cn is the group velocity, S_{yx} is the longshore component of the radiation stress (momentum-flux tensor) defined in equation (5–6), α is the angle the wave crest makes with the shoreline, and the subscript b indicates that all properties are measured at or calculated for the breakpoint of the waves.

Field data showing the stress-flux factor and the resulting longshore transport of sand measured by fluorescent tracers are shown in Figure 5–6. In this figure, circles are from Inman, et al. (1968) and Komar & Inman (1970), and open and closed squares are from Inman, et al. (1980) and White & Inman (1989). Spatial and temporal sampling techniques are described in Inman, et al. (1980).

It will be noted that the product $[C_b \cdot S_{yx}]$, referred to as the stress-flux factor (Inman, et al., 1986), enters into all of the longshore sand transport relations. This factor has units of watts per meter of beach (ft-lb_f / sec-ft) which is dimensionally identical to the immersed weight transport rate of sand, I_ℓ, in newtons per second (ft-lb_f / sec-ft). By definition, the wave energy per unit of surface area, $E = (\rho g H_{rms}^2)/8$, is calculated from H_{rms}^2, the root-mean-square wave height, which is usually assumed to be related to the significant wave height by $H_s^2 = 2H_{rms}^2$. The stress-flux factor $C_b \cdot S_{yx}$ is a dependent variable requiring interrelated values of the variables wave height, period, and direction. Further, although S_{yx} remains constant during shoaling, C does not, so their product must be calculated at the breakpoint of the waves. A relation similar to (5-9), but using significant waves, is referred to by the Corps of Engineers as the "longshore energy-flux factor", $P_{\ell s}$ (USACE CERC, 1984).

Evaluation of K_ℓ

Recent evaluations of equation (5-9) show that the value of K_ℓ varies as a function of beach slope and breaker type. In general it is found that K_ℓ varies from about 0.5 to 1.5 and for moderately sloping beaches averages about 0.8 (see Figure 5-6). K_ℓ is greater than 0.8 for steeper beaches and less than 0.8 for gentler beaches (White & Inman, 1989). Analyses of the rate of sand entrapment in Santa Barbara Harbor, California, and Rudee Inlet, Virginia, give values of K_ℓ ranging from 0.84 to 1.6 and 0.84 to 1.1, respectively (Dean, 1989). Santa Barbara appears to be nearly a total trap for longshore transport, whereas Rudee Inlet is only a partial trap.

White & Inman (1989) show that K_ℓ varies as a form of the dimensionless "surf similarity parameter", Ir, where

$$Ir = L_\infty^{1/2} \tan\beta / H_i^{1/2} \qquad (5\text{-}10)$$

and L_∞ is the deep-water wavelength, β is the beach slope, and H_i is the breaker height. Least-squared methods show that

$$K_\ell = 1.2\ Ir \quad ; \quad 0.25 < Ir < 1.2$$

Thus K_ℓ and hence the longshore transport rate increase with increasing beach slope, β.

Volume Transport Rate

In the above relations, the immersed-weight longshore transport rate, I_ℓ (newtons/sec), may be expressed in terms of the "at rest" volume transport rate Q_ℓ (m^3/sec),

$$Q_\ell = I_\ell / (\rho_s - \rho) g N_o \qquad (5\text{-}11)$$

where ρ_s and ρ are the densities of the solid grains and the water respectively, g is the acceleration of gravity, and N_o is the volume concentration of sand, equal to about 0.6 for well sorted sand at rest (Inman & Bagnold, 1963). For quartz sand ($\rho_s = 2.65 \times 10^3$ kg/m^3) transported in sea water at 15°C ($\rho = 1.026 \times 10^3$ kg/m^3) with N_o= 0.6, the denominator in equation (5-11) equals 9.55×10^3 newton/m^3. For this case,

$$Q_\ell = 1.05 \times 10^{-4}\ I_\ell$$

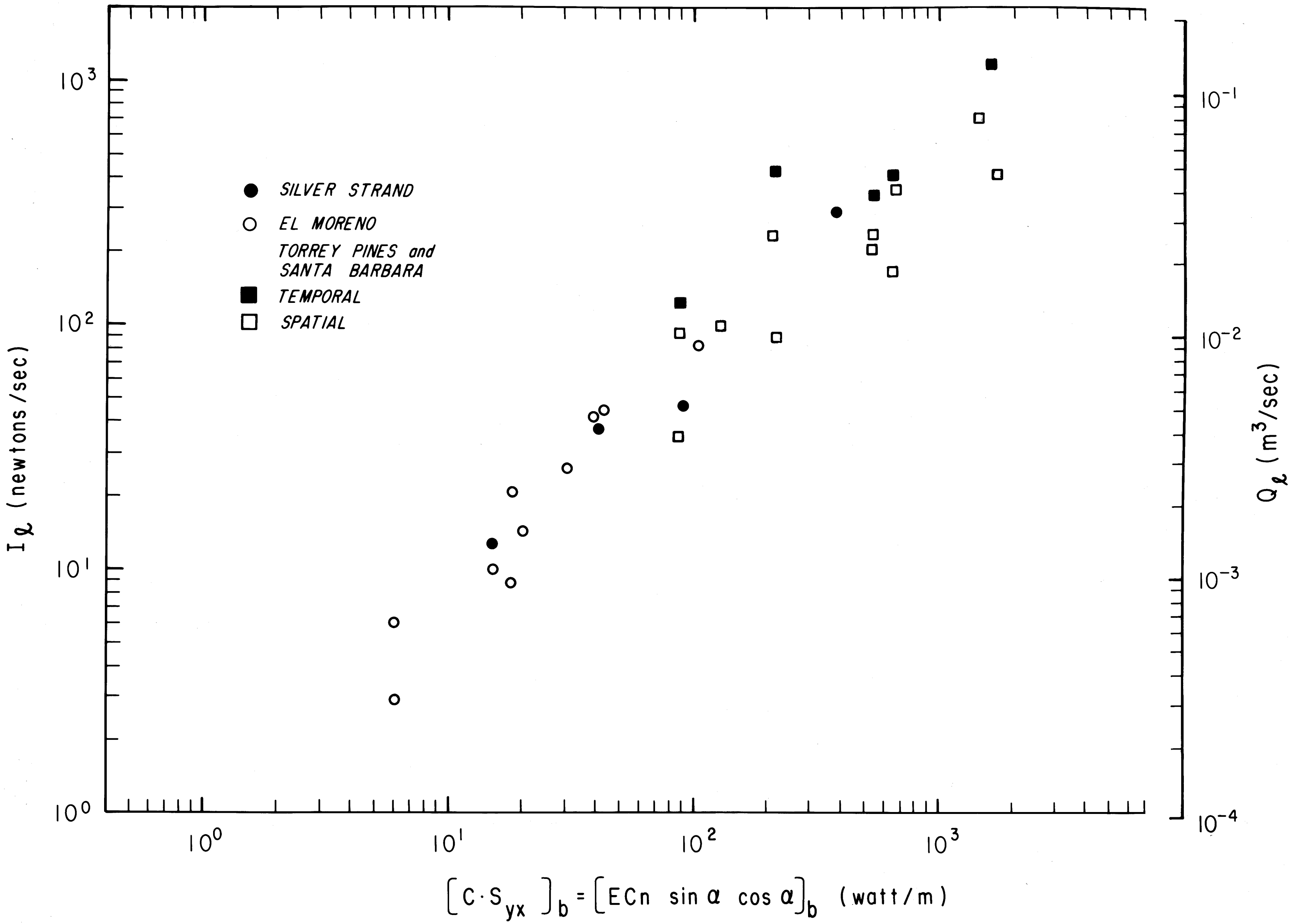

Figure 5-6. Field measurements of total immersed-weight longshore transport rate I_ℓ vs. the stress-flux $[C \cdot S_{yx}]$ as described by equation (5-9). I_ℓ is converted to volume transport rate Q_ℓ using equation (5-11). Sources of data and conversion factors are given in the text.

91.118-6

where Q_ℓ is in m^3/sec and I_ℓ is in newton/sec = watt/m. The constant of proportionality is 1.64×10^{-2} when Q_ℓ is in ft^3/sec and I_ℓ is in ft-lb_f / sec-ft.

Divergence of the Drift

Application of longshore transport relations to a straight beach with parallel bottom contours would result in a transport rate that would be the same everywhere along the beach. If the rate of supply of sand is equal to the longshore transport rate, the beach will be stable and neither erosion nor accretion will occur. On the other hand, if the coastline curves, or there is offshore topography that causes local convergence or divergence of the wave rays, then the longshore transport rate will vary along the beach. The rate of accretion or erosion will be proportional to the rate of change of the volume transport rate, Q_ℓ, with distance, ℓ, along the beach. This relation was termed the divergence of the drift by Wyrtki (1953). Using equations (5–9) and (5–11), the divergence of the drift along the beach for a given wave condition becomes

$$\nabla \cdot \mathbf{Q}_\ell = \partial Q_\ell / \partial \ell \qquad (5\text{–}12)$$

where Q_ℓ is the longshore volume flux (in m^3/yr) which is a vector quantity whose direction is taken as positive downcoast. Equation (5–12) is an application of the continuity equation treated in greater detail in Chapter 9.

It follows that the annual rate of erosion or accretion per unit beach length, ℓ, will be obtained when the net transport rate Q_ℓ is used in (5–12). Further, erosion takes place when $\partial Q_\ell / \partial \ell$ is positive. For this reason, the divergence is sometimes plotted with negative values up and positive down as in Figure 5–7.

Divergence of the drift is an important aspect of inlet dynamics and migration (Inman & Dolan, 1989). The lens effect of the ebb-tide bar off an inlet produces wave convergence near the upcoast side of the inlet and divergence downcoast (Figure 5–7). This causes the longshore transport Q_ℓ to vary across the inlet. The divergence equation (5–12) requires that there be deposition ($\nabla \cdot Q_\ell$ negative) on the upcoast side of the inlet and erosion ($\nabla \cdot Q_\ell$ positive) on the downcoast side.

Accretion/Erosion Waves

The natural supply of sand to many coasts, including those of the San Diego Region, occurs during intermittent flood events. The source is thus episodic with periods between floods ranging from months to decades. On the other hand, longshore and crosshore transport of sand, driven by waves and currents generated by storms, varies with a cyclicity of days to months. Transport

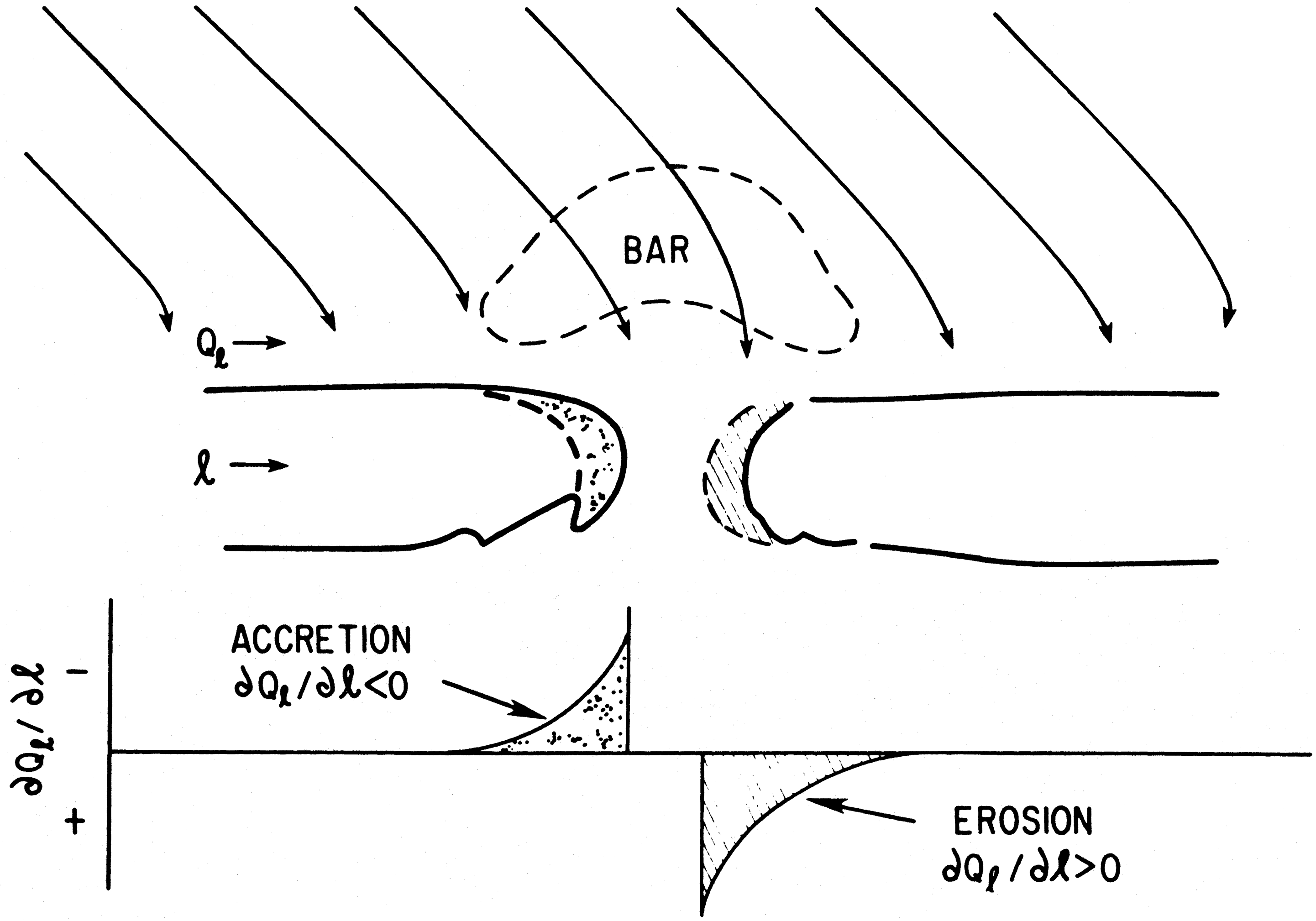

Figure 5-7. Schematic diagram of the divergence of the drift ($\partial Q_\ell/\partial_\ell$) at a migrating inlet with an ebb-tide bar (from Inman & Dolan, 1989).

events then become a nearly steady forcing in comparison to the longer periodicity of flood deposits. Consequently, flood deltas introduce perturbations on a quasi-steady longshore transport system. Perturbations also result when coastal structures are interposed in the littoral zone, interrupting the supply of sand to downcoast beaches. In both cases, the perturbations travel downcoast as accretion and erosion waves.

Accretion waves from flood deposits have been observed off many ephemeral rivers along the California coast (Hicks & Inman, 1987). Erosion waves have occurred wherever coastal structures have interrupted the longshore transport of sand (Inman & Brush, 1973), and have been well documented at Oceanside, California (Inman & Jenkins, 1983; 1985; Inman, 1985) and elsewhere (see Inman, 1987).

A model for the accretion wave produced by large sand deltas, and the erosion wave that precedes it, was developed by Inman & Bagnold (1963). If the sand bulge from the delta is relatively large compared to the width of the surf zone, then the waves reshape the bulge, forming a spit which extends downcoast parallel to but seaward of the previous shoreline. Along wave-dominated coasts, where the bulge of the delta is similar in size to the width of the surf zone, spits do not form, and waves transport the sand downcoast as an accretion wave. However, submerged longshore bars form and behave somewhat like "submerged spits." As the accretion wave travels downcoast, its amplitude is decreased by dispersion, and the effects of wave refraction on the sand bulge diminish. Thus the initial form and rate of propagation of accretion/erosion waves as observed near the initial disturbance can differ from what is observed far downcoast. The terms "near-field" and "far-field" can be used to distinguish between the two littoral regions.

In modeling the propagation of accretion/erosion waves, our concern is primarily with downcoast transport along coasts where upcoast transport is relatively small. (In the Silver Strand Littoral Cell, the downcoast/upcoast directions are reversed, but the net transport relationship holds.) Since the supply of sediment to the coast is episodic and the longshore transport rates are quasi-steady, the effects of a major accretion or erosion event at an upcoast site will arrive at a downcoast site at some later time. This lag in time will represent the time that it takes for the upcoast "disturbance" to reach the downcoast site traveling at a speed related to the mean sand transport rate. Thus the speed is that of the center of mass of the accretion disturbance or mass defect of the erosion disturbance, referred to here as accretion and erosion waves.

As these sand waves move downcoast their amplitudes decrease and they broaden by diffusion. The downcoast velocity V of the center of mass is given approximately by the longshore sand transport rate Q_ℓ divided by the volume change q_c associated with the moving center of mass,

$$V = Q_\ell / q_c \tag{5-13}$$

where Q_ℓ has the dimensions of volume per unit time (e.g., m^3/yr) and q_c is the volume change per unit length of beach (e.g., m^3/m) as defined in equation (5-1). Q_ℓ will vary with time, with local conditions of wave convergences and divergences (particularly near large disturbances such as a sand delta), and with beach slope.

Measurements at Santa Cruz, Santa Barbara (see Inman, 1987, for references), and Oceanside (Inman & Jenkins, 1983; 1985) confirm that the amplitude (amount of erosion/accretion) of the wave attenuates downcoast. As the amplitude decreases by diffusion, the travel velocity of the wave increases. In the absence of detailed knowledge of the distribution of q_c (equation 5-13) in the far-field of accretion/erosion waves, Inman (1987) found that the volume of the seasonal beach change, q_s, when substituted for q_c, gave values of V in approximate agreement with observations. For Torrey Pines Beach in the Oceanside littoral cell, where Q_ℓ = 200,000 m^3/yr (260,000 yd^3/yr) and $q_c \simeq q_s = 92$ m^3/m (110 yd^3/yd), accretion/erosion waves have a longshore velocity of about 2.2 km/yr (1.4 mi/yr). This suggests that their far-field horizontal amplitude is of order $\Delta x = q_c/Z = 5.5$ m (18 ft) which is in approximate agreement with profile changes at Torrey Pines Beach at the downcoast end of wave travel (Inman, 1976, Figure 6.5) (see also Table 5-1, equation 5-1).

5.4 YIELD FROM RIVERS AND STREAMS

A number of procedures have been used to estimate the yield of sediment to the coast from erosion of the land. These include estimates based on the erosion rate of the land in the drainage basin, estimates from sediment rating curves for individual streams, and estimates using flow equations purporting to relate sediment discharge to properties of the flow. The latter can be divided into two rather separate types: the Einstein (1942; 1950) type, based on a dimensional correlation analysis of empirical data (e.g., Brown, 1950; Gill, 1968); and the Bagnold (1966) type, based on fluid-granular mechanics and stream power (e.g., Engelund & Hansen, 1967; Ackers & White, 1973). The details of flow relations will not be discussed further here, but interested readers may refer to the original references or to texts on river hydraulics (e.g., Vanoni, 1975; Raudkivi, 1976; Richards, 1982; Chang, 1988).

Flow equations are useful in cases where stream data is insufficient for determining sediment rating curves, and in cases where high stream discharge rates greatly exceed the data leading to rating curves. Sediment yields based on sediment rating curves are generally accepted as more reliable than those based on flow equations. However, as detailed in Section 5.4.3, there are sometimes serious deficiencies in rating curves. The bedload contribution to total load on rating curves is usually assumed to be some small fraction of the measured suspended load, which is a poor assumption for southern California streams.

Studies of river sedimentation have given rise to a number of terms for the description of the material and its transport mode. Classic treatments of the subject define suspended load as that portion of the total load that is supported by fluid turbulence, while bedload is material that is placed in motion by the tangential shear stress of the fluid over the bottom and has a vertical dispersion of particles that is maintained by grain-to-grain contact and lift forces over the bed. Bedload includes transport by rolling, sliding and saltation. Total load is the sum of bedload and suspended load. Wash load consists of the fine material not present in the stream bed material. It is essentially the fine portion of the suspended load, sometimes referred to as suspended fines.

5.4.1 Erosion Rate Method

A long term estimate of sediment yield in common use by geomorphologists is based on the erosion rate of the land. In this procedure, referred to as the "erosion-rate method," the total sediment yield is calculated as the average erosion rate per unit area of land multiplied by the area of the drainage basin. The erosion rate is usually established by measuring the amount of material trapped in dams and reservoirs. Schumm (1977) shows that the yield of sediment from erosion is a function of climate and the size and topographic relief in the drainage basin. Langbein & Schumm (1958) showed that the sediment yield was a maximum for semi-arid climates, where the annual rainfall was about 30 cm (1 ft). This finding explains why the rivers in southern California produce large volumes of sand.

5.4.2 Sediment Rating Curves

For streams with gaging stations, the discharge of suspended sediment can be related to the water discharge by a sediment rating curve. There are a number of types of rating curves, the most common being the instantaneous and the annual sediment rating curves. Instantaneous sediment discharge is usually predicted from a relation of the form

$$Q_{ss} = aQ^{b} \tag{5-14}$$

obtained from simultaneous measurements of the suspended sediment discharge (Q_{ss}) and the water discharge (Q). The constants a and b are usually obtained from the measured data points in the rating relation using a simple linear regression analysis when the rating relation is expressed in the form, log Q_{ss} = b log Q + log a (Figure 5–8).

An annual rating curve relates the annual sediment yield from a stream to the annual runoff of water. This is obtained from the instantaneous rating curve by summarizing the sediment discharges over the hydrograph (flow vs. time) for each water event such as a storm. Then the sums of the sediment discharges for the year are plotted against the sums of the water discharges. Again, this gives a sediment rating curve of the form

$$V_{ss} = AV^{B} \tag{5–15}$$

where V_{ss} is the predicted annual suspended sediment yield, V is the annual water discharge, and A and B are constants determined from the data (Figure 5–9).

From Table 5–2 it is observed that for southern California rivers, the exponent b in equation (5–14) has an average value of 1.6 and ranges from 1.2 to 1.8 for instantaneous suspended load discharge. From Table 5–3, the exponent B from equation (5–15) for the annual suspended sediment discharge averages 1.5 and ranges from 1.2 to 1.6.

5.4.3 Total Load Transport and Sediment Yield

Sediment rating curves are the most reliable method of estimating the suspended yield of sediment from a river. The problem is to determine the best estimate for total load transport. Although bedload transport can be satisfactorily measured under some stream conditions using tracers, in practice there is no routine procedure for obtaining bedload. This means that it is usually taken to be a certain percentage of the suspended load, an entirely unsatisfactory procedure.

In their study of southern California rivers, Brownlie & Taylor (1981) assumed that the bedload was 10% of the suspended load. This criterion appears to be based on their more extensive experience with northern California rivers. Ten-percent appears to be too low for the drier, sandier southern California rivers. Schumm (1977, p. 110) shows that bedload increases for drier climates and for wider, shallower streams. Richards (1982, p. 106) concludes that for large rivers, bedload is normally less than 10% of the total load, but in mountain streams may reach 70%. Inman (1963, Table 8) using data from Colby & Hembree (1955) shows that

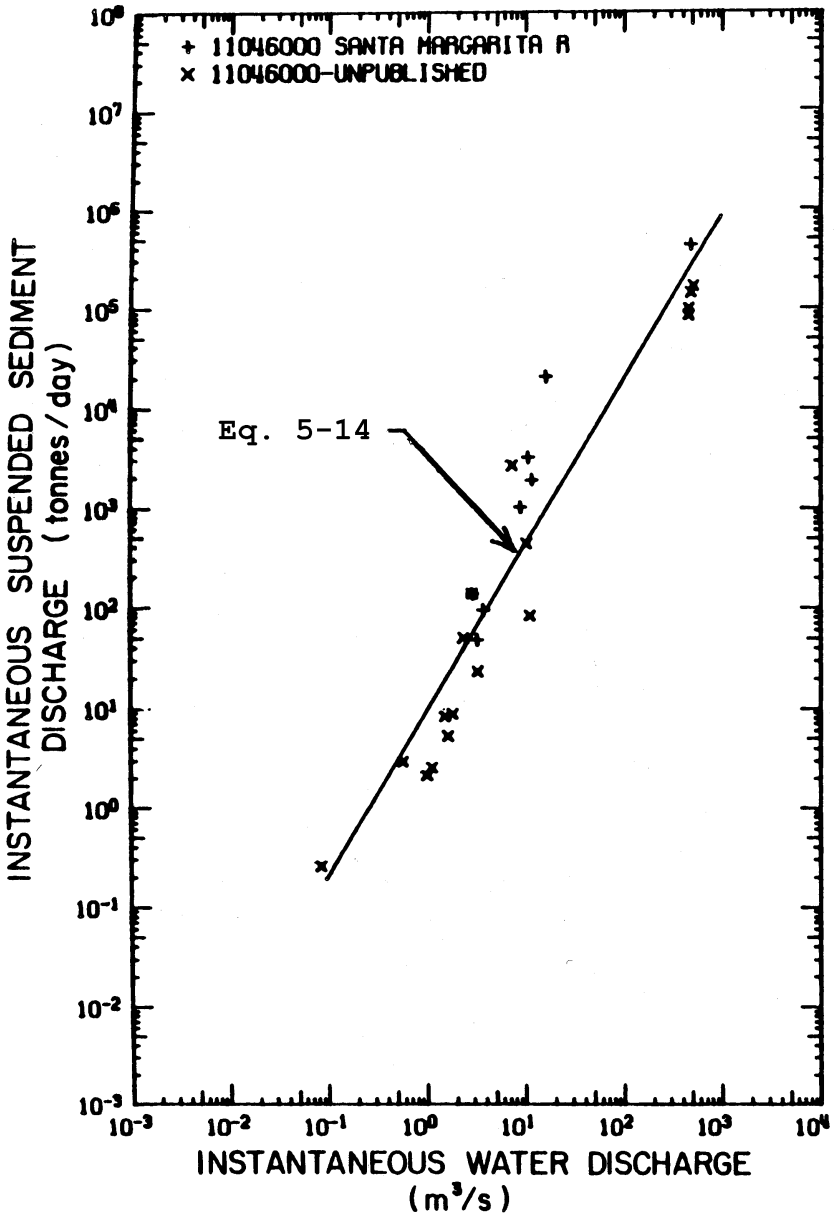

Figure 5-8. Relation of instantaneous sediment discharge to water discharge at Santa Margarita River station 11046000, 1969-76 (from Brownlie & Taylor, 1981).

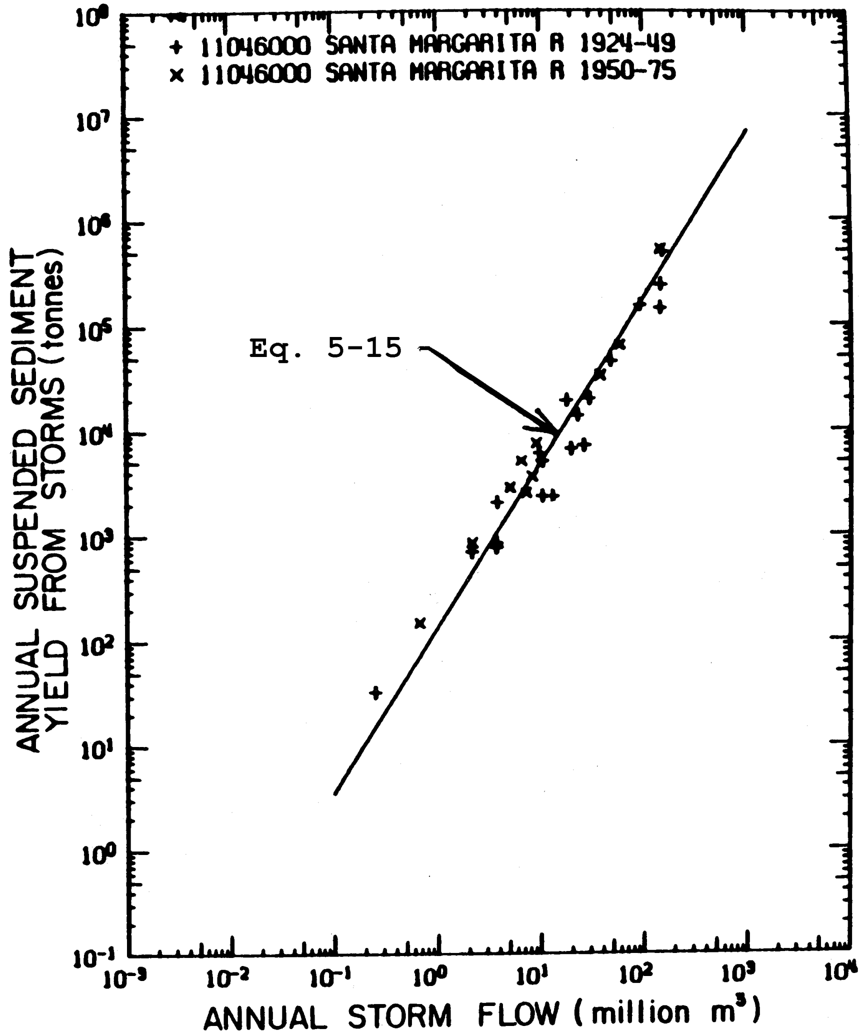

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Figure 5-9. Relation of annual suspended sediment delivered by storms to annual storm flow at Santa Margarita River station 11046000, 1924-75 (from Brownlie & Taylor, 1981).

TABLE 5-2. INSTANTANEOUS SEDIMENT RATING CURVES

USGS Station	River	No. of Samples	Highest Discharge (m^3/s)	Correlation Coeff. of Logarithms	Coeff.* a	Exponent* b
11118500	Ventura	49	555	0.978	14.2	1.83
11114000	Santa Clara	46	4,620	0.942	24.4	1.73
11046000	Santa Margarita	25	527	0.951	8.90	1.66
11042000	San Luis Rey	18	81.8	0.985	26.0	1.78
11022500	San Diego	27	33.4	0.970	8.73	1.58
11013500	Tijuana	43	3.23	0.951	255	1.22
					Ave.	1.63

* Rating curve is of the form $Q_{ss} = aQ^b$ where Q_{ss} is the predicted instantaneous suspended sediment discharge, in

TABLE 5-3. ANNUAL STORM SEDIMENT RATING CURVES

USGS Station	River	Number of Samples: Predicted*	Number of Samples: USGS Est.	Number of Samples: Total	Correlation Coeff. of Logarithms	Coeff.** A	Expon.** B
11118500	Ventura	35	7	42	0.978	588	1.52
11114000	Santa Clara	16	10	26	0.990	938	1.53
11046000	Santa Margarita	29	--	29	0.976	132	1.58
11042000	San Luis Rey	23	--	23	0.971	544	1.54
11022500	San Diego	47	7	54	0.988	139	1.43
11013500	Tijuana	27	6	33	0.996	3120	1.15
						Ave.	1.46

* Predictions based on daily streamflow data and instantaneous rating curves.

** Rating curve has the form V_{ss} (storm) = A[V (storm)]^B, where V_{ss} (storm) is the predicted annual suspended sediment yield delivered by storms, in tonnes, and V (storm) is the annual storm flow, in million m^3 (from Brownlie & Taylor, 1981).

the bedload was 40% of the total load (67% of the suspended load) for the sandy Niobrara River.

In the absence of reliable bedload measurements for southern California rivers, the above considerations led Inman & Jenkins (1983) to conclude that bedload equal to 20% of the suspended load was a more reliable estimate. Accordingly, they used 20% in their estimates for the bedload of the Santa Margarita and San Luis Rey Rivers, rather than the value of 10% adopted for the same rivers by Brownlie & Taylor (1981, Tables C6–5, C7–5).

5.5 TRANSPORT DOWN SUBMARINE CANYONS

Submarine canyons are important geomorphic features of a collision coast like that of California. Because the canyons come into shallow coastal water, they can serve as sediment sinks for the littoral cells. Under natural conditions, the submarine canyons of the southern California coast divert and channel the flow from the littoral cells into the adjacent submarine basin.

The Oceanside Littoral Cell is a well-studied example of this path-to-sink littoral transport. Scripps and La Jolla Submarine Canyons are located at the downcoast end of the cell. The canyons intersect the narrow continental shelf near La Jolla, California, and channel the sand into the deep water of the San Diego Trough. The importance of the Scripps and La Jolla Submarine Canyons is that they remain active in the transport regime for the Oceanside Littoral Cell. In contrast, Redondo and Newport Canyons to the north have been inactivated by artificial stabilization of the coastline within their littoral cells.

Scripps and La Jolla Submarine Canyons and their offshore submarine fan have been investigated by seismic-reflection profiling, drilling, coring, and direct observations by scuba and submersible divers. Scripps Canyon is narrow and deep, with nearly vertical walls cut into resistant Eocene and Cretaceous sedimentary rocks. The canyon extends for approximately 2.7 km (1.7 miles) from near the shoreline to intersect with the wider La Jolla Canyon at a depth of 285 m (935 ft). Although Scripps Canyon has three zones of varying slope and topography (see Inman et al., 1976), the overall slope from sea level to 300 m (984 ft) is about 1:10. At many localities, the canyon walls are smoothly scoured to heights of 4-5 m (13-16 ft) above the canyon floor, resulting in walls that overhang and create narrow, rocky channels that are wider at their base.

Sediments collected from the Mohole drilling project of March 1961 in a water depth of 948 m (3109 ft) and with 315 m (1033 ft) penetration into the submarine fan revealed sand layers with a similar mineral content to the beach sand found at the head of the canyon (see Inman, et al., 1976). At the mouth of the submarine canyon, the fan is transected by a channel having well-developed levees laid down by sediment emerging from the canyon. Much of the fill in the San Diego Trough is sediment discharged through the canyons.

Extensive measurements from moored instruments have yielded data on wind, waves, tide, temperature and currents for the shelf and canyons. The data can be analyzed with reference to deep- and shallow-water regimes that are dominated by specific driving forces. The prevailing motions over the slopes and deeper portions of the submarine canyons are tides, internal waves, and spin-off eddies shed from major ocean current systems. The shallow, steep-walled canyon head areas at depths less than 50 m (164 ft), however, contain canyon currents dominated by motions related to displacements of the sea surface due to gravity waves and wind (Shepard & Marshall, 1973; Shepard, et al., 1974; Inman, et al., 1976). Although deep-water driving forces extend into shallow water, the nearshore zone is dominated by surface waves and currents.

The cycle of sediment motion into the heads of submarine canyons, followed by down-canyon transport into the deeper ocean basins, is completed apparently during intense storms when winds are strong and surface waves are high. During storms, strong water motion down-canyon due to wind and wave action clears the canyon heads and nearby shelf of sediment. The primary surface driving forces are: a) the stress of the wind acting directly on the water surface causing surface currents and wind setup against the shore; b) the incident swell and sea waves that produce wave setup and longshore currents which converge on the canyons; and c) the long-period edge waves trapped by the local topography (see Inman, et al., 1976, for more extensive discussion).

5.6 TRANSPORT BY WIND

Sand transport by wind is most important along lowlying spits bordering tidal lagoons. Along the Silver Strand bordering San Diego Bay and Mission Beach bordering Mission Bay, wind blown sand has widened the spits and formed small dune fields. Wind transport is less important along the predominantly cliffed coastline elsewhere in the San Diego Region.

The relations for the transport of windblown sand were developed by Bagnold (1941) in his classic study, "The Physics of Blown Sand and Desert Dunes." The relations have been verified over horizontal beds in laboratory experiments by Kadib (1964) and Belly (1964). Several studies have evaluated Bagnold's relations in the field. Kadib (1964) evaluated the relation for sand transport by wind on natural beaches. Finkel (1959) measured the rates of migration of dunes in Peru, and Inman, et al. (1966)

evaluated the relation for a coastal dune field.

The wind transport relation is formulated in a manner similar to that for bedload in a stream. The rate of transport (discharge) of granular bed material by a fluid is found to be directly proportional to the power expended by the fluid in transporting it. If the total power available by fluid action per unit area of bed is ω, then a portion, $K\omega$, of this power is available for transporting sand, and the transport rate becomes

$$i = K\omega \tag{5-16}$$

where i is the immersed weight of sediment transported across unit width of bed in unit time. The immersed weight transport rate is converted to dry mass transport per unit time and unit width by the equation

$$j = i\rho_s/(\rho_s - \rho)g \tag{5-17}$$

and to bulk-volume transport per unit time and width q_s by the equation

$$q_s = i/(\rho_s - \rho)gN_o = j/\rho_s N_o \tag{5-18}$$

where N_o is the "at rest" volume concentration of sand, ρ_s and ρ are the densities of the solid grains and air, respectively, and g is the acceleration of gravity.

Since the density of air, ρ, is very small compared to that of the grains, ρ_s, the relation for the dry-mass transport per unit time and width becomes

$$j = i/g = K\omega/g \tag{5-19}$$

where g is gravity. The power is given by $\omega = \tau u_*$ where τ is the stress and u_* is the friction velocity obtained from the von Karman-Prandtl relation for flow over an aerodynamically rough surface. Experiments show that the coefficient of proportionality K in equation (5-19) varies with sand size and sorting. Bagnold (1941, p. 67) showed that

$$K = C(D/250)^{1/2}$$

where D is the grain diameter in microns and C is an empirical coefficient having the following values: 1.5 for a nearly uniform size sand; 1.8 for naturally sorted sands such as dunes; and 2.5 for a sand with a very wide range of grain size.

A more complete derivation of the transport relation and the method for converting wind anemometer measurements to transport rates is given in Bagnold (1941) and Inman, et al. (1966).

5.7 TIDAL LAGOONS

Most tidal lagoons have evolved during the last few thousand years when the rise in sea level has been relatively slow. They appear to have formed both by bar-building and by spit extension, although most appear to have formed by spit extension in the classic manner described by Redfield (1965). In this case, longshore transport of sand extends the spit across an otherwise open embayment, gradually closing the lagoon. As the lagoon closes, the scouring action of the flood and ebb tidal currents increases until it is concentrated in a channel that is in equilibrium with the energy dissipated from the tidal flow.

The size of the entrance channels into tidal lagoons that border sandy shorelines is related to the volume of the tidal prism of the lagoon. The situation seems analogous to that for rivers, where the sectional area is related to the stream discharge. In the tidal inlet, there appears to be a balance between the scouring actions of the tidal currents, which tend to keep the channels open, and the longshore transport of beach sand, which tends to close them. An entrance channel that maintains a constant cross-sectional area while traversing a sandy beach has attained an equilibrium with the tidal flow, and it is sometimes referred to as a scouring channel. Equilibrium does not imply a stability of channel location, but only a constancy in the relation between tidal flow and channel cross-section. It has been observed that entrance channels migrate in the direction of the dominant longshore transport unless the channels are stabilized naturally by headlands or artificially by jetties.

O'Brien (1931), using 16 harbors known to have channels that approximate scouring conditions, showed that the cross-sectional area of the entrance (below mean sealevel) was proportional to the volume of the tidal prism raised to the 0.85 power. The tidal prism was taken as the area of the tidal lagoon or bay multiplied by the diurnal range of tide (mean higher high to mean lower low water) for Pacific coast harbors and by the mean range (mean high to mean low water) for Atlantic Coast harbors. O'Brien's data ranged from San Francisco Bay with a tidal prism of 2.29×10^9 m^3 (8.1×10^{10} ft^3) to Newport Bay with a prism of 5.66×10^6 m^3 (2.0×10^8 ft^3).

Later, the equilibrium relation for the smallest harbor with two jetties, the boat basin at Camp Pendleton near Oceanside, California, was measured by Inman and Frautschy (1965). The Camp Pendleton boat basin was surveyed in 1956 following a long period without maintenance dredging during which the channel entrance was permitted to attain a condition of equilibrium with the tidal flow. The Camp Pendleton boat basin has since been modified and combined with Oceanside Harbor. Similar measurements were obtained for Mission Bay entrance four years after the completion of construction of new jetties bordering the entrance channel and prior to the commencement of maintenance dredging (Inman & Frautschy, 1965). Selected values of channel cross-sectional areas and tidal prisms are listed in Jarrett (1976).

All of the newer data are in surprisingly good agreement with O'Brien's original data, giving a relation between the cross-sectional area of the entrance channel and the tidal prism as

$$A = nP^m \tag{5–20}$$

where A is the minimum cross-sectional area of the inlet channel measured below mean sealevel, and P is the volume of the tidal prism between diurnal tides. Values of the coefficients n and m for United States inlets are listed in Table 5–4. Data for selected inlets are plotted in Figure 5–10, and those most appropriate for the small lagoons of the San Diego Region are plotted in Figure 5–11. All of the newer data for small inlets along the Pacific coast tend to follow O'Brien's original curve for scouring channels with coefficients $n = 2.05 \times 10^{-4}\ m^{-0.55}$ ($1.07 \times 10^{-4}\ ft^{-0.55}$) and $m = 0.85$ (Figures 5–10 and 5–11).

TABLE 5–4
TIDAL COEFFICIENTS FOR THE RELATION $A = nP^m$

Coast[a]	n: Metric[b]	n: U. S. Conventional[b]	m
Pacific	2.83×10^{-4}	1.19×10^{-4}	0.91
Gulf	9.31×10^{-4}	5.02×10^{-4}	0.84
Atlantic	3.04×10^{-5}	7.75×10^{-6}	1.05

[a] Data is average for all inlets of the contiguous United States.
[b] A, P respectively have units of m^2, m^3 for metric and ft^2, ft^3 for U. S. Conventional.

Source: USACE CERC (1984, p. 4-157), after Jarrett (1976).

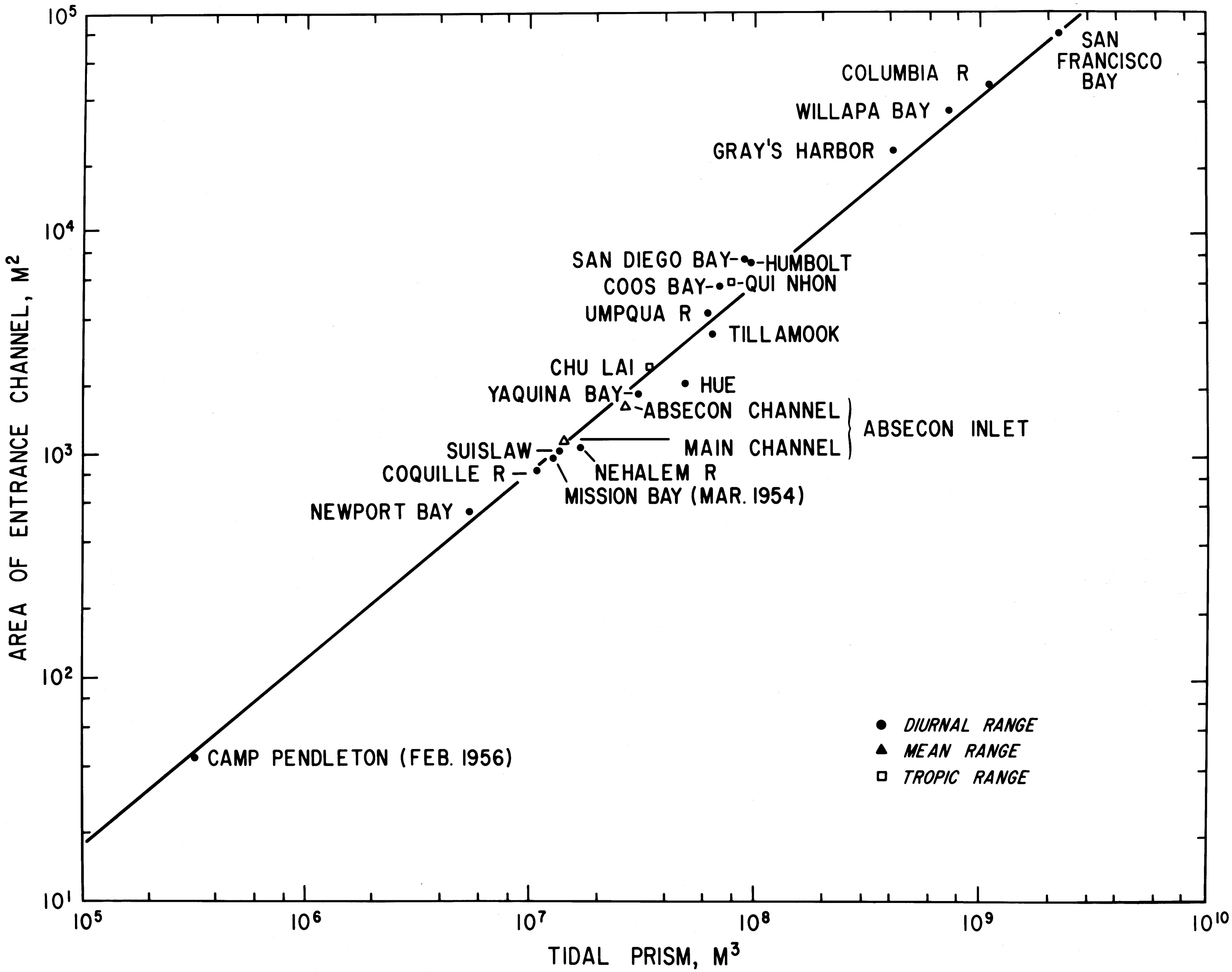

Figure 5-10. Relation between the minimum area of the inlet channel A and the tidal prism P of equation (5-20) (after Inman & Frautschy, 1965; Inman & Harris, 1966).

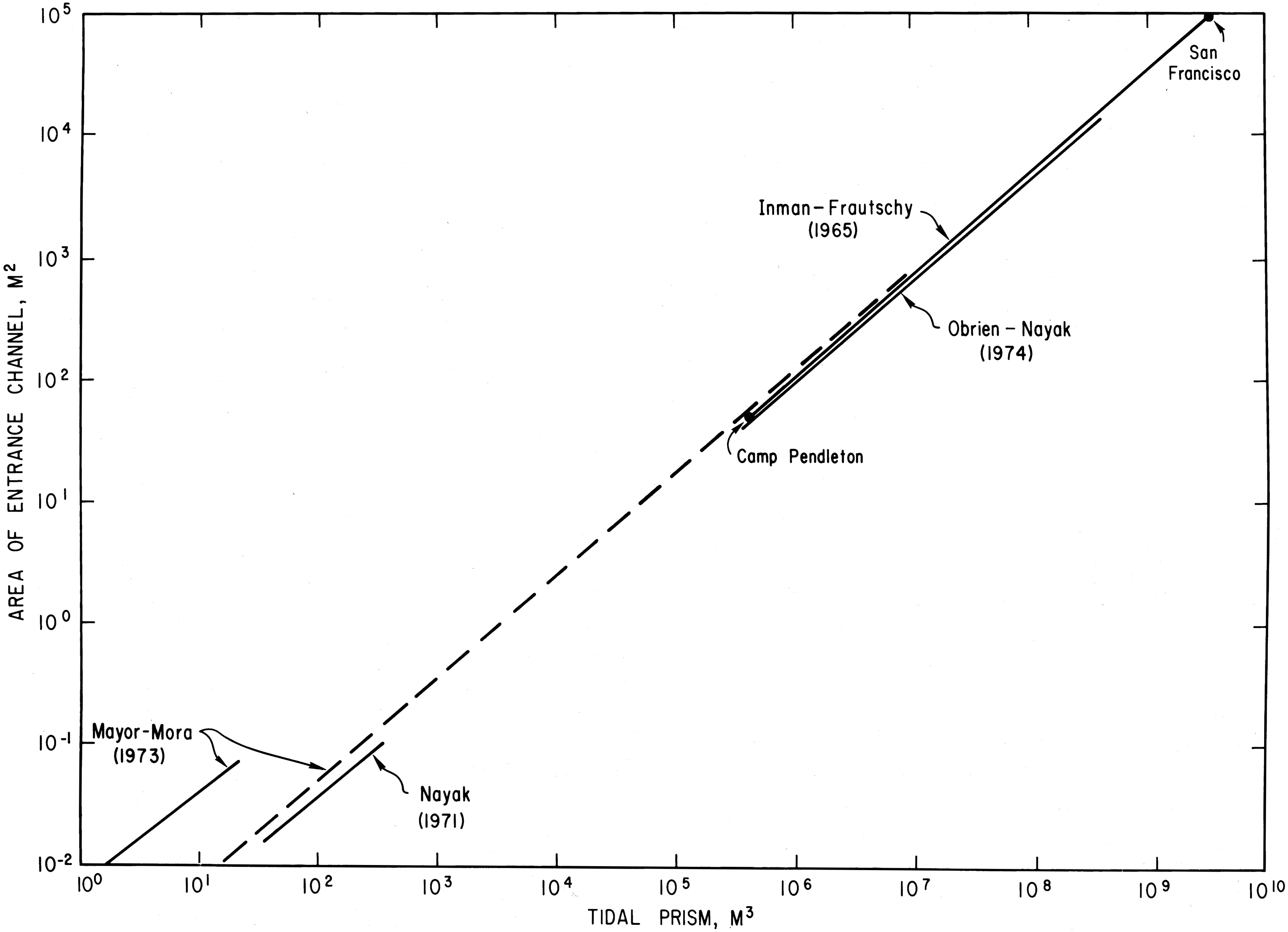

Figure 5-11. Relation between A and P of Figure 5-10 for small tidal lagoons and models. The O'Brien-Nayak curve is from Mehta & Hou (1974).

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