THE UNIVERSITY OF WISCONSIN SEA GRANT PROGRAM

EXPLORATION FOR HIGH ENERGY MARINE PLACER SITES

PART I—FIELD AND FLUME TESTS

NORTH CAROLINA PROJECT

Craig H. Everts

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SEA GRANT TECHNICAL REPORT # 10
WIS-SG-72-210

FOREWORD

Basic to developing exploration guides for use in searching for marine placers is the need to understand the processes related to placer genesis. Thus, early in the University of Wisconsin's Sea Grant Minerals Program, we initiated investigations on the lower specific gravity heavy minerals and subsequently, tests on the intermediate high density mineral types of economic interest, but at other field localities.

I chose the Oregon Inlet complex on the shore of North Carolina as our low density test site because the area is in a known heavy mineral province (black sands are mined farther south); the waves, currents and deposits could be studied and field measurements made using small craft; and, of importance to industry, this area is much like other high energy placer sites that several commercial groups are soon to explore elsewhere.

Dr. Everts extended the study to include a rigorous flume investigation of sample material he collected at the field stations, which adds considerable evidence to the interpretations and conlusions. He assumed a major lead in this approach. Together, we have reviewed the data, and as supervisor of his research, I share with Dr. Everts the responsibility for the somewhat tenuous nature of several of the conclusions. Nevertheless, we believe that the summary of our findings (pages 165-170) provides several useful clues for locating economic placers in similar coastal marine environments. Furthermore, we view much of the technical detail reported herein as necessary to <u>any</u> applied placer research program, and in this regard, the reported results will help avoid industrial redundancy.

Part II of this study -- a joint report by Drs. Moore and Everts on the gold, titanium and zirconium distribution in the study area, including sedimentary and environmental relationships -- is currently in preparation.

Dr. Everts joins me in soliciting comments and suggestions.

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TABLE OF CONTENTS

LIST OF TABLES

LIST OF FIGURES

LIST OF SYMBOLS

A = Maximum horizontal fluid displacement - oscillating flow

a = Long axis of particle

a_1 = Length used in defining gravity moment arm

a_2 = Length used in defining drag moment arm

B = Variable term

b = Channel width

C_d = Drag coefficient

c = Short axis of particle

c_1 = Coefficient referred to volume of particle

c_2 = Coefficient referred to surface area of particle

d_e = Equilibrium particle size - in feet

d_1 = Grain diameter at incipient motion - in feet

d_s = Intermediate grain diameter - sieve size

f = Darcy-Weisbach friction factor

g = Acceleration of gravity

H = Wave height

H_o = Deep water wave height - in feet

h = Depth measured from bottom to surface

k = Wave number $(2\pi/L)$

L = Wave length

L_o = Deep water Wave length

p = Pressure

p_r = Reference pressure

R = Hydraulic radius

SF = Shape factor

s_f = Specific gravity - fluid

s_s = Specific gravity - sediment

T = Wave period

$\bar{u}$ = Mean flow velocity

u_o = Velocity at a distance of $0.35d_s$ above the bed

x = Distance from origin to point of interest in a sublayer

α = Beach gradient

δ = Boundary layer thickness

η = Shape of channel plan

θ = Angle of particle repose

θ' = Imbrication angle

ξ = Measure of channel cross-section

ρ_f = Fluid density

ρ_s = Sediment density

τ = Total shear

τ_b = Bed shear

$\bar{\tau}$ = Mean flow shear

τ_{bm} = Bed shear (equation 2-5)

τ_c = Critical bed shear

τ_d = Depositional bed shear

Φ = Slope of channel profile

ϕ = Slope of the sediment boundary

ABSTRACT

Placers, of increasing importance as exploitable marine mineral resources, have been qualitatively classified in the past with reference mainly to their geomorphic location. This study provides a generalized process-response model for locating and spatially delimiting heavy mineral accumulations in the marine nearshore, estuarine and beach zones. The model is based on the dynamic mechanisms of placer formation.

A heavy mineral study area in Dare County, North Carolina was used to observe the location and conditions under which placer minerals accumulate in various marine environments. Oregon Inlet, a portion of Pamlico Sound, the Atlantic Ocean, and coastal beaches were physically monitored with respect to wave and current regimes, bathymetry, and sediment texture in order to classify existing marine placers. Surface sediment samples were collected using a pipe dredge and SCUBA methods. Laboratory analysis of 550 samples yielded median diameter, sorting, skewness, and density values.

Flume studies concerned with the critical conditions of incipient motion, transport, bypassing, and deposition of heavy and light minerals in the 2.65 to 5.18 gm/cc density range were conducted to quantify the dynamics of placer formation. Known fluid flow conditions of formation were then related to the dynamically less understood natural area.

Two types of loose boundary marine placers were observed in the field and artificially formed in the flume: (1) laminar placers formed on a planer bed and preserved above mean sea level and (2) disseminated placers formed in a ripple environment below mean water level. The laminar deposit often approaches a one hundred percent concentration of heavy minerals and contains a high ration of dense economically significant minerals. Light mineral bypassing during heavy mineral deposition accounts for the richness of the laminar placer. A disseminated-type placer may contain two to six times the heavy mineral concentration existing in the average sediment in a region. Placer minerals are deposited with their equivalent light mineral sizes in aggrading ripple troughs.

Heavy minerals are not concentrated in channels of the Oregon Inlet system, but occur as disseminated-type placers at the terminus of main tidal channels offshore and in Pamlico Sound, and on shallow shoals adjacent to steep margins of major channels in the sound. A favorable location for a laminar-type spit deposit is immediately downdrift of a coastal dune complex. Although large quantities of light minerals pass the inlet on its offshore bars, heavy minerals appear to move in a path that takes them through the inlet throat and into the sound. Ebb currents may then flush them seaward on the downdrift shore of the inlet system.

Beaches are optimum regions for concentrating heavy minerals because they provide a mechanism for creating and preserving laminar deposits. Placer minerals move preferentially on the foreshore and to a lesser extent in the surf zone. Few heavy minerals are transported parallel to the beach in the breaker zone. Stormline placers are located in the most landward storm berm, generally at the base of coastal dunes, the region providing the most opportunity for long-term preservation. Stormline laminae for at the end or near the end of a beach erosive cycle. As water level or wave intensity lessens, heavy minerals are the first to be deposited. As long as bed shear conditions remain conducive to placer grain deposition, and light mineral bypassing, the laminae will thicken. Regressive beaches are optimum sites for heavy mineral preservation.

An important offshore placer possibility is a relic lower beach, lagoon, sound, or coastal plain stream deposit that was preserved during a transgressive cycle. Placers forming under the influence of waves alone will be rare because the wide range of wave state and large bottom area precludes the steady or quasi-steady flow conditions in a limited area necessary for heavy mineral accumulation.

This contribution of information on placer formation should aid industrial exploration by placing in perspective important facts and probabilities in relation to marine sedimentation.

Research was sponsored through a Sea Grant Program grant from the National Oceanic and Atmospheric Administration to Professor J.R. Moore of the University of Wisconsin, Madison.

CHAPTER 1

INTRODUCTION

Increasing requirement for raw materials to support industry is reflected by an expanded interest in low grade mineral deposits on land and in the quest for exploitable mineral deposits on the sea floor. Considerable effort is being spent in the search for heavy mineral accumulations on the world's continental shelves (McKelvey and Wang, 1969). Since hydrodynamically formed accumulations of heavy metals, especially gold, have long been sought in rivers, it is natural that man enlarge his interest to include placers formed in the marine environment. In addition to gold, other possible placer minerals and elements include platinum, chromite (chromium), columbite (columbium), magnetite (iron), monazite (radioactive elements), cassiterite (tin), ilmenite (titanium), rutile (titanium), scheelite (tungsten), wolframite (tungsten), and zircon (zirconium), All have densities significantly higher than their common sedimentary hosts, namely quartz, feldspar, clays, and lithic fragments.

The offshore reserves of such minerals may be large; however, estimates of subsea mineral resources have not yet been established (McKelvey and Wang, 1969). Today, the bulk of the free world's tin, for example, is mined from offshore placers in Indonesia, Malaysia, and Thailand. In Alaska, offshore mining of proven gold resources at Nome is under consideration, and a platinum exploration program is in progress in the Goodnews Bay area (McKelvey and Wang, 1969). Titanium bearing minerals are mined from beach and offshore sands in Australia. The latter source currently dominates the world titanium market (McKelvey and Wang, 1969). Diamond placers have been worked off the mouth of the Orange River in South Africa; however, the project is no longer economic.

Occurrences of various heavy minerals are common to all clastic deposits of the world. (Cf, Preliminary map, Geologic and physiographic provinces, subsea underground mines, and coastal placer deposits, in McKelvey and Wang, 1969). In general, the presence of deposits in deep water is unknown. The potential of any marine placer mineral is only as bright as the explorationist's ability to locate it in economic concentration.

Current interest in the location and exploitation of marine placers largely emphasizes the engineering technology of prospecting and of mining rather than scientific facts, statistical probabilities, and relevant concepts concerning the origin of placer deposits (Emery and Noakes, 1968). Because favorable areas for the sea floor occurrence of placer minerals are, at present, difficult to outline, prospecting today is largely confined to exploring areas offshore of known or producing coastal deposits. The search for marine placer deposits must first be closely guided by sound geological and oceanographic understanding of events leading to and resulting in a placer accumulation. Offshore geophysical investigations accompanied by a non-random sampling program will logically follow this initial study. Random sampling, on the other hand, can be a waste of time and money.

OBJECTIVES

Although a marine prospecting program for heavy mineral placers should recognize the odds are against the occurrence and more so against the discovery of commercial deposits, there are several approaches for improving these odds.

One method is to extrapolate from a known deposit to an unexplored area on the basis of similar dynamic and sedimentary characteristics. While this may be satisfactory in certain instances, it does not allow for expansion to areas of significantly different oceanic, depositional, and provenance conditions. Correlation of environmental variables without an understanding of their interrelationships does not favor carrying observations related to placers between dissimilar localities. A rational approach to constructing a general framework to aid in the search for placers must, therefore, account for the mechanics of their formation.

It is the objective of this study to provide a generalized process-response model for placer accumulations in the marine nearshore, estuarine, and beach zones.

Because of the great complexity of processes active in any individual locality, it will be impossible to precisely define the dynamics resulting in a specific placer, but a general knowledge of the various dynamic mechanisms responsible for placer formation will aid in outlining areas worthy of further detailed(and expensive) sampling. The model will also assist in defining the spatial extent of a placer.

Such a contribution of information on placer formation will aid industrial exploration by placing in perspective important facts and probabilities in relation to marine sedimentation. This scientific approach combining, as it does, the geologic understanding of present marine phenomena with the quantitative knowledge gained of sedimentary hydraulics under controlled conditions should enhance this and subsequent research by providing critical answers and questions pretaining to the relationship of placer formation to the overall field of nearshore marine sedimentation.

PREVIOUS RESEARCH - PLACERS

Extensive research has been done on heavy minerals as indicators of sediment source, as exemplified by van Andel(1959). However, very little has been accomplished concerning the hydraulic flow regime responsible for heavy mineral accumulations. Rittenhouse (1943) in a classic paper was the first to attempt a description of the textural associations of heavy and light minerals and their hydraulic significance. Since 1943 (to the author's knowledge) the physical aspects of heavy mineral concentrating mechanisms have not been investigated.

A general description of marine placers and the geological problems inherent in locating them was treated by Emery and Noakes (1968). They qualitatively covered transgressive and regressive beach cycles and concluded offshore marine placers were unlikely to prove as economically significant as present-day beach deposits and regressive placers now located above sea level.

Specific offshore heavy mineral accumulations have been described by Pomerancblum (1966) who studies the Mediterranean continental shelf of Israel; and Moore and Silver(1968) and Clifton (1968) who described surficial gold distribution on the Oregon continental shelf. Klum and others (1968) conducted a magnetic investigation of heavy mineral occurrences off southern Oregon.

Hill and Parker (1970) discuss tin and zirconium in marine sediments around the British Isles. They observed tin values to be low even off the Cornwall mineral district. Veenstra (1965) cities anomolous heavy mineral values in the Dogger Bank area of the North Sea.

Tanner *et.al.* (1961) studied a possible masked placer in a shoal complex off the mouth of the Apalachicola River, Florida. Bates (1963) extended this investigation in a heavy mineral reconnaissance of western Florida beaches and the adjacent continental shelf. Poole (1958) discussed heavy mineral variations in several bays of the central Texas coast.

Marine placers have been observed and described in the most detail on beaches. Griggs (1945) and Pardee (1934) discuss coastal placers of Oregon. Bradley (1957) describes the variation in heavy minerals in the beach sediments of Mustang Island, Texas. while Neiheisel (1962) studied beach, barrier island, and Coastal Plain relic Pleistocene placers in Georgia. Rao (1957) examined and explained heavy mineral deposits in conjunction with beach erosion in India.

Pirkle and Yoho (1970) report on the Trail Ridge, Florida, heavy mineral ore body. Their studies indicate the deposit was formed during a sea level stillstand in a marine beach environment. Huston and Murphy (1962) studied an indurated titaniferous black sandstone deposit in Wyoming and attributed its origin to beach processes. Willden and Hotz (1955), Anderson (1960), Reid (1960), Savage (1960), Stanley (1961), and Stoll (1961) describe continental placers formed in response to the hydraulic action of rivers and streams.

OCEAN PLACER MINERALS STUDY

This investigation is one in a continuing marine placer minerals program of the Marine Research Laboratory of the University of Wisconsin. In 1967 as a prelude to the present Sea Grant supported research, a detailed study of classical placer environments was conducted off the coast of Ireland. Findings of this earlier research provided insight to the high energy system, and showed that investigations in the current and sediment processes, which actually form the placer accumulation, are needed to provide reliable search clues in complex systems, particularly in areas of swift tidal and coastal currents. This need formed the reason to pursue the present investigation, which had been previously proposed by Professor Moore for Sea Grant support at the University of Wisconsin.

In July, 1970, a graduate student began dissertation research on the non-propriety sedimentary aspects of an Inlet Oil Corporation platinum prospect on the coast of the Bering Sea. With access to samples, cores, and data, he is working on a "real" exploration problem.

To further test the results of applied placer mechanics research, personnel of the Marine Research Laboratory will, in 1971-72, join a British team for a cooperative study of the famous Cornish tin sands off southern England and other placer sites in British waters.

ACKNOWLEDGMENTS

This research was sponsored by the National Science Foundation under a University of Wisconsin Sea Grant Program grant to Professor J. Robert Moore. Appreciation is extended to the United States Coast Guard who provided a vessel and crew for offshore sampling work, and the Department of Civil Engineering of the University of Wisconsin for the use of a flume in their hydraulics laboratory. I express my thanks to Dr. J.R. Moore of the Marine Research Laboratory at the University of Wisconsin for making this investigation possible. I am also indebted to Dr. Peter Monkmeyer and Dr. James Villemonte of the Department of Civil Engineering for valuable advice concerning the hydraulic aspects of the study. Appreciation is extended to Mr. Robert Blaedel for his assistance in the field, and to Miss Martha Wright for her laboratory aid. Finally I thank my wife Paulette for her understanding, consideration, and assistance during the entire project.

APPROACH

In resolving the problem of predicting the location of placer deposits and the related questions pertaining to their origin, it was necessary, due to a dearth of information on such deposits, to conduct several preliminary studies. The first was a natural model study intended to observe the conditions under which heavy minerals accumulate in various marine environments. Secondly, a controlled laboratory investigation of the limited hydrodynamic conditions of placer formation was necessary to quantify the dynamics of their origin. Known fluid flow conditions of formation may then be related to a less dynamically understood natural area.

NATURAL MODEL STUDY

An ideal natural study area is present along the Atlantic coast in Dare County, North Carolina, and was selected for the study. Roanoke Island in the center of the study area is approximately 100 miles south of Norfolk, Virginia, and the mouth of Chesapeake Bay, and 40 miles north of Cape Hatteras, North Carolina (Figure 1-1). This coastal zone which is part of an elongate barrier island system known as the Outer Banks is an extension of the Atlantic Coastal Plain Province.

The reasons for selecting this area as a general field model are as follows: The area is easily accessible and the sediments contain a sufficiently large percentage of heavy minerals to enable recognition of local selective concentrations. This modern sedimentary province has no offshore topographic obstructions that could hinder or complicate shoaling gravity waves approaching the coast. The beach is unobstructed and straight for 80 miles north to the Chesapeake Bay. The dominant southerly littoral drift results in an equilibrium beach with local conditions which are reasonably stable.

The study area is well monitored physically with a Coastal Engineering Research Center wave and tide gage at Jenette's Pier (offshore profile B, Figure 1-1). This is one of five such recording stations along the Atlantic coast. The Wilmington District of the U.S. Army Corps of Engineers has also monitored topographic changes in the Oregon Inlet area and has maintained a tide gage there for the past several decades (Vallianos, 1967). Shelf water circulation has been well documented which provides insight into offshore sediment patterns (Harrison *et. al.*, 1967).

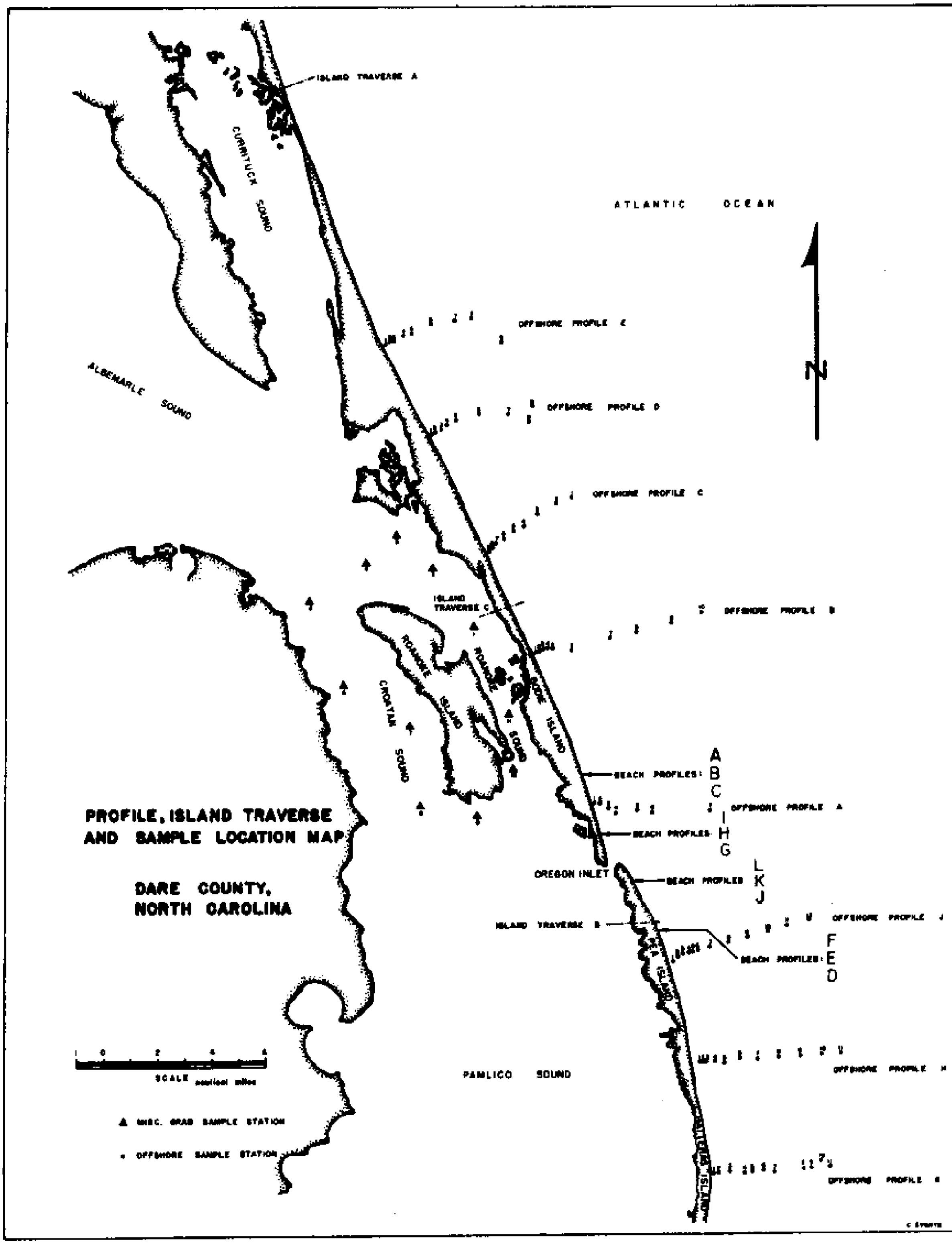

ISLAND TRAVERSE A
CURRITUCK SOUND
ATLANTIC OCEAN
N
OFFSHORE PROFILE E
ALBEMARLE SOUND
OFFSHORE PROFILE D
OFFSHORE PROFILE C
ISLAND TRAVERSE C
OFFSHORE PROFILE B
ROANOKE ISLAND
ROANOKE SOUND
CROATAN SOUND
BODIE ISLAND
BEACH PROFILES: A B C
OFFSHORE PROFILE A
BEACH PROFILES I H G
PROFILE, ISLAND TRAVERSE AND SAMPLE LOCATION MAP
DARE COUNTY, NORTH CAROLINA
OREGON INLET
BEACH PROFILES L K J
ISLAND TRAVERSE B
PEA ISLAND
BEACH PROFILES: F E D
SCALE nautical miles
PAMLICO SOUND
MISC. GRAB SAMPLE STATION
OFFSHORE SAMPLE STATION
HATTERAS ISLAND

Figure 1-1

Previous Research - North Carolina Coastal Region

The coastal sediments, topography, stratigraphy, and flow regimes of the North Carolina Coastal Region have been well documented. Gorsline (1963) and Field and Pilkey (1969) studied the bottom sediments of the Atlantic shelf. Size distribution, mineralogy, and relic characteristics were reported. Whitmore and others (1967) studied various indicators, including elephant teeth, of Pleistocene coastal plain sedimentation in deposits now found on the continental shelf.

Uchupi and Tagg (1966) and Uchupi (1970) describe surface microrelief and the shallow structure of the shelf. Their work is based on precision depth profiling and seismic traverses off the east coast of the United States including North Carolina. Drake *et. al.* (1959) examined the continental margin north of Cape Hatteras.

Harrison and others (1967) report shelf water circulation off the Chesapeake Bight to move in a generally southwesterly direction. Sea bed drifters released on the shelf were often recovered in estuaries and inlet regions.

It was considered instructive to review some of the previous research done in the general area. Harrison *et. al.* (1965) report evidence for a possible late Pleistocene uplift of the Chesapeake Bay entrance. Newman and Rusnak (1965), on the other hand, present evidence indicating a Holocene submergence of the coast immediately north of the Chesapeake Bay mouth. Pierce (1969) studied the sediment budget of the barrier chain south of Cape Hatteras. He concluded that considerable sediment was being supplied to the beaches from an offshore source. Giles and Pilkey (1965) investigated beach and dune sediments along the east coast. Richards (1967) reviews Atlantic Coastal Plain stratigraphy between Long Island and Georgia. White (1966) relates drainage asymmetry across the Coastal Plain to the form of the Carolina Capes.

Field Procedure

The sedimentary aspect of field sampling consisted of collecting samples of 100 to 400 grams from the offshore, beach and inlet environmental zones plus dune samples, miscellaneous grab samples from the sounds, and samples at selected intervals on three island traverses (Figure 1-1). Offshore samples, collected by SCUBA (self contained underwater breathing apparatus) or bottom dredge, have been taken on eight traverses on a spacing of five miles. Traverses were normal to the coast with sample stations at every ten foot depth change or three-fourths of a mile, whichever came first. Sample station locations were extended to a distance of six miles from shore, and ripple topography and orientation were observed at all SCUBA stations. Sediment sampling was initiated according to the Nyquist interval, the maximum spacing resulting in total information recovery equal to two or more points per cycle of the highest assumed frequency (Nyquist, 1928).

Approximately 300 bottom samples were collected from a small boat in the vicinity of Oregon Inlet (Figure 1-1). Samples of the upper two centimeters of bottom were taken by swimmer, and any bottom topography and its orientation were noted. The close sampling program in the inlet is in response to the shorter wave length flow fields in this critical environment, whereas offshore, the dominant dynamic sedimentation mechanism should be shoaling gravity waves with a nearly normal coastal approach vector dependent upon refraction. Offshore this appears to result in fairly long linear sediment patterns parallel to the coast. Currents in the inlet, however, are not as easily defined, and the close sample grid is to ensure that small concentrations of placer minerals and high frequency variations in sediment patterns are recognized. (One purpose of the survey was to establish spatial limits of placer deposits)

Twelve beach profiles were surveyed and sampled from the back beach dunes to 300 feet seaward of the berm. Approximately 15 samples were taken on each profile with emphasis on sampling each textural change on the beach slope. A transit-and-stadia survey was made onshore. Offshore samples were collected and depths recorded by a swimmer connected to a shore base with a line of known length. Pertinent wave and breaker characteristics as well as longshore drift velocities at the time of sampling were recorded. The sampling was conducted above and below the inlet (Figure 1-1) in an attempt to discover what influence the inlet has on beach sedimentation. This aspect of the field study provides the location of heavy mineral accumulations and defines zones of sediment movement in this high energy environment. Dune samples, grab samples from the sounds, and island traverse samples were taken to tie sedimentation in the sounds to that on the ocean side.

Following collection each sample was placed wet in a 12 by 18 inch previously marked polyethylene bag and the bag was sealed. Field sample numbers were assigned to each sample and recorded along with the collection location, water depth, water temperature, and local flow at the time of collection. Additionally, bed configuration and ripple orientation as well as bottom vegetation and organisms, if any, were noted.

Laboratory Procedure

A prerequisite for establishing a practical sedimentary process model is a series of easily obtained, commonly used input elements. A brief introduction to these elements and the means utilized in obtaining them sets the state for an examination of their interrelationships when influenced by dynamic processes.

Upon receipt at the Marine Research Laboratory, supplementary numbers were assigned the samples in order to locate stations readily on the charts.

For each sample a 50 ml fraction of unwashed sediment was bottled and sealed. This was to be saved as a permanent reference sample. The remainder of the sample was washed in fresh water to remove all salt, then dried at 50 to 80° C. The dried sediment was disaggregated, if necessary, and split into two 50 ml portions, one to be used in the light mineral sieve analysis, the other for the heavy mineral analysis. The remainder of the dried sample was subsequently used in conducting flume tests.

SIZE ANALYSIS

Mechanical analyses were made on a 50 ml portion of each sample according to the method described by Folk (1968). Eight inch sieves were used on a one-fourth phi interval. Each sieve fraction was weighed to 0.01 gm on a beam balance. Sieve fractions were computed as percent total sediment weight and plotted as a weight accumulation curve. Diameters representing the 5, 16, 50, 84, and 95 percentile intercepts were used in computing textural statistics.

PARTICLE SHAPE

Sediment particle shape defines the geometrical properties of a grain. A shape factor SF should aid in describing the volume, surface area, and orientation of a particle in a sediment bed. Each of these parameters is critical in that it partially defines resisting and flow forces on the grain in a fluid.

Small protrusions on the surface of a grain are difficult to measure, and their influence on the grain behavior is considered negligible in this model. This assumption is considered valid because most natural sediment grains were observed to be well worn.

McNown *et al.* (1951), in studying the effect of shape on settling velocity, noted there is little difference in settling velocity between ellipsoids and observed values for various other shapes with equivalent principal axis lengths. This indicates the ratios of the major axes of natural sediment grains may be obtained from settling velocity tests. The ellipsoidal shape provides an easily treated and conceived geometrical form when dealing analytically with natural particles.

Settling velocity tests were conducted on all sieve fractions. A settling tube 10 cm in diameter and 122 cm long was mounted in a vertical bracket and secured to a wall. The tube was filled with fresh water, and the temperature was maintained at 20° to 22° C. The temperature was taken from a thermometer fixed inside the tube and recorded after each test. Rings were scribed around the cylinder at 15 cm from the top and 25 cm intervals thereafter for a distance of 100 cm. A white background was placed behind the tube to aid in monitoring dark heavy mineral movement. A black background was placed at right angles to this to aid light-colored light mineral measurements.

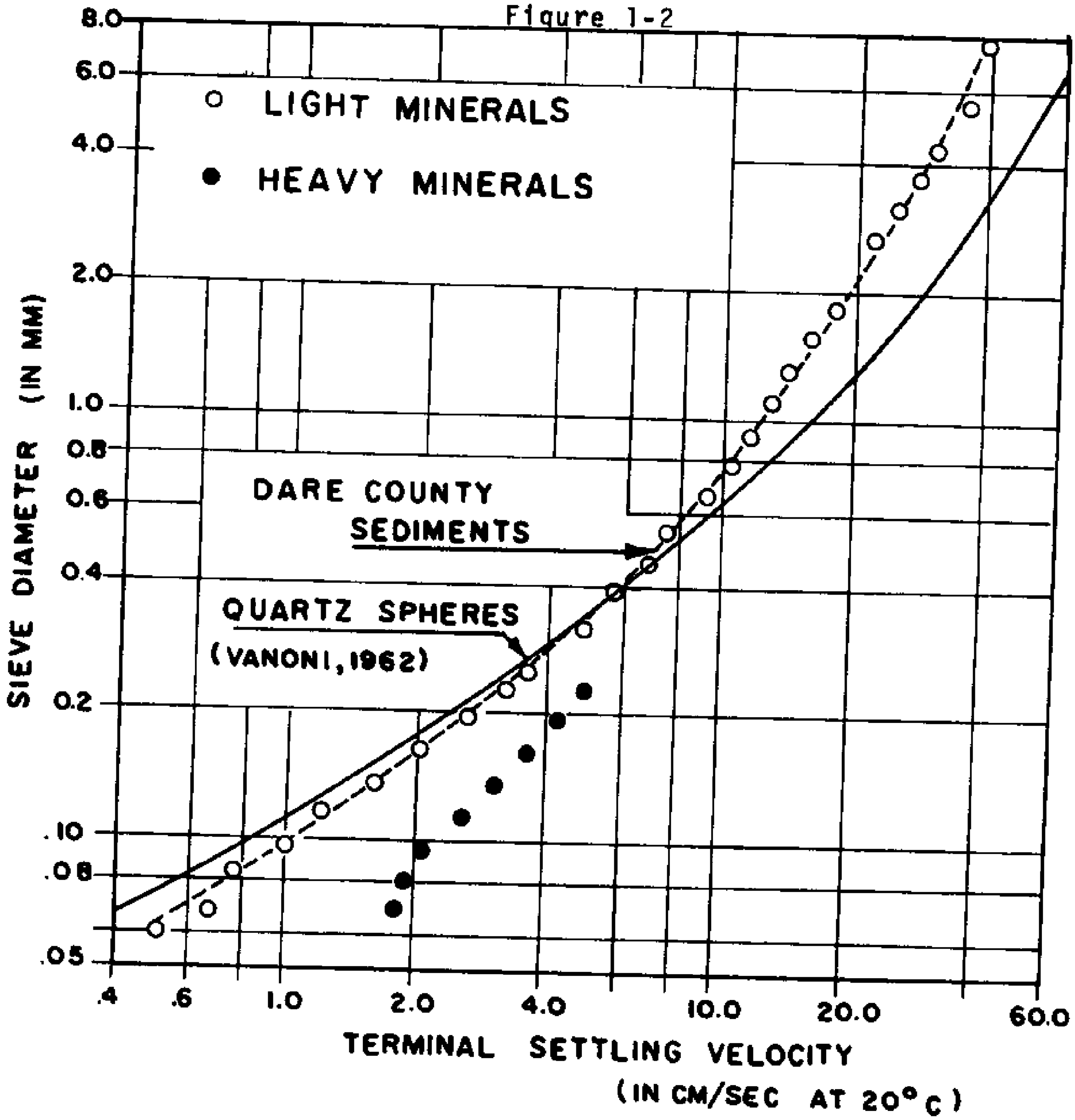
The above procedure for obtaining settling velocities was rapids, rarely taking in excess of one minute for each sieve fraction; indeed settling velocity measurements were made while the next sieved sample was still being shaken.

A small spatula was wetted and dipped into the dry sample until a few dozen grains were trapped in the wet bead. Only a small amount of sediment was used because, in concentrated quantities, interference between grains affects their fall velocities. The spatula was slowly dipped into the free surface of the tube with the wetted sediment bead face downward. The falling sediment reached terminal velocity when recording began at the first scribed ring. A stopwatch accurate to 0.1 sec. was then tripped. The grains fell for at least 10 sec. before the timer was terminated at a lower ring. The centroid of the settling sediment mass was recorded, thus ensuring a fall velocity of the mean sample fraction. Time and fall distance data were recorded and then converted to settling velocity in cm/sec.

According to Vanoni *et al.* (1960) and to provide easily defined parameters in the model, sieve diameter and settling velocity are utilized to define shape while assuming an ellipsoidal form. The values for the three perpendicular axes are obtained from an experimental plot of sphere and sedimentation diameter. The latter is defined as the diameter of a sphere of the same particle density and the same terminal settling velocity as the given particle in the same sedimentation fluid (Vanoni *et al.*, 1960). Knowing the settling velocity of quartz spheres and the settling velocity of the natural sediment of the same density as obtained in still water at a temperature of 20° to 30° C, the sedimentation diameter is obtained from Figure 1-2. A plot of sieve diameter versus settling velocity for a composite of all sediments from Dare County is shown in this figure. The mineralogy of natural material was obtained petrographically.

Grains settling at particle Reynolds numbers less than 18 were noted to fall at whatever orientation they had when released. This accounts for the discrepency in settling velocity between spheres and natural sediments. Stable motion with the short axis parallel to flow motion involved a maximum particle Reynolds number of approximately 100. Between Reynolds numbers of 100 and 1000 instability was observed to develop dependent on particle shape.

Figure 1-2

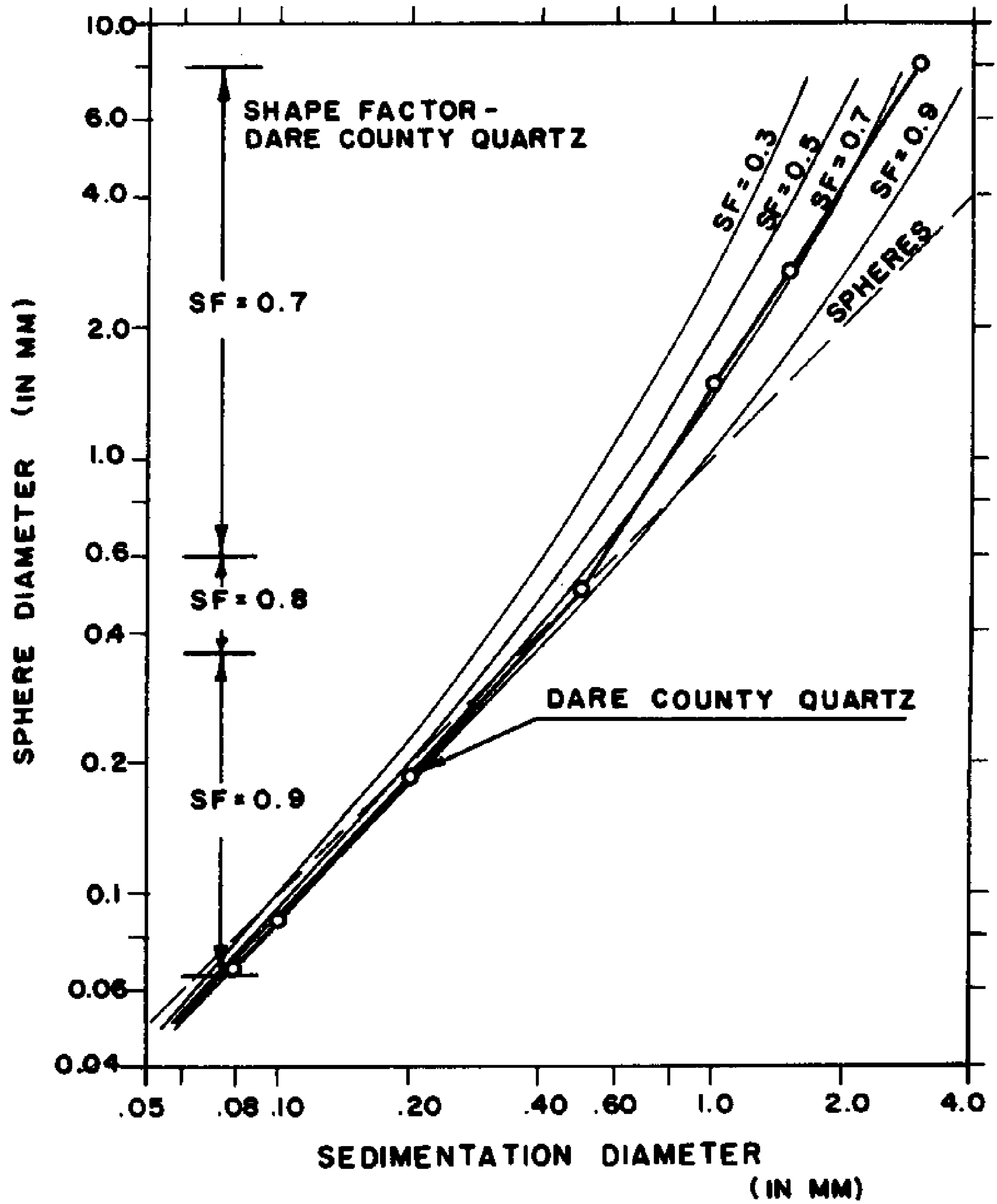


TERMINAL SETTLING VELOCITY - QUARTZ SPHERES & NATURAL SEDIMENT *

(AFTER VANONI ET AL., 1962)

*Precise limits of turbulent and laminar conditions have not been defined in this study.

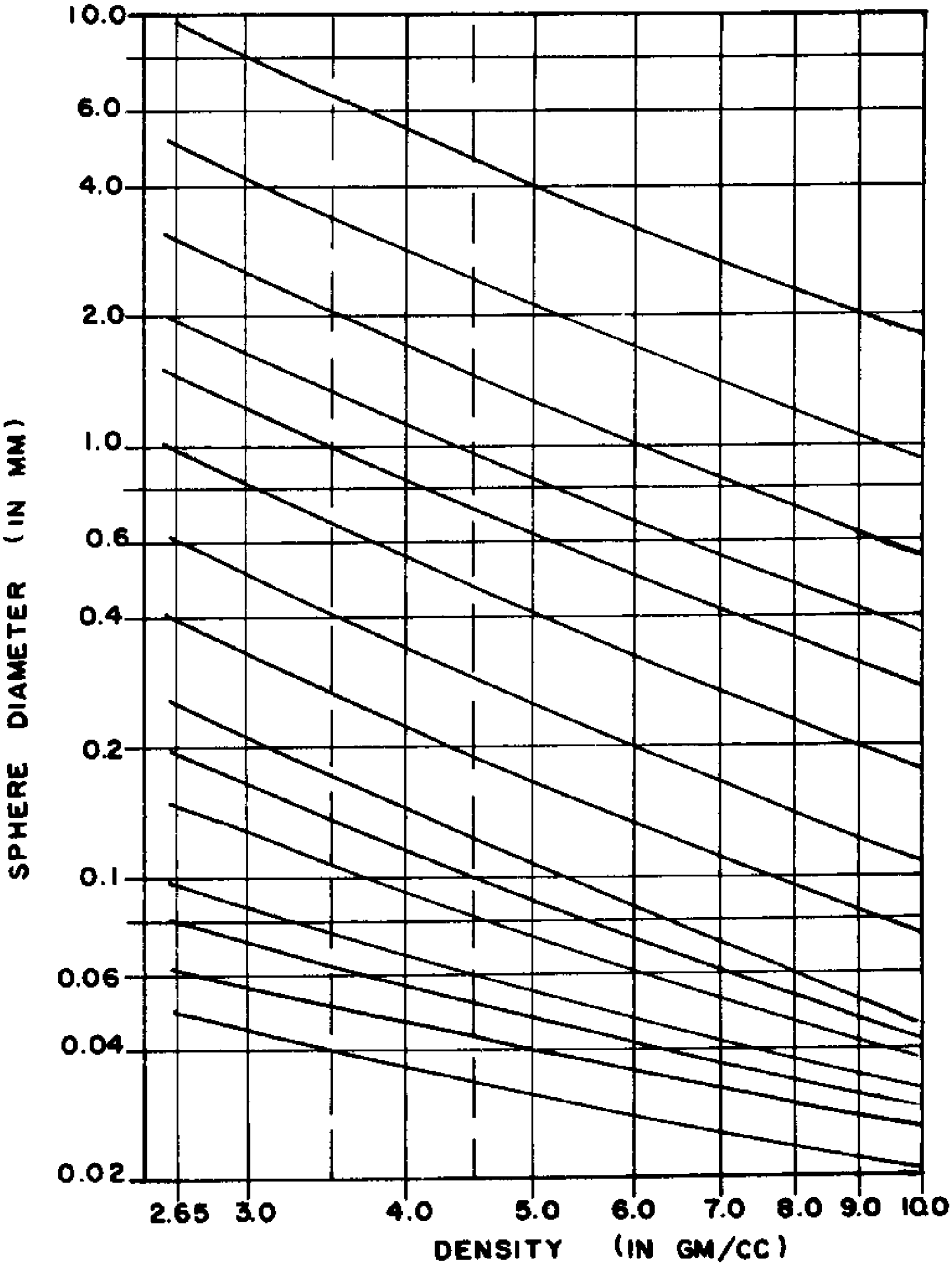
Figure 1-3



SHAPE FACTOR - DARE COUNTY SEDIMENTS

(FROM VANONI ET AL., 1962)

Figure 1-4



EQUIVALENT SPHERE DIAMETERS BASED ON TERMINAL SETTLING VELOCITY

(MODIFIED FROM TOURTELOT, 1968)

In figure 1-3 the sedimentation diameter plotted versus sphere diameter for the Dare County samples is compared to various Corey shape factors. The shape factor formula is $\underline{SF} = c/(ab)^{1/2}$ where a is the long axis, b, the intermediate axis, and c, the shortest axis (Vanoni *et al.*, 1960). The result is the ratio of the three axes as a function of sphere diameter which is reconverted to sieve diameter with the aid of Figure 1-2. Assuming the length of the b axis equal to sieve diameter d_s, the ratio of the a and c axes is obtained. This ratio defines the length limits of each axis if c is to be less than b and a greater than b. The final dimensional values are inferred to lie half way between maximum and minimum values. In the case of particles of much different shape than an ellipsoid, various ratios of c^2/a can be obtained by direct observation, however, most sediment grains are satisfactorily approximated by the ellipsoid. The average shape factor of natural sediment particles is approximately 0.7 (Kennedy and Koh, 1961), N.B., Cf, herein, p. 18.

In order to find the Corey shape factor for natural sediments with densities other than 2.65 gm/cc, Figure 1-4 may be used to obtain equivalent sphere diameters (quartz) of each known heavy mineral based on terminal settling velocities. A curve then constructed on Figure 1-2, plotting equivalent quartz sphere diameter against settling velocity, results in a method of obtaining sedimentation diameter analogous to that previously described. The shape factor is taken from Figure 1-3.

Although Rittenhouse (1943) and other investigators have indicated shape to be of minor relative importance in sedimentation, in a complete theoretical treatment of sediment motion, the shape factor must be included.

PARTICLE DENSITY

The difference in densities of particles in the sediment as a whole is the main cause of selective hydraulic sorting and the concentration of placer minerals. The common rock forming minerals such as quartz and feldspar have densities near 2.65 gm/cc. These form the bulk of most sediment in the silt and larger size range. Some heavy minerals have a density near 20.0 gm/cc, as evidenced by natural gold and platinum, including admixed elements. Table 1-1 lists the maximum, minimum, and average densities of common placer minerals.

TABLE 1-1

PLACER MINERAL DENSITIES (in gm/cc)

Mineral	Min. Density	Max. Density	Ave. Density
rutile	4.18	4.25	4.22
chromite	4.10	4.90	4.50
ilmenite	4.50	5.00	4.75

monazite	4.90	5.33	5.12
magnetite	5.18	5.18	5.18
scheelite	5.90	6.10	6.00
columbite-tantalite	5.30	7.30	6.30
cassiterite	6.80	7.10	6.95
wolframite	7.00	7.50	7.25
platinum	14.00	19.00	17.00
gold	15.00	19.30	17.20

The individual ranges are indicative of variations caused by impurities in the natural minerals.

A rapid means of obtaining the approximate average density of an individual heavy mineral fraction is to plot the sieve diameter of the heavy mineral against the equivalent light mineral sieve diameter as based on equivalent terminal settling velocities (Figure 1-4). This approach assumes a constant shape factor for each density. Using this method the density of a composite of each heavy fraction of the Dare County samples is given in Figure 1-5. The Dare County heavy minerals exhibit a narrow range of sieve diameters, and the lower density fractions are a complex mixture of many mineral species.

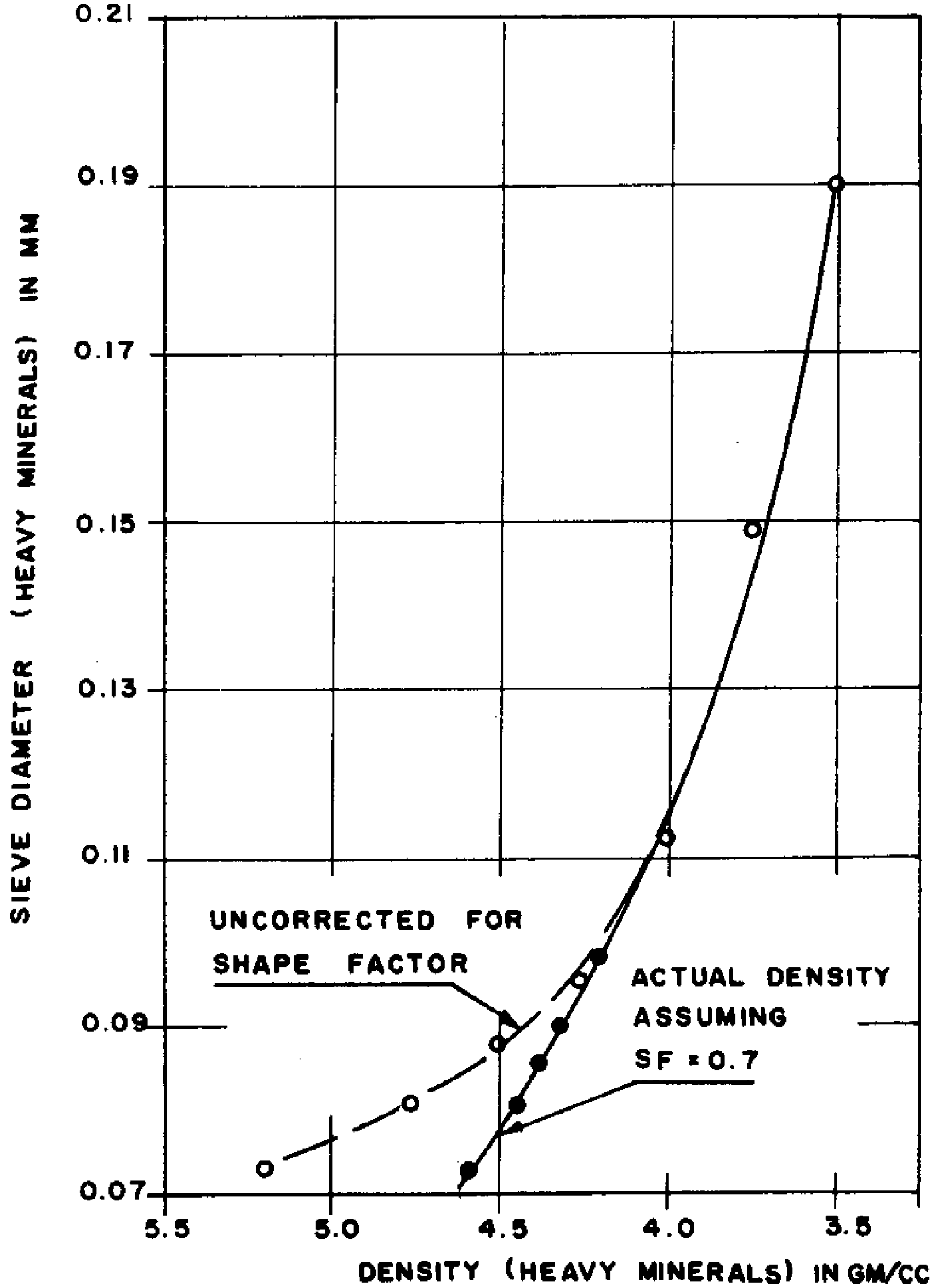
HEAVY MINERAL ANALYSIS

Light and heavy minerals were separated from a total sample volume of 50 ml. A small amount of tetrabromethane(density = 2.94 gm/cc) was added to a 500 ml separatory flask in a ventilated hood. A weighed sediment was next added to the flask followed by more heavy liquid to a total of 250 ml. The flask was then stoppered and shaken and allowed to remain motionless for one to two hours depending on the turbidity of the fluid. When all grains had settled either on the surface of the fluid or in the thin lower neck of the flask, the heavy minerals were withdrawn onto a filter paper in a funnel and washed in acetone to remove all the heavy liquid adhering to their surface. Heavy minerals and the filter paper were then removed from the funnel and allowed to dry at room temperature.

The remaining heavy liquid was removed from the flask. and the light minerals in the flask were washed with acetone. Miscible wash mixtures of acetone and tetrabromethane were placed at the rear of the hood where the acetone evaporated.

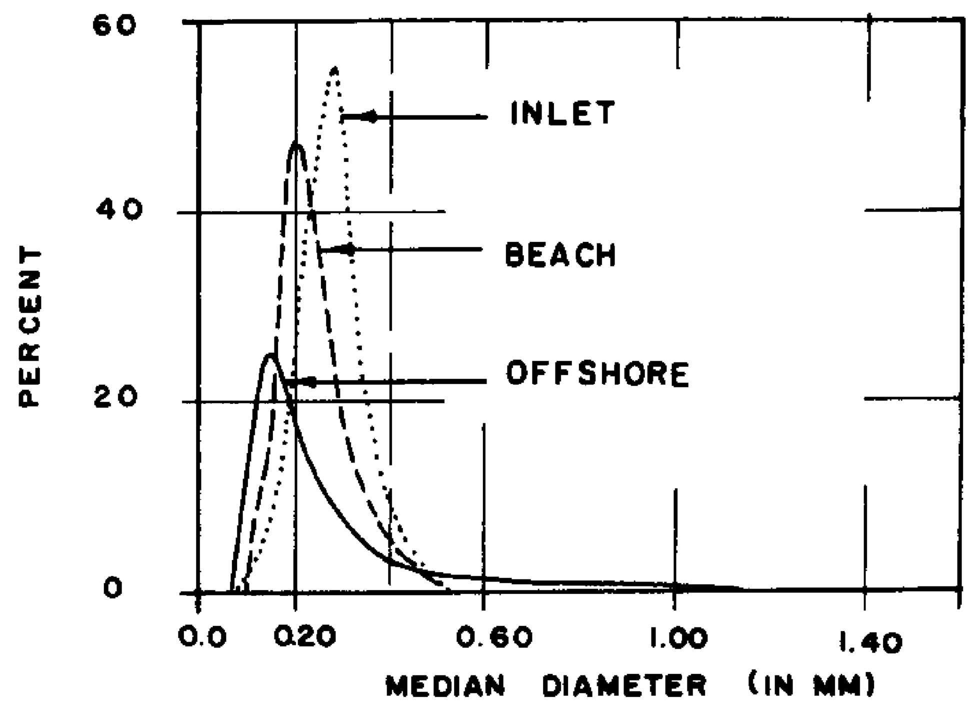
Once the heavy minerals were dry, they were weighed and the total weight percent heavy mineral value computed. One hundred sixty-one selected heavy mineral samples were sieved and the 5, 16, 50, 84, and 95 percentile intercepts calculated.

Figure 1-5

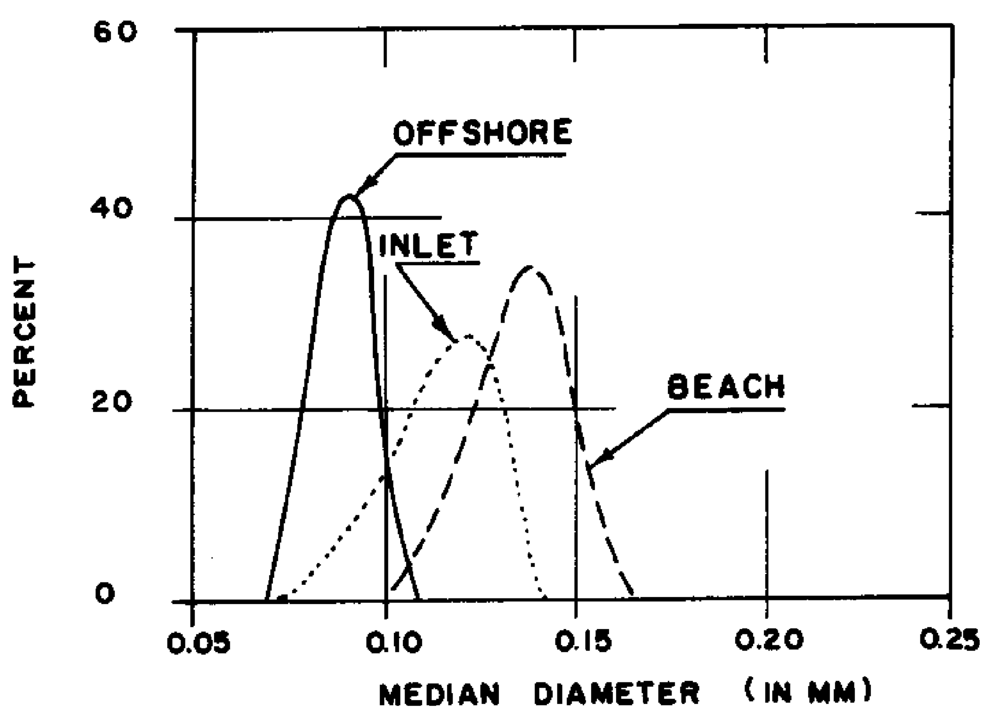


DENSITY OF DARE COUNTY HEAVY MINERALS

Figure 1-6



RANGE OF AVAILABLE LIGHT MINERAL SIZES



RANGE OF AVAILABLE HEAVY MINERAL SIZES

SEDIMENT AVAILABILITY - DARE COUNTY

SEDIMENT AVAILABILITY

Sediment availability (SA) is defined as the range and total percentage of each grain size and density available to forces effecting change within a system. Without the knowledge of absolute availability, it is impossible to properly evaluate textural parameters and sediment dispersal.

Sediment availability, as such, is dependent on an investigator's concept of a system, including this one. A system is defined as any localized region characterized by a known or inferred sediment group subject to known or inferred flow conditions. The system may be either closed with no sediment passing its boundaries or open with a known sediment input and loss across its boundaries. A system is chosen to facilitate a description of the responses to fluid/sediment interactions in a region. It may be as small as a ripple wavelength or as large as a continent. In any case a system is chosen as a spatial-limiting boundary which aids in solving a specific problem.

Figure 1-6 illustrates the percentage of the various heavy and light mineral fractions available in Dare County environments or systems. This limited size range is the culmination of all the mechanisms of transport and wear that have acted upon the sediment from its parent source to its present location. By knowing the sediment range available to the system today and in the past, it may be possible to define the equilibrium dispersal patterns that have resulted. Sediment availability is the foundation upon which any understanding of sediment dispersal is based.

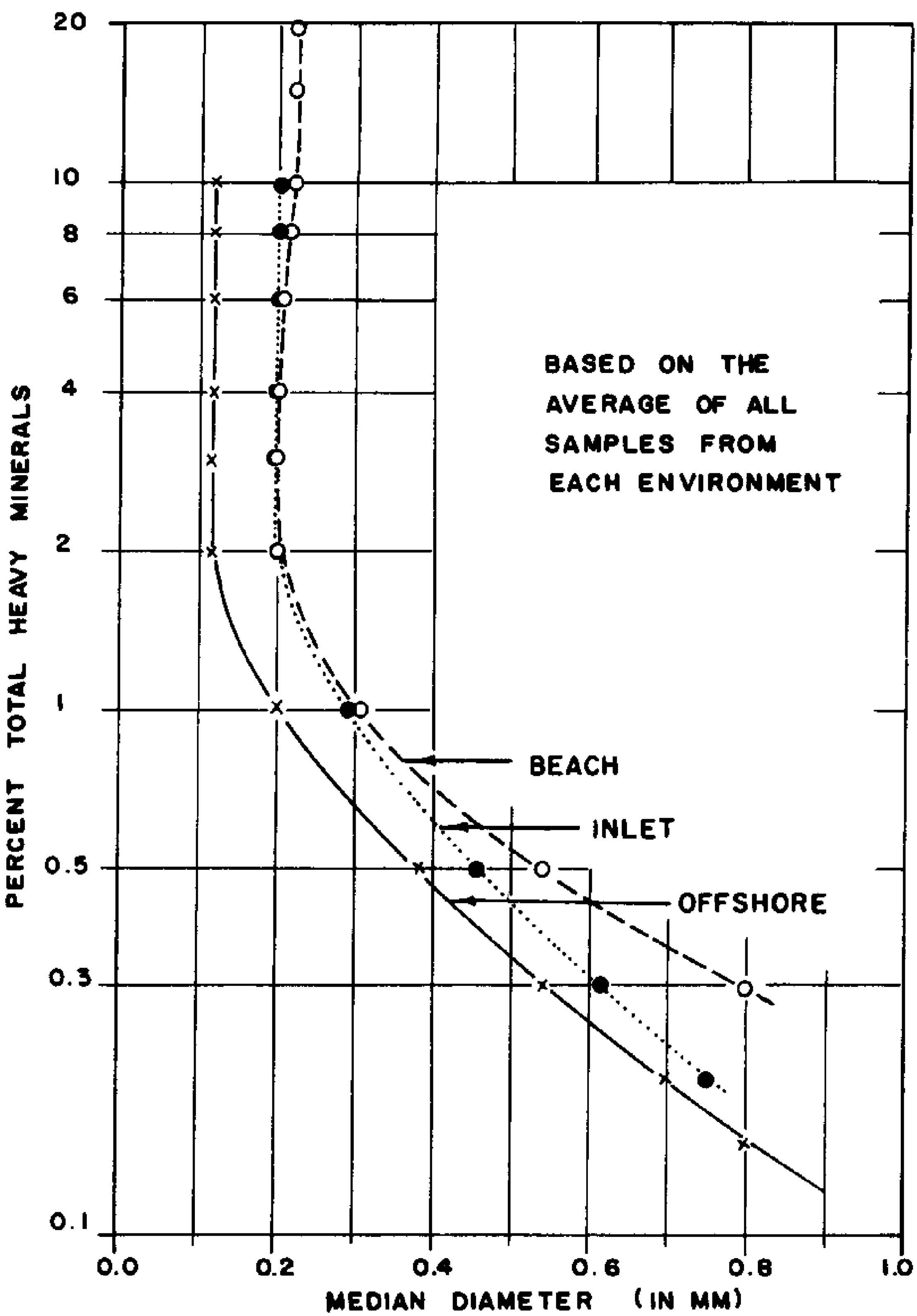
Data on sediment availability in modern sediments may be obtained by procuring a few representative samples of the system in question. Samples should be taken from the extremes of flow energy regions as well as the average. For instance, in considering a beach system, samples recovered from the berm, swash zone, and plunge point would fit these criteria and generally cover the total range available. Inferrences based on the geologic history of an area which may or may not be based on physical data will provide a clue to past sediment availability.

MEDIAN DIAMETER

Median sieve diameter, M_d, is the size at the 50th percentile intercept when all fractions of a sample are plotted on a cumulative frequency graph. The range of M_d for each Dare County environment* is given in Figure 1-6. Due to a paucity of silt and finer sized material in all the environments, a very sharp termination of M_d is noted. A plot of percent total heavy minerals versus M_d in a total sample is given in Figure 1-7. Although heavy mineral percentages are a function of available heavy and light minerals below two percent, above that value median diameter remains constant. This is indicative of other influences involved in placer formation.

*See page 8.

Figure 1-7



TOTAL HEAVY MINERAL PERCENT AS A FUNCTION OF MEDIAN DIAMETER

In defining sediment dispersal patterns as well as discussing the behavior of a sediment, in a fluid flow situation, M_d is very much a function of sediment availability. Any deviation from the average M_d in a system is interpreted as the result of differential behavior of the sediment in response to dynamic influences within the system. Because the nucleus of the sediment population centers around a small M_d, any deviation in heavy mineral concentration and sediment dispersal as the result of dynamic influences should be observed in the arrangement of the total sediment population around the median diameter.

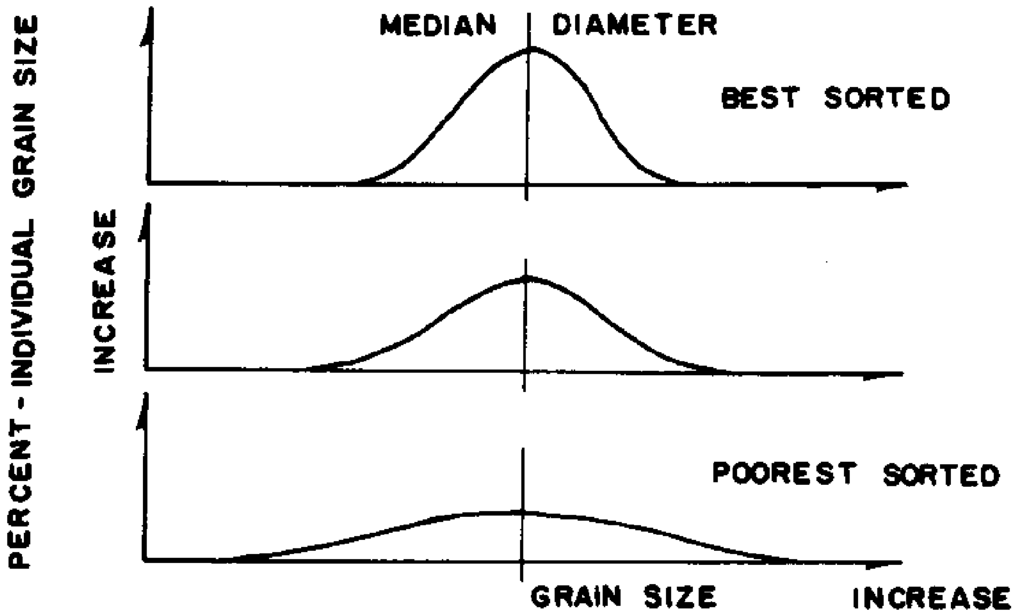
SORTING AND SKEWNESS

Sorting $\underline{\sigma}$ and skewness $\underline{Sk}$ are two textural parameters describing variations in a sediment population around the median diameter. Sorting describes the range of various grain fractions, that is, the total spread of sediment sizes in the sample. Sorting as defined by Inman (1952) is $(\phi_{84} - \phi_{16})/2$ with ϕ_{16} and ϕ_{84} the 16th and 84th phi percentiles respectively. The σ value increases as the spread increases. This is illustrated schematically in Figure 1-8a. Skewness, on the other hand, defines the distortion or deviation of the frequency distribution curve from the symmetrical. Again, following Inman (1952), skewness is given as $(\phi_{16} + \phi_{84} - 2\phi_{50})/\sigma$. Figure 1-8b illustrates variations in the frequency distribution curve due to skewness. Each of these parameters is of critical importance in assessing the magnitude of the various dynamic processes active during sedimentation.

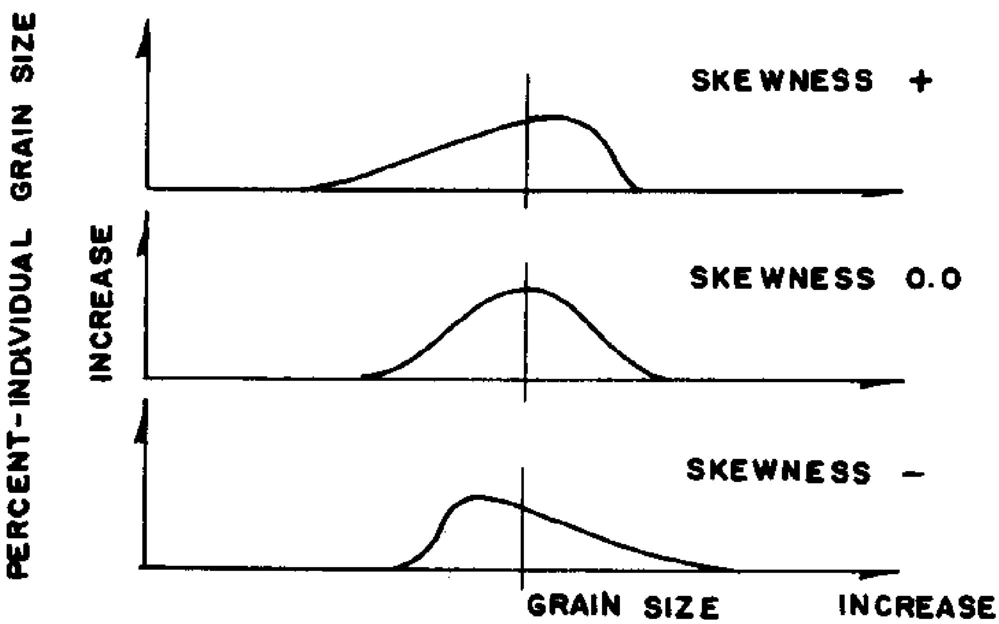
The percentage of heavy minerals as a function of sorting is shown in Figure 1-9 for several environments (systems) in Dare County. In each case the lower sorting values correspond to greater heavy mineral concentrations. A definite cutoff point is observed above which percent heavy minerals is no longer a variable of sorting. Figure 1-10 clarifies the importance of median diameter and sediment availability on sorting. Near the mean median-diameter for a region (Figure 1-6) sorting is best corresponding to the most available sediment fraction. With M_d equal to approximately 0.20 mm for the average of all samples, sorting understandably decreases near this value because this represents the size of the largest available fraction. The sorting values of heavy minerals follow the same trend.

Skewness in the Dare County samples exhibits the same dependence on sediment availability and median diameter as does sorting. Heavy mineral percentages are seen to increase as skewness becomes positive (Figure 1-11). Skewness in both the total and heavy mineral fraction are plotted against median diameter in Figure 1-12. Zero skewness is observed at M_d values near 0.20 mm for beach and inlet samples. The slightly lower value for offshore

Figure 1-8



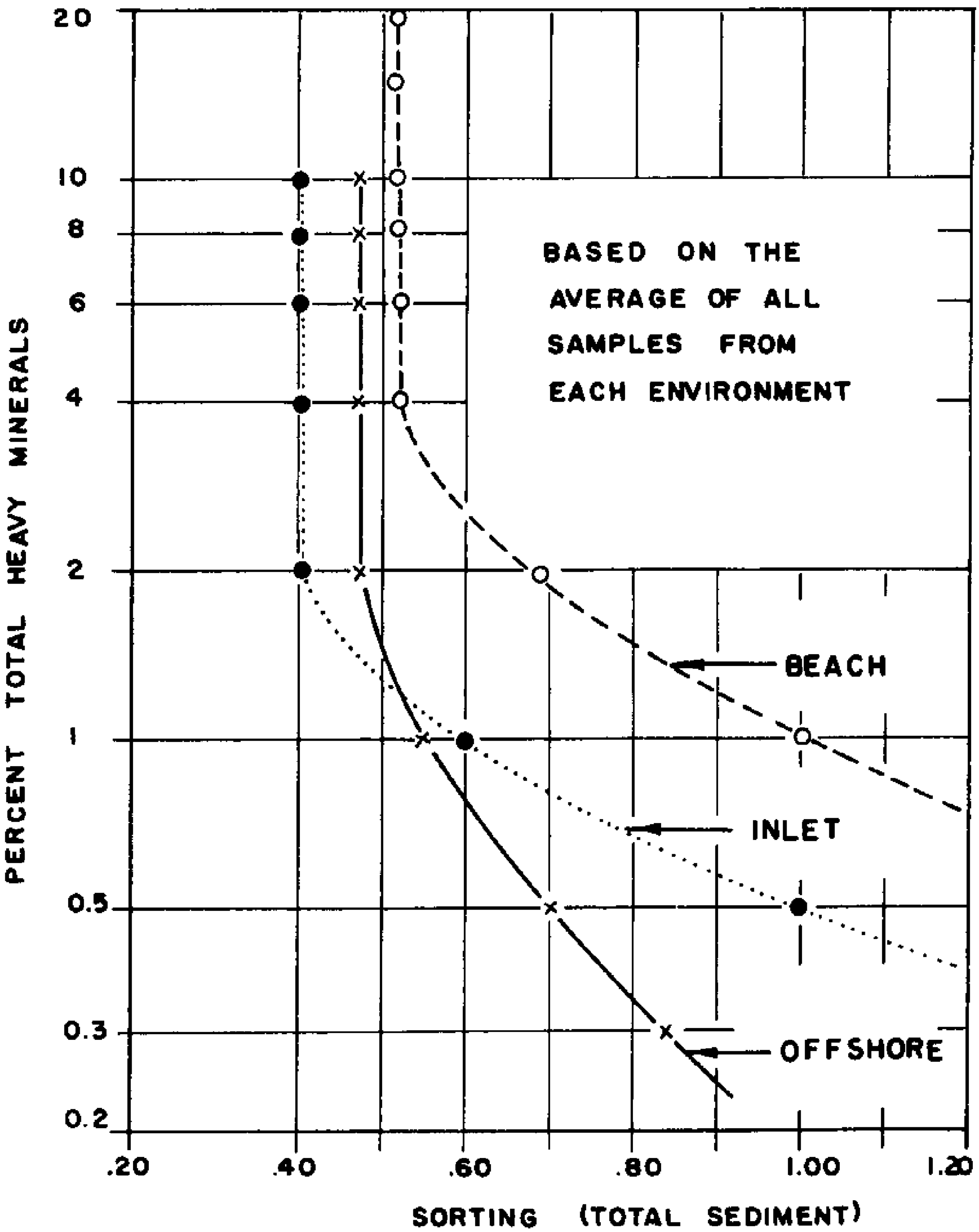
SORTING



SKEWNESS

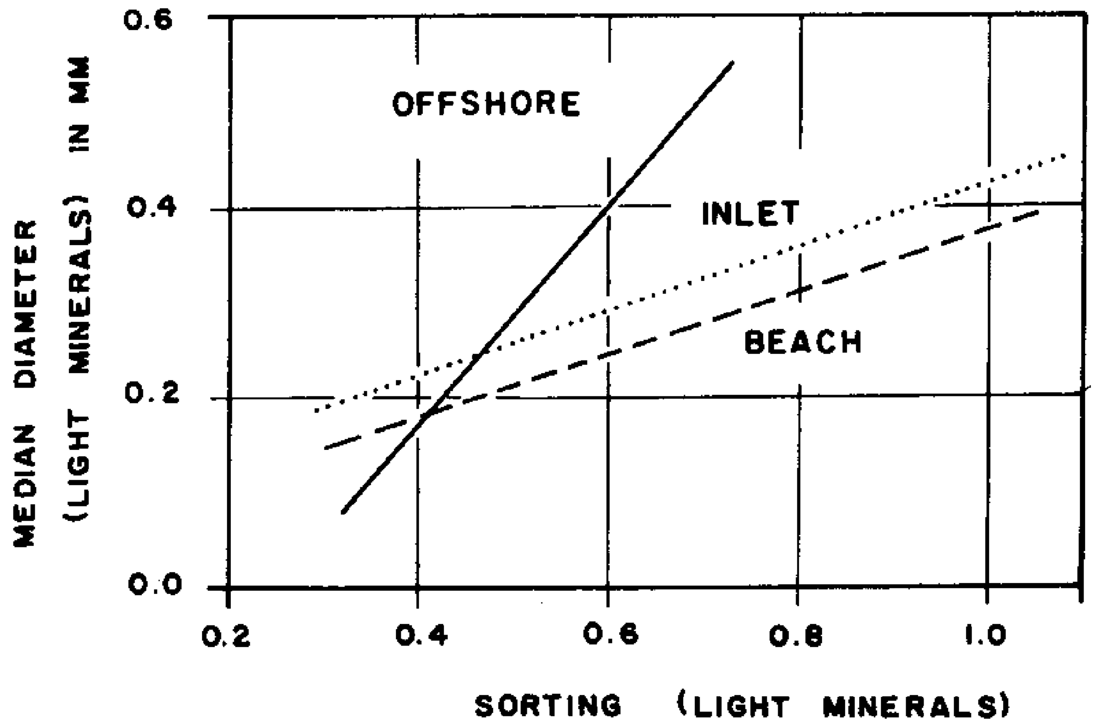
CURVE GEOMETRY - TEXTURAL PARAMETERS

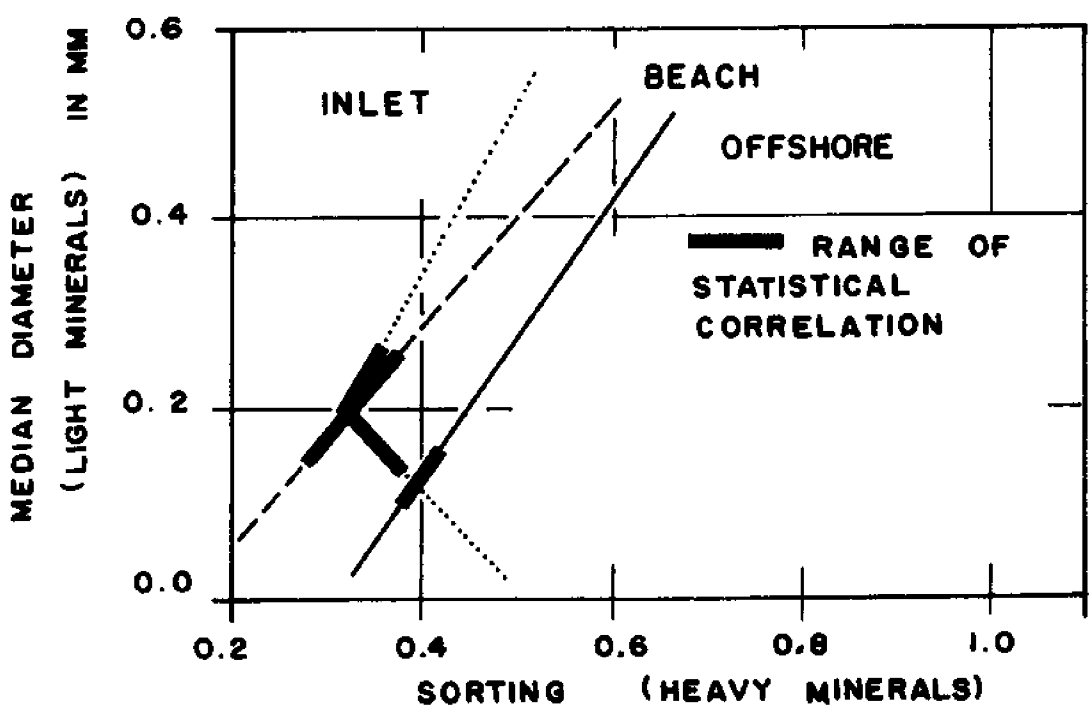
Figure 1-9



TOTAL HEAVY MINERAL PERCENT AS A FUNCTION OF SORTING

Figure 1-10





SORTING - DARE COUNTY SEDIMENTS

samples is due to the somewhat lower average size of offshore samples. Skewness as a function of sorting and median grain size is illustrated in Figure 1-13. A knowledge of the strong interdependency of these parameters is not only helpful to the geologist who uses them in identifying ancient environments, but is essential to a sedimentologist studying the effects of fluid flow on natural materials.

Sediment Composition

Light minerals in the offshore, beach, and inlet systems of Dare County are predominantly quartz. Although quartz is the principal mineral, feldspar frequently comprises a substantial part of samples with a median diameter in excess of 1.0 mm. These are presumably relic sands and often contain 10 to 20 percent feldspar.

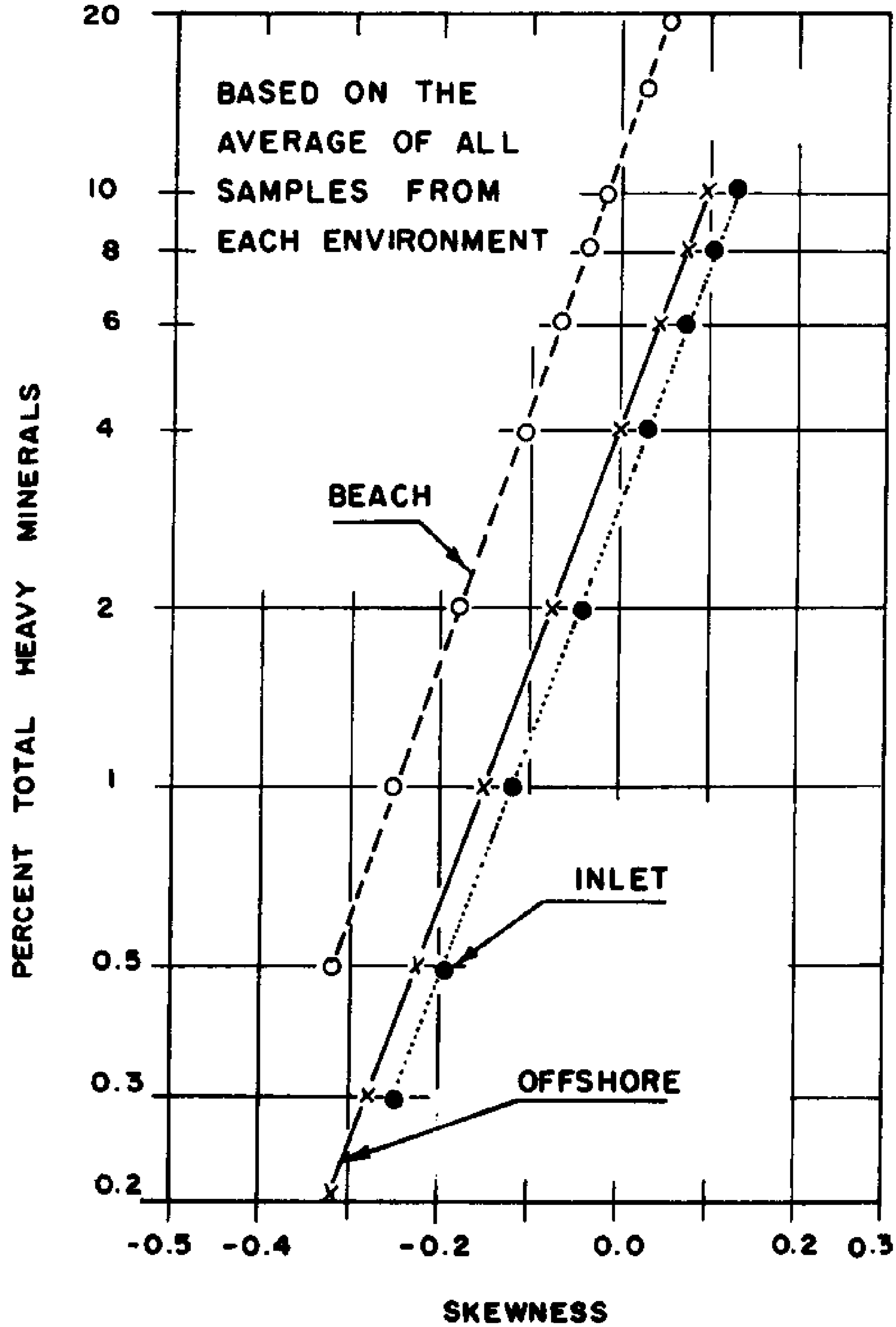
Heavy mineral species in the study area have been identified by Professor D.J.P. Swift of Old Dominion University in Norfolk, Virginia. (Swift, personal communication). He observed the following percent relationships based on nine sampling profiles:

MINERAL	OFFSHORE%	BEACH%
epidote	4.5	5.5
tourmaline	4.0	2.8
staurolite	5.5	4.4
amphibole	16.0	19.2
kyanite	4.0	4.4
garnet	14.0	14.8
opaque	50.0	45.0

Opaque minerals are primarily magnetite, ilmenite, and leucoxene (leucoxene is not always opaque). They generally form small well-rounded particles and have a higher specific gravity than the non-opaque minerals.

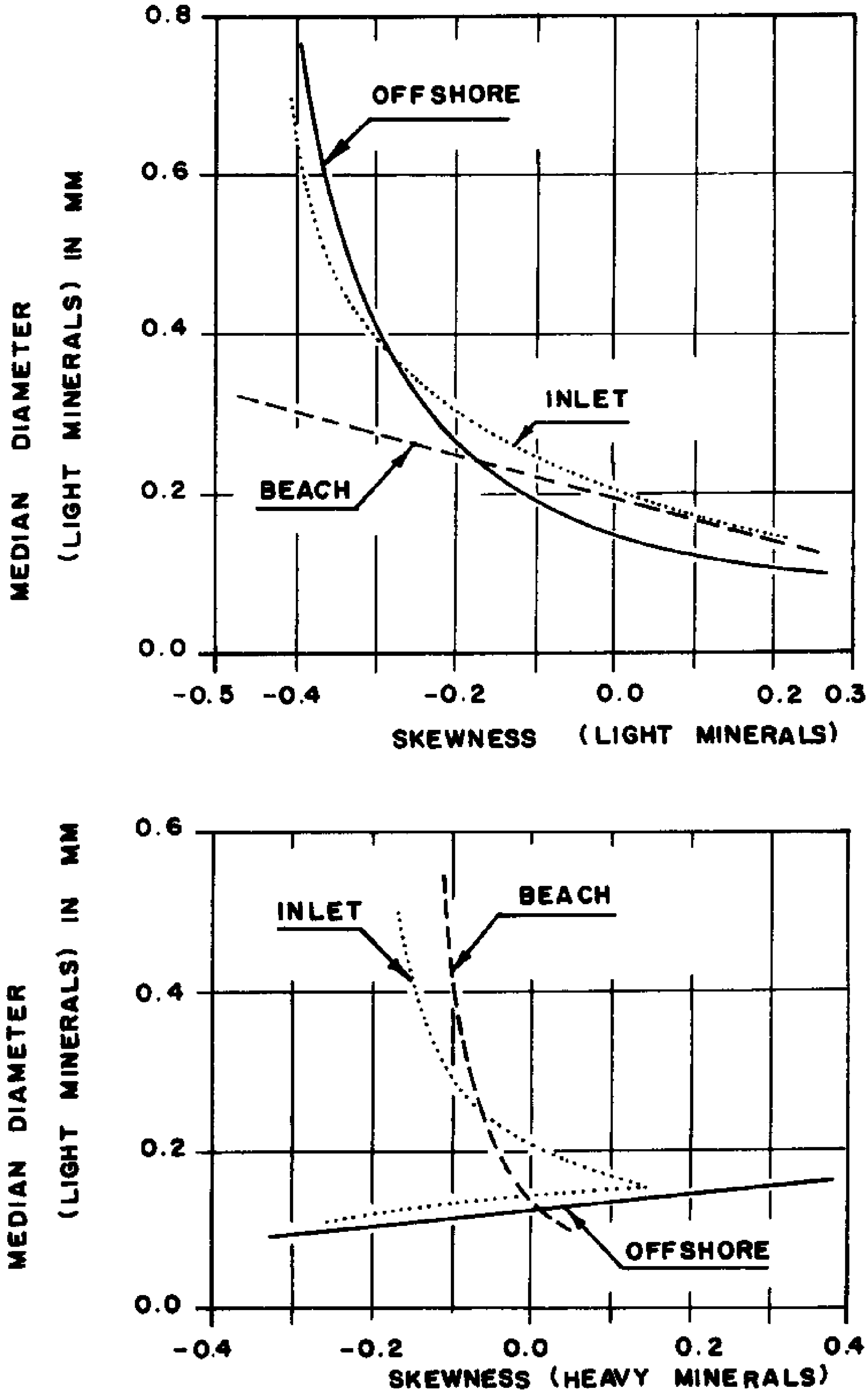
The percentage of opaques and the total percentage of heavy minerals were observed to be closely dependent on each other. This is reasonable because the greater the density difference between light and heavy minerals, the greater the possibility of dynamic separation. Judge (1970) and Martens (1935) observed the same phenomena in beach deposits in California and the eastern United States respectively. Since the economically valuable minerals have higher specific gravities than epidote, hornblende, and other common non-opaques, this suggests the more concentrated deposits will also be richer in valuable heavy minerals.

Figure 1-11



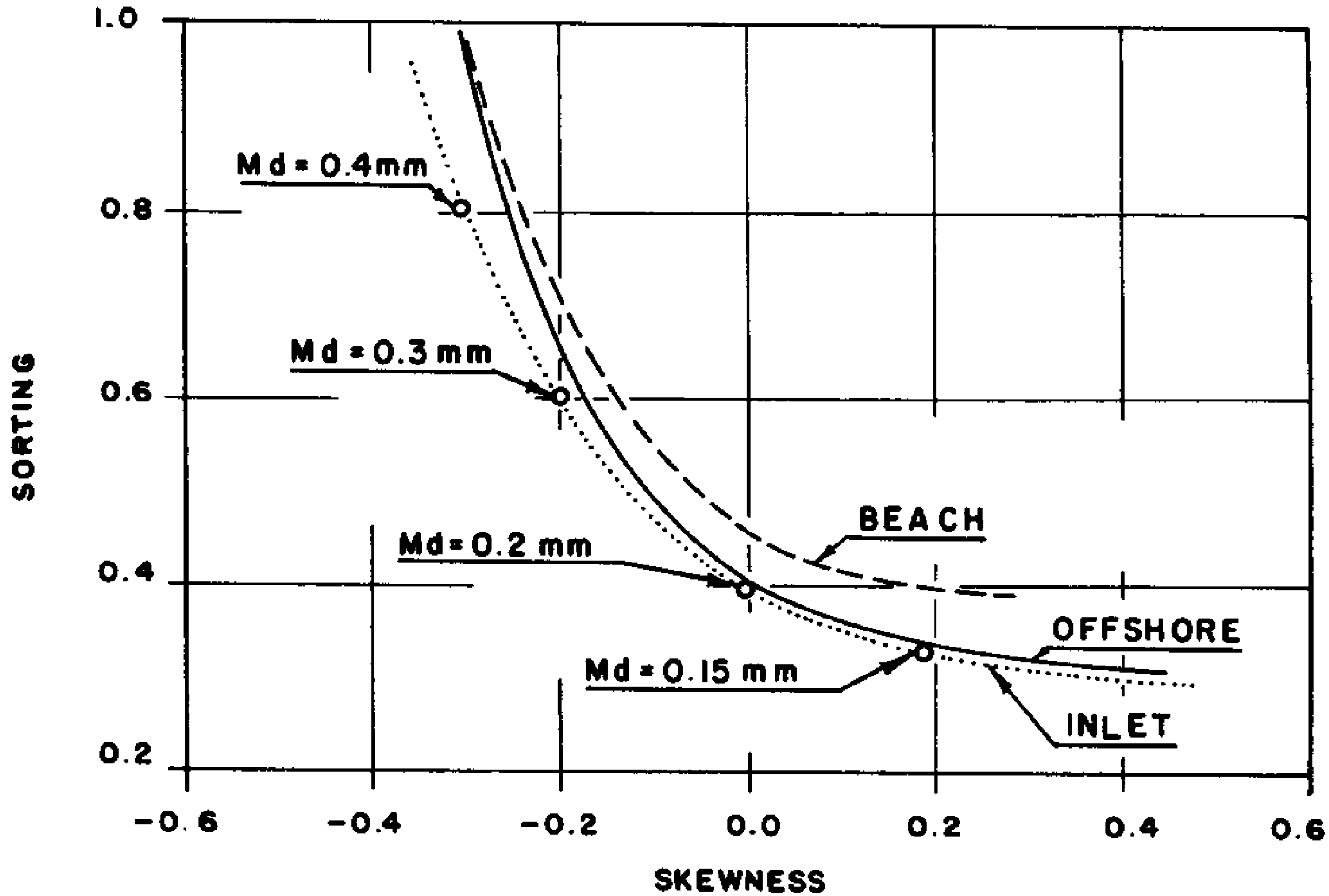
TOTAL HEAVY MINERAL PERCENT AS A FUNCTION OF SKEWNESS

Figure 1-12



SKEWNESS - DARE COUNTY SEDIMENTS

Figure 1-13



SORTING AS A FUNCTION OF SKEWNESS
DARE COUNTY SEDIMENTS

CHAPTER II

GENERAL PLACER MODEL

INTRODUCTION

"Sans la theorie la pratique n'est que la routine donnee par l'habitude"
- PASTEUR

Heavy mineral concentrations, whether their form is a highly concentrated laminae, a less concentrated bulk deposit, or an accumulation on or within a non-movable surface, all represent the final result of a complex interrelationship of geological factors, hydraulic conditions and the sediments themselves. An understanding of these relationships must ultimately be based on a detailed physical knowledge of the complicated fluid-sediment interraction. This chapter is devoted to formulating and evaluating a rational model of heavy and light mineral behavior based on this interaction. Subsequently, the model will be interfaced with geological variables and used as a foundation for predicting and delimiting natural heavy mineral accumulations on loose boundaries.

The general model accommodates sediments in the sand size range and larger (0.62 to 4.00 mm), the so-called cohesionless sediments. Grain densities applicable to the model vary from those of quartz (2.65 gm/cc) and feldspar(2.65 gm/cc) which are light minerals, through ilmenite(4.75 gm/cc) and magnetite (5.18 gm/cc) classified as heavy minerals.

DYNAMICS OF SEDIMENTATION

As water flows over a natural sediment bed it has a singular and often dissimilar effect on each mineral particle of given grain size and density that it encounters. It is this distinguishing characteristic of heavy and light minerals which accounts for the formation of placers. The purpose of this section is to investigate the behavior of sediment particles in a movement cycle beginning with first motion and culminating in deposition. The ability to understand the interrelationships of variables involved in the cycle sets the stage for a later approach to the sediment behavior involved in natural placer formation.

Uniform sediment boundaries were used in all flume experiments because field data (Figure 1-9) indicates that high heavy mineral concentrations are well sorted and essentially uniform in both their light and heavy mineral fraction.

INITIATION OF SEDIMENT MOTION

Uniform sediment grains lying on a flat bed and exposed to a gradually increasing flow velocity were observed to begin motion at a specific average velocity and flow depth for a given grain size and density. The specific conditions of incipient motion are fundamental to acquiring an understanding of erosion and accretion in a sedimentary situation.

Applications of the conditions of incipient motion to heavy mineral accumulation centers about the preservation of a sediment surface and the conditions under which a given particle may be entrapped in that surface. If a deposit is to increase in volume, shear on the fluid/sediment boundary must remain less than that required to initiate motion of the boundary. If the surface composition is known, critical flow and depth conditions may be obtained, corresponding to first motion of the bed. It will be shown that preservation and continued construction of a heavy mineral deposit is very much a function of the critical conditions of incipient motion. The mechanisms involved in initial motion form the basis of the discussion of depositional processes.

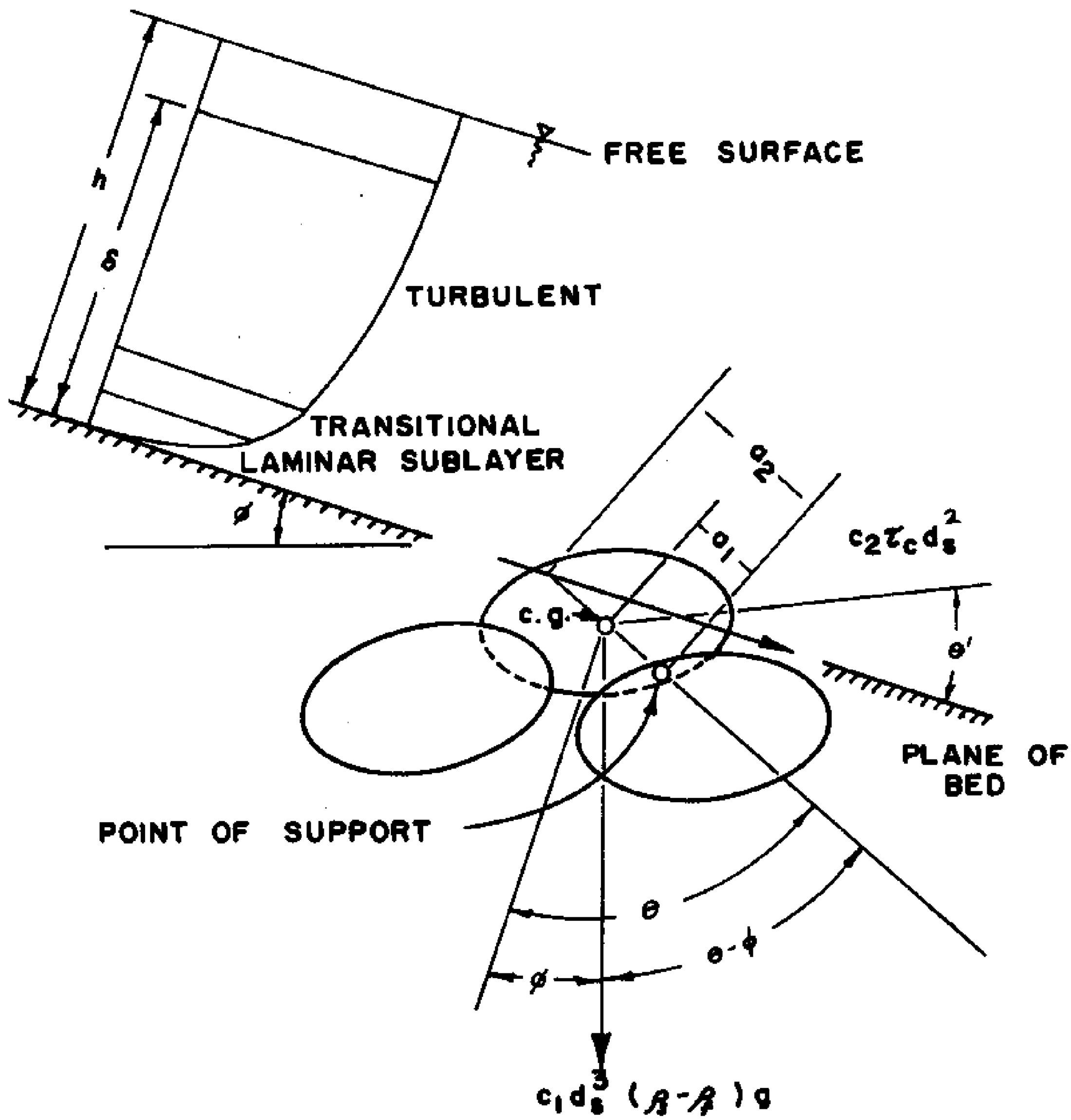
The derivation of an expression containing parameters involved in incipient conditions for a sediment grain on a uniform bed roughly follows that of Vanoni *et al.* (1967). They discuss the initiation of motion problem from its historical inception to contemporary philosophy and include an excellent reference listing.

ANALYSIS - UNIFORM PARTICLES ON A FLAT BED

Assuming a non-cohesive sediment bed over which a fluid is flowing, the forces acting on each surface grain are gravity forces, hydrodynamic drag parallel to the bed, and lift normal to the bed. Lift depends on the same variables as drag and is automatically considered in most theoretical equations because the constants are experimentally determined.

Forces on an ideal grain are shown in Figure 2-1 in which ϕ defines the slope of the bed; θ, the angle of repose of the grain; and θ', the imbrication angle. The imbrication angle is the inflection between the long axis of the particle and the plane of the bed. The lengths a_1 and a_2 are used in defining moment arms, c_1 and c_2 are coefficients referred to the volume and surface area of the bed particle, d_s is the intermediate grain diameter, g is acceleration of gravity, and ρ_s and ρ_f are the sediment and fluid densities respectively.

Figure 2-1



FORCES ON SEDIMENT GRAIN IN BED OF SLOPING CHANNEL

(MODIFIED FROM VANONI, 1966)

At critical conditions of incipient motion, bed shear attains the value T_c and the particle becomes unstable and begins to roll around its point of support. The result of equating moments of gravity and drag force about this point is

$$c_1(\rho_s - \rho_f)\ g d_s^3 a_1 \sin(\theta - \phi) = c_2 T_c d_s^2 a_2 \cos\theta \qquad (2\text{-}1)$$

where $c_1(\rho_s - \rho_f) g d_s^3$ is the gravity force due to weight and buoyancy with $c_1 d_s^3$ being the volume of the grain. The critical drag force is $c_2 T_c d_s^2$. The quantity $c_2 d_s^2$ is the effective projected surface area of the particle exposed to bed shear at conditions of incipient motion. The moment arm for the gravity force is $\alpha_1 \sin(\theta - \phi)$ and that for critical drag is $\alpha_2 \cos\theta$. Critical shear may therefore be given as

$$T_c = \frac{c_1 a_1}{c_2 a_2} (\rho_s - \rho_f)\ g d_s \cos\phi(\tan\theta - \tan\phi) \qquad 2\text{-}2$$

and for a horizontal bed

$$T_c = \frac{c_1 a_1}{c_2 a_2} (\rho_s - \rho_f)\ g d_s \tan\theta. \qquad 2\text{-}3$$

Resisting Force

The gravity force may be obtained by assuming the particle to be an ellipsoid. In this case the gravity force would be $4/3\pi a d_s c(\rho_s - \rho_f)g$ where a and c are the long and short axes of the particle respectively. The lengths af the axes are a function of d_s and the particle shape factor given in Figure 1-3. If the particle is further assumed to be a prolate spheroid with $a > d_s = c$, the gravity correction factor c_1 is as follows for the various shape factors:

c_1 =	2.0	for SF = 0.5
	1.7	0.6
	1.5	0.7
	1.3	0.8
	1.1	0.9

The gravity force is therefore:

$$c_1 4/3\pi d_s^3(\rho_s - \rho_f)g \qquad 2\text{-}4$$

At sea level the acceleration of gravity is approximately a constant, 980 cm/sec^2.

Resisting Moment Arm

The resisting force moment arm for a flat bed $a_1 \sin\theta$ depends on the orientation of the particle commensurate with that resulting from critical shear stress at incipient motion. This is because a particle will continually reorient itself until it reaches its most stable position relative to critical shear conditions. Any displacement from this position will result in a force couple and tend to restore equilibrium. Since fluid forces acting on the particles are proportional to their cross-sectional area, a particle in its somewhat random journey over the bed will sometime come to rest with its smallest projected cross-sectional area exposed to the flow. This position with the long axis parallel to the flow will be its point of maximum stability. Results of several series of flume experiments related to orientation of sand grains in a unidirectional flow system showed every sample run to have significant orientation in the flow direction (Rusnak, 1957; and Curray, 1956a).

Particles in an end-on position in the flow regime have been observed to be inclined in a direction dipping into the flow (Rusnak, 1957). This angle of inclination, known as the imbrication angle, θ', is shown in Figure 2-1. The free body diagram indicates fluid forces on the grain surface will inhibit lift in this position. Were θ' inclined with the flow, lift would be aided and the grain would not be in its most stable orientation. The imbrication angle increases as surface shear increases; thus the resisting moment arm is balanced. Rusnak (1957) experimentally proved this to be true and noted a correlative increase in θ' at incipient conditions, varying from 8 degrees for a 0.10mm quartz sediment to nearly 18 degrees for a 1.00mm grain size.

The angle θ is generally considered in analytic treatments to be equal to the angle of repose of the grain or 30-35 degrees for sand-sized sediments. However, in a uniform natural sediment, this angle is also a function of orientation and imbrication angle, as previously described; these two are, in turn, dependent on particle shape and grain packing. The additional complication of including θ' in the θ of equations 2-1 to 2-3 is obvious, especially since θ' is difficult to define. It can be shown though, that as the imbrication angle increases, θ will increase only slightly.

The gravity force moment arm will be included in an experimentally determined coefficient involving the fluid force moment arm and other undetermined variables. Orientation angle and imbrication angle would be important in a discussion of the behavioral patterns of particles of significantly different shape factor, and although not explicitly included in the preceding equations defining incipient motion, they are a factor. It should be noted all Dare County sediments have a similar shape factor of approximately 0.7 to 0.8.

Shear Stress

The dynamic fluid forces opposing and equal to the resisting force of a sediment particle at conditions of incipient motion are known, collectively, as bed shear. This is simply a group of forces acting in the direction of flow that results from the pull of the fluid on the wetted bed. Individually, this grouping can be broken into inertial forces, static pressure forces, virtual mass force, and lift and drag forces. As there is difficulty in separating these forces, the general term bed shear for all of them is most frequently used in sedimentation (Silberman,1963). Studies concerned with specific aspects of the boundary friction problem include those of Rouse (1965), and Chow (1959), Schlichting (1968), Yen and Liou (1969), and Sternberg (1968).

The problem in utilizing the shear approach for defining sediment motion arises when determining the energy gradient of total friction loss for a given fluid flow and bed condition. Pioneering work in anything but a purely empirical approach to this problem has its origin in pipe flow investigations where the following expression for total shear

$$T = \rho_f \bar{U}^2/8 \; f = \rho_f \bar{U}^2/2 \; C_d \qquad (2\text{-}5)$$

was derived by equating frictional losses to the measured pressure loss (Moody, 1944). An experimentally derived dimensionless coefficient f, the Darcy-Weisbach friction factor, accounts for the dimensions and roughness of the pipe. The drag coefficient C_d is also of empirical origin. The mean flow velocity is $\bar{U}$.

A better approach to defining shear is the velocity profile method that includes the flow regime near the bed and consequently incorporates the concept of the boundary layer in its makeup. Because much more has been done concerning the first approach, experimental data covering a wide range of conditions are available and will greatly enhance the range of the velocity profile approach to bed shear.

Fundamental to understanding the fluid dynamic foures acting on a movable sediment bed requires a comprehension of flow conditions in the fluid boundary layer near the bed. Here, frictional forces retard the motion of the viscous fluid and the velocity increases from zero at the boundary to its full value corresponding to external frictionless flow. The thickness of this retarded layer and the associated bed shear are governed by surface roughness and the form of the bed, the kinematic viscosity of the fluid, and the velocity profile.

Boundary layers may be laminar or turbulent. A turbulent layer has a laminar sublayer near the bed (Figure 2-1). The thickness of the sublayer with respect to the roughness height of bed particles governs whether a turbulent or laminar velocity profile exists at the particle face. If the sublayer is so thin that bed grains protrude through it, the shear on the particle will be a consequence of a turbulent velocity profile.

In general, the turbulent boundary layer can be considered fully developed in a natural channel (Chow, 1959). This means that the turbulent layer occupies the entire channel and δ, the boundary layer thickness, reaches the free surface. Subsequent discussion of the boundary layer will assume a fully developed condition; however in cases of developing flow or instances where a large change in bed roughness occurs, this assumption may be incorrect. The distance required for a new boundary layer to become fully developed, or the thickness of the layer a given distance from it's origin may be obtained from nomographs relating flow conditions to bed roughness dimensions as reported by Schlichting (1968).

The velocity profile in a laminar, transitional or turbulent boundary layer is given by the semi-empirical logarithmic expressions (Schlichting, 1968)

$$\frac{U(h)}{U_*} = 2.5 \ln \frac{R}{k_s} + B \qquad (2\text{-}6)$$

with k_s equal to an arbitrary roughness height, R the hydraulic radius equal to the depth in wide channels, and U_* the shear velocity, where

$$U_* = \sqrt{T/\rho_f} \qquad (2\text{-}7)$$

B assumes different values depending on the roughness regimes given below:

(1) Hydraulically smooth regime:
All protrusions are contained in the laminar sublayer. Friction is due to laminar skin resistance which is a function of $\frac{U_* k_s}{\nu}$. Conditions for a smooth boundary are

$$0 \leq \frac{U_* k_s}{\nu} \leq 5 \qquad (2\text{-}8)$$

and B is defined as

$$B = 5.5 + 2.5 \ln \frac{U_* k_s}{\nu} \qquad (2\text{-}9)$$

The thickness of the laminar sublayer is

$$\delta = 11.6\ \nu/U_* \qquad (2\text{-}10)$$

When the development of this layer is of consequence, Schlichting (1968, p. 217) states that the thin layer will remain hydraulically smooth if

$$\frac{100\ \bar{u}}{\nu} > k_s \qquad (2\text{-}11)$$

when the Reynolds number $\frac{\bar{U}x}{\nu}$ is less than 10^6. The distances from the origin of the sublayer to the point of interest is the distance x. This formula gives only one admissible k_s for the entire distance and does not account for changing depth. Schlichting (1968), however, shows equation 2-11 to be accurate for the limits set on the Reynolds number.

(2) Transitional regime:

In this overlap region protrusions extend partly outside the laminar sublayer, and additional resistance, as compared to smooth flow, is mainly due to form drag on protrusions in the boundary layer. Turbulent eddies also penetrate to the bed to a certain extent in this regime. Friction is a function of $\frac{U_* k_s}{\nu}$ and $\frac{B}{k_s}$. The value of B for a uniform sand-roughned surface may be obtained from figure 2-2.

(3) Completely rough regime:

All protrusions reach outside the laminar sublayer and by far the largest resistance to flow is due to form drag acting on the surface projections. B is no longer a function of viscosity and now assumes a constant value of 8.5. The completely rough regime is defined when

$$\frac{U_* k_s}{\nu} > 70.0 \qquad (2\text{-}12)$$

The validity of the logarithmic velocity profile has been established in sea floor observations by Lessor (1951), Revelle and Flemming (in: Sverdrup *et al.*, 1942), Sternberg (1966 and 1968), and the author's observations in a tidal inlet at Oregon Inlet, North Carolina. Additionally, logarithmic laws are asymptotic expressions for very large Reynolds Numbers and can be, therefore, extrapolated to arbitrarily large values beyond the range of experiments.

In maintaining the criteria for developing as practical a sediment model as possible, bed shear must be given as a function of known or readily derived parameters. When solving problems the most common quantities available are mean velocity, flow depth, and sediment size and texture. The commonly used dimensionless roughness quantity k_s is a function of these parameters(Schlichting, 1968). Critical bed shear may be defined by the following function

$$\tau_c = \tau_c\,(\bar{u}, h, k_s) \qquad (2\text{-}13)$$

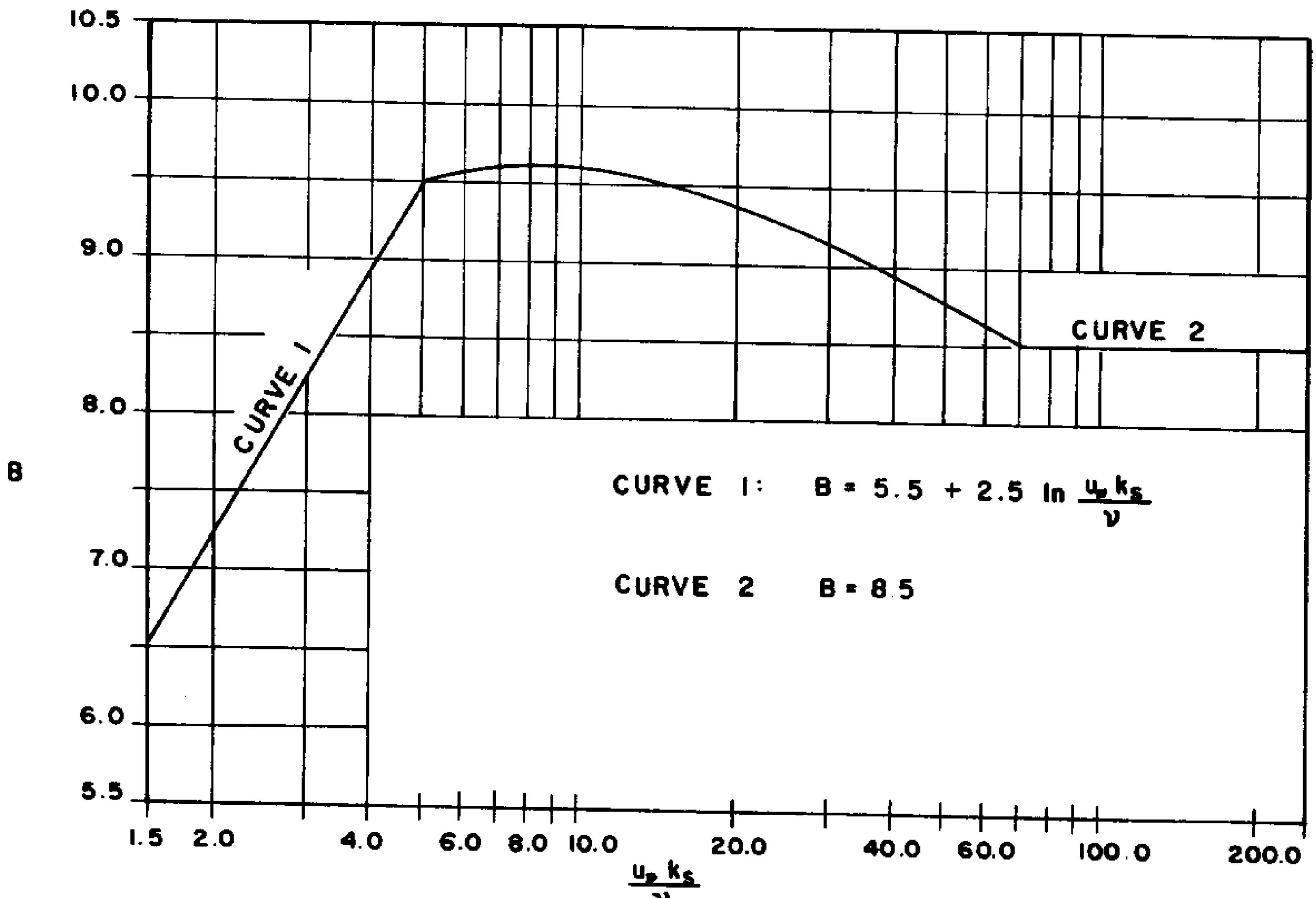
If the mean velocity is assumed to occur at six-tenths of the flow depth or 0.4h, when h is measured from the bed, and R=h, assuming a wide channel, the result is an expression incorporating equation 2-6 for mean shear velocity

$$u_* = \frac{\bar{u}}{2.5 \ln \frac{0.4h}{k_s} + B} \qquad (2\text{-}14)$$

Many determinations of vertical velocity in natural streams have established the mean velocity as that measured at 0.4h (Grover and Harrington, 1966). Although stream gaging, as practiced by the U.S. Geological Survey, commonly obtains mean velocity from an average of two depths, 0.2h and 0.8h, the single measurement used in the past is obviously more adaptable to bottom shear problems. Using the criteria of equation 2-7 and 2-14, mean flow shear $\bar{\tau}$ is defined as

$$\bar{\tau} = \rho_f \left(\frac{\bar{u}}{2.5 \ln \frac{0.4h}{k_s} + B} \right)^2 \qquad (2\text{-}15)$$

Figure 2-2



ROUGHNESS FUNCTION B (FROM SCHLICHTING, 1968)

Equation 2-15 shows that if two flows have different depths but identical bed sediment and same mean velocity, the shallowest flow will exhibit the largest mean shear. Mean velocity itself, therefore, is not sufficient to express bed scour. Depth is a necessary quantity which increases in importance as h approaches the value of d_s(or k_s).

This derivation, utilizing the aforementioned simplifying assumptions, is also based on the assumption that mean shear is a constant fraction of bed shear. In the presence of strong secondary currents, bed shear T_b and mean shear would not be expected to be in agreement. Because secondary currents are difficult to predict in a cross-section they will be disregarded and the assumption made that

$$T_b = K\,\overline{T} \tag{2-16}$$

where K is an unknown empirical constant. Silberman *et al.*(1963) note that for channels with homogeneous boundary material equation 2-16 is approximately correct. The value of K is unimportant in discussing relative critical shear values.

In the complicated and almost wholly experimental study involving values of k_s, the relationship of k_s, f, and C_d is

$$T_b = K\rho_f\left[\frac{\bar{u}}{2.5\ \ln\frac{0.4h}{k_s} + B}\right]^2 = \rho_f U_*^2 = \rho_f\frac{\bar{u}^2}{8}f = \rho_f\frac{\bar{u}^2}{2}C_d \tag{2-17}$$

The arbitrary measure of geometrical roughness properties k_s is probably a function of at least the following variables

$$k_s = k_{s1}\ (h, \bar{u}, d_s, \xi, \Phi, \eta, \rho_f, \nu, \partial h/\partial t) \tag{2-18}$$

where ξ is a measure of the channel cross-section, Φ, the shape of the channel profile, η, the shape of the channel plan, ν, the kinematic viscosity, and $\partial h/\partial t$, an unsteady term. Rouse(1965) combined h, $\bar{u}$, and ρ_f with the other variables to obtain the dimensionless quantities

$$k_s = k_{s2}\ \left(\frac{\bar{u}h}{\nu}, \frac{d_s}{h}, \xi, N, \frac{\bar{u}}{\sqrt{gh}}, \frac{\partial h}{\partial t}/\bar{u}\right) \tag{2-19}$$

The first three terms relate to surface resistance and in principle are common to all conduit flow. The first term is a Reynolds number relating inertial to viscous forces. The fourth and fifth terms involve cross-sectional non-uniformity and define form and wave resistance respectively. N is a combination of Φ and η. The Froude number plays no part in surface resistance in uniform flow since there is no change in free surface conditions; however, it is important in non-uniform free surface flow. The last term defines the departure of the frictional factor due to unsteady conditions. Both surface and form resistance may vary in unsteady flow. It is obvious the solution of frictional problems in rough channels is frustrated by the fundamental difficulty that the number of parameters describing roughness is extraordinarily large.

Hydraulic resistance is also affected in two interrelated ways due to the presence of a movable bed. The boundary may be modified with the formation of different bed forms which result in a corresponding change in the character of the vertical velocity gradient. Resistance may also vary due to an increase or decrease in drag on the bed form itself.

If it were possible to solve the Navier-Stokes equations of viscous motion and the equation of continuity, it would be possible to describe the motion of the fluid and sediment. Even if the initial and boundary conditions are known, analytic solutions of these equations could not be obtained because of their highly-non-linear nature and the present lack of knowledge on the nature of turbulence in the boundary layer. It is because of this difficulty, even when approaching the problem from a purely analytical standpoint, that we must resort to various schemes using experimentally derived values of k_s or f.

Experiments concerning friction as a function of uniform roughness elements on a flat bed indicate roughness can be expressed with the aid of a single roughness factor k_s/R, known as the relative roughness. The equivalent roughness k_s is obtained experimentally. Schlichting (1968) shows that k_s is equal to $0.63d_s$ for the most dense arrangement of spherical particles. He obtained this value by measuring the velocity distribution and shear stress on a rough wall, then with the aid of equation 2-6, he was able to determine k_s. The Darcy-Weisbach friction factor f can be equated to k_s using equation 2-17.

Shear Stress Moment Arm

The derivation of the fluid force moment arm $a_2\cos\theta$ suffers from the same difficulties inherent in solving for the resisting moment arm. Each is dependent on particle orientation, imbrication angle and θ. The angle of repose of the sediment is in actuality dependent on packing which itself is partially a result of flow conditions. Due to the seemingly small possibility of success in analytically determining $a_2\cos\theta$ for a natural sediment, it will be experimentally obtained in combination with the resisting moment arm and other undetermined variables. The experimental determination of unknown quantities will complete the solution of the incipient motion problem for uniform sediment grains in the size and density range existing in the Dare County marine environment.

Experimental Study

1. Equipment

Experiments were performed in a 55 foot flume in the Hydraulics Laboratory of the Civil Engineering Department at the University of Wisconsin, Madison.

The loose sediment test section of the flume was a well eight feet long located 32 feet from the intake box. The flume width was 0.76 feet and the trough had glass walls 0.6 feet in height. The slope of the flume was adjustable from horizontal to about 0.02 feet per foot. A large sump located below floor level contained the water supply for the flume. From this sump the water was pumped to the intake box at the upstream end of the flume with the amount of water regulated by a valve and a semi-quantitative measuring device in the system. Absolute discharge was measured by gaging the water level rise in a constant area volume tank located at the downstream end of the flume. A simple but accurate computation of discharge was obtained of volume increase as a function of time. A straight point gage accurate to 0.001 foot was used to measure depth.

Special equipment used in the sediment studies included five eight foot by nine inch aluminum sections to which a commercial grade of roofing was bonded. The diameter of the sand grains fixed to the roofing was approximately 0.3 mm. A fully developed flow over the test section was insured to be turbulent by utilizing this upstream roughened bed. Three-fourths inch wooden blocks fastened beneath the roughened strips effectively raised the bed of the flume one inch. Four strips were connected and placed upstream of the test section. They were bonded with a watertight sealer to the sidewalls to prevent fluid from flowing beneath the artificial bed. An aluminum strip eight feet long placed one-half inch above the flume bed was next cemented in place. This was the test section. The size of this section was regulated by the available sediment, especially the heavy mineral fraction.

Bonded to the test section on its downstream margin was the fifth roughened section. The experimental apparatus in the flume, therefore, consisted of a 32 foot roughened bed upstream of an eight foot sediment well which was followed by an additional eight foot roughened section.

2. Sediment

Sediment used in all flume experiments was obtained during field sampling in Dare County,North Carolina. This natural sediment was the composite of all samples taken in the field and is entirely representative of the sediments in this area. The heavy mineral samples used in the flume were taken from several natural heavy mineral concentrations located in the field study area.

3. Procedure

The procedures involved in all flume experiments were essentially the same. The flume was adjusted to zero slope, the tailgate raised, and water allowed to enter the channel to a depth of four to five inches. Sediment was poured into the well of the test section and agitated to remove any entrained air. The amount of sediment necessary to fill the test section well was approximately 0.2 cu. ft. Next the bed was flattened to a sediment elevation equal to the elevation of the fixed bed.

Flow was slowly increased while the tailgate was correspondingly lowered. Either depth or discharge was varied as the other remained constant. Uniform flow conditions were maintained at all times throughout the test section by a combination of adjustments in slope and tailgate height. All sediment grains in the bed were allowed to readjust and reorient themselves at near-critical flow conditions until movement was no longer observed. The bed was then assumed to be equivalent to that of a natural channel at critical conditions.

As the critical point was attained, the discharge, depth, free surface slope, and water temperature were measured and critical phenomena were observed for 120 seconds to ensure that the critical point was reached. Often the critical conditions had not been attained and a further reorientation of the particles was observed. The time lag allowed for a separation of the unsteady conditions of reorientation from actual critical steady state phenomena.

Side wall effects on computed shear velocity have been studied by Vanoni and Brooks (1957) who developed a sidewall correction procedure for adjusting the hydraulic radius, and Williams (1970) who developed an empirical correction factor. Williams noticed shear stress decreased as flume width widened relative to depth. Over a large range of flow and sediment bed conditions he defined a shear stress correction factor to be

$$\tau_{wide\ channel} = \tau_{flume}/(1.0 + 0.18\ ^{h}/b) \qquad 2\text{-}20$$

where be is the channel width. This correction factor was used in all experimental work to correct flume side wall influences on calculated shear values.

Incipient Motion Experiments

A series of experiments were conducted with uniform sediments on a flat bed in uniform fresh water flow. This was done in order to obtain the value of $\overline{a_2 c_2}$ tanθ defining the moment arms and surface area coefficient as well as undetermined influences on first motion in equation 2-3. The complexity of the term requires an empirical solution.

Experimental procedure involved (1) varying depth and discharge in the flume, and (2) observing incipient motion at various depths for each uniform grain size. The entrainment process is statistical in nature because there is no exact critical point of incipient motion. As observed by Sutherland (1966), the motion of grains is unsteady and non-uniformly distributed over the bed near critical conditions. The motion occured in bursts and incipient motion was recorded when a small number of particles was in motion in an isolated area. The number of grains in motion was small, thus they could be readily counted.

Table 2-1 lists the result of each run for the various light and heavy mineral sizes. A density of 4.5 gm/cc was assumed for each heavy mineral fraction. The density of light minerals is 2.65 gm/cc. Discharge has been converted to mean velocity. The sediment diameter listed is the average of the sieve size on which the sediment was retained and the next largest sieve. In this case the size was based on a 0.5 phi sieve interval for the light minerals and a 0.25 phi interval for the heavy minerals.

Several observations of the bed prior to and at critical conditions are noteworthy. When the natural ellipsoidally-shaped sediments were first placed in the flume they were randomly distributed in the bed. Immediately after flow was introduced, a reorientation of particles was observed which continued as long as the flow increased. When the flow was uniform, all appreciable realignment was accomplished in 60 seconds, after which the bed remained immobile, indicating that grains in a uniform bed will continually readjust to flow conditions. Their orientation, therefore, is a factor of possible importance in defining conditions of incipient motion.

A random commencement of motion in isolated areas was observed for all grain sizes. The surface area covered by the individual bursts increased as the grain size of the bed increased, or perhaps more physically correct, as the flow intensity at incipient motion became stronger. Apparently the bursts are the result of turbulent eddies impinging on the sediment bed. For the smaller sized sediment beds (0.127 to 0.395 mm), there is a laminar or transitional sublayer predicted according to equations 2-8 and 2-12 that was observed to be periodically penetrated by the bursts. This would indicate that incipient motion, even for the smallest grains, takes place under a rough or transitional boundary regime and the assumption of smooth conditions of laminar incipient motion does not strictly hold.

Two methods were utilized in the determination of critical shear stress: 1) the solution of equation 2-5 with f obtained from an experimentally determined nomograph (Figure 2-3), and 2) the logarithmic method of equation 2-15. The results of each are given in Table 2-1 where the subscript cm indicates the first approach and the subscript cv connotes the second method. Results are also plotted on Figure 2-4 as a function of d_s.

Figure 2-3 is a version of the Moody Diagram (Moody, 1944) which was constructed originally for pipes of varying roughness; however, it has been modified for flat bed conditions according to Schlichting (1968). Advantages of using the Moody Diagram method include the ease of obtaining f and the simplicity of using equation

Table 2-1

DATA ON CRITICAL SHEAR STRESS

Exper. No.	Mean Sed. Size (d_s) in mm	Density (ρ_s) in gm/cc	Depth (h) in cm	Mean Vel. ($\bar{U}$) in cm/sec	Free Surf. slope $\times 10^{-3}$	Water Temp. (°C)	(τ_{cm})	(τ_{cv})
4a	3.57	2.65	6.01	48.25	4.4	24.6	13.10	11.10
			4.42	50.25	6.8	24.6	15.45	13.50
			1.86	46.00	15.2	24.6	16.50	16.00
4b	1.79	2.65	6.34	38.50	3.6	24.8	6.56	5.20
			4.45	38.80	3.1	24.8	7.29	5.86
			3.14	32.45	4.6	25.0	5.54	4.56
			2.25	28.80	5.5	25.0	4.81	4.07
4c	.895	2.65	8.80	26.58	2.0	24.6	2.51	1.78
			5.79	28.30	2.0	24.6	3.05	2.55
			4.02	24.00	2.8	24.8	2.56	1.80
			2.25	24.58	3.0	24.8	2.86	2.20
4d	.508	2.65	8.14	22.12	0.7	24.8	1.62	1.14
			5.58	23.90	1.6	24.8	2.00	1.49
			3.68	22.10	1.8	24.8	1.84	1.36
			2.59	22.40	2.0	25.0	2.05	1.51
4e	.359	2.65	7.53	23.04	2.3	24.6	1.70	1.24
			5.67	21.10	1.65	24.6	1.50	1.12
			4.11	21.00	1.65	24.6	1.59	1.19
			2.41	22.12	2.1	25.0	2.05	1.52
			0.94	15.72	2.65	24.6	1.33	1.03
4f	.254	2.65	8.03	20.65	2.8	24.8	1.33	0.99
			5.85	20.93	2.1	24.8	1.43	1.10
			3.29	21.90	2.5	24.8	1.69	1.40
			1.13	20.60	2.5	24.8	1.95	1.62

Table 2-1

DATA ON CRITICAL SHEAR STRESS (continued)

4g	.180	2.65	7.56	19.20	0.00	24.8	1.13	0.62
			3.44	19.68	0.70	25.0	1.35	0.86
			0.85	18.10	2.10	25.0	1.59	0.92
4h	.127	2.65	7.95	18.91	2.20	24.8	1.07	0.88
			5.24	18.10	1.85	24.9	1.07	0.89
			4.27	14.28	1.20	25.0	0.72	0.62
			0.98	13.12	1.75	25.0	0.82	0.72
2a	.180	4.50	5.27	35.28	0.70	25.0	3.65	2.88
			3.68	34.20	2.70	25.0	3.68	2.96
			2.80	36.20	2.80	25.3	4.05	3.32
2b	.127	4.50	7.24	30.70	4.00	24.5	2.66	2.10
			6.24	31.10	3.60	24.8	2.75	2.22
			4.73	30.35	2.90	24.9	2.74	2.25
			3.56	30.31	3.90	24.8	2.89	2.43
2e	.090	4.50	6.92	25.80	3.40	24.5	1.87	1.85
			4.39	27.70	2.60	24.6	2.30	1.92
			3.68	24.20	2.80	24.8	1.89	1.61
			2.25	26.80	4.00	24.8	2.65	2.16

2-5. The logarithmic method on the other hand is solvable only by trial and error for the transitional and smooth cases and, therefore, more complex an equation to use.

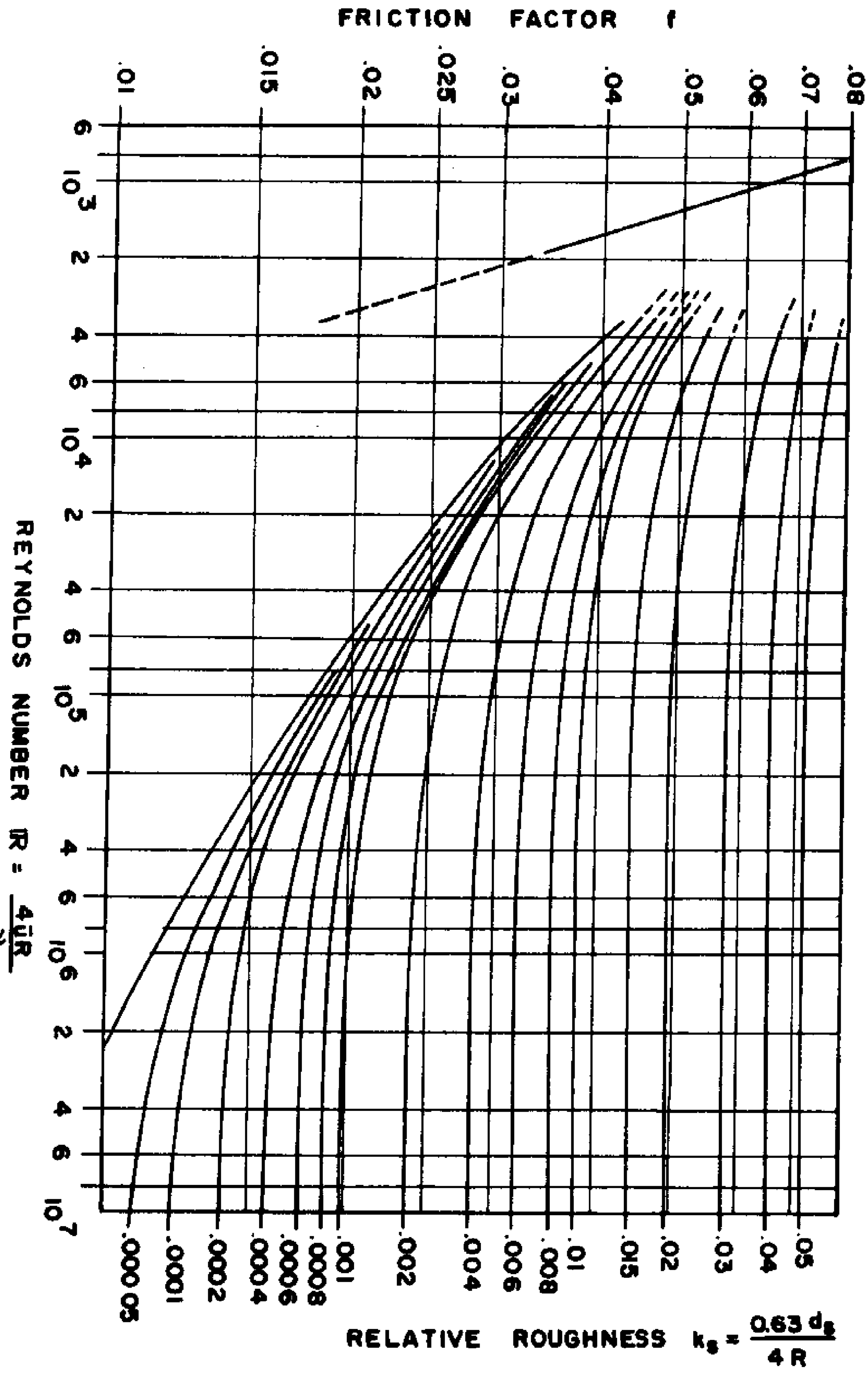
Observation of sediment beds indicated smooth conditions are not found during incipient motion. The reason is that the sublayer appeared always to be penetrated by impinging turbulent eddies. Vanoni (1966) in similar experiments observed fluctuations of 50 percent in the mean velocity of the flow in the vicinity of the bed. Because shear stress is approximated to the velocity squared, shear pulses of more than twice the average intensity are expected. This indicates neither method produces an absolute shear value at the bed during first motion, but it is to be expected that the relative shear values between the various grain sizes and densities are correct. This explanation is based on turbulent measurements taken near a roughened bed under increasing flow intensities which indicated eddy intensity increases as total flow intensity becomes greater (Raichlen, 1967). If this is the case, relative bed shear would remain a function of the flow and bed variables used in either of the two methods.

The curves plotted for each method are alike and τ_{ev} is shown to be a constant function of τ_{em}. The constant of proportionality between the two is 1.25 when k_s is taken to be $0.63d_s$. The reason for this similarity is that the Moody Diagram was based on a logarithmic velocity profile such as defined in equation 2-6. The discrepancy in values arises in the assumption of mean shear based on $\bar{u}$ and 0.4h. Because neither method results in absolute bed shear each is of equal validity in obtaining relative shear, although the method using equation 2-5 is simpler to use. The two methods were discussed to assess the importance of the boundary layer-velocity profile method with respect to the experimentally verified Moody Diagram technique.

Having established a means of obtaining bed shear and the resisting force based on simple flow and sediment parameters, it remains to ascertain the value of $\frac{a_1}{a_2 c_2}$ tanθ using the limited experimental data available. Values of this term plotted against d_s for the two experimental densities used in the flume are given in figure 2-5. A fourfold increase in $\frac{a_1}{a_2 c_2}$ tanθ or C is noted for the light mineral sediments between grain diameters of 0.9 and 0.1mm. The trend of the plotted heavy mineral curve appears to be the same. As the shape factor varied only slightly over this size range, and, indeed, increased slightly for the smaller grains, it is unlikely that the large increase is related to a relative change in moment arms or the effective surface area. In fact, it is unlikely this term will change appreciably over the entire range of uniform sediment sizes. If this hypothesis is correct, another reason must be found for the variation.

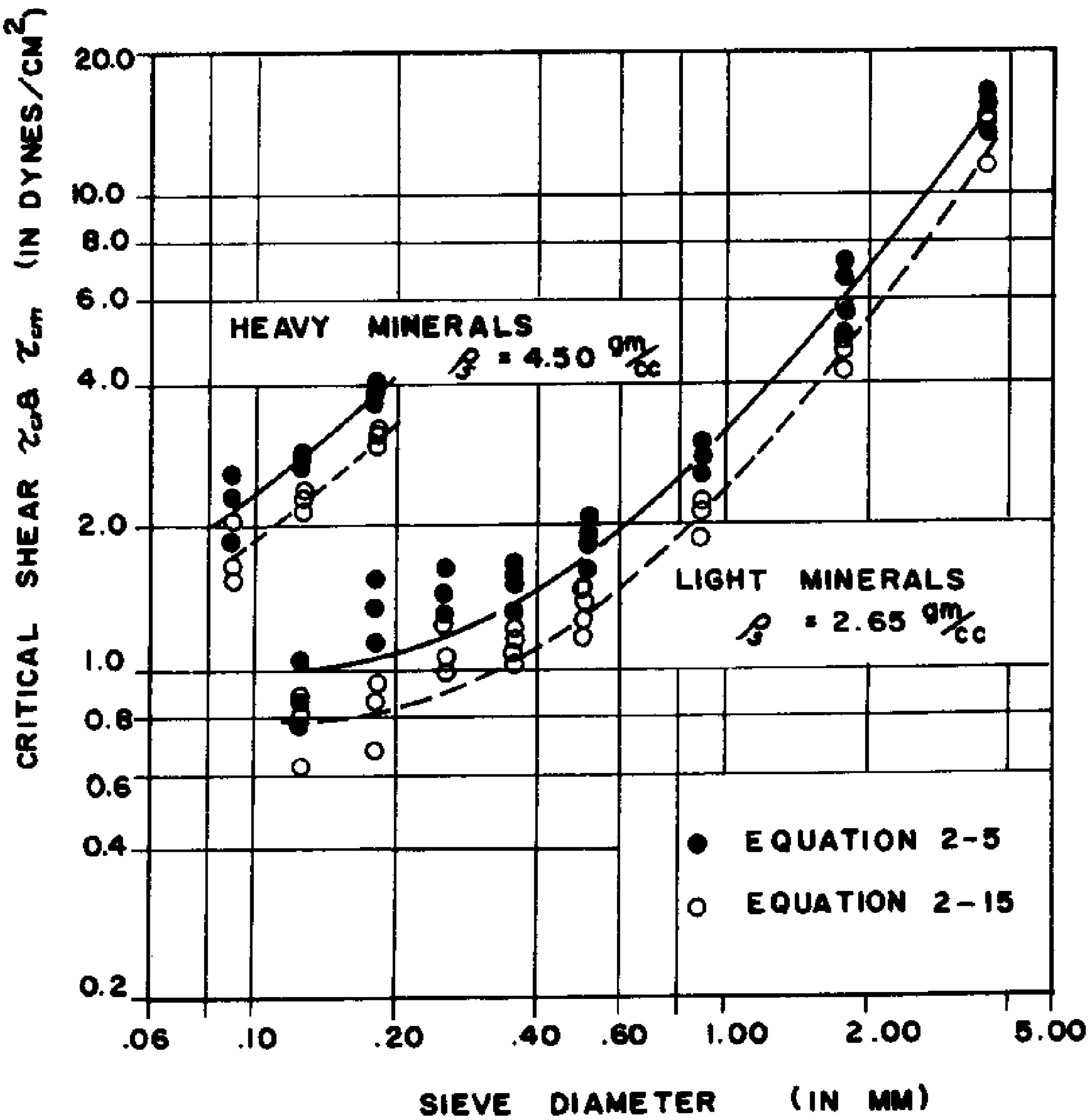
Henceforth, $\frac{a_1}{a_2 c_2}$ tanθ will be referred to as a correction factor C divorced from any moment arm, shape factor, or effective surface area connotation.

Figure 2-3



MOODY DIAGRAM (MODIFIED FROM STREETER, 1962 p.218)

Figure 2-4



CRITICAL BED SHEAR - UNIFORM BED

Noting the similar shapes of the light and heavy mineral curves in Figure 2-5, an axis through the point of zero $^{dC}/_{dd_s}$ and drawn connecting the two curves. Figure 2-6 shows this axis to be one of constant τ_c . This is a significant result as it indicates the undetermined factor C is a function of bed shear as it has been defined at critical conditions. Intermediate density values have been extrapolated for various sizes with the curves drawn on Figure 2-6. Experimental curves are shown as thick solid lines.

The shape of the curve above τ_{cv} values greater than 3.00 dynes/cm^2 could possibly be straightened to assume a vertical attitude. That is, C may be constant for each density above this specific critical shear value because the k_s used in computing τ_{cv} might have been too large as a result of the much wider range of absolute particle sizes in the larger sieve-sized experimental fractions. In any case the relative variation in C above 3.00 dynes/cm^2 is slight compared to that below this value.

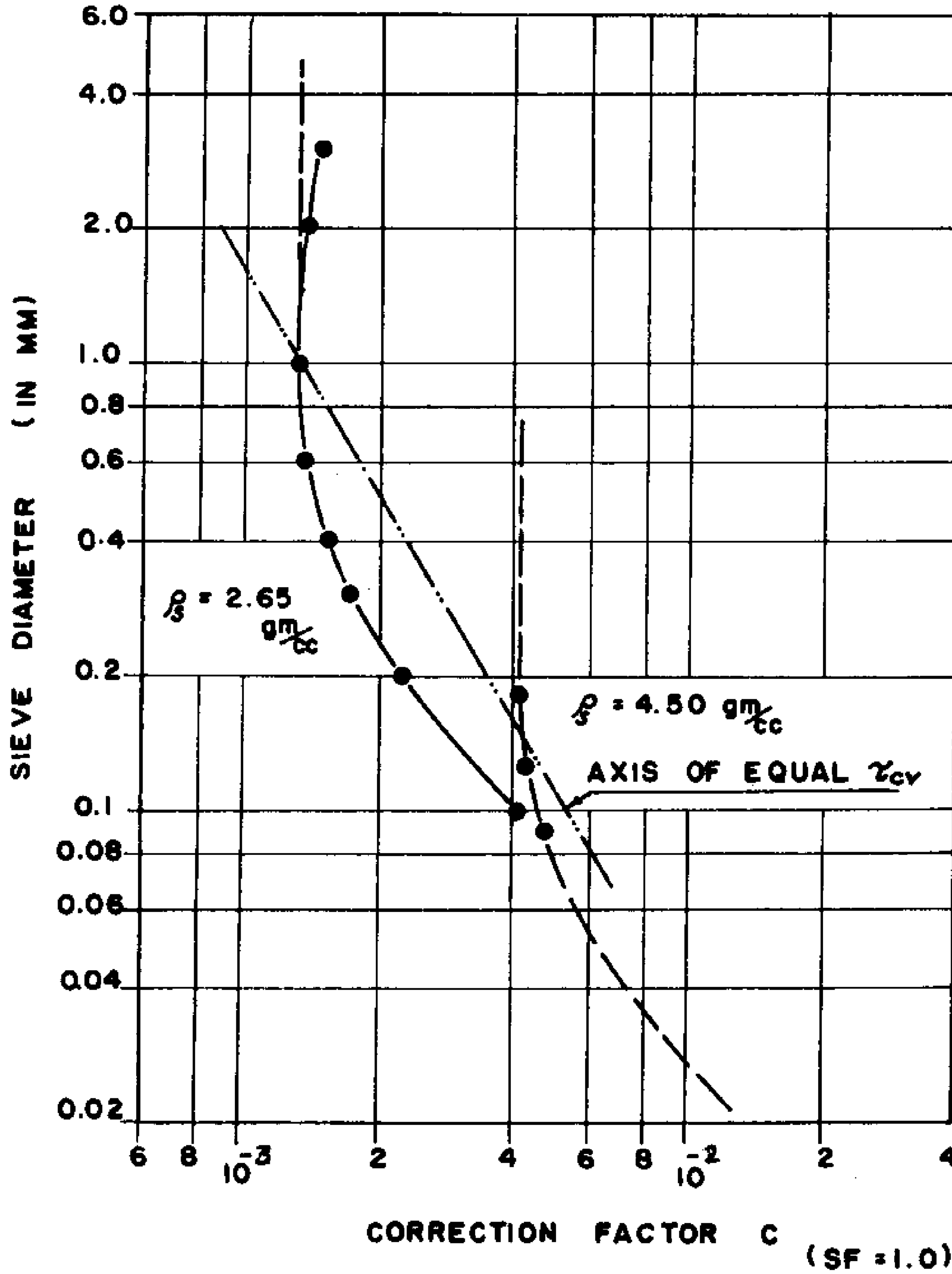
Two factors must be accounted for in dealing with the correction factor. The first is the variation of C with τ_{cr} when ρ_s is constant, and the second is the variation of C with a constant τ_{cr} as ρ_s changes. Addressing ourselves to the first factor and stipulating the C value to be constant for τ_{cr} greater than 3.00 dynes/cm^2, C is observed to increase markedly as critical shear stress falls below 3.00 dynes/cm^2. Noting τ_{cr} to be partially a function of d_s, we may assume the increase to be partly due to a decrease in the grain size of the bed. The maximum variation in the correction factor between particles of different densities occurs at critical shear values above 3.00 dynes/cm^2(Figure 2-6). In this range the C variation is constant and a function of density only. Below a critical shear of 3.00 dynes/cm^2 the variation in C continually decreases until at a d_s value of approximately 0.1mm, density is no longer a factor (Figure 2-5).

For sediments of any density with a critical shear stress in excess of 3.00 dynes/cm^2, the variation in C is explained through deviations in the individual density members of the gravity force term $c_1 d_s(\rho_s - \rho_f)g$. This explanation is available because of the elimination of one degree of freedom through the use of sediments of different density values. Therefore, anomalous values of C with respect to the gravity force term are assumed to occur only for the small values of critical shear stress. In this lower range, as elsewhere, shear is a function of mean flow velocity, depth, and grain size.

The equality of C values for sediments of any density when the grain diameter is less than 0.1mm shows that flow velocity and depth are important factors as well as grain size because the τ_{cr} values remain dissimilar even though d_s is equal.

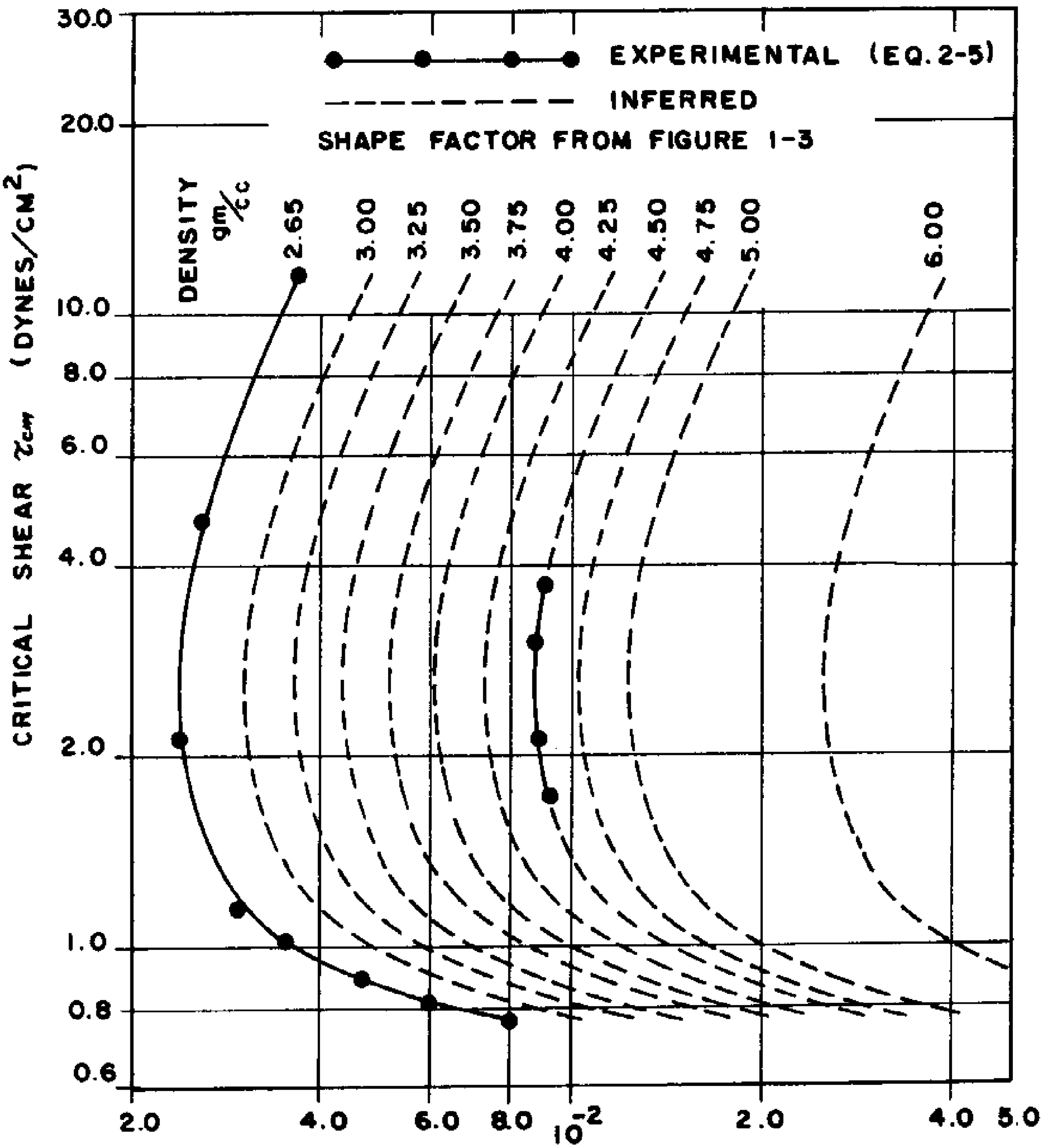
The explanation of this phenomenon is found in our definition of fluid forces. In the beginning of this discussion it was assumed that shear included hydraulic lift as well as drag. Further, it was assumed these fluid forces would remain in constant proportions

Figure 2-5



CORRECTION FACTOR AS A FUNCTION OF GRAIN SIZE

Figure 2-6



CORRECTION FACTOR AS A FUNCTION OF CRITICAL SHEAR (Based on limited data. Intermediate curves subject to correction.)

under any circumstances. This is true for critical bed shear values greater than 3.00 dynes per square centimeter, in which case equation 2-3 is correct and the offset of the curve in Figure 2-6 is solely due to differences in density. This explanation does not satisfy equation 2-3, however, for bed shear values less than that specified above because in this range the effect of lift decreases as the bed particles become smaller and bed velocity lessens. Thus, the drag/lift ratio increases, and to make up the lifting force deficit, τ_{cr} must also increase.

The decrease in lift at critical conditions in the smaller sediment particle sizes is an important factor in explaining the behavior of fine-grained sediment in a flowing fluid. The lift decrease becomes a critical factor in assessing conditions for initial movement when the sediment is a mixture of different grain sizes and densities. Such familiar diagrams as that of Hjulstrom (in Vanoni, 1966), equating critical mean velocity to sediment size, exhibit minimum intensity critical conditions at a d_s of 0.1 to 0.2mm. Sediments larger and smaller require a greater shear to initiate motion. Cohesion between the grains is considered the important factor in defining the critical velocity, and although this is no doubt true for the very fine-sand, silt and clay sizes, it appears unlikely cohesion due to the electrochemical environment plays a dominant active role in bonding sediments greater than 0.1mm. Einstein and Krone (1961)observed an increase in bed structure and a corresponding increase in resistance to movement as more flat contacts are made in clay-sized sediments. This further strengthens the belief that with ellipsoidal particles that are much larger than clay sizes, electrochemical bonding is minimal.

Einstein and El-Samni (1949) measured lift on large spheres (6.8 cm) bonded to a flat bed. The variation in pressure measured vertically across the grains was found to be

$$\Delta p = 0.178 \, \rho_{f/2} \, u_0 \qquad (2\text{-}21)$$

where u_0 is the velocity at a distance of $0.35d_s$ above the bed. This equation should hold true for shear values above 3.00 dynes/cm^2 where lift is not a function of grain diameter. Sutherland (1966) qualitatively describes lift as a major moving force in the initiation of motion of a bed particle. Chepil (1962) found the drag/lift ratio for grains in air to be 0.4 or lift to have 2.5 times the effect on grain motion as drag. Each investigator indicates lift to be a dominant factor in sediment movement and any factor that affects lift would be important in the overall understanding of sedimentation.

Lift is due to the velocity gradient in the flow near the bed, the stagnation pressure under the grain, and the upward velocity components resulting from the reflection of turbulent bursts at the bed surface. As bed velocity decreases each of these factors decrease. However, as sediment size decreases, the effect of lift lessens significantly because the space between the grains decreases until the rapid fluctuation in pressure resulting from turbulent eddies is no longer transmitted to both sides of the grain. The ability of the bursts to penetrate the grain interstices is obviously a function of the particle size, but it is also dependent upon near-bed velocity because at a grain size of 0.1 mm the density of the particle no longer affects the correction factor, although critical shear values are different. In this case, the only inconsistent variables in equation 2-3 are depth and mean velocity, balancing the effect of density.

Lift, or the lack of it, is an important factor in initiating motion in small sediments. The heretofore constant drag/lift ratio increases as does the apparent resistance of the particle to motion. In effect, this increase reflects the loss of lift as a viable moving force. In short, the decrease in lift is a consequence of decreasing pore space in sediments of the upper bed layer and lessening bed velocity.

Conditions of incipient motion IM for a uniform flat bed are defined according to

$$IM = IM(d_s, \rho_s, \bar{u}, h) \qquad (2\text{-}22)$$

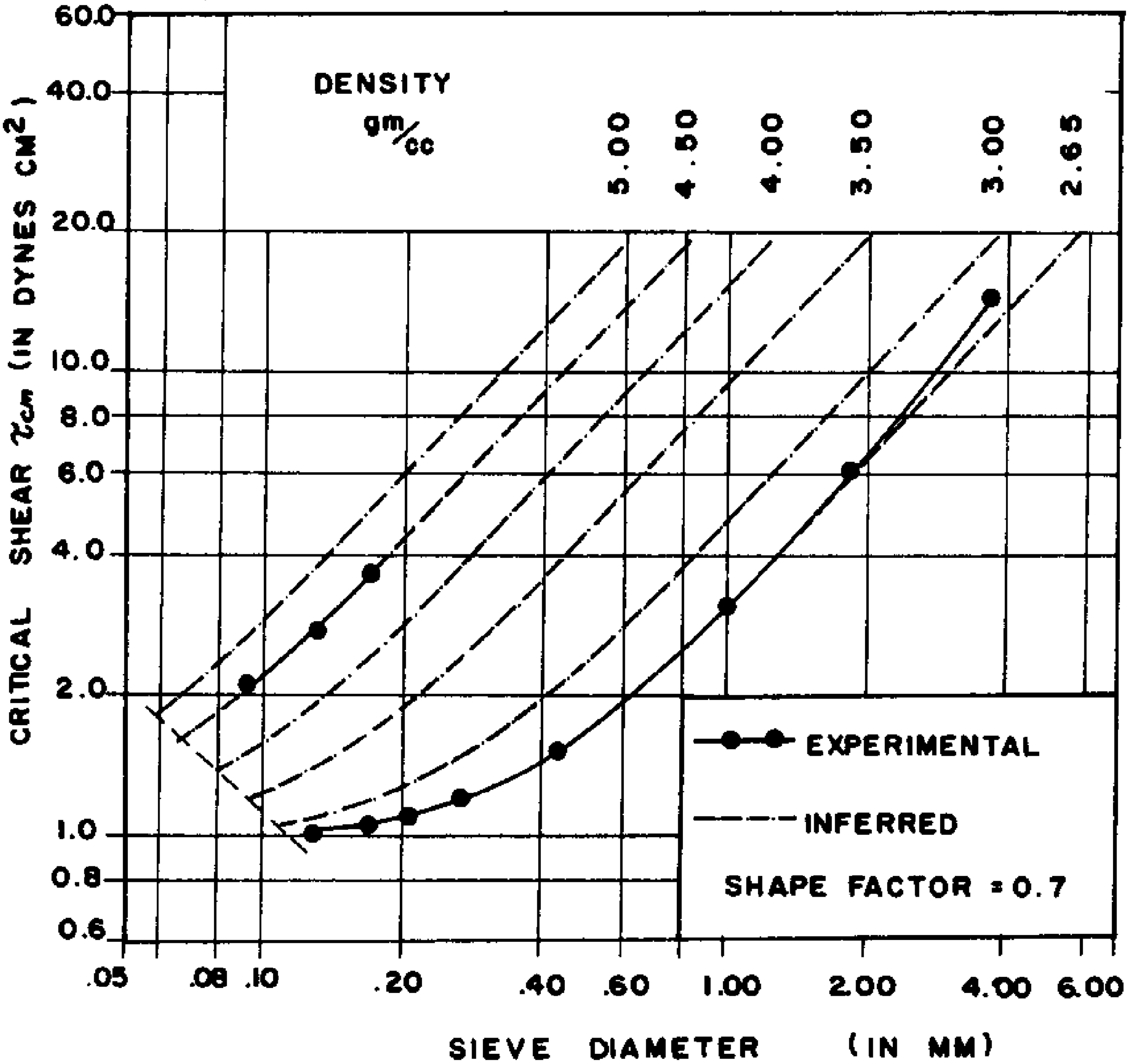
using the modified Moody Diagram (Figure 2-3). Figure 2-6 for the drag/lift correction factor, and equation 2-3, any combination of mean velocity, depth, grain size, or grain density can be applied to conditions of incipient motion for a uniform flat bed. Critical shear stress plotted as a function of density and particle size is given in Figure 2-7. The dashed line indicates the probable cutoff point between cohesionless and cohesive sediments. The principles and relationships established using uniform sediments will provide a basis for defining conditions on a nonuniform bed, the mixed situation common to natural sediments.

SEDIMENT TRANSPORT

As critical conditions of particle stability are exceeded, individual grains commence to move as bed load. With increasing flow intensity, rolling or sliding bed grains may be projected from the boundary into the flow itself in a trajectory that rapidly returns them to the bed. As a grain is restored to the boundary, it may remain stationary or begin motion as bed load until conditions of eddy impingement or particle impact again launch it in this saltating mode of transport. A further increase in the strength of flow may hinder the saltating particle from returning to the bed so that its suspended motion exists independent of contact with the boundary. Whatever the manner of transport, an understanding of transport mechanics is fundamental to understanding heavy mineral accumulations in nature.

The importance of transport phenomena centers on the relative behavior of heavy and light mineral grains during deposition. A consideration of where each grain moves with respect to its size and density in the sediment/fluid column and why, may lead to predictions for their preferred separation during transport. Also,

Figure 2-7



CRITICAL SHEAR AS A FUNCTION OF SIEVE DIAMETER the sediments available for deposition at a site depend on the opportunity each has of arriving at that location.

Although transport is a difficult problem to approach analytically, the rewards of at least a qualitative attempt warrant the effort. This section deals with heavy and light mineral movement over boundaries of uniform roughness. The purpose of the discussion is to define conditions amenable to light mineral transport, while heavy minerals are trapped in the bed.

Modes of Transport

The significance of the various modes by which a sediment grain travels in a fluid exists in the particle's interaction with the bed over which it is moving. Obviously, a grain in suspension cannot be trapped at the sediment interface, but under specific circumstances a saltating or creeping grain may come to rest.

Bed Creep

Surface creep in the sediment/fluid interface begins as soon as critical conditions are exceeded. Importantly, creep is also the mode of transport most likely to precede deposition because in this transport state the moving particle is in constant contact with the boundary.

The point of contact of a particle with the bed is its pivot point (Figure 2-1), and any displacement from stable equilibrium results in a couple. When the particle rolls, the magnitude of the fluid force moment arm is wholly dependent on the bed elements over which it rolls. Should this moment arm vary in length due to a different bed roughness or change in particle orientation, a change in transport strength may be necessary to continue its mode of transport; conversely, entrapment in the bed could ensue. Such a mechanism of movement emphasizes the importance of bed roughness in trapping or passing moving bed particles.

An ellipse will pivot about its small diameter and roll with its major axis normal to the line of motion. This has been observed experimentally in the flume by Hunzicher (1930) and under natural conditions by Lane and Carlson (1954). Flow orientation of a grain moving on the bed relegates the description of orientation to secondary importance in comparison to bed roughness with respect to rolling particles. Sediment grains rarely slide, and when they do only in the case of flat sided particles moving on a smooth bed.

Flume Experiments

A series of experiments was conducted for the purpose of defining those bed and flow conditions over which a given range of particle sizes and densities will pass the bed without disturbing it. The reason for performing this particular sequence of tests was to simulate observed conditions of heavy mineral concentrations on beaches while the associated light mineral fraction bypassed the depositional boundary.

(1) Procedure

The flume procedure employed was similar to that described previously. Tests began with a flattened uniform bed of known critical shear as defined in the section on incipient motion. Particles of a diameter differing from that of the bed were placed on the boundary at near-critical uniform flow conditions τ_{cm}. The flow intensity was slowly decreased, either by increasing the depth or decreasing the mean velocity, until the transported particles ceased moving and became elements of the boundary. Particles were placed on the boundary in concentrations of less than two percent to ensure minimum interference between moving grains. The small concentration also guaranteed that bed shear τ_{bm} , based on the relative roughness k_s of the uniform bed, would be unaffected by a concentrated moving layer.

(2) Results of Flume Tests - Bypassing Sediments

The results of the tests are shown on Table 2-2 and in Figure 2-8. The vertical axis of Figure 2-8 is uniform bed diameter, ranging from 0.127 to 3.57 mm for beds with a density of 2.65 gm/cc, and 0.127 to 0.180 mm for a heavy mineral bed of 4.50 gm/cc. Transported grain diameter is given on the horizontal axis. In all cases the bypassing grains were light minerals. Heavy particles only moved when the boundary itself moved. The range of moving grain sizes capable of bypassing a given light mineral bed is illustrated by the solid block on the figure. Dotted lines (Figure 2-8) bound the range of light mineral sizes capable of bypassing a heavy mineral boundary. Whether a particle will move on a given bed is dependent on the relative magnitude of the fluid force on the moving grain with respect to the critical resisting force controlling bed stability. Once τ_{cm} for the bed is reached, the entire uniform boundary moves, and selective deposition and bypassing is no longer possible.

In all cases the bed shear required to move light mineral grains over a light mineral bed was 80 percent or more of the critical bed shear. Figure 2-8 illustrates this differential in the blocked out areas as the percent given by $\left(\frac{\tau_{cm} - \tau_{bm}}{\tau_{cm}}\right) \times 100$. The values given on the upper scale are the mean for each test sequence. The lower scale refers to light minerals bypassing a heavy mineral bed. Boundaries for this bed shear-bypassing shear difference are the dotted lines near the bottom of the figure. In the case of a heavy mineral boundary, the percent difference is as great as 78 percent.

Table 2-2

DATA ON SEDIMENT-BYPASSING SHEAR STRESS

Exper. No.	Mean Sed. Size-Bed Gr. (d_s) in mm	Mean Sed. Size-Moving Gr. (d_s) in mm	Density-Bed (ρ_s) in gm/cc	Depth (h) in cm	Mean Vel. ($\bar{u}$) in cm/sec	Free Surf. Slope $\times 10^{-3}$	Water Temp. (^{o}C)	(τ_{bm})
3a	3.57		2.65	(No bypassing of smaller sediments observed on this bed)				
3b	1.79	3.57	2.65	6.34	38.50	3.60	24.8	6.56
				4.52	38.25	2.80	24.8	7.09
				3.17	32.50	4.60	24.8	5.54
				2.29	28.62	5.20	25.0	4.76
		.895		6.56	37.15	3.00	24.8	6.12
				4.67	37.00	2.50	24.8	6.64
				3.26	31.20	4.60	25.0	5.11
				2.26	28.70	5.40	25.0	4.79
3c	.895	1.79	2.65	8.80	26.58	2.00	24.6	2.51
				5.79	28.30	2.00	24.6	3.05
				4.02	24.00	2.80	24.8	2.56
				2.25	24.58	3.00	24.8	2.86
		.508		9.02	25.95	1.70	24.6	2.40
				5.98	27.35	1.50	24.6	2.85
				4.08	23.60	2.70	24.8	2.47
				2.32	23.95	2.60	24.8	2.72
3d	.508	1.79	2.65	8.44	21.32	0.20	24.8	1.50
				5.63	32.65	1.15	24.8	1.95
				3.65	21.88	1.65	25.0	1.80
				2.65	21.90	1.60	25.0	1.97

Table 2-2

DATA ON SEDIMENT-BYPASSING SHEAR STRESS (continued)

		.895		8.68	20.68	0.10	24.8	1.41
				5.92	22.58	0.85	24.8	1.78
				3.78	21.20	1.10	25.0	1.70
				2.86	20.30	1.10	25.0	1.69
		.359		8.28	21.70	0.30	24.8	1.55
				5.61	23.82	1.20	24.8	1.98
				3.68	21.70	1.40	25.0	1.78
				2.68	21.68	1.55	25.0	1.93
3e	.359	1.79	2.65	7.90	21.92	1.60	24.6	1.54
				5.67	21.08	1.60	24.6	1.50
				4.12	21.02	1.60	24.6	1.58
				2.41	22.12	2.10	25.0	2.05
				0.94	15.72	2.65	24.6	1.33
		.895		7.80	22.20	1.80	24.6	1.57
				5.79	20.65	1.05	24.6	1.43
				4.27	20.30	1.30	24.6	1.49
				2.44	21.80	2.05	25.0	1.99
				0.94	15.72	2.65	24.6	1.33
		.508		7.59	22.80	2.10	24.6	1.65
				5.97	20.00	0.90	24.6	1.35
				4.30	20.18	1.40	24.6	1.47
				2.48	21.52	1.60	25.0	1.94
				0.94	15.72	2.65	24.6	1.33
		.254		7.56	22.90	2.20	24.6	1.67
				5.76	20.80	1.35	24.6	1.46
				4.15	20.86	1.60	24.6	1.48
				2.44	21.80	1.90	25.0	2.00

Table 2-2

DATA ON SEDIMENT-BYPASSING SHEAR STRESS (continued)

3f	.254	1.79	2.65	8.03	20.65	2.80	24.8	0.99
				5.85	20.93	2.10	24.8	1.10
				3.29	21.90	2.50	24.8	1.40
				1.13	20.60	2.50	24.8	1.62
		.895		8.29	20.17	2.20	24.8	1.27
				6.00	20.65	1.60	24.8	1.40
				3.42	21.10	2.05	24.8	1.58
				1.25	16.58	2.00	24.8	1.27
		.508		8.26	20.20	2.20	24.8	1.28
				6.03	20.60	1.60	24.8	1.39
				3.38	21.30	2.10	24.8	1.61
				1.22	17.00	2.05	24.8	1.34
		.359		8.17	20.40	2.50	24.8	1.30
				5.91	20.83	1.90	24.8	1.42
				3.32	21.72	2.35	24.8	1.67
				1.22	17.00	2.15	24.8	1.34
		.180		8.17	20.40	2.50	24.8	1.30
				5.88	20.89	2.05	24.8	1.37
				3.32	21.72	2.40	24.8	1.67
				1.13	18.37	2.40	24.8	1.56
3g	.180	.895	2.65	7.56	19.20	0.00	24.8	1.13
				3.44	19.68	0.70	25.0	1.35
				0.85	18.10	2.10	25.0	1.59
		.508		7.62	19.04	0.20	24.8	1.11
				3.57	19.08	0.70	25.0	1.27
				0.95	18.10	2.10	25.0	1.59

Table 2-2

DATA ON SEDIMENT-BYPASSING SHEAR STRESS (continued)

		.359		7.74	18.72	0.15	24.8	1.08
				3.54	19.20	0.40	25.0	1.29
				0.95	16.36	1.80	25.0	1.35
		.254		7.71	18.80	0.15	24.8	1.09
				3.47	19.54	0.60	24.8	1.34
				0.92	16.92	2.00	25.0	1.39
		.127		7.62	19.04	0.20	24.8	1.11
				3.45	19.70	0.70	25.0	1.36
				0.86	18.10	2.10	25.0	1.59
3h	.127	.895	2.65	7.95	18.91	2.20	24.8	1.07
				5.24	18.10	1.85	24.9	1.07
				4.27	14.28	1.20	25.0	0.72
				0.98	13.12	1.75	25.0	0.82
		.508		7.98	18.76	1.95	24.8	1.06
				5.27	18.00	2.00	24.9	1.05
				4.27	14.26	1.15	25.0	0.71
				0.98	12.74	1.50	25.0	0.77
		.359		8.05	18.64	1.95	24.8	1.04
				5.27	18.00	2.15	24.9	1.05
				4.33	13.96	0.90	25.0	0.68
				1.07	12.00	1.25	25.0	0.68
		.254		8.16	18.36	1.90	24.8	1.01
				5.39	17.63	2.05	24.9	1.00
				4.33	13.98	0.90	25.0	0.68
				1.03	12.34	1.40	25.0	0.72

Table 2-2

DATA ON SEDIMENT-BYPASSING SHEAR STRESS (continued)

		.180		8.10	18.48	2.00	24.8	1.02
				5.33	17.78	1.85	24.9	1.03
				4.26	14.04	1.00	25.0	0.69
				1.01	12.74	1.55	25.0	0.77
1a	.127	1.79	4.50	5.55	29.40	3.20	24.6	2.92
				4.66	22.00	3.15	24.8	1.70
				2.77	20.30	3.10	24.8	1.56
		.895		8.13	20.05	4.20	24.6	1.24
				5.67	18.10	1.60	24.8	1.10
				4.18	17.97	2.05	24.8	1.17
		.508		8.26	19.75	3.80	24.6	1.18
				6.09	16.85	1.45	24.8	0.94
				4.63	15.88	2.45	24.8	0.88
		.359		8.80	18.53	3.20	24.6	1.02
				6.25	16.45	1.45	24.8	0.88
				4.73	15.54	2.30	24.8	0.84
		.254		8.92	18.02	2.90	24.6	0.97
				6.30	16.27	1.40	24.8	0.86
				4.88	15.08	2.25	24.8	0.79
		.180		8.80	18.53	3.00	24.6	1.02
				6.43	15.90	1.20	24.8	0.82
				4.81	15.24	2.05	24.8	0.82
		.127		8.80	18.53	3.00	24.6	1.02
				6.40	16.06	1.30	24.8	0.84
				4.94	14.90	2.30	24.8	0.78

Table 2-2

DATA ON SEDIMENT-BYPASSING SHEAR STRESS (continued)

1b	.180	1.79	4.50	5.85	19.41	2.60	24.5	1.32
				3.75	23.52	3.10	25.0	2.08
				2.50	23.10	4.20	25.0	2.07
		.895		6.49	17.49	1.55	24.5	1.00
				4.84	17.04	1.85	25.0	1.02
				3.53	16.34	2.55	25.0	0.97
		.508		6.68	17.00	1.10	24.5	0.93
				4.96	16.62	1.60	25.0	0.96
				3.62	15.94	2.40	25.0	0.92
		.359		6.68	17.00	1.15	24.5	0.93
				5.15	16.00	1.50	25.0	0.88
				3.57	16.20	2.80	25.0	0.96
		.254		6.37	17.82	1.75	24.5	1.04
				5.27	15.70	1.40	25.0	0.83
				3.47	16.60	2.60	25.0	1.01
		.180		6.37	17.82	1.75	24.5	1.03
				4.63	17.79	0.90	25.0	1.10
				3.05	18.95	2.85	25.0	1.35
		.127		7.16	15.79	0.60	24.5	0.88
				5.15	16.82	1.50	25.0	0.98
				3.05	18.95	2.85	25.0	1.35

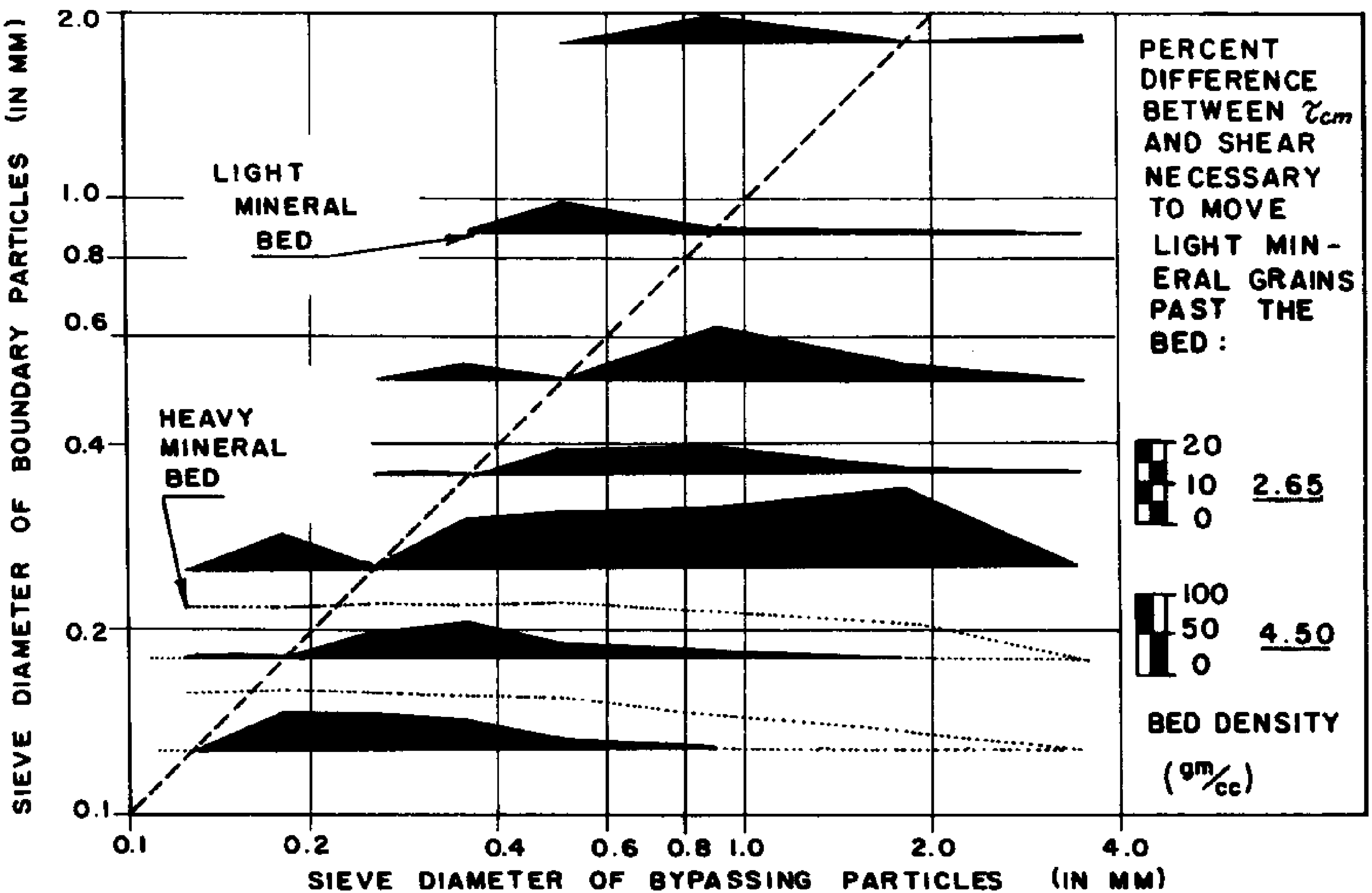
The variance in percent of critical shear required to move a sediment with a light versus a heavy mineral bed shows that the range widens as the bed becomes surfaced with heavy minerals. So as the percent coverage of heavy minerals on the surface increases, its ability to pass light minerals also increases, and in this manner the heavy mineral surface propagates itself. This is very likely part of the mechanism of heavy mineral lamina formation where the lamina is formed under nearly uniform flow conditions.

For a large-grained bed, the range and shear difference between bypassing grains and the bed grains is largest for bypassing particles smaller than the bed. The opposite is true for smoother boundaries where the range of bypassing sizes extends to ten times the boundary grain diameter. Figure 2-7 is useful in explaining this phenomena. For larger bed sizes, especially those greater than 1.0 mm, the difference in critical bed shear between adjacent grain diameters used in the tests is appreciable. For the smaller sized boundaries between 0.1 and 0.6 mm, bed shear varies very little between the different sizes. Large particles rolling over beds with a grain diameter in excess of 1.0 mm have a strong tendency to become a part of the bed because the shear required to move them out of coarse bed interstices is great. Particles transported over beds composed of grains smaller than 1.0 mm are moved by significantly less bed shear than that required to initiate motion in the bed. This keeps them in motion because there is considerable lift on coarse beds.

Fine-grained boundaries inhibit lift forces and therefore lessen the chances for particles smaller than the bed to bypass it. No matter what the bed size is, grains smaller than bed interstices tend to become entrapped. On the smaller grained beds there is little lift to inhibit this entrapment, while the larger boundaries not only have a higher critical shear for the bed with respect to the entrapped particle, but also exert a maximum lift on any particle becoming trapped in the bed spaces.

Fine-grained beds pass a large range of sediment sizes because the tendency of a grain to be deposited is correlative to its shear at incipient motion. The small bed sizes all have much the same critical bed shear; therefore, a larger moving particle will not come to rest as easily as on a coarse bed because its resisting force, based on contact points with the boundary, is relatively smaller on a smoother surface. It was observed that the largest surface bed particles in the size range less than 1.0 mm moved first during non-uniform boundary tests in the flume. This was because the larger sizes always project a greater surface area to the flow. On smooth fine-grained boundaries, where lift, which always aids small-particle bypassing, is minimal, the critical bed shear is nearly constant so that surface drag on the moving particle surface becomes the most important moving factor. The increase in drag forces for grains larger than the bed is at a greater rate than that for resisting forces when the particles are moving across a fine boundary.

Figure 2-8



RANGE OF SEDIMENTS BYPASSING A UNIFORM BED

As the bed becomes smoother the difference between critical bed shear and the shear needed to move a bypassing particle increases. It reaches a maximum at the light mineral bed size of 0.254 mm.

Table 2-2 indicates that the τ_{bm} required to move a bypassing particle decreases as h decreases for particles passing a bed composed of smaller grains. An increase in the bypassing shear values is noted as h decreases when the particles are passing a bed of larger grains. This may be explained in relation to the projected surface area of the bypassing grain. When the bed is constructed of grains smaller than the bypassing particles, shear on the moving grain increases as depth decreases because the velocity profile is compressed. This suggests that a smaller bed shear is capable of moving the same grain. When the bypassing grain is smaller than the bed grains, shear must increase as depth decreases because the roughness ratio k_s/R increases at a greater rate for the boundary than for the moving grain.

Heavy minerals were trapped in all light mineral test-beds and in no instance were they bypassed. Likewise, they did not bypass a heavy mineral boundary, and all sizes became trapped regardless of the bed size. The reason heavy minerals will not bypass each other is that their critical shear values are significantly different even where their diameters are nearly the same (Figure 2-7). This indicates that the sorting of a heavy mineral concentration caused by the light mineral bypassing mechanism will be poorer under ideal conditions than that of a light mineral bed. It also suggests that most heavy minerals will be deposited and most light minerals will pass as long as the range of deposited heavy minerals does not significantly alter the bed roughness. For the conditions tested, bypassing is a one-way mechanism with light minerals being the only material capable of being bypassed.

Although the shear range over which the bypassing mechanism is effective is narrow, the range of particle sizes bypassed may be very large. Under ideal conditions a moving sediment with a sorting coeffecient of two or greater could selectively deposit a fraction of its size range. The sorting of the deposited fraction would be much less than two. Sorting of less than 0.5, as illustrated in Figure 1-9, is probably the optimum expected in an ideal natural case where sorting is probably more a function of flow and bed conditions than sediment availability. The shear range over which selective deposition takes place is narrow. In nature, if this mechanism is to be effective in concentrating appreciable amounts of heavy minerals, uniform flow conditions in this range must prevail over a considerable surface area and for a reasonable period of time to establish a third dimension to any deposit.

Saltation

Saltation is a transitional mode of transport between creep and suspension. Jumps from the bed experienced by a particle vary from a few grain diameters in length to a few hundreds of diameters. The maximum saltation height has been variously given as $13d_s$, where d_s is sieve diameter as defined by Nordin (1964), 8 to 10 d_s by Kalinske (1941) and a few tens of d_s by Danel (1953). Thus, as would be expected, a grain's vertical travel is much less than its horizontal journey. Visher (1969) estimated saltation to be a dominant process of movement in most fluid environments. His conclusions of transport processes were based on truncation points taken from log-probability curves for a large number of light mineral samples. Although only an approximate range of sizes for each transport mode is given, and nothing was observed concerning causitive forces, the values of conditions at deposition are of interest in interpreting heavy mineral accumulations.

Saltating heavy mineral particles would be expected to be trapped in the bed for several reasons: 1) a saltating heavy grain by its momentum will preferentially settle in the bed on impact and eject a lighter grain and 2) as the bed becomes covered with heavy minerals, a light particle impacting upon it would tend to recochet.

The question of bed stability is important if a heavy mineral deposit is to form. For accretion to progress, the net movement of transported material must decrease throughout the depositional locality. Therefore, shear τ_{bm} must remain less than the τ_{cm} required to move the bed. Most light mineral particles larger than the heavy mineral bed particles were observed to move as creep in the flume, and most smaller grains moved in suspension or by saltation. The heavy minerals moved as mildly saltating grains or as creep. The implication of this is that to concentrate heavy minerals, the smaller light fractions must by absent from the moving sediment or not be in contact with the bed. In the latter instance saltating particles would come in contact with the bed much less frequently than creeping grains and particles in suspension will never by deposited.

A combination of bed creep and mildly saltating grains was observed on a natural beach at Dare County, North Carolina as a dense traction carpet. Within this blanket, small heavy particles were shielded as they impacted and dropped to the bed. On the bed they moved by creeping, or thus were deposited.

Suspension

Turbulent sediment suspension may be visualized as an advanced stage of bed load creep and saltation. Moss (1963) defined the maximum grain size of light minerals capable of being suspended under natural conditions immediately before deposition as 0.07 mm. The maximum value predicted by Visher (1969) is nearer 0.10 mm. If these observations are correct, it is likely few heavy minerals in the Dare County field study area are moved in suspension just prior to deposition. All have a grain size in excess of 0.06 mm (Figure 1-6). Several samples of suspended sediment were taken from the Oregon Inlet tidal channel and each contained one to two percent heavy minerals. The light minerals exhibited an average grain diameter of approximately 0.25 mm, or more than 2.5 times the diameter of the heavy minerals. The entire bed was moving during the sampling,and bed shear exceeded 30 dynes/cm^2. The implication is that although heavy minerals surely move in suspension, they are not being carried in this manner immediately prior to deposition. In the case of light minerals deposition may directly follow suspension.

Suspension, per se, has only a secondary effect on heavy minerals but it may remove the opportunity that fine light minerals have of being deposited in the bed. In the Mississippi River appreciable bed load was observed only during high flow conditions (Peyronnin, 1961). The reason for this bypassing is probably twofold: 1) the suspended sediment moving in the river fills the interstices of the coarser bed material thus raising the critical bed shear necessary to move bed sediments and 2) as the amount of suspended load increases, channel resistance decreases because the energy provided by turbulence is damped and reduced in intensity as an effect of the added sediment (Vanoni and Nomicos, 1956). Either bypassing mechanism affects heavy minerals in that it is a means of moving large quantities of fine material in suspension without disturbing the bed.

SEDIMENT DEPOSITION

The importance of a marine environment as a possible placer accumulation site lies in its ability to preferentially entrap and preserve heavy minerals while bypassing light minerals. Aspects of preservation have been previously described with respect to incipient motion (Figure 2-7). It remains to define the flow and and bed conditions resulting in light and heavy mineral deposition. Maximum depositional bed shear has been defined for light minerals moving across uniform boundaries of size and density characteristics differing from the transported particles (Figure 2-8). The following section deals with the depositional traits of single-sized sediments becoming entrapped in boundaries of similar composition.

Transport Conditions - Prior to Deposition

Bed creep was observed to be the mode of transport prior to deposition for all light and heavy mineral sizes in the available sediment range of samples obtained from Dare County. For light minerals the range is 0.15 to 3.57 mm while the range is 0.09 to 0.18 mm for heavy minerals. This assumes an aggrading boundary and relatively uniform flow conditions as well as a flat bed composed of the same sizes as the moving grains. On a rippled bed, sediment particles move as creep, saltation, or in suspension; however, immediately prior to deposition on an aggrading boundary, the grains are trapped in ripple trough interstices as they move as bed creep.

The reason bed creep is the most important transport mode preceding the deposition process is that in this mode the moving particle has the greatest opportunity for contact with the bed. Rolling was observed to occur more often than sliding because the sand-sized particles, both light and heavy minerals, were ellipsoidal in shape. Shell fragments, however, moved by sliding. Rolling was accomplished by shear on the grain due to the velocity gradient near the bed which produced a moment around the pivot point. This force couple occurred because the center of gravity is below the point of force application. On a perfectly smooth bed a particle would be expected to roll under the influence of the smallest velocity. As the bed roughness increases, the resistance to rolling movement across the bed also increases. This is well documented in Table 2-2.

Saltation prior to deposition is also a significant transport mode; however, saltation in a uniform flow requires a greater bed shear to maintain it than surface creep. As the bed shear decreases to that value critical for deposition, saltation often is replaced by bed creep as the dominant transport mode.

Saltation requires a reflecting surface. When this surface was smaller than the moving particles, saltating grains were observed to initiate surface motion. In this case, the boundary was erosional and is of little interest in placer formation. When the surface is composed of particles much larger than the saltating grains, saltation ceased because the moving grains were lost in surface interstices.

Deposition directly from saltating grains is a function of the ratio of impacting material which ricochets, launches another or many particles, or is retained in the bed. A net gain to the bed results in accretion. In no instance in either flume experiments or field observations were sand-sized particles directly deposited in the bed from saltation.

Suspended particles require an even greater shear to maintain them in suspension and are only deposited in conditions of radically changing flow regime or bed configuration.

It is necessary to consider sediment availability as well as hydraulic availability in defining depositional processes. Availability of a given grain size or density at any location within a region is governed by its distance from the system input point and the time-dependent flow and bed variables acting on the particular grain in the region. The summation of all processes acting on the sediments will dictate whether they will pass or be deposited before reaching the possible area of heavy mineral accumulation.

Experiments - Deposition

A series of flume tests were run to determine the difference between the critical bed shear required to initiate motion in a bed particle and the shear existing at the time of deposition from a creeping bed layer. Incipient shear is given in Figure 2-7. The following tests deal with deposition only. All analytical techniques are the same as defined at the beginning of this chapter.

Procedure

The procedure for the depositional tests is much the same as described previously. The bed was composed of the same sized particles as were transported across the bed. Critical shear was measured for the bed and it was established that the boundary was stable at that shear. Grains of the same size as the boundary were then introduced by hand to the flow system above the test section. They were allowed to move past the stable bed as either depth or velocity was slowly decreased until the moving particles ceased moving. All movement was bed creep. Uniform flow conditions were maintained throughout the tests.

Results

The results of the tests are shown in Table 2-3. Depositional bed shear is given by τ_{dm}. Figure 2-9 illustrates the percent decrease in bed shear τ_{bm} between the critical shear at incipient motion and shear at the time of deposition. This figure refers to a light mineral bed being passed by light mineral grains.

Bed shear differences increase from essentially zero for a bed size of 3.57 mm to approximately 10 percent for a boundary of 0.5 mm or smaller. The mean value, 10.1 percent for sizes between 0.127 and 0.508 mm, appears to be relatively constant for this size range.

Table 2-3

DATA ON DEPOSITIONAL SHEAR STRESS

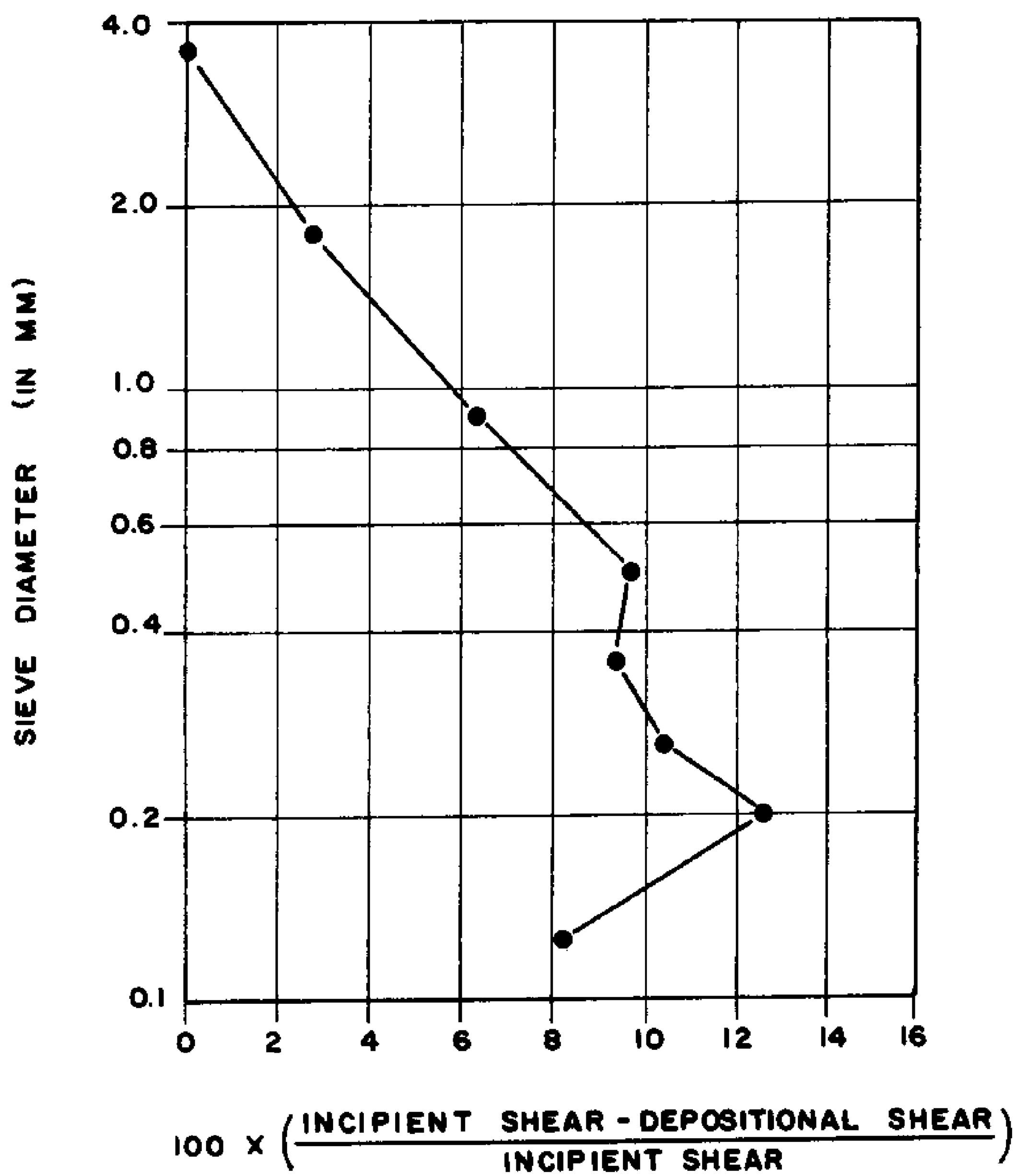
Exper. No.	Mean Sed. Size (d_s) in mm	Density (ρ_s) in gm/cc	Depth (h) in cm	Mean Vel. ($\bar{u}$) in cm/sec	Free Surf. Slope $\times 10^{-3}$	Water Temp. (oC)	(τ_{dm})
5a	3.57	2.65	6.01	48.25	4.40	24.6	13.10
			4.42	50.25	6.80	24.6	15.45
			1.86	46.00	15.20	24.6	16.50
5b	1.79	2.65	6.38	38.35	3.55	24.8	6.50
			4.49	38.40	3.00	24.8	7.16
			3.22	31.70	4.60	25.0	5.28
			2.31	28.10	5.45	25.0	4.59
5c	.895	2.65	9.14	25.60	2.05	24.6	2.34
			6.02	27.22	2.20	24.6	2.82
			4.11	23.49	2.95	24.8	2.46
			2.32	23.88	3.05	24.8	2.71
5d	.508	2.65	8.66	20.80	0.65	24.8	1.43
			5.80	23.02	1.45	24.8	1.86
			3.88	20.96	1.70	24.8	1.66
			2.70	21.48	1.90	25.0	1.89
5e	.359	2.65	7.84	22.15	2.05	24.6	1.56
			5.91	20.20	1.55	24.6	1.38
			4.29	20.10	1.50	24.6	1.46
			2.50	21.33	2.00	25.0	1.91
			0.98	15.10	2.50	24.6	1.23
5f	.254	2.65	8.38	19.80	2.60	24.8	1.25
			5.99	20.42	2.00	24.8	1.36
			3.52	20.50	2.35	24.8	1.48
			1.18	19.70	2.40	24.8	1.79

Table 2-3

DATA ON DEPOSITIONAL SHEAR STRESS (continued)

5g	.180	2.65	8.02	18.05	0.00	24.8	1.00
			3.65	18.55	0.60	25.0	1.20
			0.89	17.30	2.00	25.0	1.46
5h	.127	2.65	8.32	18.05	1.95	24.8	0.98
			5.42	17.50	1.70	24.9	1.00
			4.45	13.73	1.05	25.0	0.66
			1.04	12.38	1.60	25.0	0.73

Figure 2-9



PERCENT BED SHEAR DIFFERENCE - INCIPIENT MOTION VERSUS DEPOSITION

Lift becomes an increasingly important influence on the percent shear difference for sizes less than 1.0 mm because the lift influence becomes greater on the moving grain than the bed particles. The bed grains are well packed and lift is partially a function of interstice volume in this size range. As the pore space decreases, lift influence on the bed grains also decreases. This does not hold for grains moving across the surface, however, because the velocity differential above and below the particle is maintained in open flow. This suggests lift is of greater magnitude on the moving grain; therefore, the bed shear necessary to move the particle is less than that to begin motion of a similar grain from the boundary. Sediment sizes in excess of 1.0 mm exhibit a decrease in percent shear difference because lift on the moving grain and lift on the bed grain are similar.

Heavy Mineral Deposition

Heavy minerals were deposited on a heavy mineral bed under the same bed conditions observed for the heavy mineral incipient motion (Figure 2-7). Observations of deposition were difficult due to the dark color of the bed and moving grains. This may partially account for the similar results.

Heavy minerals were observed to collect on a stable light mineral bed in all flume tests. They were deposited on large light mineral grains and in interstices. Little turbulence was observed in the pores. The heavy minerals did not move until the entire bed began motion. Heavy minerals moving across a fine light mineral bed in association with coarse light minerals were also trapped in the bed. The implications of these observations is that heavy minerals only move when the entire flat light mineral bed moves although they may by selectively deposited prior to light mineral deposition.

CHAPTER III

TRANSGRESSIVE SHELF MODEL

INTRODUCTION

Placer exploration on continental shelves encounters many problems resulting from a disequilibrium fluid-sediment interface. Much of the surface material on the shelves was originally deposited in continental and littoral environments. Thus, the sediments are relic of past events, events that occurred when relative sea level differed from present sea level. Heavy minerals may be located in economic quantities in the shelf region, but their formative history may in no way be reflected by contemporary dynamic processes. Therefore, a placer formed on a beach, in an estuarine environment, or on a coastal plain may now be preserved offshore as a consequence of marine encroachment of the original depositional site.

The placer prospector is confronted with the problem of 1) reconstructing the original placer setting and 2) relating the primary deposit to destructional and preservative processes active during sea level fluctuations and stillstands. Since many contemporary surficial shelf-sediments are relic of past depositional environments and past dynamic events, the explorationist is doubly challenged in his offshore placer search.

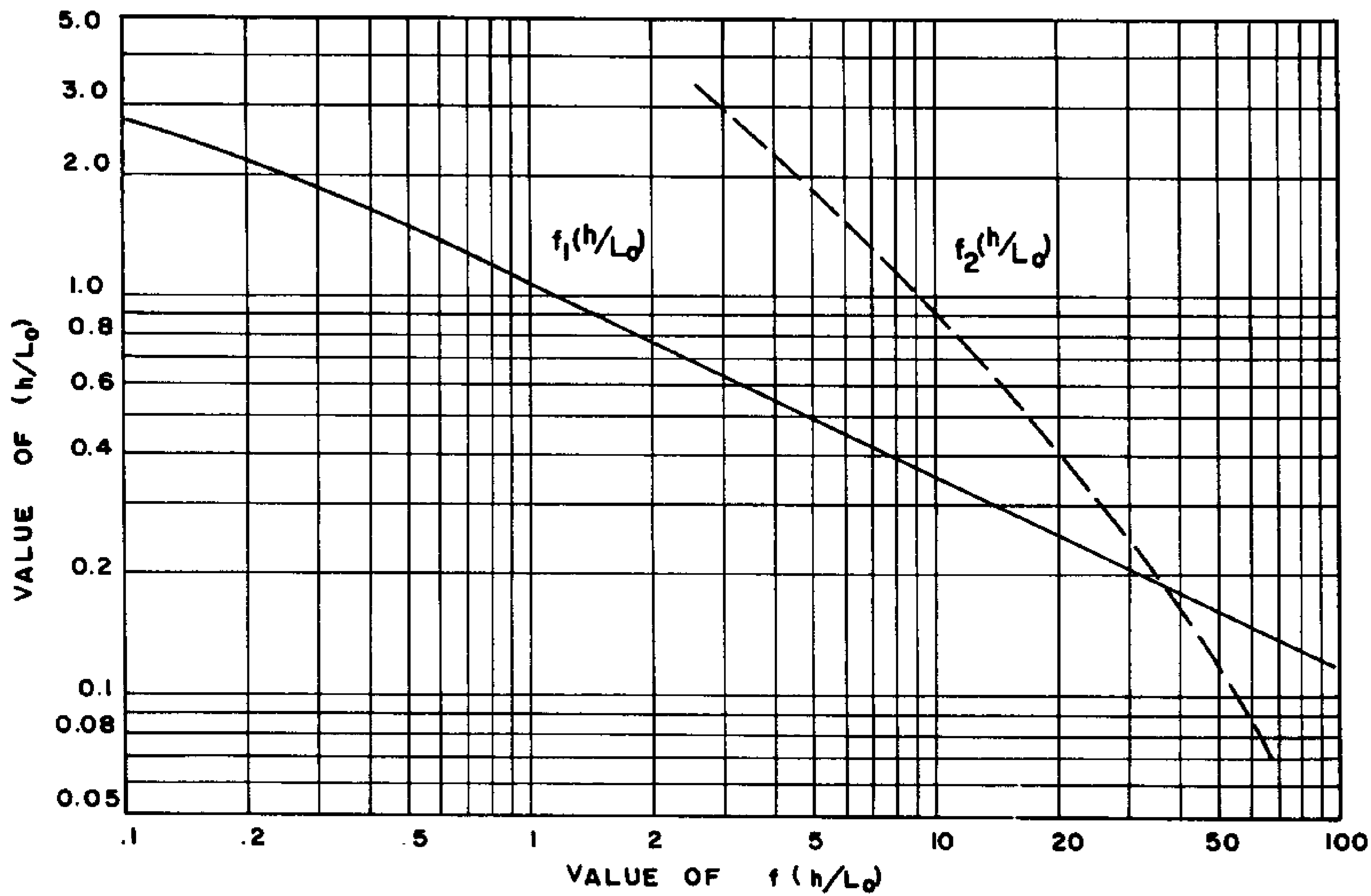
The purpose of the chapter is to develop a conceptual transgressive sedimentation model capable of placing the heavy mineral aspect of the phenomena in perspective. Its objective is to aid the prosepector in visualizing events leading to relic placer preservation on the shelf.

WAVE PROCESSES

Waves impinging on a coast constitute the dominant dynamic process responsible for changes in topography and sediment distribution above and below sea level. Waves acting on the fluid-sediment interface may modify and redistribute bed materials in a manner such that the entire beach progrades or retreats. The intensity of beach erosion in response to waves and the lateral rate of beach movement dictate whether a land-derived placer will be preserved in the shelf environment.

As a shoaling gravity wave moves shoreward, the oscillating fluid velocities and pressure gradients produced near the ocean bottom will increase as the water depth decreases. At some point they may become sufficiently large to result in incipient motion of a given sediment size. This condition of unstable equilibrium

Figure 3-1



SEDIMENT MOTION FUNCTIONS (FROM EAGLESON ET AL., 1963) is defined by Ippen and Eagleson (1955) to be

$$d_i = 258.7 \frac{\nu^{1/2} H_o}{T^{3/2} g} \left(\frac{S_f}{S_s - S_f}\right) \frac{f_2(h/L_o)}{1.3 + \sin\alpha} \qquad (3\text{-}1)$$

where $f_2(h/L_o)$ is plotted in Figure 3-1 as an aid to computation. The grain diameter d_i at incipient motion for the given deep water wave conditions, wavelength less than 0.5 times the depth, is given in feet. The wave height is expressed as H_o in feet, the wave length is L_o in feet, and T is the wave period in seconds. The subscript denotes deep water wave characteristics. The slope between the beach gradient and horizontal is the angle α. Specific gravities of fluid and sediment are S_f and S_s respectively. Equation 3-1 assumes zero reflection and normal incidence.

Shoreward of the point of incipient motion, assuming the beach slope remains constant, a given particle size will exhibit oscillating or quasi-oscillating equilibrium because sometime during each wave cycle, fluid forces will exceed resisting forces (Eagleson and Dean, 1961). There may be a mass transport and drift of sediment particles just as there is of fluid particles, the direction and magnitude of which depends on the relative values of the opposing forces. When the resisting (or gravity) force is greater, net bed motion will be offshore; in the case the fluid force is the larger, net motion is onshore; when they are equal, the particle oscillates with no net motion. The null or equilibrium condition for any particle size d_e is given by Eagleson _et al_. (1958) as

$$d_e = \left[13 \frac{H_o^2 \nu}{g T L_o} \left(\frac{S_f}{S_s - S_f}\right) \frac{f_1(h/L_o)}{\sin\alpha}\right]^{7/6} \left(\frac{\pi}{\nu T}\right)^{2/3} \qquad (3\text{-}2)$$

where $f_1(h/L_o)$ is plotted in Figure 3-1.

Equations 3-1 and 3-2 were developed using a theoretical laminar near-bottom velocity distribution. Higher velocities might be expected in both fluid and sediment because field conditions are likely turbulent. A direct application of the equations to heavy minerals is probably unrealistic due to the low heavy mineral concentration anticipated in an offshore environment. However, the mechanisms of transport and deposition discussed in Chapter II suggest light mineral characteristics may be extrapolated to equivvalent heavy mineral sizes, if they are present.

Equations 3-1 and 3-2 provide a means of answering questions concerning equilibrium sediment distribution and the equilibrium offshore profile in the wave-active offshore zone. Additionally, we can also define disequilibrium sediments and their direction of transport as well as their final resting place using these formulae. The equations also allow a qualitative discussion of profile changes in response to sea level fluctuations. This is important because it enables us to predict the modification of beach and backshore placers during a transgression.

THE EQUILIBRIUM OFFSHORE PROFILE - STABLE SEA LEVEL

The rate of bed load movement is a function of its position on the wave-active face (Figure 3-2). The profile will vary unless d_e is reached for all sediments on the active slope; therefore, the profile configuration is expected to be governed simultaneously at all points by d_e when the predominant means of sediment transport is a wave mechanism. Eagleson et al. (1963) observed that the use of d_i produced unrealistically large slopes.

The seaward limit of profile modification is shown in Figure 3-2. To maintain a stable profile, local sediment sizes must be greater than d_i because no movement commences until $d_s = d_i$. Under the present definition of established motion, the water depth h_i at the point of incipient motion may be greater or less than h_e, the depth for which $d_s = d_e$.

For the case of h_i greater than h_e for a given grain size, an offshore net sediment motion will be partially involved in forming a new equilibrium profile (Figure 3-3). Miller and Zeigler (1958), in observations of Cape Cod beaches, reported this motion to be weak and of limited extent although such motion accounts for the gradation of sizes offshore. When fine-grained sediments are moving on the wave-active face, the foreward wave surge may place them in suspension while the outward reverse surge moves them offshore and downslope. Equations 3-1 and 3-2 do not explain this as they account solely for bed load. The quick shoreward wave stroke and slower return stroke favors shoreward tractive load transport and seaward suspended motion.

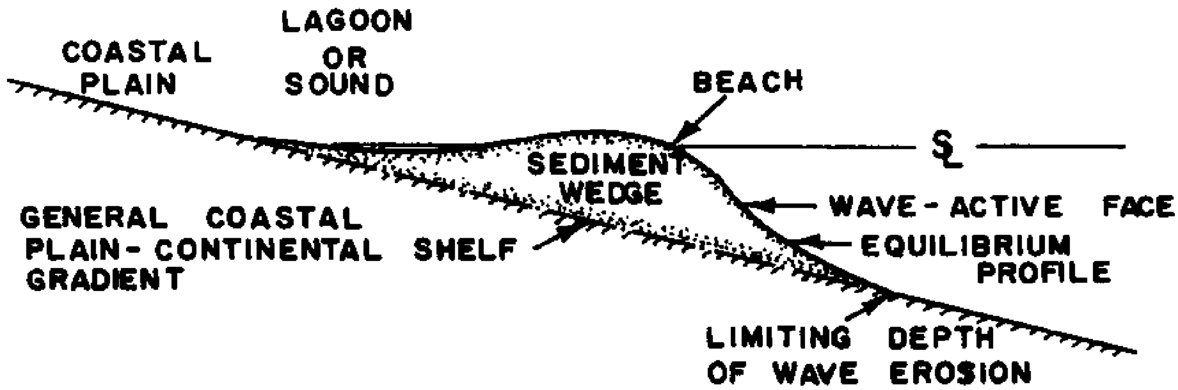
The zone onshore of the null point in Figure 3-3 is aggrading, being nourished by a hypothetical sediment source at the null point. On a natural slope composed of many sizes, the null point is different for each size. For this reason the profile may be considered as a redistributor for its sediments, and in a concave upward form.

For a given initial beach profile with h_e greater than h_i (Figure 3-3), the offshore limit of profile modification must be the point of incipient motion. This corresponds to a degrading profile immediately shoreward of h_i. Because the point of incipient motion is onshore of the point of established motion (null point), profile modification is due entirely to onshore net bed sediment motion. Offshore scour results in aggradation in the onshore zone of profile modification.

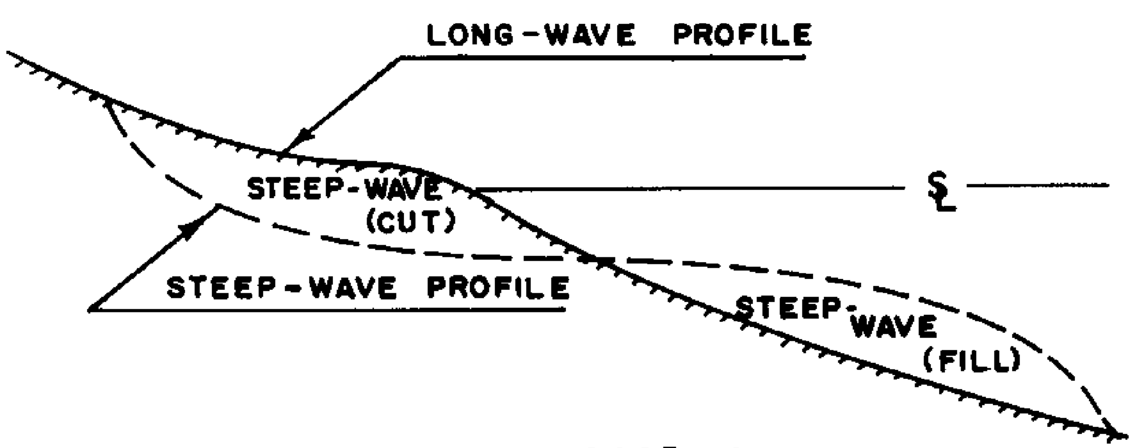
An equilibrium slope results when $d_s = d_e$; therefore, in order to get an even slope, a consistant gradation of increasing sediment sizes shoreward is necessary.

Figure 3-2

VERTICAL EXAGGERATION X 20



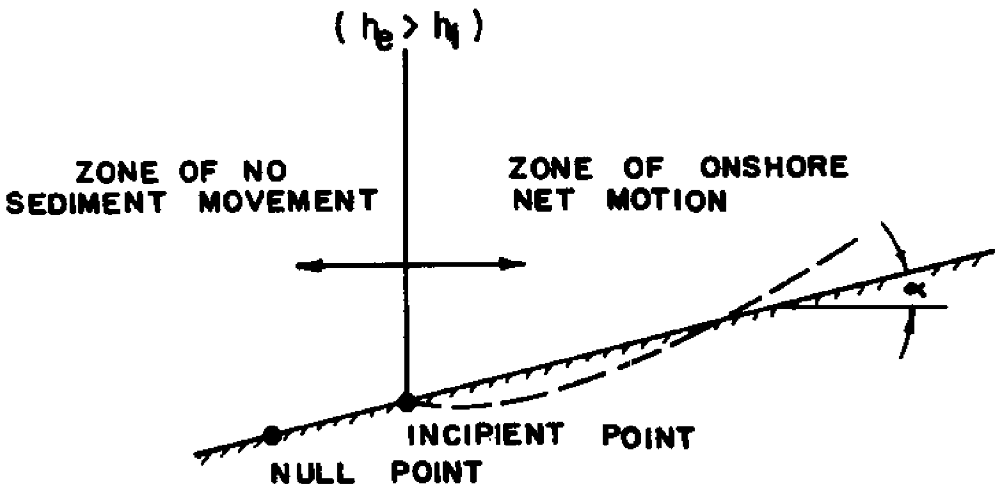
EQUILIBRIUM OFFSHORE PROFILE



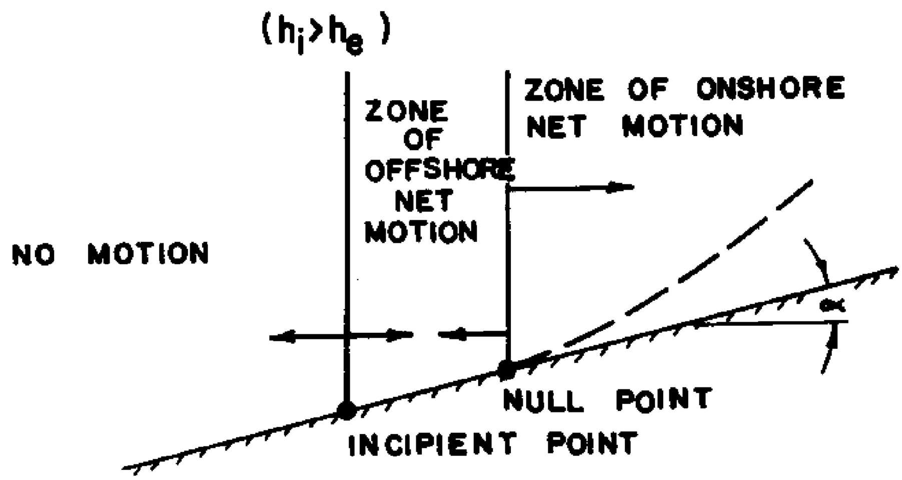
VARIATION IN BEACH PROFILES

BEACH PROFILE RESPONSE TO WAVE CLIMATE

Figure 3-3



ERODING PROFILE



AGGRADING PROFILE

OFFSHORE PROFILE MODIFICATION

(MODIFIED FROM EAGLESON ET AL., 1963)

Theoretical profiles in uniform sediments were proven to be qualitatively correct, but, generally, equation 3-2 predicts too steep a slope (Eagleson and Dean, 1961). This indicates that the mass transport velocities predicted by theory are larger than the actual velocities. Bagnold (1947) notes that the rippled surface common on the wave-active face causes a reduction in the foreward mass transport, however.

Assuming a general gradient for an area, the wave-active sediment face is taken as that portion of the offshore and beach zones between the point of no significant offshore bed transport and the highest high water mark on the beach (Figure 3-2). The general gradient is conceptual in nature and represents the sloping coastal plain-continental shelf surface upon which the sediment wedge migrates. It is a valuable conceptual aid in evaluating sediment and profile changes in response to a transgressing sea.

With a constant sea level, equations 3-1 and 3-2 predict offshore sediments will be moved onshore. They also forecast a recession of the wave-active face if the upward moving sediments are moved out of the beach zone. If sufficient sediments are supplied to the beach and offshore zones, the retreat may be counteracted, thus, resulting in possible beach progradation.

THE TRANSGRESSIVE COAST - RELATIVE SEA LEVEL RISE

Sediments on the continental shelf are either actively modified by contemporary shelf processes or inactive and relic of formative conditions of the past. In the former case, materials undergoing modification often are relic sediments also deposited at lower sea levels. Whichever condition prevails, it is important for the placer prospector to know the original depositional conditions as well as their chances of preservation during a transgressive period and the present sea level cycle. Heavy minerals concentrated on land may now be located offshore in sufficient concentrations to warrant exploration.

As the sea level rises, the beach zone and certain portions of the sediment wedge may retreat landward (Figure 3-4). The amount of wedge that moves is dependent on (1) the intensity of wave attack (2) the form of the sediment wedge, (3) the quantity of sediment supplied to the wave-active face, (4) the general gradient of the coastal plain-continental shelf region, and (5) the relative rate of sea level rise. A variation in any of the five conditions could result in either the formation of a bypassed offshore-relic deposit or an eroded segment of the shelf base.

The expenditure of wave energy on a given coast regulates the depth to which wave action will erode the wave-active face. This depth controls the selection of members of the sediment wedge which will be eroded or preserved as the coast and wave-active face retreat.

The form of the beach wedge is also an important factor in governing the erosional history of a transgression. If the upper beach surface region of the sediment wedge is topographically high, erosion of the wave-active face will be hindered because: (1) wave-activated sediments moving landward with the transgressing wave-active face will have no outlet and therefore will remain in the wave zone and inhibit erosion, and (2) a greater volume of sediment must be moved to accomplish beach retreat. Because the energy expended on erosion is finite for a given coast, and because much of this energy is utilized in moving sediments from the beach foreshore regions, a high coastal profile will absorb more energy in affecting its retreat than a low profile coast. The energy loss per unit time in removing coastal material from high profile coasts will therefore cause a decrease in transgressive retreat and allow for a deeper wave-cut profile.

Sediment availability is a controlling factor in beach retreat because erosion is only possible when an outlet exists for the removal of eroded material. If new sediments are entering an area or if longshore and backshore removal of material is inhibited, beach retreat, even during a sea level rise, may slow, or in certain instances, stop. The beach face could prograde if sediment input exceeded removal.

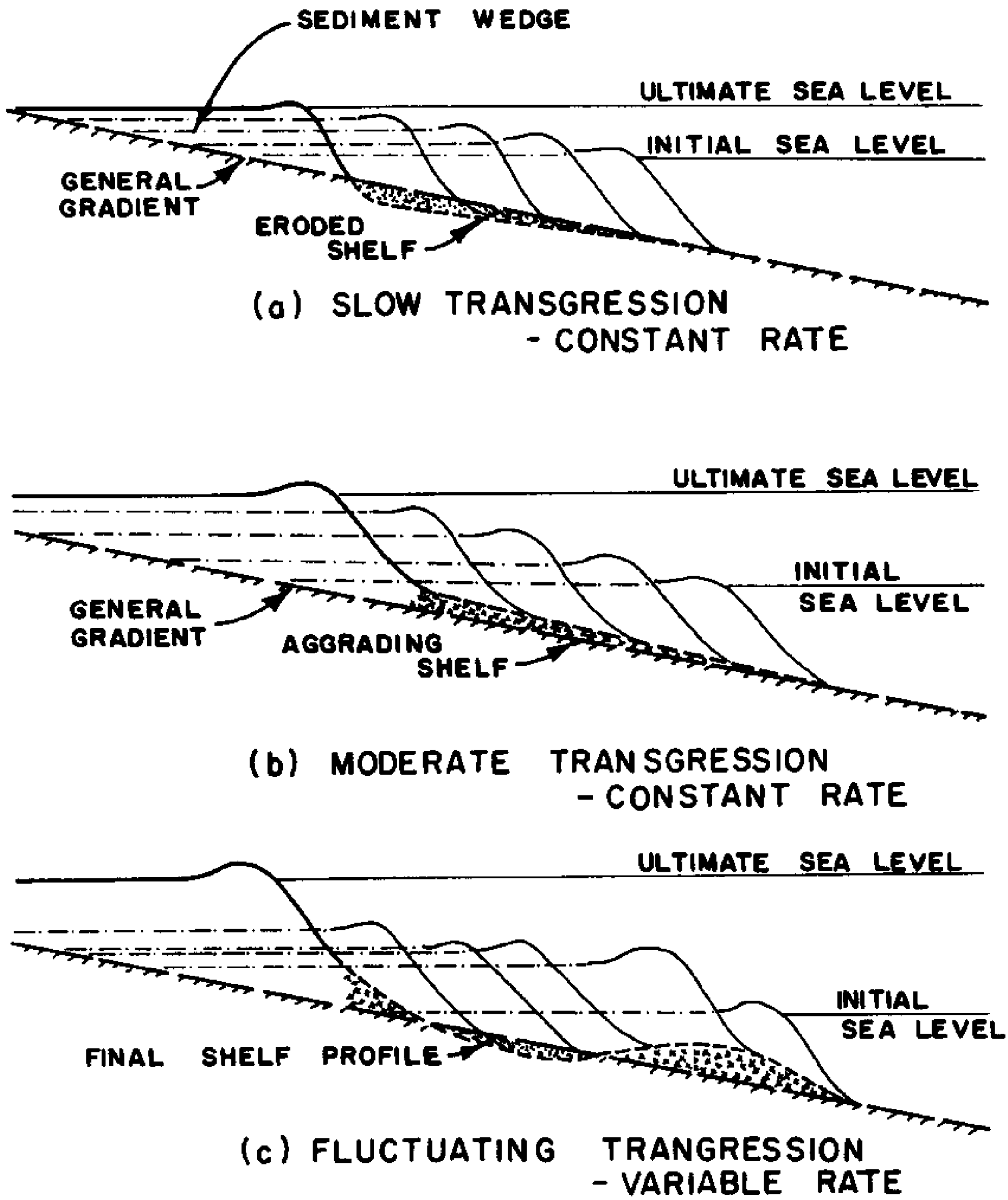
Steepness of the general coastal plain-continental shelf gradient partially dictates the amount of erosion or deposition extant on this surface during a transgression. A slope steeper than the wave cutting plane of the transgression will erode the regional gradient surface. Conversely, a mild regional gradient may become everywhere a depositional surface.

Several types of transgressive sequences are illustrated in Figure 3-4. Assuming all other variables to be constant, changes in the location and form of the wave-active face are a function of the relative rate of sea level rise. During a very slow rise or at the time of a stillstand, the shelf may actually erode as wave scour penetrates the surface of the general gradient (Figure 3-4a). In this sequence material is introduced into the sediment wedge from a shelf floof relic of past regressive or stable cycles.

If the transgression rate increases at a constant rate, the shelf may aggrade (Figure 3-4b). For a fluctuating rate of rise or in the case of variable conditions of sediment input or loss, wave intensity, or general gradient, a stepwise manner of shoreline retreat may ensue (Figure 3-4c). This is probably the more common situation as it results in an undulating relic topography and disequilibrium surface sediment distribution on the shelf.

Transgressive sequences commonly rest disconformably on underlying strata (Swift, 1968). They may be a legacy of earlier cycles of marginal marine and coastal plain deposits of the syn-

Figure 3-4



SCHEMATIC OF TRANSGRESSIVE SEQUENCES

chronous cycle. Wells (1960) reported that the transgressive sequence in a cyclical record was observed to be generally thinner than the regressive sequence. This suggests, to me, that wave action moves a considerable amount of sediment landward with the retreating wave-active face.

Surveys of continental shelves using precision depth recorders have established an undulating surface. Emery (1958, 1960), Heezen and others (1959), and Uchupi and Tagg (1966) observing these remnants of the lower sediment wedge, described them as secondary shelf terraces, and noted that they were typical of all continental shelves. Stetson (1938) and Gorsline (1963), among others, have confirmed the presence of disequilibrium relic surface sediments at many depths on continental shelves.

Any offshore relic deposit owes its existence to a given set of dynamic geological and oceanographic variables. If an investigator wishes to locate relic placers, he must understand the relationship of these variables to the inferred land-formed placer he wishes to trace offshore. It is not the purpose of this discussion to define the specific types of sedimentary deposits that might be preserved as offshore relics, but to introduce the processes culminating in their preservation.

CHAPTER IV

OFFSHORE MARINE PLACERS

INTRODUCTION

Heavy mineral exploration in the offshore zone, defined as that region seaward of the breaker zone to the continental shelf is presented with the unique and complex problem of determining the effect sea level changes have had on coastal sediments. The placer situation in further complicated by the present lack of detailed knowledge concerning contemporary sediment dispersal mechanisms on the ocean floor.

This chapter deals with the mechanics of passing material into the offshore region and the effect that offshore flow systems may have on these sediments. The objective is to relate heavy mineral accumulations, their location, and their spatial extent by relating wave-activated sediment transport mechanisms to the region.

For the purpose of this discussion, offshore placers are broadly categorized into three types based on the time and location of their original formation:

1. Placers forming today from present-day heavy mineral input into the offshore region.

2. Placers forming today from a redistribution of heavy minerals introduced to the shelf in the past.

3. Placers relic of past heavy mineral input and past formative processes.

The first category is briefly discussed with respect to a stable sea level and an equilibrium offshore profile. The second case is covered in detail in the discussion of the natural study site. The third class of placers is important especially along prograding or emergent coasts, coasts subject to low intensity wave motion, or coasts experiencing a rapid transgression. This class, however, will be dealt with only cursorily as it is the subject of a present Marine Research Laboratory offshore placer investigation in Alaska. Its inclusion in this discussion is to maintain continuity because each placer type may grade into the other, thus a knowledge of the existence of each is important in an evaluation of an offshore region.

PLACER AVAILABILITY TO THE OFFSHORE SYSTEM

All placer minerals are derived from crystalline rocks. McKelvey and Wang (1969), using published literature, have plotted worldwide offshore occurrences of important placer minerals with respect to these rock types. The importance of source in this discussion in not mineral provenance, rather the site of its insertion into the offshore dispersal system.

Movement to the point of insertion may be as straightforward as a river emptying heavy minerals directly offshore or high velocity ebb-tidal currents passing them from an inlet or estuary. Some placer minerals may be trapped in the beach zone and moved in the direction of predominant littoral drift. Their secondary origin with respect to offshore nourishment lies at that location offering an opportunity for introduction offshore. For very steep nearshore gradients, such as those encircling oceanic islands, placer minerals may, in the absence of an unconsolidated beach zone, weather from the mountain slopes and pass directly offshore.

As the sea level rises and the beach zone transgresses the shoreline, many land-formed placers may be incorporated into the offshore region. Some placers may be subsequently redistributed in a type two offshore placer while others remain preserved in their original form, essentially as type three offshore placers. Obviously, familiarity with coastal plain morphology and continental placer-forming processes is of importance.

Rivers, Estuaries, and Inlets

Sediments, including heavy minerals, moving in a river, estuary, or inlet may enter the offshore zone, may be halted at the coastline and incorporated in the beach transport system, or may never reach the coast. The behavior of a heavy mineral particle as a function of flow conditions is critical to its movement offshore.

Assuming a placer mineral reaches the coast, it will be transported offshore if the river, estuary, or inlet current velocity at its mouth is sufficiently high to thwart landward-moving tidal currents. With a high volume to transported sediments or a steep coastal profile, similar offshore transport may prevail. In each case waves impinging on the outlet must be insufficient to counteract offshore movement. A plentiful sediment supply may result in the formation of an offshore delta or shoal. Moreover, sediments may pass beneath the influence of waves and settle offshore if the coastal gradient is steep.

An offshore shoal will contribute sediment as it erodes, some of which may pass into deep water. During a relative sea level rise, a shoal may be transgressed losing its source of nourishment. Because of its large sediment volume in a cross-section normal to the coast, transgressive scouring by waves impinging on its seaward edge may be incomplete. The remaining material may constitute an important offshore source of placer minerals if they were initially present on the shoal.

Tanner et al. (1961) and Bates (1963) reported a possible masked heavy mineral deposit in the submerged Apalachicola River delta and its offshore shoal systems in the Florida Panhandle. This is an area of very low wave energy. Upon reaching the coast, heavy minerals are deposited or transported seaward by currents and redistributed by storm waves. The placer minerals may have been deposited in the present offshore region by the Apalachicola River at a lower sea level stage. Processes of heavy mineral concentration on shoals are discussed in Chapter VI.

If wave and longshore drift conditions are of sufficient intensity, all or most river, estuary, or inlet-borne sediment will become incorporated in the beach zone there to be moved by littoral drift or to remain as the constituents of a prograding beach. In either case, they are lost to the offshore distribution system.

Sometimes, in the case of drowned river systems, sediments never reach the coast. This is the situation for Chesapeake Bay today. Tidal currents may move sediments into the estuary mouth further influenced by saline bottom currents. The result is a meeting of seaward moving river sediment and incoming tidal current-borne sediment forming a shoal inland of the coast.

Beach

The beach system, defined in this investigation as the region between the breaker zone and the high water mark, is an effective transporting medium when a net longshore drift occurs. Many heavy mineral deposits are located immediately shoreward of this zone. The journey of these minerals between their coastal input source and the placer location, oftentimes hundreds of miles apart, could only have been accomplished in the beach conduit.

The question of concern is whether or not placer minerals will pass out of this zone into the offshore region. Under normal conditions this is unlikely as sediment movement is predominantly onshore, entering the beach system from offshore sourses (Ingle, 1966; Duane, 1970). Heavy minerals were observed by this writer to travel chiefly on the foreshore and surge zones, neither of which is on the beach-offshore boundary.

Under certain wave and beach conditions, rip currents may form and move beach material offshore through the breaker zone. The maximum distance is probably 1000 meters. Placer minerals moved by rip currents will be dispersed in a thin, fan-shaped veneer immediately beyond the breakers if, indeed, they do reach the breaker zone. Unless rip currents are common to a given coast, it is unlikely that they are major offshore feeding agents. Even when they are common, rip currents can only move heavy minerals into relatively shallow water where they are further dispersed and probably moved onshore by waves.

During storms or hurricanes, a large eroding undercurrent may be imposed on the beach. This is a consequence of water piling on the beach face. With no means of egress other than beneath the incoming surge, continuity is satisfied by a shoreward surface flow and a seaward bottom current. Murray (1970), during Hurricane Camille, August 16-18, 1969, monitored an undercurrent using a current meter mounted one meter from the bottom in 6.3 meters of water off the Florida Coast. As wind velocities increased on shore and waves began piling up, he observed an increase from a mean zero offshore bed current component to a maximum velocity of 160 cm/sec. At that point the meter was rendered inoperative, and no further readings are available. His observations suggest that on coasts subject to extreme storm and hurricane conditions, coastal erosion accompanied by sediment transport offshore may be a major factor in shelf nourishment. Harrison and Wagner (1964) and Urban and Galvin (1969), among others, have reported changes in sand volumes between aggrading long-wave dominated beaches and storm modified beaches; however, most sediment moved offshore is assumed by the investigators to again replenish the beach during moderate sea conditions.

Wind Transport

Wind, as an agent in moving coastal sediments outside the beach system, is of little significance. The author observed wind transported sand grains to move by saltation with bounces rarely exceeding two or three feet. The maximum jump observed at velocities of 40 to 50 miles per hour was 15 feet for 0.20 mm sand grains. As soon as the saltating layer encountered a water boundary, it no longer had a surface from which to ricochet and was trapped. Entrapment occurred in the first 5 or 10 feet of water surface it reached.

OFFSHORE SEDIMENT TRANSPORT

Waves, the predominant offshore dynamic force, result in an oscillatory bottom flow regime. Bed flow may or may not be accompanied by a superposition of unidirectional shelf currents. Although currents on an open continental shelf are rarely strong enough to singularly initiate sediment motion (Ippen, 1966), they may move significant volumes of material already in motion due to wave action on the bottom.

The oscillatory flow regime affects sand-sized surface sediments by regulating bed configuration, bed form orientation, and heavy and light mineral movement. The same transport mechanisms existing on a flat bed are active on a rippled bed; however, the net direction of movement in each mode may vary in a reversing flow field.

Bed Configuration and Orientation

Oscillation ripples were noted on all SCUBA (Self Contained Underwater Breathing Apparatus) observations of the offshore bottom. Wave ripples reflect the scale and velocity of oscillating motion with the ripple spacing nearly equal to the horizontal displacement of water particles near the bottom (Inman, 1957). A ripple index of four to six was observed off the Dare County coast. Ripple wavelengths were observed to be always greater in coarse beds (median diameter greater than 0.5 mm). Inman (1957) examined ripples on the Southern California shelf and reported the wavelength of natural ripples to vary from four to sixteen times the ripple height. Harms (1969) observed the average height of wave ripples in a flume to be consistently one-sixth of the spacing. This was duplicated in the laboratory by Carstens et al. (1969) who noted a ripple index range of 5.0 to 6.5 for two-dimensional ripples.

In most instances the swells extant during diving operations had a wave height of three to five feet and a period of eight to eleven seconds, enough to disturb the bed at a depth of 30 feet according to equations 3-1 and 3-2.

In all cases ripples were observed to be two-dimensional, that is, they exhibited a very long crest dimension compared to their wavelength. The crests in some instances were not oriented parallel to wave crests existing at the time of the survey. For example, in 30 feet of water, the linear direction of a five inch ripple system on a coarse bed was 060 degrees while the direction of wave approach was 120 degrees. A smaller ripple system was superimposed on the surface and oriented consistent with the surface wave pattern.

Carstens et al. (1969) reported bed forms as a function of surface grain size d_s and the maximum horizontal bottom fluid displacement A. Their conclusions were that bed forms would be two-dimensional if A/d_s were less than 775, three-dimensional when A/d_s ranged from 775 to 1700, and the bed would be flat when A/d_s exceeded 1700. According to small amplitude wave theory, A is (Ippen, 1966):

$$A = \frac{H}{2 \sinh kh} \qquad (4\text{-}1)$$

where H is the wave height and k is the wave number equal to 6.18/L. L is the wave length. For shallow water waves, h/L less than 0.05, sinh kh is equal to kh.

Sediment Dispersal on Bed Forms

The most outstanding feature of the observed bed forms was the relatively coarse character of the ripple crests with respect to the troughs. In all instances the coarsest available particles existed at the crests. This phenomena is also noted by Inman (1957). Heavy minerals, when present in visible quantities, were always situated in ripple troughs in association with the finest light mineral sizes. This correlation existed in both coarse (d_s is greater than 0.3 mm) and fine-grained beds although heavy minerals were relatively rare on beds with a median diameter in excess of 0.4 mm.

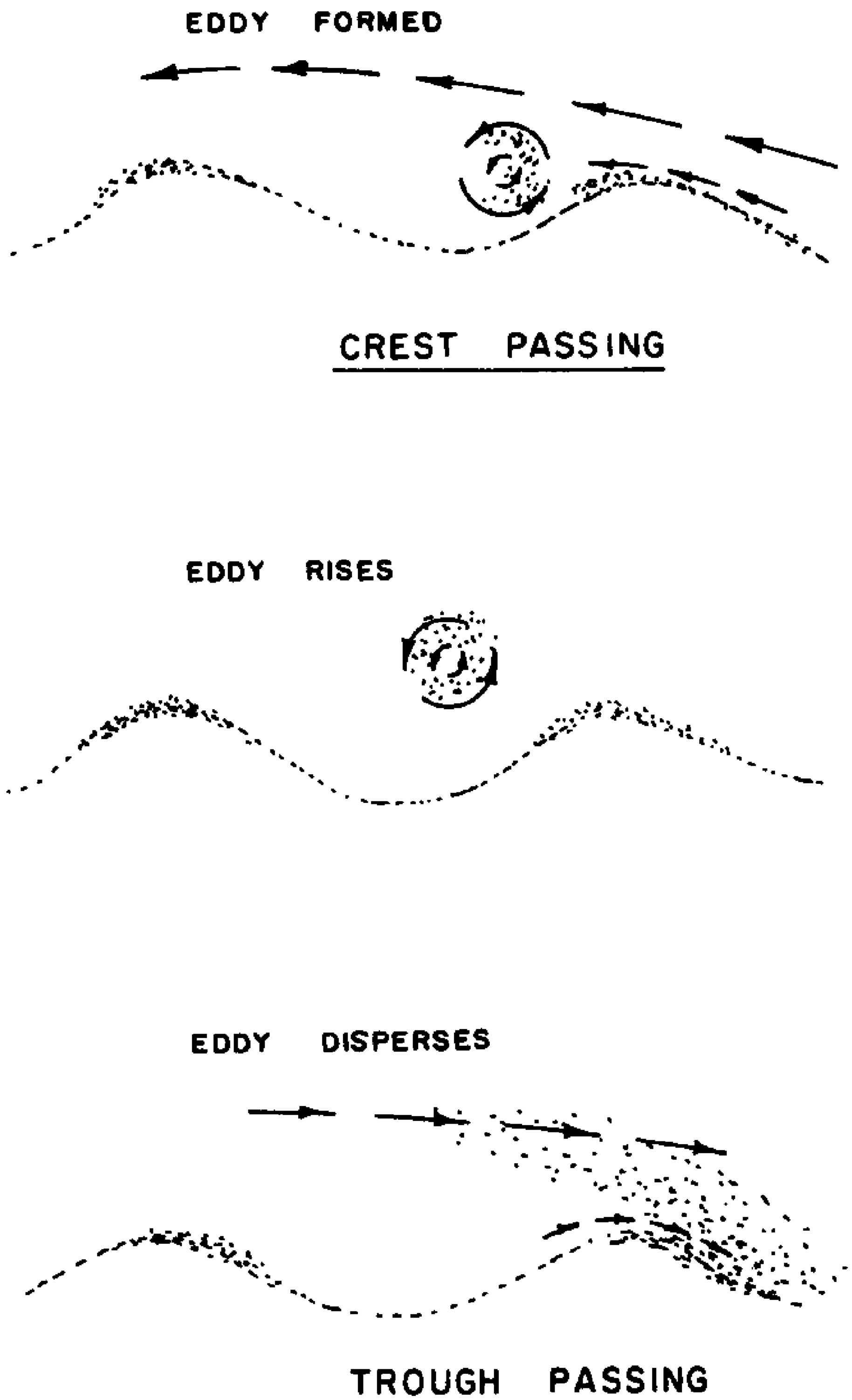
As a wave crest passed on the free surface, scour was observed on the upstream face of the ripple, this being an area of accelerating flow as the fluid moves towards the crest. Sand grains were rolled as bed creep up the ripple face (Figure 4-1). Under sufficiently intense conditions, flow separated from the boundary at the crest and formed a vortex in the downstream region. Leeward of the crest region, the eddy moved the smaller particles in suspension while the larger grains settled to the lee face. The separated flow rejoined the bed in the trough and formed a stagnation line where surface particles remained nearly at rest.

As the trough reverses due to a passing free surface wave, the cycle is repeated in the opposite direction (Figure 4-1). Coarse grains were observed to move slowly shoreward as the maximum velocity in that direction appeared to be stronger but of a shorter duration than the reversed seaward flow. The finer particles, those remaining in suspension, tended to move in a net seaward direction.

Because turbulence and lift are most intense at the crests, the coarsest grains are located there. These constitute the moving fraction in an aggrading environment. In water depths of 30 to 50 feet coarse light mineral grains were observed to pass over fine heavy minerals located in ripple troughs. In shallow water, where computed maximum bed velocities exceeded 100 cm/sec, the entire bed, including heavy minerals was observed to be in motion. The placer grains appeared to move slower than the light minerals.

Heavy minerals on boundaries that appeared to be aggrading tended to be bypassed and buried. The mechanism of concentrating and preserving the placer grains is analogous to flat boundary bypassing where the larger grains move across the smooth, well packed fine bed surface. In an oscillatory ripple environment the ripple crests appear to migrate over the trough sediments.

Figure 4-1



SAND MOVEMENT OVER BED FORMS

(FROM KEULEGAN, 1948)

Heavy minerals in several instances were observed to roll in the trough in response to passing surface waves. When the gravity wave crest was not oriented parallel to the ripple crest the heavy minerals moved in a zig-zag pattern parallel to the ripple axis and parallel to the coastline.

OFFSHORE SEDIMENT DISTRIBUTION - STABLE SEA LEVEL

Assuming an equilibrium sediment distribution, equation 3-1 predicts a particle size decrease with an increasing depth offshore in response to constant wave characteristics. The trend is well established in shallow depths. For example, Bascom (1951) observed linearly decreasing grain sizes with depth for several west coast beaches to a depth of 35 feet. At these depths, the bed reacts rapidly to changes in wave energy. At depths greater than approximately 30 feet, the bed is still a function of contemporary waves, but the frequency of sediment change is much lower than the frequency of wave variation; thus, the deeper regions are in a state of quasi-equilibrium, i.e., bottom sediment disturbance in deep water is in response to the net wave state over a long period of time.

In the ideal static situation, assuming sediments enter the offshore region only through the breaker zone, heavy minerals would not be expected to migrate offshore and be deposited with their equivalent light mineral size. Flume experiments discussed in Chapter II indicate large light mineral sizes may bypass smaller heavy mineral sizes. Although depositional light and heavy mineral sizes are similar, the heavy minerals move in company with the largest available light mineral fraction. If the larger light minerals are deposited on the upper portion of the wave-active face (Figure 3-2), the placer minerals will move no further.

Considering a wide range of wave conditions, large storm waves can and do scour the beach and move most sediment sizes offshore. Later during normal conditions, the large light minerals move onshore again while the placer minerals remain immobile offshore. Johnson (1949) recognizes and defines two distinctive beach profiles, each the result of specific wave characteristics. The principal governing factor is the incipient wave steepness H_0/L_0. For long waves H_0/L_0 is defined by Johnson to be less the 0.0025; for steep eroding waves, it is greater than 0.0025.

With an offshore source, heavy minerals would be expected to move slower than any large light mineral moving upslope. In this case, the heavy minerals would move until deposited with their equivalent light mineral size. In the case of sediment input from either a beach or offshore source, however, heavy minerals would be unlikely to accumulate in sufficient concentrations to warrant exploration. The reason for this is that waves cover a considerable range of intensities, and any location responding to the range would deposit a wide range of sediment sizes, thereby diluting the heavy mineral deposit.

FIELD INVESTIGATION

In order to apply the general concepts of a transgressing beach wedge (Chapter III), an offshore study area was selected in a heavy mineral province off the North Carolina coast. The purpose of the field investigation was to account for offshore topography, sediment distribution, and especially heavy mineral disposition with respect to the general concepts previously discussed.

Location

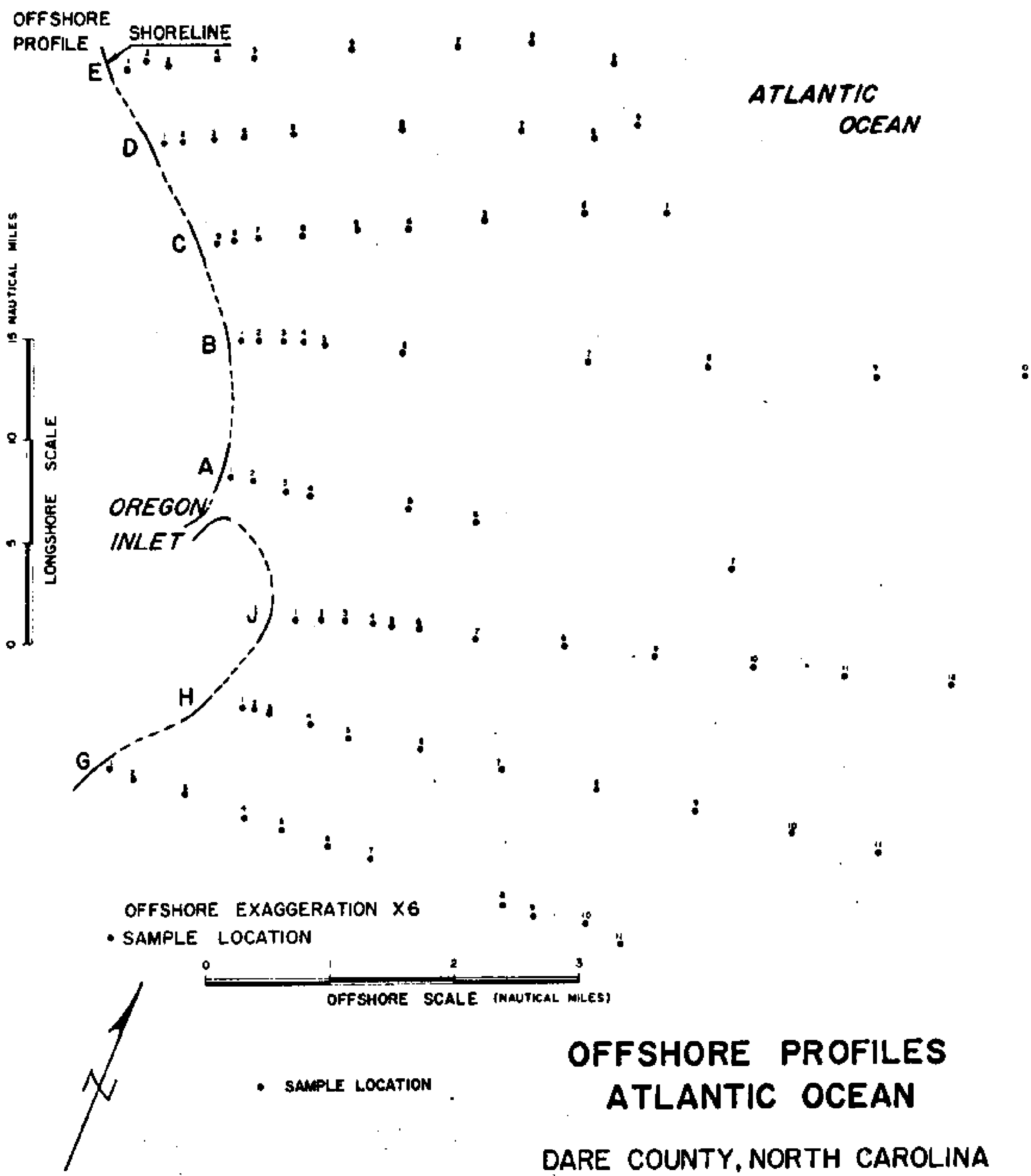
The northern boundary of the study area is situated 60 miles south of Cape Henry, Virginia. The southern margin lies 45 miles further south with the western limit being the coastal breaker zone. The eastern margin is approximately six miles offshore (Figure 1-1). This region faces the Atlantic Ocean and is unrestricted from the shoreline to the shelf break. Width of the continental shelf decreases from 44 miles in the northernmost portion of the study area to 30 miles in the region south of Oregon Inlet.

A 44 foot motor vessel was provided by the United States Coast Guard for use in the collection of bottom sediment samples and determining bed character. Station positioning was accomplished using a Bendix Radar System and three-point triangulation fixes on prominent shore features. Angles were obtained to a 20 second precision.

Samples were taken along eight profiles extending seaward from the 10 foot depth contour, normal to the beach (Figure 4-2). The maximum offshore distance surveyed was six miles, and the distance between profiles was also six miles. Samples were obtained at each 10 foot depth increment, or three-forths mile horizontal distance, whichever came first. It was reasoned that changes in sediment texture and bottom topography would be more pronounced normal to the beach. This assumption, made on the basis of the strong wave activity in the area, was proven correct. The number of stations was limited by vessel availability.

Bottom samples were obtained utilizing a pipe dredge operated from the vessel, and collected on the bottom using SCUBA. The dredge was rigged to entrap the upper two inches of bottom sediment. At stations sampled employing SCUBA methods, the upper two inches of sediment were scooped into a small can by hand. Additionally, the orientation and characteristics of bed forms were noted, as was the locus of any heavy minerals.

Figure 4-2



OFFSHORE PROFILES
ATLANTIC OCEAN

DARE COUNTY, NORTH CAROLINA

Sea Conditions

The predominant yearly wave direction off this part of the Atlantic coast is from the northeast as illustrated on the wind rose in Figure 4-3. Villianos (1967) reports wave conditions for a site near Oregon Inlet as follows:

TABLE 4-1

	Direction of approach	Period (seconds)	Wave height (feet)
Average	NE	9	3
Maximum		13	28
Average	E	7	3
Maximum		21	22
Average	SE	7	3
Maximum		13	23

The largest waves and those of longest duration are the result of northeasterly winter storms.

Records of hurricanes indicate that 34 hurricanes have affected this area since 1905 (United States House of Representatives, 1966). Because of this high incidence of tropical cyclones, extreme sediment dispersal events must be important. The maximum probable wave height during a hurricane is 40 feet with a wave period of 11.5 seconds assuming a water depth of 50 feet (Villianos, 1967).

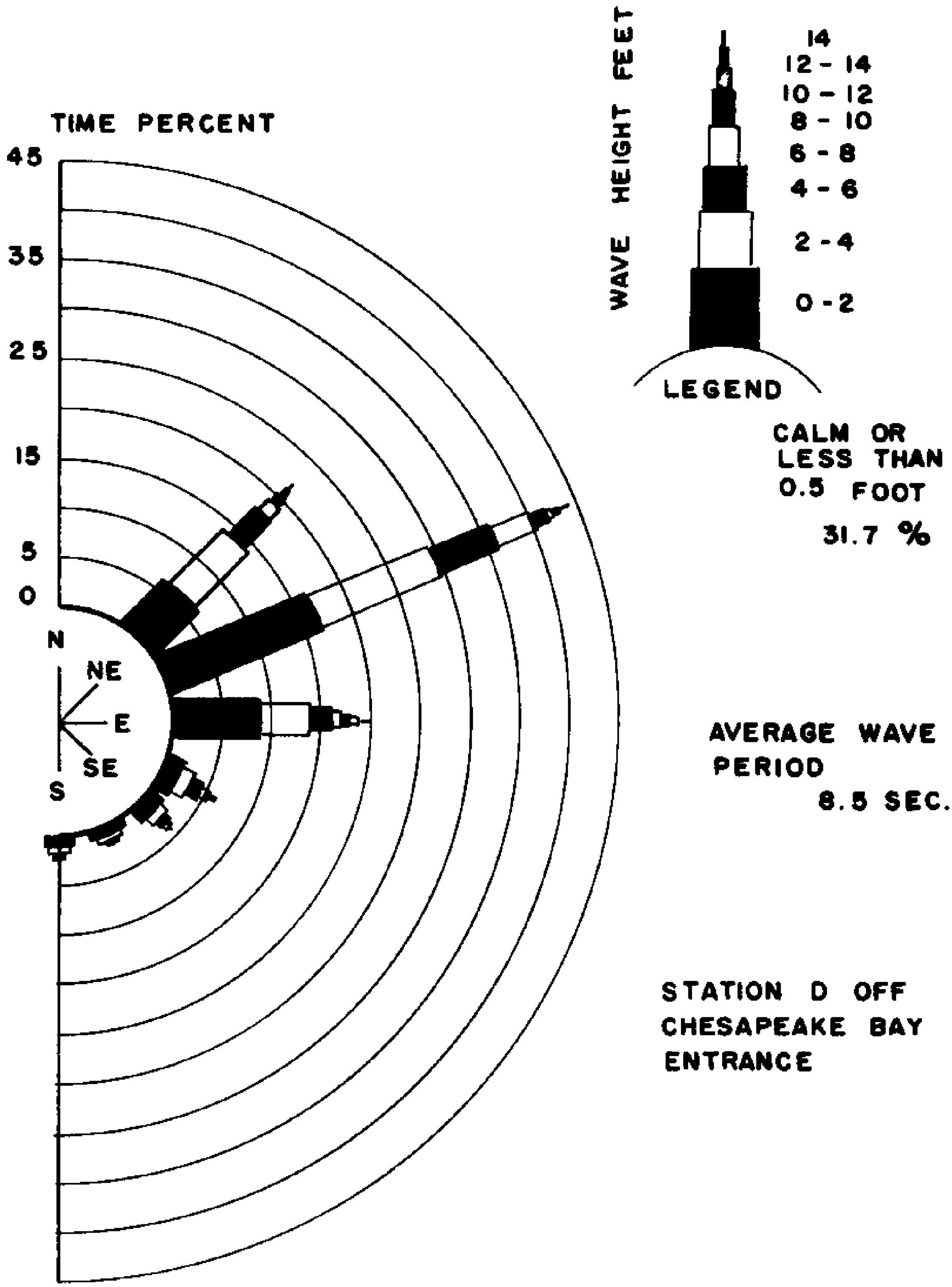
Tides observed along this stretch of coast during the summer of 1969 averaged a diurnal variation of two to three feet.

Bathymetric Features

Figure 4-4 illustrates bathymetry obtained using a continuous sonic depth recorder on each profile. The offshore scale on this chart is exaggerated by six in order to reduce page dimensions.

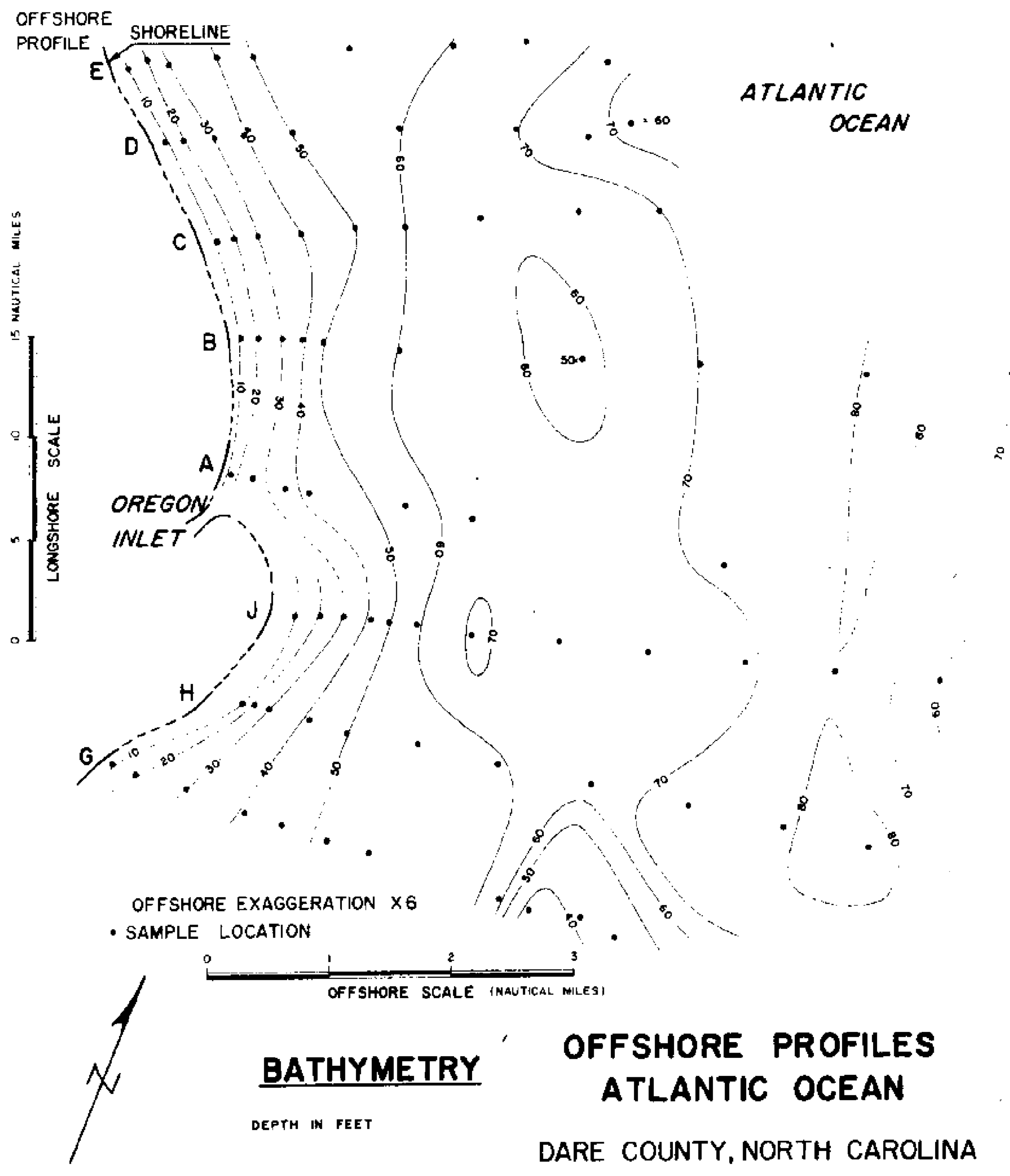
The most dominant bathymetric feature of the offshore study region is the steep nearshore slope terminating at a depth of 50 feet. An erosional nature for this grade is suggested by its concave upward form. The slope gradient increases from 0.0065 in the vicinity of profile E to nearly 0.0080 near Oregon Inlet. South of the inlet the gradient again decreases, becoming less than half the inlet value at profile G (Figure 4-4).

Figure 4-3



WAVE ROSE (MODIFIED FROM LANGFELDER ET AL., 1968)

Figure 4-4


OFFSHORE PROFILE
SHORELINE
ATLANTIC OCEAN
OREGON INLET
LONGSHORE SCALE
15 NAUTICAL MILES
OFFSHORE EXAGGERATION X6
• SAMPLE LOCATION
OFFSHORE SCALE (NAUTICAL MILES)
BATHYMETRY
DEPTH IN FEET
OFFSHORE PROFILES
ATLANTIC OCEAN
DARE COUNTY, NORTH CAROLINA

The profile north of Oregon Inlet is probably near equilibrium with respect to its slope, although this is not true for the sediment texture. The present stillstand has been effective for 5,000 years (Curray, 1964). In the absence of appreciable amounts of sediment entering the wave-active zone from onshore, the slope should be stable as a function of a relatively constant offshore sediment input, wave intensity, and general gradient. Due to the similar topography of each profile which results in uniform wave intensity on each profile at any given depth, the rate and volume of shoreward-moving sediments at each location should be alike and the wave-active face should exhibit a constant gradient. These conditions will not hold true south of the inlet because of sediment input to the offshore zone by inlet currents. The dominant southward coastal current system is induced by the northeasterly wave trains impinging on the shore.

Seaward of the 50-foot contour, the bottom is gently undulating with a regional gradient of 0.003, or one-twentieth the slope of the wave-active face. The regional gradient of the coastal plain is the same, assuming, of course, that the surface of the regional slope lies at a minus 50-foot elevation two miles from shore.

Another conspicuous feature is a trough with a depth of 70 to 80 feet located approximately five miles from shore. Several non-laterally-continuous rises occur landward and seaward of the depression, and most of these are parallel to the coast.

The wave-active slope can be shown to be transporting sediment landward at all depths according to equations 3-1 and 3-2 when employing the wave conditions of Table 4-1. This corresponds to Pierce's (1968) observations, i.e., the beaches immediately south of the study area are being replenished from offshore sources. Langfelder _et al_. (1968) in a reconnaissance of coastal erosion in the study area noted erosion to be dominant. This suggests that the incoming sediments are removed from the beach zone and are insufficient to prevent coastal retreat.

Sediment Distribution

Median grain size decreases in an offshore direction from the coast as shown in Figure 4-5. The northernmost profile exhibits this decrease to a depth of 50 feet and at least a mile offshore. The same trend further south appears to terminate at 30 feet, or less, against a zone of very coarse (greater than 0.3 mm) material. The 0.15 mm median diameter contour correlates with 15 foot depths suggesting equilibrium sediment conditions prevail on the shallower portion of the wave-active face.

Figure 4-5

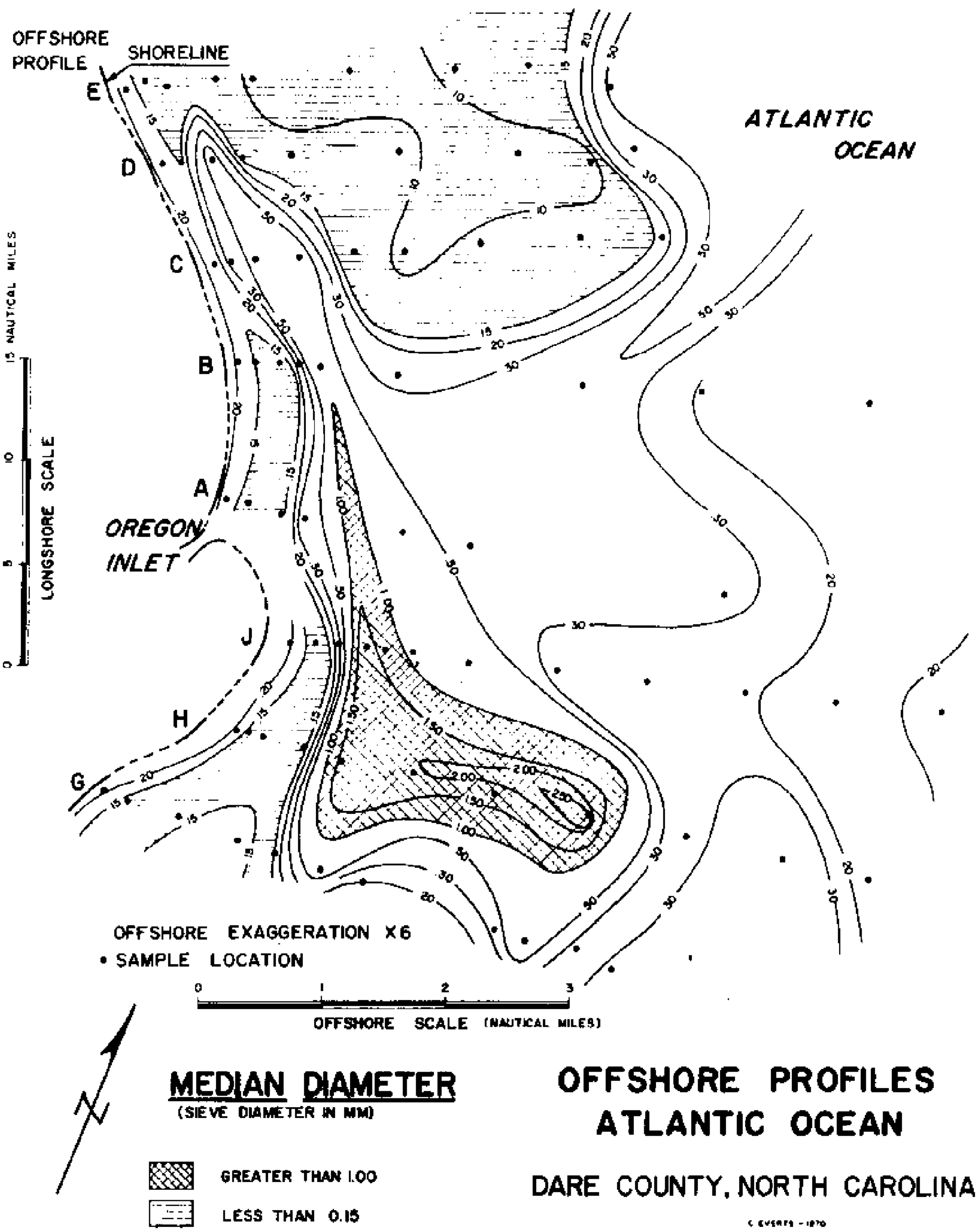

OFFSHORE PROFILE
SHORELINE
ATLANTIC OCEAN
OREGON INLET
E
D
C
B
A
J
H
G
15 NAUTICAL MILES
LONGSHORE SCALE
OFFSHORE EXAGGERATION X6
• SAMPLE LOCATION
OFFSHORE SCALE (NAUTICAL MILES)
MEDIAN DIAMETER
(SIEVE DIAMETER IN MM)
GREATER THAN 1.00
LESS THAN 0.15
OFFSHORE PROFILES
ATLANTIC OCEAN
DARE COUNTY, NORTH CAROLINA

Figure 4-6

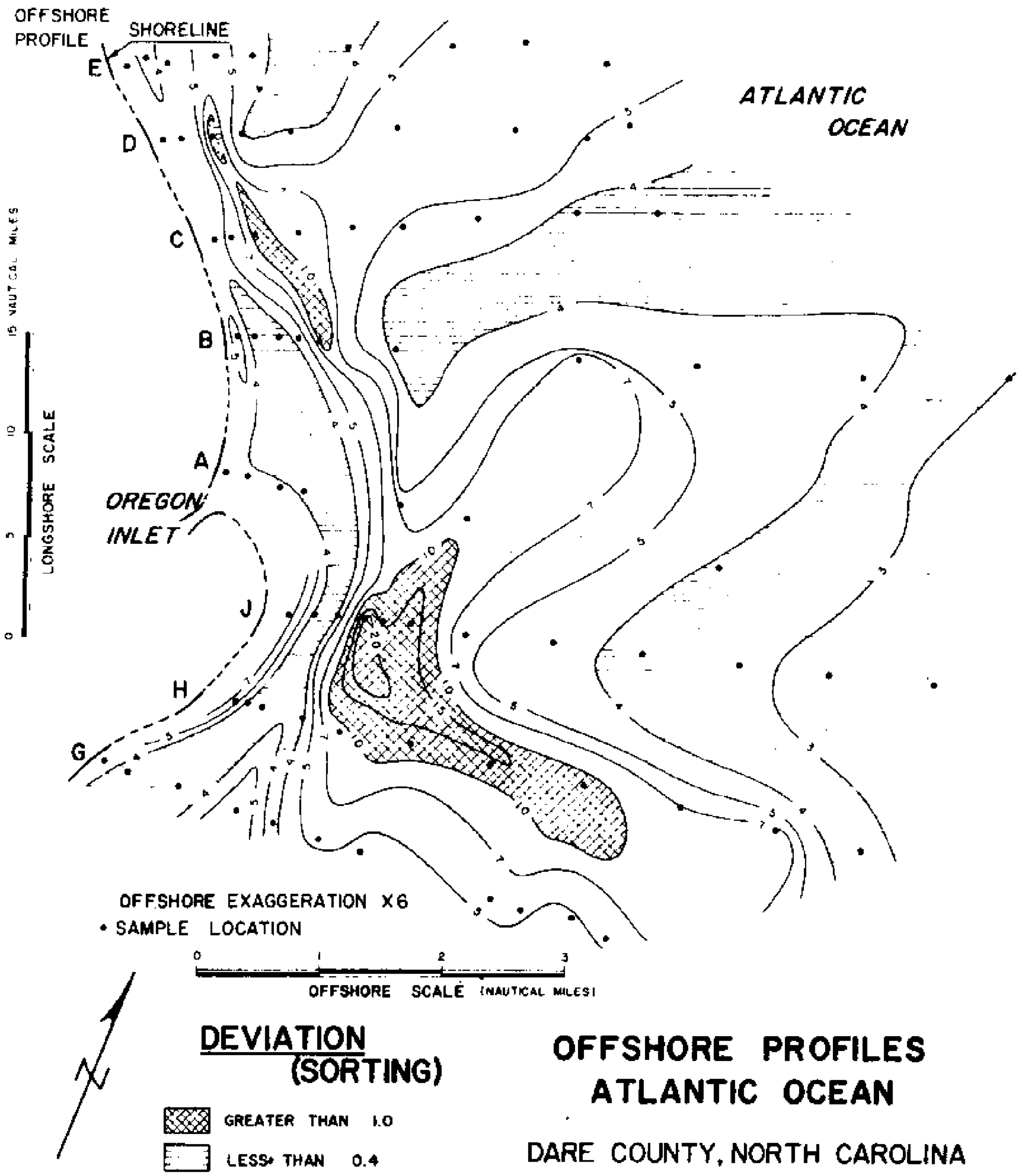
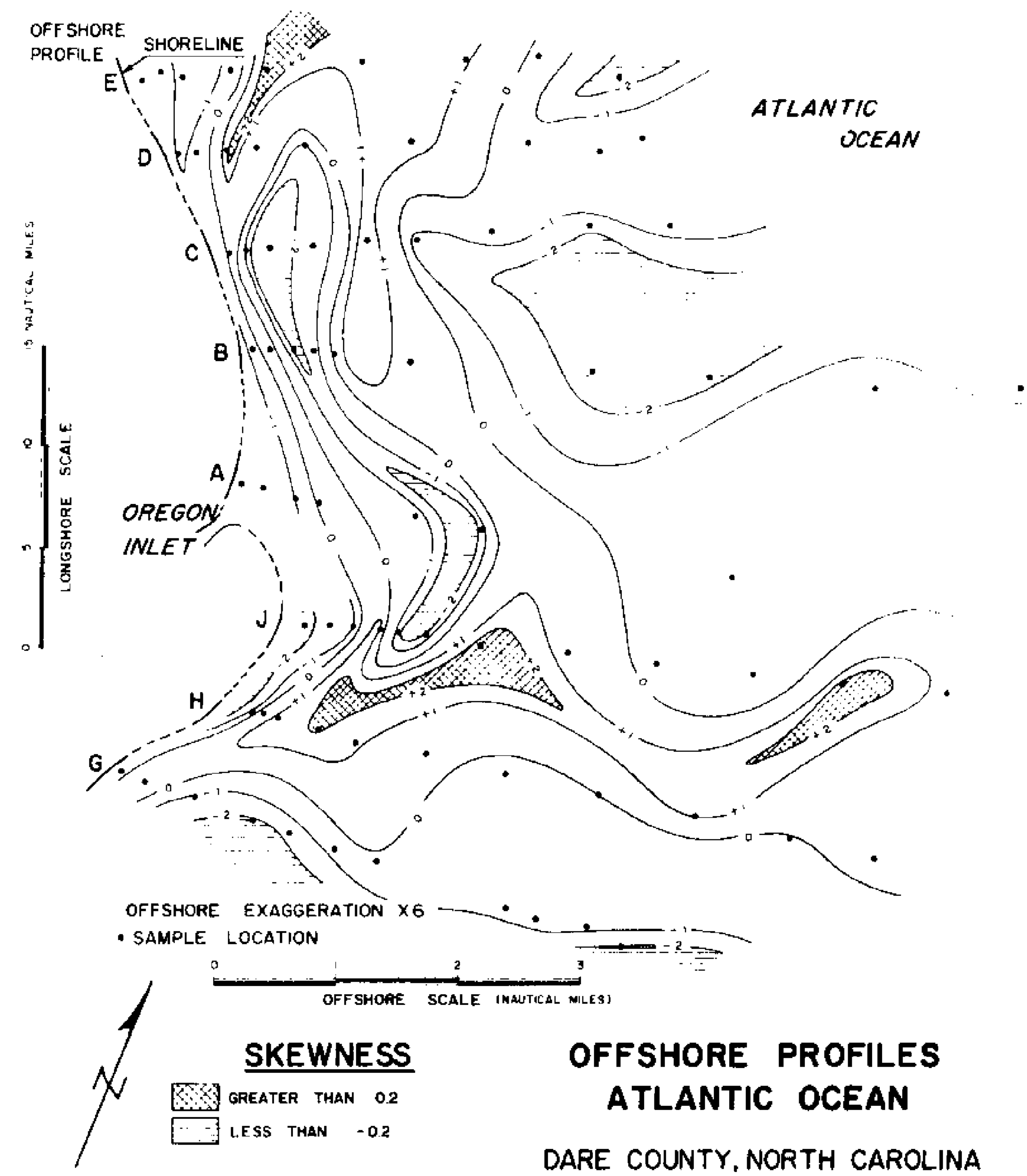

OFFSHORE PROFILE
SHORELINE
ATLANTIC OCEAN
OREGON INLET
LONGSHORE SCALE
NAUTICAL MILES
OFFSHORE EXAGGERATION X6
• SAMPLE LOCATION
OFFSHORE SCALE (NAUTICAL MILES)
DEVIATION (SORTING)
GREATER THAN 1.0
LESS THAN 0.4
OFFSHORE PROFILES
ATLANTIC OCEAN
DARE COUNTY, NORTH CAROLINA

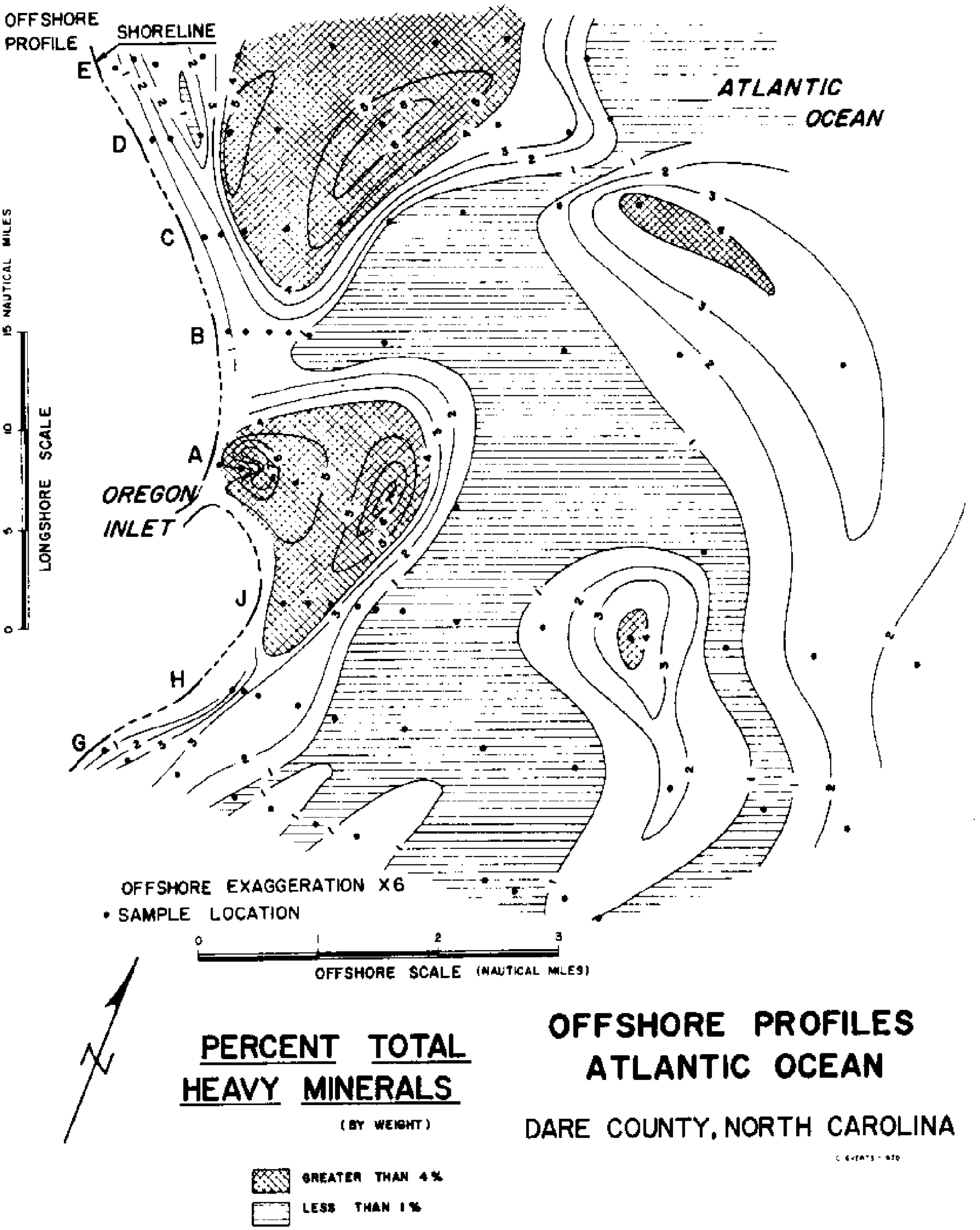
Figure 4-7



OFFSHORE PROFILES ATLANTIC OCEAN

DARE COUNTY, NORTH CAROLINA

Figure 4-8


OFFSHORE PROFILE
SHORELINE
ATLANTIC OCEAN
E
D
C
B
A
OREGON INLET
J
H
G
15 NAUTICAL MILES
LONGSHORE SCALE
0
5
10
OFFSHORE EXAGGERATION X6
• SAMPLE LOCATION
0
1
2
3
OFFSHORE SCALE (NAUTICAL MILES)
N
PERCENT TOTAL HEAVY MINERALS
(BY WEIGHT)
GREATER THAN 4%
LESS THAN 1%
OFFSHORE PROFILES
ATLANTIC OCEAN
DARE COUNTY, NORTH CAROLINA

Sorting values (Inman, 1952), as illustrated in Figure 4-6, decrease slightly seaward in this region in response to the increasing constancy and steadiness of the processes affecting the bed. This is to be expected because wave variations increase in the shallower depths due to relative depth changes during a tidal cycle. The relatively high sorting values (0.7 and greater) inshore and immediately south of Oregon Inlet reflect the wave and longshore current mixing of sediments issuing from the inlet.

Skewness, shown in Figure 4-7, also suggests equilibrium conditions in this region. In all cases, negative skewness values were determined for sands in the shallower regions, while positive values are reported for deeper water sands. A negative value indicates a paucity of fine sizes, grains less than 0.15 mm, while a positive value indicates a termination of coarse grains with respect to the median diameter of the sample. Skewness also reflects an equilibrium upper wave-active face, because as bed intensity decreases downslope, the ability of the bed sediment to entrap and hold fine sizes increases. Plotted skewness trends (Figures 1-12 and 1-13) clearly reveal this.

The general concepts utilizing equations 3-1 and 3-2 agree with our descriptions of the region, i.e., it is in equilibrium with summer wave conditions. As this is the case, the distribution of surface heavy minerals should also reflect equilibrium conditions and indicate the direction of placer mineral input into this zone. Figure 4-8, illustrating the weight percent of heavy minerals in the sediment, suggests a slight increase with depth in the equilibrium nearshore region. There appears to be a correlation between heavy mineral percentages in the wave-active zone and offshore sources. For instance, the nearshore low placer mineral values on profile B reflect a heavy mineral-poor offshore region suggesting a landward transport of heavy grains.

The most conspicuous sediment pattern in the study region is the narrow, linear body of coarse sand just seaward of the equilibrium nearshore zone (Figure 4-5). This feature is 40 miles long, and from 0.7 to 3.0 miles wide. The axial depth varies from 30 feet or less in the north to more than 60 feet south of Oregon Inlet, indicating that the deposit lies predominantly on the wave-active slope. Beaches north of profile C were observed to become increasingly coarse (sand sizes exceeding 0.3 mm). The coarse trend on the beach culminated at profile E where the beach is characterized by the same textural and mineralogical components as the offshore deposit. This evidence, as well as that predicting onshore movement, suggests the sand body has migrated landward and impinged on the northern beaches. Moreover, as the coast north of Oregon Inlet trends northwest, while the coast south of the inlet has a more northerly orientation, the dominant northeasterly wave trains move sediment with a stronger component normal to the northern beaches. This may account for the depth distribution of the coarse deposit.

Bruun (1967) noted that eight or nine miles south of the present position of Oregon Inlet, another inlet was once open, possibly for centuries. The inlet was closed in 1922, but reopened in 1924 and termed New Inlet. It was again closed in 1930. The location of this inlet corresponds to the widest dimension of the coarse-textured offshore deposit. Although inlet flushing may not supply sediments to this coarse offshore layer, flushing probably inhibits them from moving onshore under the influence of waves.

The source of the coarse sand is unknown, but all evidence points to landward migration from an offshore relic source. Hoyt and Henry (1967) observed lateral inlet migration to be a significant form of coastal erosion in areas of dominant longshore drift. Inlets may exceed the depths of other barrier features by 30 to 50 feet along the Dare County coast. The barrier may effectively dig its own trench and then fill it at its lowest levels with coarse inlet channel sand as the inlet migrates. Thus, the sediments at depth beneath a barrier, those most likely to be preserved during a transgression, could be coarse. A shoreward retreat of the wave-active face might result in the offshore preservation of this facies.

Heavy mineral percentages offshore exhibit no correlation with equivalent light mineral sizes in the coarse, poorly sorted deposit. Skewness, however, is generally negative in regions of high heavy mineral concentration in these coarse sediments, especially along the C,A, and J profiles. The region characterized by sediments with negative skewness is centered on profile C and it represents the leading edge of a heavy mineral lobe at the landward margin of the coarse light mineral body. The orientation of the two bodies suggests they are not moving sympathetically. According to the concepts outlined previously in this chapter, light minerals preferentially bypass a heavy mineral bed. This is likely the case for this region, suggesting, of course, that the coarse layer at its northern extremity forms only a thin blanket as it moves shoreward. The high skewness values for sediments on its leading edge are consistent with this process, as all fine sizes would be expected to move at a lesser rate than coarse grains, and to concentrate on its seaward periphery.

The positive skewness values observed for sediments sampled near the 50 foot contour of profile E suggest the excess fine sizes one would expect offshore on an equilibrium profile.

Negative skewness 1.5 to 2.0 miles offshore and south of Oregon Inlet near the outer edge of a heavy mineral zone reflect the effect of current and wave action. In this region, waves disturb the bed while inlet tidal currents, swinging south with the longshore outer current, remove fine particles. These fine sediments are moved south and trapped in the coarse sediments between profiles J and H resulting in highly positive skewness values for sediments sampled in that region.

The heavy mineral zone outside and south of the inlet mouth represents the flushing capability of tidal currents. Once on the outer edge of influence of the inlet flow regime, the dominant northeast wave and current system moves the flushed heavy minerals back to the beach. Observations of the beach south of the inlet emphasized this process as heavy mineral concentrations increase slightly south of profile J although it appears some heavy minerals are trapped with their equivalent light mineral sizes in 20 to 30 feet of water and moved south outside the breaker zone.

The large surface zone of very fine sediments bisecting profiles E, D, and C, is well sorted with a dominant positive skewness. This suggests a depositional environment possibly close to summer wave equilibrium. The offshore warp in the 60 foot depth contour supports the depositional hypothesis.

The northern area is of interest due to the large southward-trending heavy mineral lobe extending 15 miles downcoast in a southerly direction. The beach system appears to be fed placer minerals from this source as the beach near profile B was observed to contain significantly greater concentrations of heavy minerals than beaches further north.

The origin of this northern lobe is unknown; however, it may be a transported remnant of the large Pleistocene shoal off Cape Henry, Virginia. Should heavy minerals have been deposited on these shoals at lower sea level, they could provide a source which is migrating south today. Because heavy minerals only move when the entire oscillating bed moves, and since light minerals move at a higher rate than heavy particles, the placer grains would tend to form a surface lag deposit, probably of limited depth, but of great areal extent.

A linear region of heavy mineral concentration of less than one percent parallels the present coast from between three and four miles offshore. This is also a trough region implying an erosional bed during a recent sea level rise. Under such conditions there would be little opportunity for relic deposits of any type that had originated in the sediment wedge to be preserved offshore.

CHAPTER V

BEACH PLACERS

INTRODUCTION

Laminar heavy mineral concentrations are common to clastic, non-carbonate beaches of the world. Almost every type of heavy mineral has, at one location or another, been commercially obtained from beach placers. Placer mineral types range from gold, previously mined in littoral terraces at Nome, Alaska, to diamonds, presently being extracted from the beaches of South Africa. In many regions beaches are the only marine environment containing economic quantities of heavy minerals. For instance, only the beaches of Cornwall, England, with the possible exception of a few offshore stream extensions, or relic strands, are believed to contain significant tin to warrant production there (Hill and Parker, 1970).

Contemporary beach placers, especially those with heavy minerals in the 3.5 to 6.9 grams per cubic centimeter range, are often referred to as stormline deposits. "Stormline" signifies the sorting and accumulation of placer minerals at the landward margin of the backshore in response to extreme storm conditions (Rao, 1957). Bradley (1957), one of the many authors who have described beach placers, observed present-day stormline concentrations at the seaward base of active dunes on Mustang Island, Texas. Pirkle and Yoho (1970) discussed the relic placer deposits of Trail Ridge, Florida, and interpreted this titanium, monazite, and zirconium ore body as a stormline concentration formed during a Pleistocene sea level regression. Another, and much older beach deposit described by Huston and Murphy (1962), is an elongate and lenticular indurated ancient beach placer containing potentially economic amounts of zirconium and hafnium. This zirconium-rich depositional unit is a member of a clastic Cretaceous formation in the Rocky Mountain region.

The advantages of mining placers on modern beaches include (1) the lack of sediment cover, (2) ease of access, and (3) the significant concentration of heavy minerals resulting from the present sea level stillstand. Often these advantages are negated by the high recreational land value placed on many beaches. Although an educated prediction of placer location on present beaches is somewhat academic, an understanding of active processes leading to their formation and preservation may warrant extrapolation to past beach environments for the purpose of locating and delimiting buried placers.

The objective of my discussion of beach placers is to define active placer-concentrating mechanisms. Ultimately, this knowledge should enable an explorationist to predict optimum placer locations and the probable three-dimensional extent of relic deposits in relation to hydrodynamic and geologic variables. In order to attain this objective, natural beach process and sedimentary responses were observed on four Dare County beaches, the results of which have provided the basis for establishing a model for beach placer formation.

FIELD INVESTIGATION

A necessary beginning in establishing the significant processes involved in creating a beach placer is to describe first the sedimentary characteristics and heavy mineral accumulations on clastic sand beaches. The flow regime responsible for the formation of a beach is complex and unsteady, and for this reason difficult to measure and even more difficult to treat analytically. Although the fluid dynamics involved in placer formation provide the actual mechanism of genesis, it is easier to anticipate these processes by concentrating on the results, namely the sediments themselves.

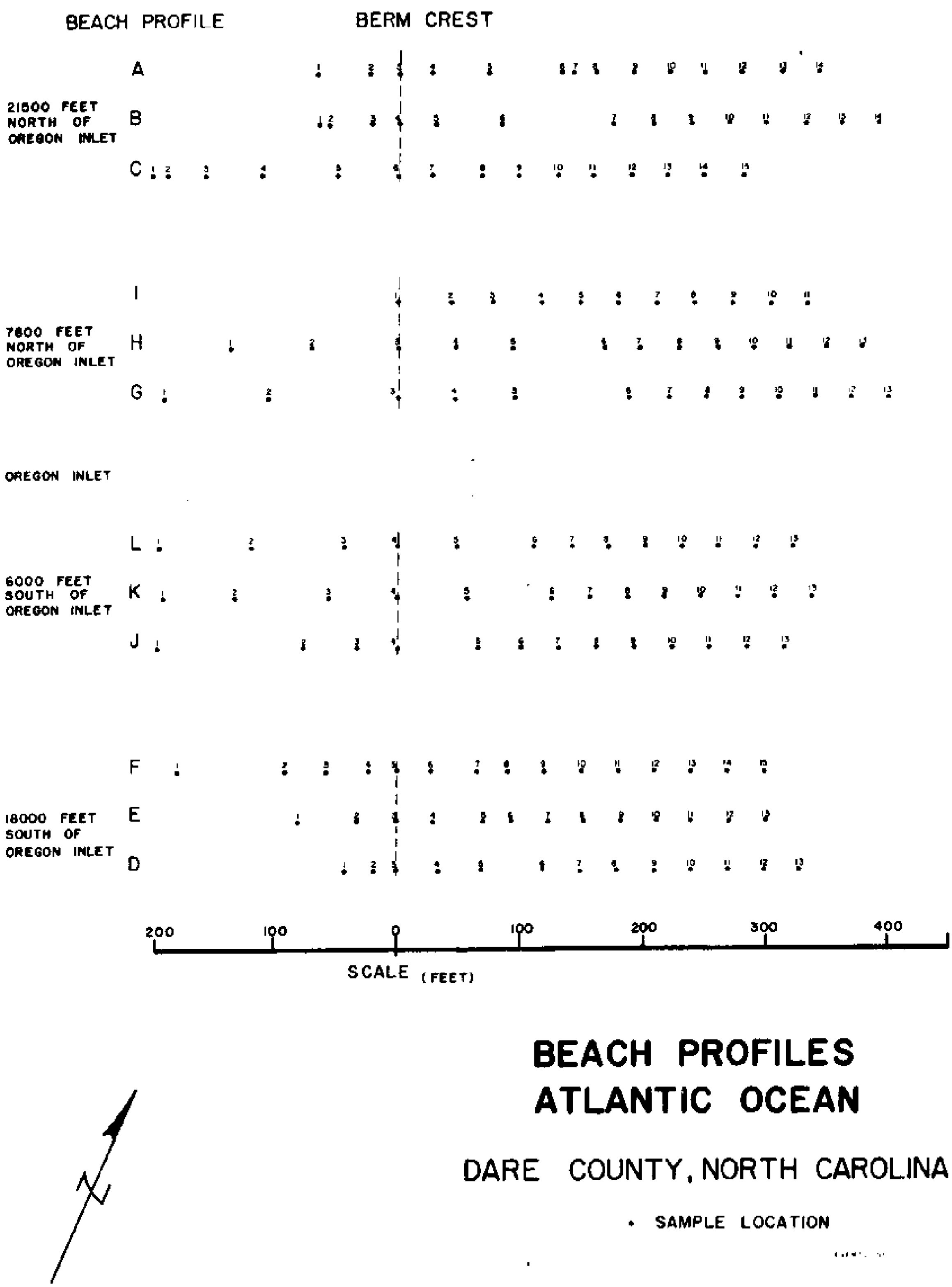
All beach placers are formed as a result of predominantly the same variables transformed in the same manner, although, perhaps, in varying magnitudes. Beach placers are created above the high tide line, consequently their formative processes are alike regardless of whether the coastline is advancing or retreating. Sea level fluctuations are, however, important in the ultimate preservation and, thus, the location of the placer.

Location

The locations of field survey beaches reported in this study are shown in Figure 1-1. Note that two beaches are north and two are south of Oregon Inlet, each is approximately 2.5 miles from its neighbor. Beach sites were located in this fashion in order to define the influence of inlet processes on the nearby coast. A second, and less important reason for choosing these locations, was their accessibility for study.

The northern beaches, represented by profiles A, B, C, I, H, and G, are little influenced by the inlet due to the predominantly southward moving littoral drift in this region (Villianos, 1967). The next breach in the barrier to the north is the entrance to Chesapeake Bay, approximately 90 miles distant. Beaches north of Oregon Inlet should be as nearly ideal as could be expected in nature, influenced only by shoaling gravity waves and the resultant longshore drift. All four beaches are as free as possible from undefinable offshore influences. Therefore, inferences concerning placers based on these beaches should be applicable, with some modifications, to most clastic beaches.

Figure 5-1


BEACH PROFILE
BERM CREST
A
B
C
21500 FEET NORTH OF OREGON INLET
I
H
G
7800 FEET NORTH OF OREGON INLET
OREGON INLET
L
K
J
6000 FEET SOUTH OF OREGON INLET
F
E
D
18000 FEET SOUTH OF OREGON INLET
200
100
0
100
200
300
400
SCALE (FEET)
BEACH PROFILES ATLANTIC OCEAN
DARE COUNTY, NORTH CAROLINA
• SAMPLE LOCATION
N

Sampling Procedure

Three sediment sample and elevation profiles were made normal to the coastline at each beach (Figure 5-1) and spaced 1500 feet apart. All three profiles on an individual beach were obtained during a single day to lessen the undesirable influence of changing wave and surf conditions. The three profiles provide a cross-check on survey and sediment sample uniformity parallel to the shore. All beach sampling was done during the first ten days of August 1969. A total of 160 sand samples was obtained.

The region encompassing the backshore, berm, foreshore, and a portion of the surf zone was surveyed by transit and elevation rod. Distances were obtained by stadia. The berm crest elevation was taken as the zero reference base for both horizontal and vertical control. This was the only point common to each beach and individual profile.

The summer berm represented the upper beach response to wave conditions at the time of these surveys. The backshore, especially that portion near the coastal dunes, was generally relic of previous summer storm events.

Sediment samples were obtained of the upper two centimeters of surface material. Ideally, sampling should be of the immediate surface grains only as this would ensure a sample of grains deposited during a single dynamic event. Sampling limitations, particularly in the surf and breaker zones, preclude such precision, nevertheless the samples are considered representative for such a study as this one. Samples taken to a depth of two centimeters assume the surface bed layer is modified to this depth or deeper during a tidal cycle. This assumption is most applicable seaward of the berm (Schwartz, 1968).

Samples, and elevations relative to the summer berm reference base, were taken in the surf and breaker zones by a swimmer. The transit located on the berm crest, and a stake with its horizontal and vertical position established seaward of the zero reference base, served as a lead line. A polyethylene line, with floats fastened at ten foot intervals, was carried seaward by the swimmer. The floats provided the necessary bouyancy to keep the line visible at all times thereby ensuring a straight line as well as accurate distance measurements at the time each sample and depth was taken. Sediment samples and water depths were obtained at thirty foot intervals away from the stake. The swimmer's distance was regulated by line control from the stake. Depths were obtained by sounding with a lead line, and sediment samples were taken of the upper two centimeters of bottom by the swimmer. Ripple orientation and the geometrical characteristics of the bed were also obtained by the swimmer. Following the collection of each sample the swimmer returned to shore, recorded the depth and bottom configuration, and assisted in packaging the sample. He then returned to occupy the next sample station. Individual collection procedures required three to six minutes depending on the distance from shore. An entire profile required an average of one and one-half hours to complete.

Sea Conditions - Beach Zone

The dominant direction of wave approach in the study beach region is from the northeast (Figure 5-2). Average breaker heights at Nags Head, North Carolina, approximately ten miles to the north, is four feet (Helle, 1958). These data were collected daily over a period of several years by coastguardsmen stationed at the Nags Head Lifeboat Station. High surf conditions are primarily associated with northeasterly storms which accounts for the large net southerly littoral drift in the area (Villianos,1967).

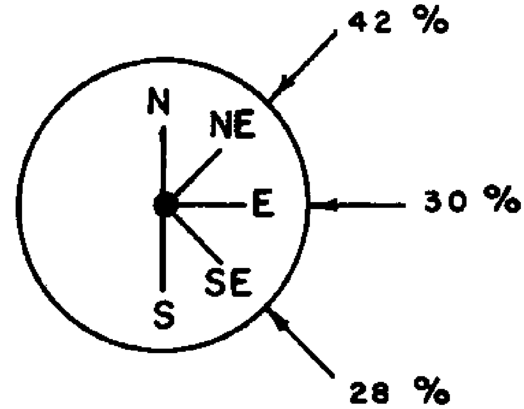
The maximum recorded tide at Norfolk, Virginia, 90 miles north of Oregon Inlet, was 7.4 feet above mean sea level (Harris, 1958). This was registered for a period of six hours during the hurricane of August 23, 1933. At Okracoke, North Carolina, about the same distance south of Oregon Inlet, the maximum recorded tide was 5.8 feet. We assume from these data that storm surge along the Oregon Inlet coastal region should reach five feet or more during intense storms. Between 1905 and 1955 Dare County suffered coastal damage as the result of 20 storms. In some areas high water reached a maximum 15 feet above mean sea level (Harris, 1958). The duration of high water was generally less than six hours, although in areas of poor drainage it persisted much longer.

The scale of shore erosion is critical in approaching preservation possibilities for heavy mineral laminae formed on various regions of the beach. Although a placer is created under depositional conditions, it will retain its integrity only while it remains erosion-free. Schwartz (1968) defined various degrees of erosion from small scale swash scour to long term sea level changes. Rao (1957) observed a two foot seasonal sea level change on the eastern coast of India and noted placer formation and preservation to be the result of storm conditions and spring tides during the high water period of the seasonal cycle. The erosive power of storm surges coupled with large steep waves and possible wave focusing along sections of the North Carolina coast result in all near-surface sedimentary deposits located seaward of the coastal dune complex being solely transitory features.

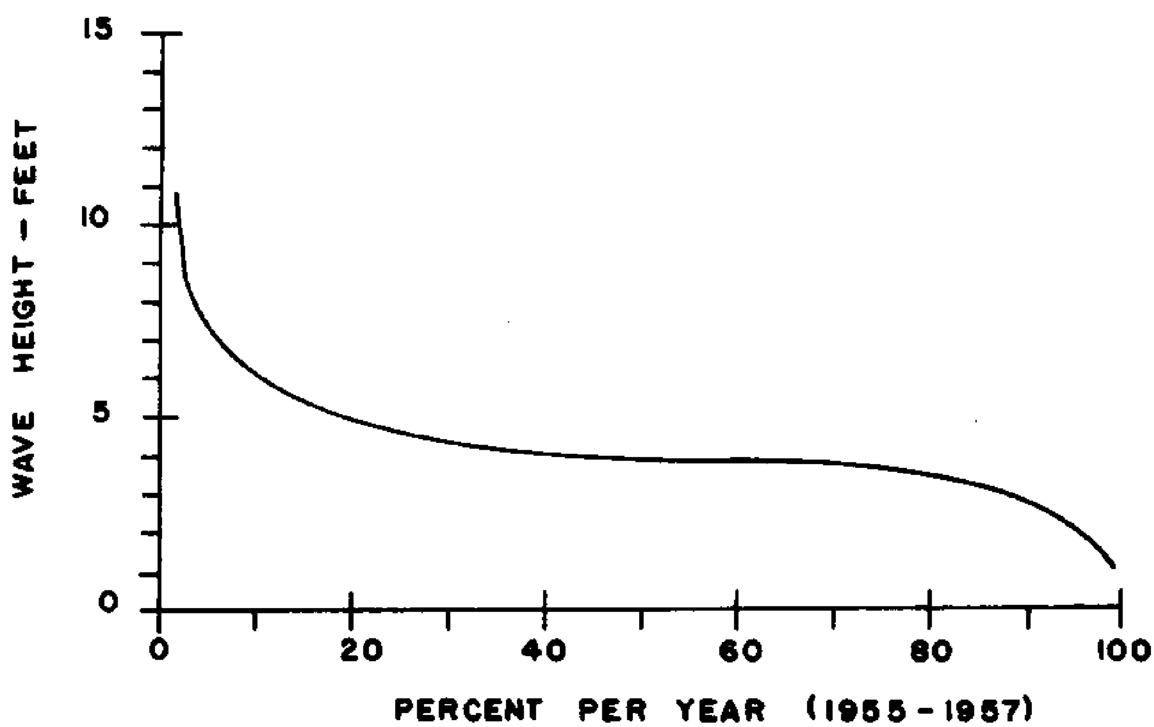
Zeigler et al. (1959) observed the effects of nine storms and two hurricanes on beaches in the Cape Cod area. They noted that storms arriving on the coast at the time of spring tides do considerably more cutting than other storms, and that all storms shift and rearrange large amounts of sediment. Figure 3-2 illustrates the change in a beach profile in response to moderate long wave sea conditions and steep storm wave conditions.

The only surface region on an active beach that is not modified during seasonal changes is the winter berm at the landward edge of the storm profile. This is the unique beach location: it retains its character from year to year. It is also the location of stormline placers observed on Dare County beaches. During an extreme event, such as a hurricane surge, the winter berm and coastal dune line may be moved landward.

Figure 5-2



TIME PERCENTAGE OF WAVE APPROACH DIRECTION
NAGS HEAD, NORTH CAROLINA



CUMULATIVE FREQUENCY OF SURF AT
NAGS HEAD, NORTH CAROLINA
(FROM HELLE, 1958)

Sediment dispersal patterns and beach profiles discussed in the following sections are sympathetic to waves of eight to ten second period and a two to four foot breaker height. Dispersal patterns represent the sediment response to moderate wave conditions. Tides recorded at a Coastal Engineering Research Center gage at Jeanettes Pier in Nags Head indicated a two-foot diurnal variation during the beach sampling program.

Beach Topography

Figure 5-3 illustrates the topography of the four study beaches. The berm crest is taken to be a reference base of zero elevation. All elevations shown seaward of the base signify vertical distances less than the berm elevation. Contours landward indicate elevations above the berm crest. Not illustrated because of the two foot contour interval, is an ever-present backshore trough, 0.5 to 1.5 feet lower than the berm crest. The trough is the static result of summer berm upbuilding. The berm slope is convex upward indicating aggrading conditions (Miller and Zeigler, 1958).

Included in Figure 5-3 is the average zero water depth contour taken at the time of the survey. This contour is commonly six to eight feet beneath the summer berm crest. In most instances surveys were made during low tide conditions in order to reach a greater offshore profile distance.

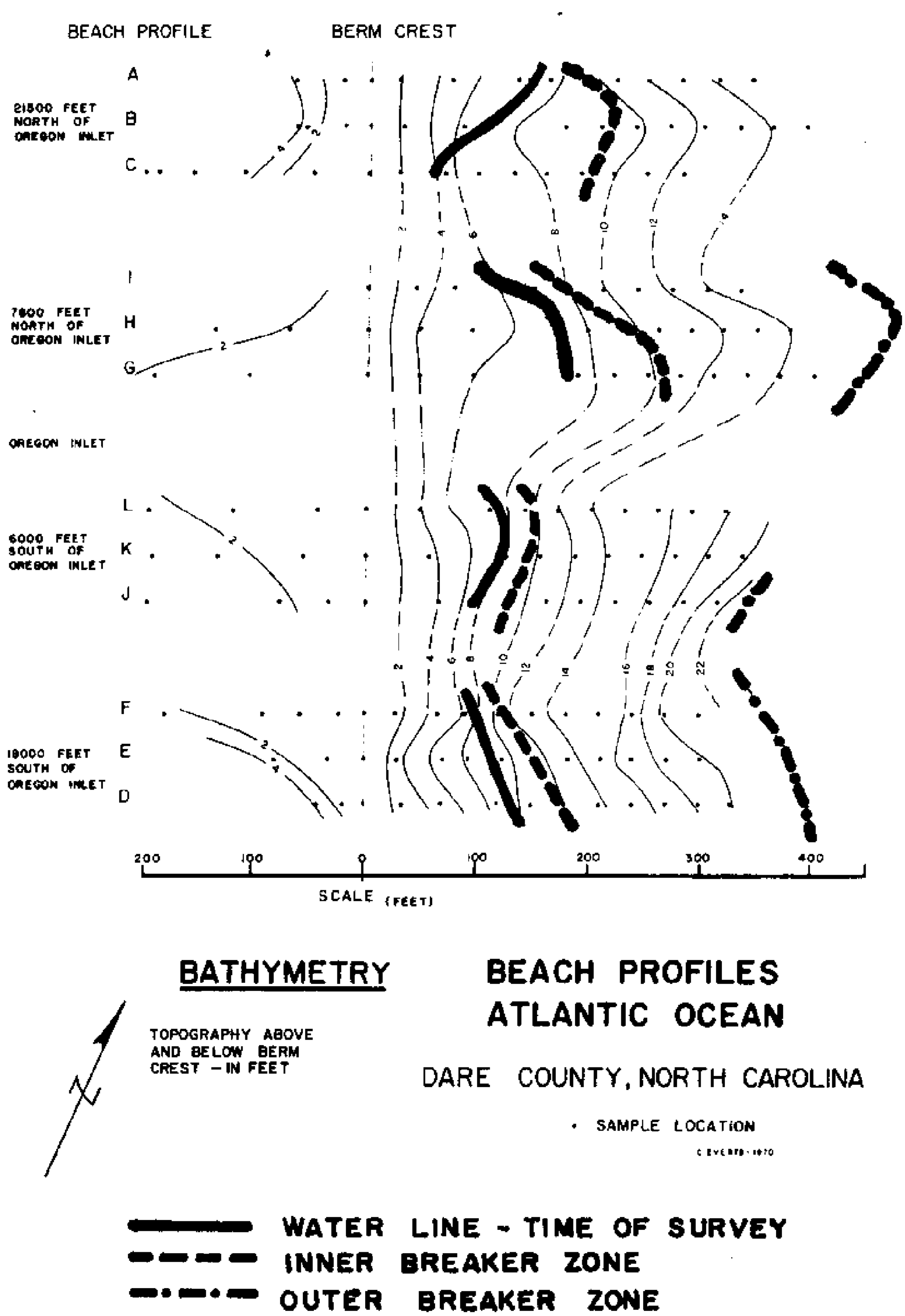
The dashed line indicates the location of inshore plunging breakers while the alternately dashed and dotted line marks the region of outside spilling breakers. The latter breaker zone was not present during surveying operations on profiles A, B, C, L, and K. Out side breakers generally spilled over offshore bars projecting to within six to eight feet of the water surface.

Beach gradients north of Oregon Inlet are approximately 0.04ft/ft while the two southern beaches are much steeper with slopes averaging 0.07 ft/ft. Predominant southward-moving longshore wave activated currents do not account for this topographic anomaly because wave and littoral drift conditions on all beaches were similar at the time of each survey. The ebb-tidal flow from the inlet throat was very strong in both a northerly and southerly direction parallel to the coast, while the tidal flood enters Oregon Inlet from the north and east. Flood and ebb currents on northern beaches act in opposition over a complete tidal cycle resulting in little net tidal scour on these beaches. In the south, however, the ebb current, superimposed on the wave-activated littoral drift, is sufficient to scour the beaches. Maximum flow velocities of 1.2 m/sec were observed moving in a southerly direction across profiles L, K, and J.

Sediment Distribution

The dynamic processes active within the breaker, surge, foreshore, berm and backshore regions of a beach are reflected in the sediment textural values obtained from analysis of samples collected at each of these locations. An understanding of the total sediment distribution picture is valuable in placing limits on the regions of heavy mineral transport and deposition.

Figure 5-3


BEACH PROFILE
BERM CREST
A
B
C
21500 FEET NORTH OF OREGON INLET
I
H
G
7600 FEET NORTH OF OREGON INLET
OREGON INLET
L
K
J
6000 FEET SOUTH OF OREGON INLET
F
E
D
19000 FEET SOUTH OF OREGON INLET
200
100
0
100
200
300
400
SCALE (FEET)
BATHYMETRY
TOPOGRAPHY ABOVE AND BELOW BERM CREST - IN FEET
N
BEACH PROFILES
ATLANTIC OCEAN
DARE COUNTY, NORTH CAROLINA
• SAMPLE LOCATION
WATER LINE - TIME OF SURVEY
INNER BREAKER ZONE
OUTER BREAKER ZONE

The most striking features of Figure 5-4, illustrating median sediment diameter, are the linear trends parallel to the coast. They indicate the dominance of shoaling gravity waves on sediment dispersal patterns.

The outside breaker zone of profiles I, H, G, J, F, E, and D is reflected in a coarse (greater than 0.3 mm) sediment layer near the offshore limit of the survey profiles. The inner plunge zone is an exact image of this coarse trend. Lack of a coarse layer at those seaward profile margins without an outside breaker line illustrates the rapidity by which the sediment surface of a beach responds to a dynamic environment. In the case of profiles K and L, breakers were observed in the offshore regions within a 24-hour period preceding the survey. The slight coarsening of sediments in some of these regions also indicates past breaker activity. This is especially true of the outer margins of profiles A, B, and C.

Sorting values in the breaker zones are large, indicating sediment mixing (Figure 5-5). Skewness is generally negative (Figure 5-6), and suggest an absence of fine-grained material.

Longshore sediment drift observed in the breaker regions during the sampling program was dependent on the angle of wave attack. Particles of all sizes were lifted in suspension, sometimes reaching the surface immediately after a wave broke. The amount of horizontal sediment movement was dependent on the settling velocity of the grains. Larger particles settled rapidly while the finer grains tended to move away from the breaker zone toward a less turbulent region. Fine sand often moved in an offshore direction. Although the largest maximum surge velocity in the breaker region was observed to be in a shoreward direction, the less intense seaward surge was of longer duration. This was observed to favor net offshore transport of highly suspended particles. Coarse sand was observed to migrate landward and seaward with the breaker zone in response to tidal fluctuations and varying sea conditions, such as variations in wave period or wave height.

The surge, or surf, region between breaker zones exhibits a very small median diameter, generally less than 0.20 mm. These are well sorted fine-grained sands with a slightly more positive skewness value than for those found in the breaker zones. Ripples observed in the surge zone were small with current ripple indexes (Inman, 1957) of from three to four. The surge zone was the only area of the entire beach consistantly surfaced with ripples. The bottom seaward of the outer breakers was also ripple covered, but it is defined here as a part of the offshore zone. The fact that surge zone ripples were asymmetrical indicated the translational flow nature of the region. Ripple configuration generally suggested a net shoreward sediment transport. The opposite was observed to be true under storm conditions when strong bottom currents satisfy mass continuity in removing wave-ponded water from the foreshore. Heavy minerals were not observed in ripple troughs or elsewhere in the surge zone. Most sediments were perceived to move across this section by saltation and bed creep.

Figure 5-4

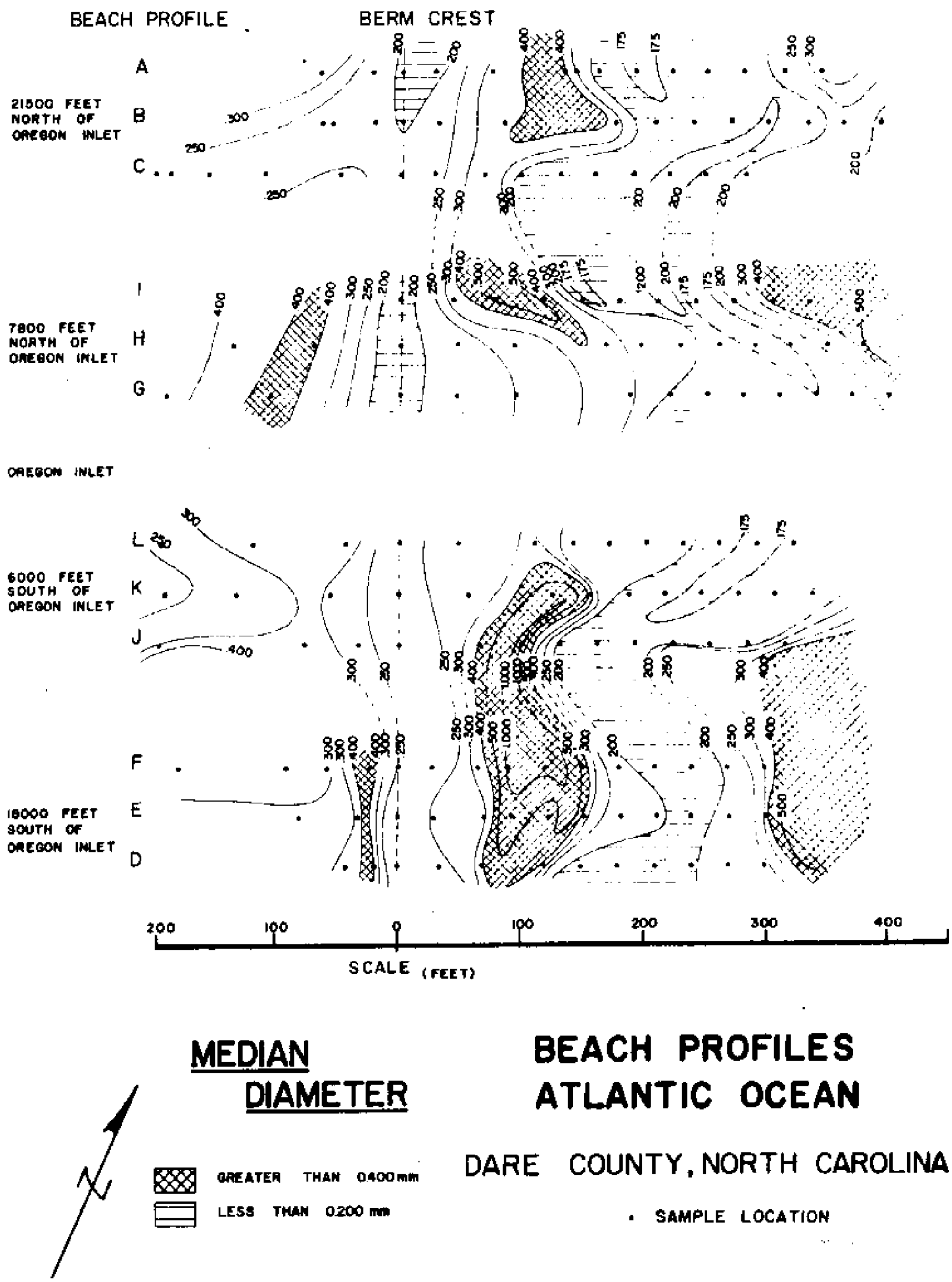

BEACH PROFILE
BERM CREST
21500 FEET NORTH OF OREGON INLET
7800 FEET NORTH OF OREGON INLET
OREGON INLET
6000 FEET SOUTH OF OREGON INLET
18000 FEET SOUTH OF OREGON INLET
A
B
C
I
H
G
L
K
J
F
E
D
200
100
0
100
200
300
400
SCALE (FEET)
MEDIAN DIAMETER
GREATER THAN 0.400 mm
LESS THAN 0.200 mm
BEACH PROFILES ATLANTIC OCEAN
DARE COUNTY, NORTH CAROLINA
SAMPLE LOCATION

Figure 5-5

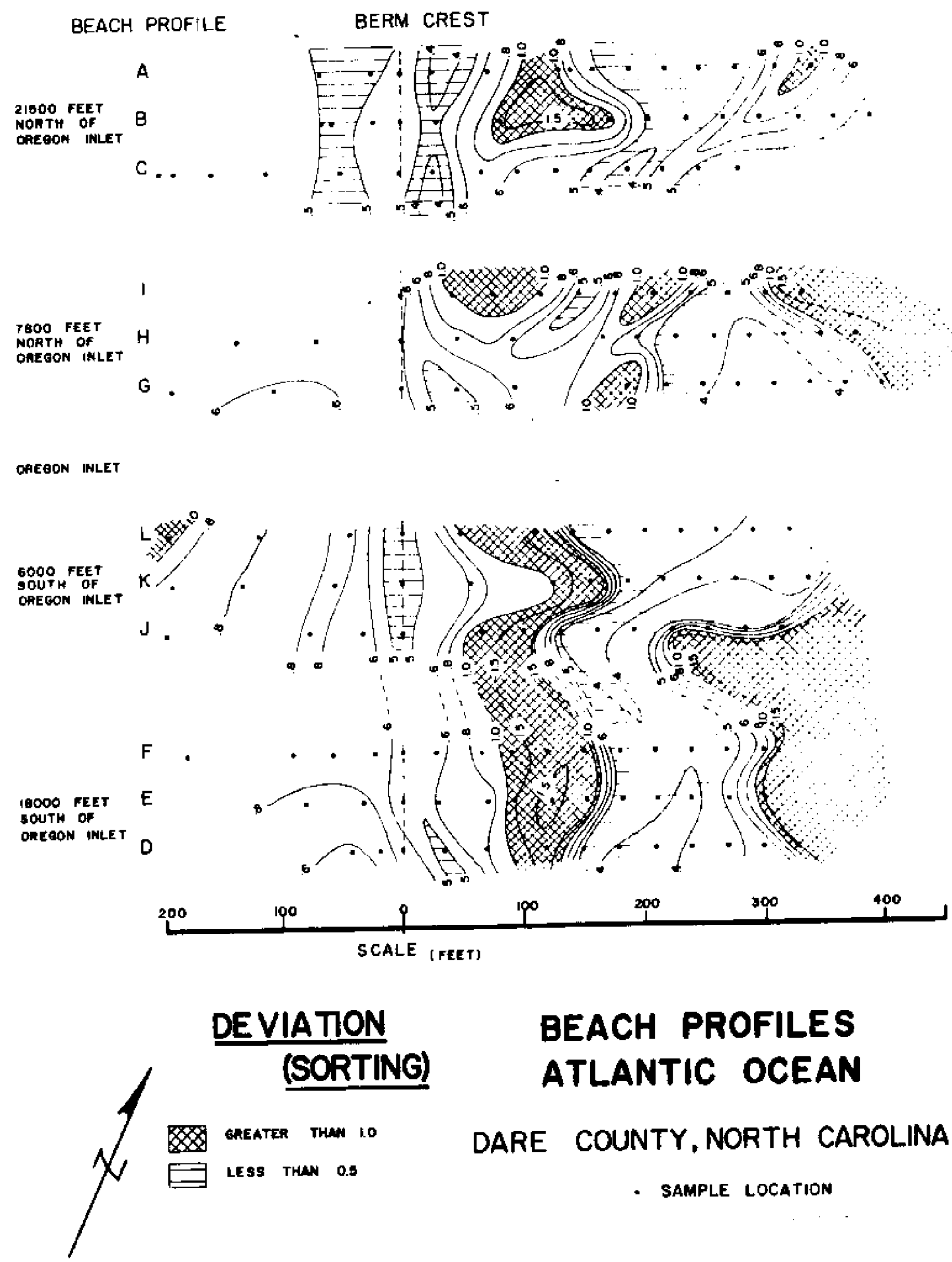
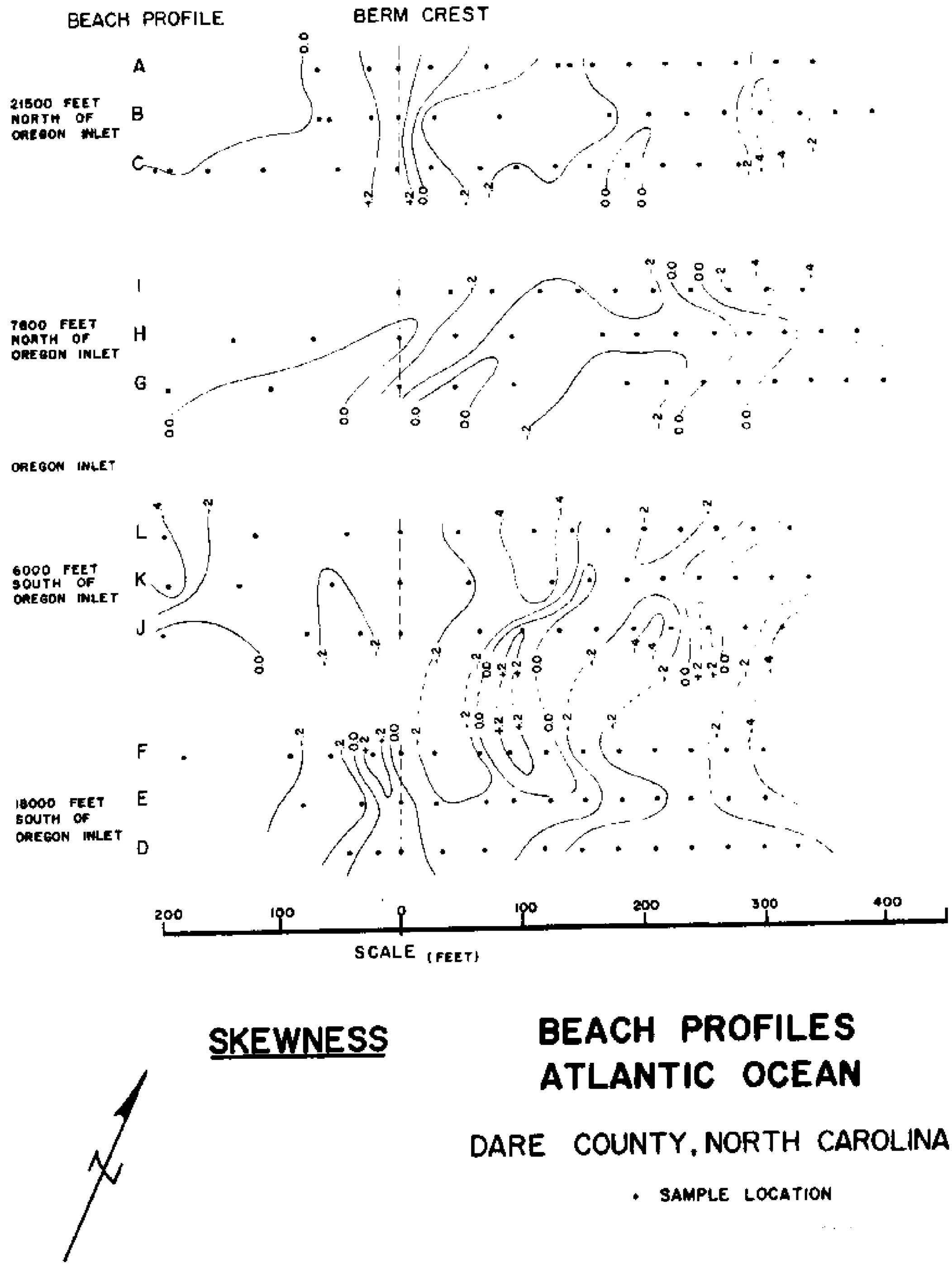

BEACH PROFILE
BERM CREST
A
B
C
21500 FEET NORTH OF OREGON INLET
I
H
G
7800 FEET NORTH OF OREGON INLET
OREGON INLET
L
K
J
6000 FEET SOUTH OF OREGON INLET
F
E
D
18000 FEET SOUTH OF OREGON INLET
200 100 0 100 200 300 400
SCALE (FEET)
DEVIATION (SORTING)
GREATER THAN 1.0
LESS THAN 0.5
BEACH PROFILES
ATLANTIC OCEAN
DARE COUNTY, NORTH CAROLINA
• SAMPLE LOCATION

Figure 5-6



BEACH PROFILE
BERM CREST
A
B
C
21500 FEET NORTH OF OREGON INLET
I
H
G
7600 FEET NORTH OF OREGON INLET
OREGON INLET
L
K
J
6000 FEET SOUTH OF OREGON INLET
F
E
D
18000 FEET SOUTH OF OREGON INLET
200
100
0
100
200
300
400
SCALE (FEET)
SKEWNESS
BEACH PROFILES
ATLANTIC OCEAN
DARE COUNTY, NORTH CAROLINA
• SAMPLE LOCATION
N

The foreshore region displays a uniform decrease in median particle diameter from the point of zero water depth to the berm crest. Sorting improves upslope (from 1.2 to 0.4) while skewness changes in value from -.2 to +.2.

The median diameter of sediment at the berm crest is the smallest of any region above mean sea level, including coastal dunes. The berm crest is also the beach location of best sorting and most positive skewness values.

Median diameter increases in the backshore region to a high located between the berm crest and the coastal dunes. Sorting values increase toward the dunes as skewness becomes slightly negative. Coarse sands, granules, shells, and driftwood were concentrated at the base of the dunes. The dune face itself often contained heavy mineral laminae. Sediments in the far backshore region represent the distributive results of storms, not the moderate sea state extant during the survey.

Dynamic Beach Processes - Sediment Responses

A trend summation of the various textural parameters for a typical summer beach is shown in Figure 5-7. The trends are based on a composite of all 12 beach profiles studied.

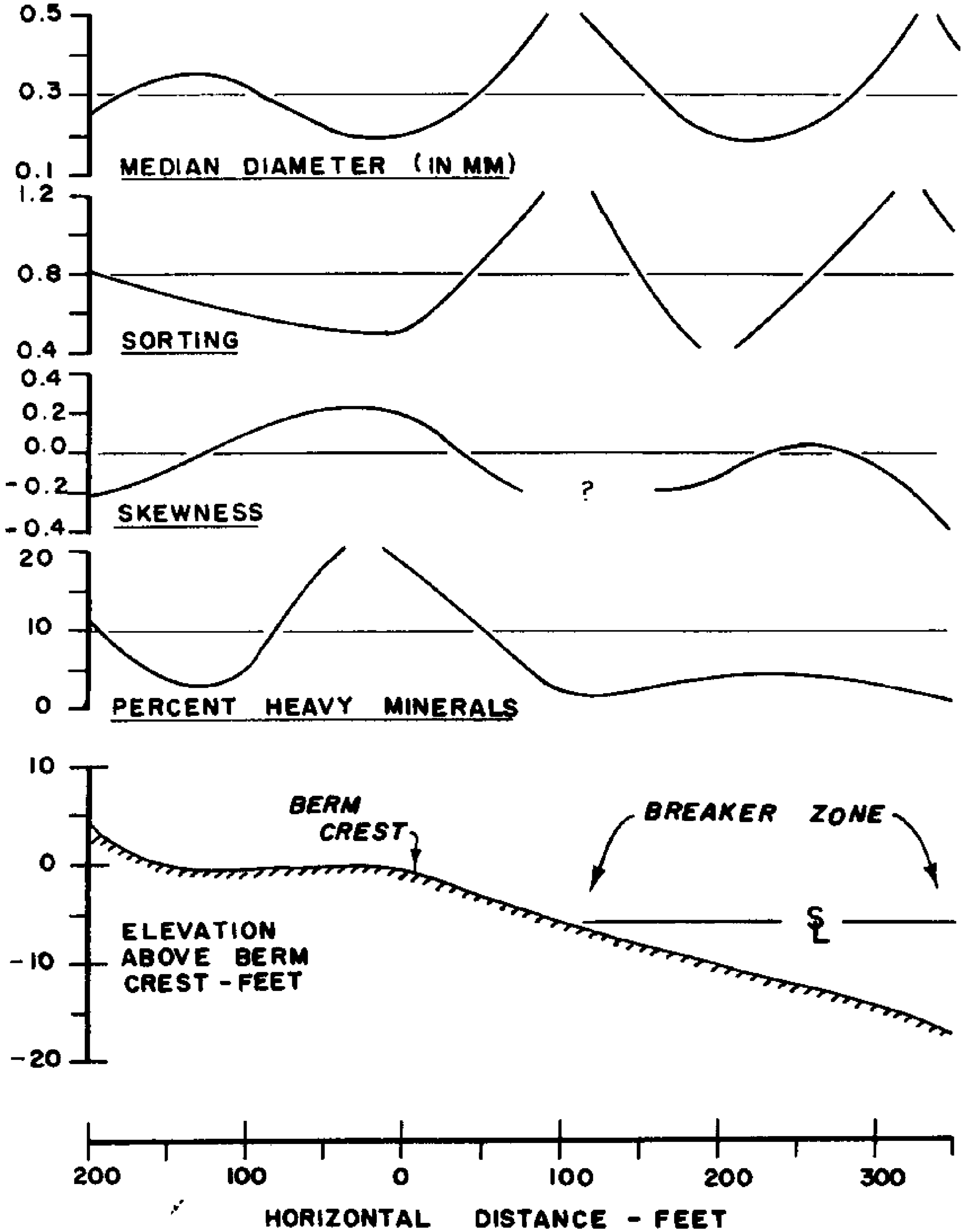
Breaker Zones

The outer breaker zone is characterized by a very coarse, poorly sorted, negatively skewed sand. This is largely an area of a suspended mode of net transport where finer fractions of the total sediment are moved either into the surge zone or offshore. The time duration of particle suspension is dependent upon local turbulence and the settling velocities of individual grains. Larger, more rapidly settling particles, remain beneath the breakers and migrate with the breaker zone. During storms even the coarsest grains may be moved seaward.

Poor sorting evidenced in the breaker zone sediments emphasizes the unsteadyness of the region while the negative skewness values reflect the relative absence of grains significantly finer than the median diameter of the total sediment. The absent fine grains are those moved through the zone in suspension, but not deposited.

The net bed load transport direction under normal conditions was observed to be shoreward through the breaker regions. Suspended fine or very-fine-grained sands moved either landward or seaward. The direction of heavy mineral transport was onshore because placer grains do not remain in suspension during the outward flow component beneath a breaking wave. Placer minerals would, however, be expected to move seaward during storms in association with medium and coarse sands. All breaker zones are lateral conduits for the transport of large quantities of suspended sediment when a longshore drift is superimposed on a breaking flow regime (Ingle, 1966; Duane, 1970).

Figure 5-7



TEXTURAL PARAMETERS - TYPICAL SUMMER BEACH

Surge Zone

The summer surge zone, located between breaker zones or between the inner breaker zone and the foreshore, represents a region of fine (0.15 to 0.20 mm), well sorted sediment. The predominant mode of transport in this region was observed to be bed creep and saltation. Suspended sediment moving in from a breaker region tended to settle in this zone because of the relative lack of bottom turbulence. The low bed turbulent intensity was further manifested by the bottom compactness in contrast to the loose bed observed beneath breaking waves. Coarse particles often move between breaker zones by rolling across this hard smooth bed.

The surge zone is a trap for fine suspended grains moving in from offshore and out following foreshore and berm erosion. It is also an area of low net longshore transport when compared to the breaker and foreshore regions (Duane, 1970). Heavy minerals moving shoreward in association with medium and coarse-grained light minerals will tend to be trapped if the surge zone is aggrading. Because breaker zones continually migrate across the nearshore beach region, however, placer grains will be preferentially carried to the foreshore. This precludes selective deposition of heavy minerals in the surge zone.

The inner breaker region is characterized by much the same conditions existing in outer breaker zones. The nearshore breakers pass heavy minerals to the foreshore, but in many cases will not move the same placer minerals seaward again.

Foreshore

Foreshore and berm regions are of special interest because they are zones of heavy mineral deposition on the summer beach. The basic mechanism of summer and winter foreshore deposition are the same, only the depositional sites differ.

In many instances a step forms at the base of the foreshore when the seaward terminus of the foreshore is also the location of the inner plunge zone (Figure 5-3). The step is important in sorting the sizes available for movement up the foreshore.

In an upslope direction the foreshore is a region of uniformally changing textural parameters; the median particle size decreases, sorting values decrease, and skewness becomes positive. Maximum uprush velocities observed near the midpoint of several summer foreshores were 220 cm/sec. Beginning at the base of the foreshore, an average upslope decrease in water depth of 0.3 in/ft was noted. The depth decrease appeared to be dependent upon percolation losses, the foreshore slope, and the midpoint velocity. The maximum uprush velocity was observed approximately one-third the way up the foreshore.

Backrush maximum velocities were observed at the step. The maximum backrush value at the midpoint was 90 cm/sec. The backrush flow depth decreased seaward, generally between 0.05 and 0.15 in/ft. Uprush flow depth at the foreshore midpoint was usually between four and six inches, which was always two or three times the backrush depth and oftentimes much greater.

Variation in median grain diameter on the foreshore is a partial function of the foreshore slope. The change in M_d per upslope surface foot on beaches south of Oregon Inlet is -0.0075 mm/ft on a slope of 0.08 ft/ft. On northern beaches with a slope of approximately 0.06 ft/ft, the change in M_d is -0.0015 mm/ft, or one-fifth the gradient of that in the north. The textural range of sediments available is the same on all beaches. The steeper slopes, therefore, concentrate and segregate given particle sizes in smaller zones normal to the direction of swash. This attribute of the slope is important for defining and delimiting beach placers.

The mode of sediment transport on the foreshore surface changes in an upslope direction. At the lower step most of the moving sediment load is in suspension. As the fluid begins movement upslope, the largest suspended particles return to the bed. Because the foreward flow velocity increases from essentially zero at the step to a maximum at a point from one-third to one-half the way upslope, the coarser particles on reaching the bed near the step remain immobile and do not move as bed load. Further upslope, saltation, then creep, becomes the dominant transport mode as flow velocity increases and depth decreases. As suspension gives way to saltation, progressively smaller particles may be trapped in the bed during uprush, however, in each case it is the largest grains which are lost first. Uprush scour exceeded deposition during most foreshore observations suggesting backwash as the more important foreshore depositional agent.

The backwash process is the reverse of uprush deposition because the backwash flow regime deposits fine particles first. The deposit becomes coarser in a seaward direction because the larger grains bypass the upper fine-grained smooth bed.

Shear on a bed particle is basically a function of the surface area of the grain exposed to the flow, the flow velocity acting on the particle, and the flow depth. As the flow depth decreases the velocity gradient near the bed increases and shear on the particle surface increases. In a very shallow flow regime the difference in surface shear between various sized particles resting or moving on the boundary may be much greater than the resisting force acting to deposit them. Depositional bed shear (Figure 2-9) was observed to be 90 percent of the critical bed shear for both heavy and light grains. The maximum bed shear required to keep a particle in motion was established under conditions of depositional equilibrium where resisting forces balanced shear forces (Figure 2-8).

Because the bed shear required to initiate motion in a uniform bed for particle sizes in the foreshore range of 0.15 to 0.50mm is similar (Figure 2-7), the main criterion of the bed to pass larger sizes is its smoothness. Figure 2-8 illustrates the bypassing efficiency of uniform beds. The uniform light mineral bed will pass light minerals up to nine times the diameter of the bed grains. A heavy mineral bed will pass an even greater range of light mineral sizes. Grain sizes smaller than the bed size tend to concentrate in bed interstices.

Figure 5-8 shows the increase in shear on a light mineral particle as depth decreases. The light mineral size is given on the horizontal scale, particle shear on the vertical scale, and the four curves represent depths ranging from one meter to 0.1 cm. The bed is a 0.10 mm heavy mineral boundary with a 4.20 gm/cc density with a critical shear of 1.80 dynes/cm^2. Flow velocity was regulated to maintain the critical shear on the bed. This figure graphically illustrates the increase in shear on large particles in shallow flow regimes as they cross smaller uniform beds. It should be noted that the sediments on the foreshore are quite uniform.

The main point to note here is that as the bed becomes finer it becomes smoother, and larger particles will roll, as bed creep, with relative ease across its surface. Also, as flow depth increases, shear on the larger moving particles may become large with respect to shear acting on the bed, thus the bed may remain immobile as larger light mineral grains bypass it.

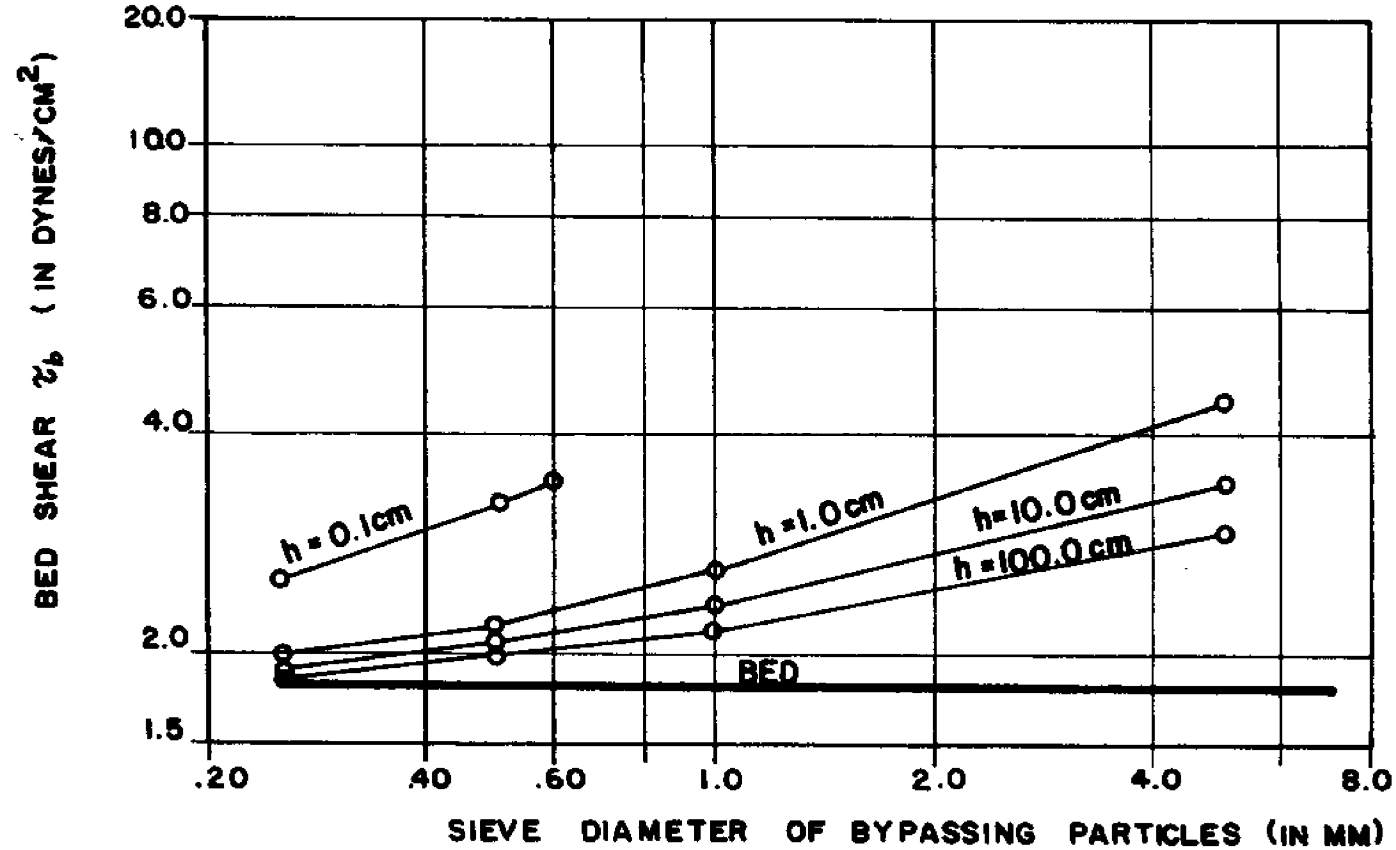
The upslope paucity of grain sizes in excess of 0.25 mm is reflected in the low sorting values in this region. Many fine particles are trapped during the swash cycle and fill interstices in the lower foreshore boundary, accounting for its poor sorting. As the bed becomes finer, the size range of particles lost in interstices decreases and sorting improves. It should be noted there is a finite minimum medium size available on the foreshore: about 0.175 mm. Sizes smaller than this have all been moved offshore in suspension, if indeed they were ever present on the beach.

The progressively more positive skewness values in the upslope direction reflect the loss of large particle sizes near the step. Skewness values are representative of the sizes lost prior to a given depositional site as well as the range of sizes available at the site and the particles actually lost there.

Coarse grains in the swash regime are either moved to the step with backwash flow, or moved over the berm during uprush to be deposited on the backshore. In either case, bypassing plays a key role in maintaining a uniform fine sediment on the upper foreshore and berm.

The foreshore dramatically illustrates the importance of transport modes in sediment dispersal. In suspension, the largest grains, those with the highest settling velocities, are deposited

Figure 5-8



SHEAR ON A BYPASSING LIGHT MINERAL GRAIN
(HEAVY MINERAL BED)

first. As transport becomes dominantly rolling or sliding, the larger grains may move past a finer bed. In the foreshore and berm case, the mode of transport regulates deposition.

Berm Crest Area

Moving landward, the median grain diameter decreases over the foreshore and reaches a minimum at the berm crest. This is the location of lowest flow velocity, and importantly, the smallest flow depth. The berm crest is the most effective bypassing region on the beach. The bypassing mechanism accounts for small sorting values and highly positive skewness. Skewness is positive on the berm crest because bed shear decreases and velocity decreases, while shear on bypassing grains often increases as depth decreases. The lower bed shear enables particles finer than those comprising the bed to be deposited in boundary interstices.

Backshore

The summer backshore is a trap for berm bypassed material. This accounts for its larger median sizes with respect to the berm. Since the sizes and amounts of sediment passing the berm vary as wave conditions change, the backshore depositional area becomes poorly sorted and more negatively skewed.

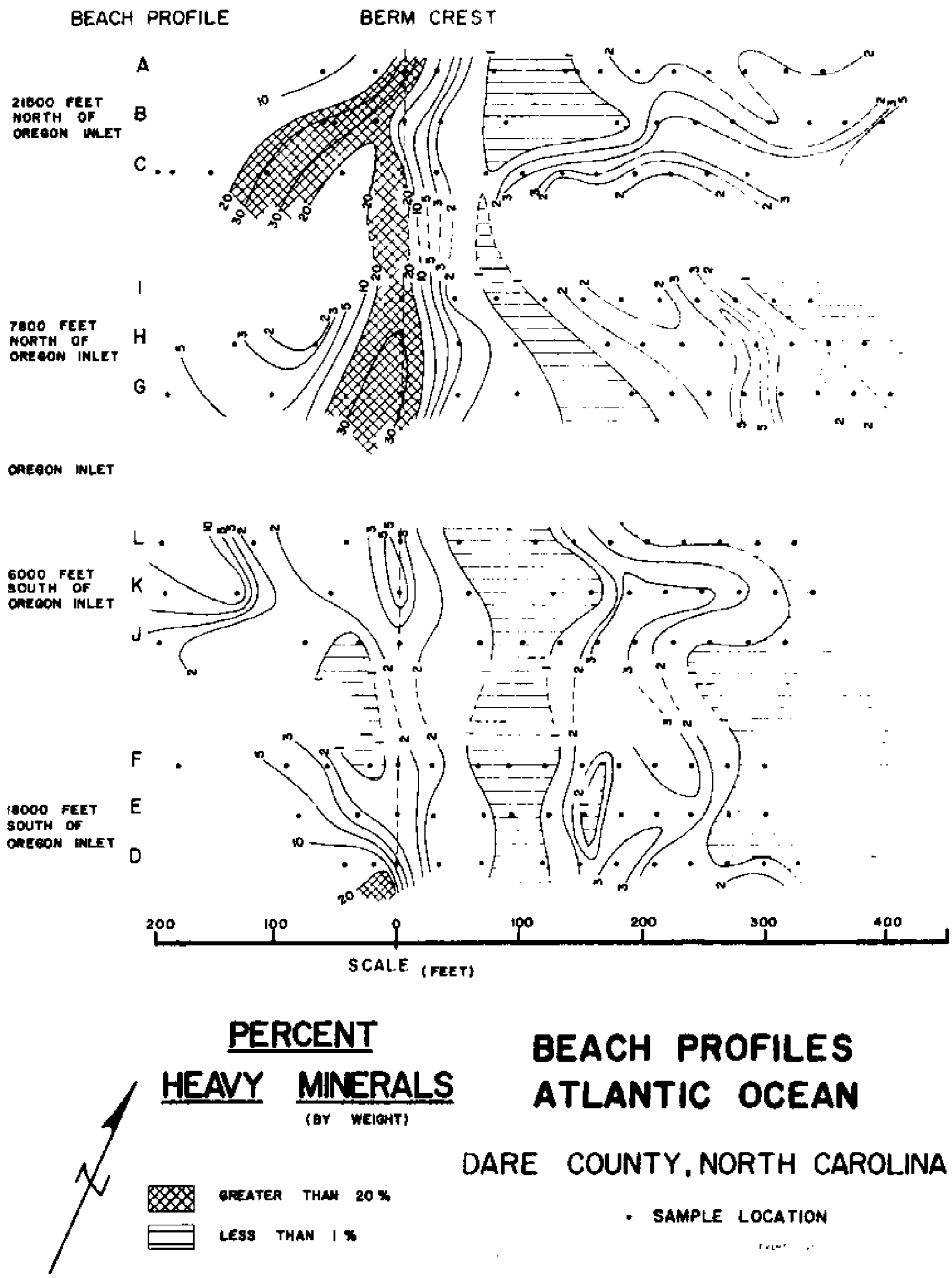
Heavy Minerals on the Beach

Figure 5-9 represents the percent of heavy minerals in the total beach sediment. The largest heavy mineral percentages are found on the upper foreshore and berm. With the exception of profiles L, K, and J, which have a high winter-deposited placer mineral concentration near the coastal dunes, all profiles represent summer depositional conditions. Breaker regions are everywhere poor in placer minerals. Heavy mineral concentrations are located in regions where the light mineral median diameter is roughly 1.5 times that of the placer mineral diameter. Heavy minerals in the beach zone were determined to have an average density of 4.0 gm/cc.

An important question concerning the formation of beach placers is: What is their transport residence in the various regions of the beach zone? The answer affords an insight into heavy mineral availability with respect to deposition as a beach placer. Although the outer breaker region transports a large total quantity of material, the small relative volume of placer minerals located in this region suggests it is not an important longshore path for heavy minerals.

Generally, two or three percent heavy minerals is contained in surface sediments of the surge zone, or about the average for the beach as a whole. Ingle (1966) and Duane (1970) have shown total longshore transport in this region to be less than in other beach zones, therefore, heavy mineral transport will also be small. Surge zone heavy mineral transport is generally shoreward, the grains moving in from the outer breaker regions.

Figure 5-9


BEACH PROFILE
BERM CREST
A
B
C
21800 FEET NORTH OF OREGON INLET
I
H
G
7800 FEET NORTH OF OREGON INLET
OREGON INLET
L
K
J
6000 FEET SOUTH OF OREGON INLET
F
E
D
18000 FEET SOUTH OF OREGON INLET
200
100
0
100
200
300
400
SCALE (FEET)
PERCENT HEAVY MINERALS
(BY WEIGHT)
GREATER THAN 20 %
LESS THAN 1 %
BEACH PROFILES
ATLANTIC OCEAN
DARE COUNTY, NORTH CAROLINA
• SAMPLE LOCATION

The step area, or the region of the inner plunge point, is very low in heavy minerals because its high turbulent intensity moves everything but the largest grains landward. Foreshore backwash may again move the same grains seaward, however, the heavy minerals are often temporarily concentrated in the foreshore, berm, and backshore regions of shallow flow.

The foreshore is an important conduit for transporting sedimentary material in a longshore zig-zag pattern parallel to the beach. This environment, due to its high percentage of heavy minerals, is the most important region of placer grain transport. The zig-zag journey a heavy mineral takes as it moves on the foreshore allows the placer grain to cover a large area normal to the coast as well as parallel to it. On a depositional beach the heavy mineral zig-zag pattern was observed to exist on the upper foreshore. Because of this mechanism, heavy minerals rarely returned to the step through backwash action.

Strahler (1966) observed swash on a rising tide to first deposit small quantities of material on the foreshore. With further tidal rise the swash scoured the foreshore sc that during most of the period of rising tide the foreshore was moving and little deposition was extant. This is essentially the sequence of events noted on Dare County beaches. As the tide retreated, depositional conditions were observed to exist and heavy minerals were preferentially and intermittantly lost to the foreshore transport conduit. Heavy minerals moved in a longshore direction at a slower rate than light minerals because they tended to move in an area further up the foreshore than the foreshore sediment as a whole.

Net deposition or scour on the foreshore is a function of the relative capacity of uprush and backwash to transport heavy and light minerals. The first foreshore deposition on a receding tide was coarse sand followed by fine sand. Under the latter condition, placer minerals were entrapped in the bed and often thin heavy mineral laminae were formed.

Beaches north of Oregon Inlet contain much greater concentrations of placer minerals than the more southerly beaches because the steeper southern beaches inhibit heavy minerals from moving out of the surge zone. Most of the placer grains in the south are moving in from an offshore source and are not yet in the foreshore conduit (Figure 4-8).

The two northern beaches contain consistantly high heavy mineral values on the foreshore because these beaches are gentle and slightly progradational as well as being supplied with placer minerals from beaches further to the north. While heavy mineral movement to the foreshore region of southern beaches is from offshore through several breaker zones and a surge zone, the northern beaches are supplied directly by the foreshore littoral conduit.

The heavy minerals observed on summer berms on Dare County beaches are ephemeral and will be lost during storms; however, they do reflect the amount of placer minerals being transported as well as the effect beach processes have on them.

Placer forming Mechanisms - Beach Laminae

Several assumptions may now be made concerning placers forming on a beach: 1) Heavy minerals are not concentrated below mean sea level, although they may be transported through the breaker and surge zones and to some extent in a longshore direction within them. 2) The foreshore is the region of most significant longshore heavy mineral transport on the beach. 3) The beach area above sea level offers the only location conducive to placer concentration and preservation. 4) Placer minerals accumulate on the foreshore, berm, and backshore as very concentrated laminae or diluted amounts within light mineral laminae.

The laminar concentrate type of placer is either a unidirectional or a reversing flow, shallow water accumulation formed when bed shear is too great to deposit fine light minerals and the boundary is too smooth to trap large light grains.

Heavy mineral laminae were sampled on two Dare County beaches. Samples were taken from a summer lamina on the foreshore, and a relic, partially buried storm berm stormline lamina at the landward margin of the backshore. Each was deposited under similar mechanical conditions.

Summer Foreshore Lamina

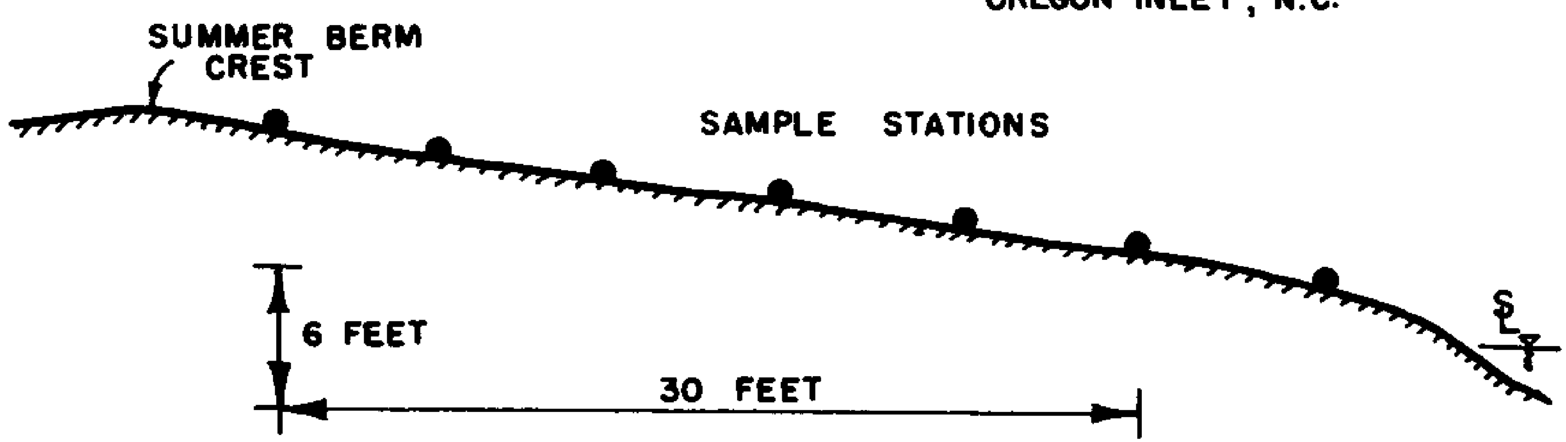
The summer lamina, illustrated in Figure 5-10, is typically concentrated on a relatively gentle slope, in this case the slope is 0.2ft/ft. Lamina thickness is greatest at the central five sample location points, thinning from eight to zero mm immediately beyond the outer stations. The length of the deposit parallel to the coast is 50 feet and its upslope width is 36 feet. The heavy mineral concentrate has an elliptical form beginning about four feet from the step and extending nearly to the berm crest. Surface undulations on the lamina are almost nonexistent. Surface smoothness is broken only by small rills a few grain diameters in height and about three feet long indicating flow in a seaward direction. Sediments beneath the lamina are clean foreshore light mineral sands with a mid-foreshore median grain diameter of 0.24 mm.

Textural parameters from the foreshore lamina exhibit trends significantly different from those of the foreshore light minerals. The median placer particle diameter, for instance, increases upslope, just the opposite of the light mineral trend. Sorting values remain relatively constant on the foreshore while skewness values increase upslope.

Figure 5-10

SORTING	.58	.53	.57	.59	.57	.45	.61
SKEWNESS	.172	.264	.210	.170	-.211	-.155	.148
MEDIAN GRAIN DIAMETER (mm)	.281	.264	.210	.235	.167	.163	.247

SURFACE LAMINA 8 mm THICK
PROFILE NORMAL TO COAST
LOCATION: THREE MILES NORTH OF OREGON INLET, N.C.



FORESHORE HEAVY MINERAL LAMINA

Stormline Lamina

Stormline laminae are located in the vicinity of coastal dunes. The location of a Dare County stormline placer is shown in Figure 5-11. The berm in this figure is a summer berm present at the time of sampling, but not present when the stormline deposit was formed. Individual sample locations are shown in the lower schematic of Figure 5-11.

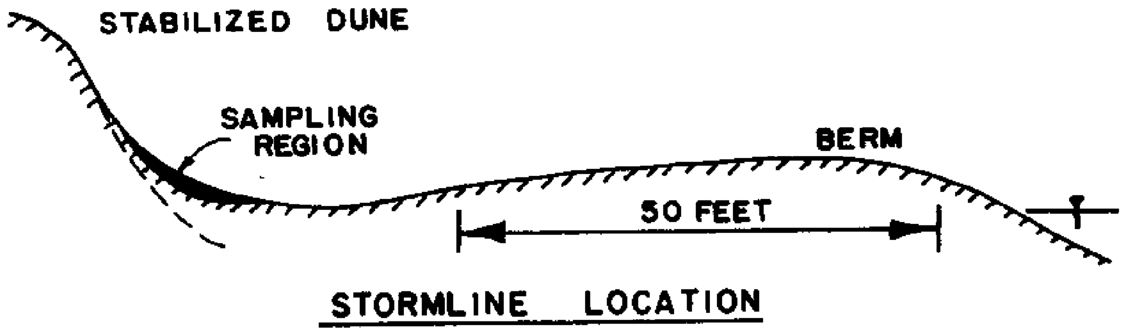
The stormline placer slope is 0.4 ft/ft or twice that of the foreshore deposit. Individual lamina thicknesses range from 50 to 100 mm. Several sequences of laminae were observed in a vertical section. The sampled lamina was located nearest the surface. It began at the dune base and terminated approximately 12 feet landward, pinching out rapidly at each end of the sampled section. Deposit dimensions parallel to the coastal dune face were not defined because the placer is largely buried by wind blown sand. The heavy mineral surface was very smooth and the light mineral contact above and below the lamina was observed to be very sharp. The light minerals are in the M_d range of 0.22 to 0.30 mm, which is a much smaller range than that observed on the foreshore.

Although the heavy mineral size range is the same for both foreshore and stormline laminae, the latter range is compressed into one-third the horizontal distance. This suggests that the steeper depositional surface will concentrate available heavy minerals into a thicker lamina, a desirable attribute of a placer. The steeper surface will also aid in decreasing the swash flow depth, which in turn increases the light mineral bypassing ability of the flow over the placer.

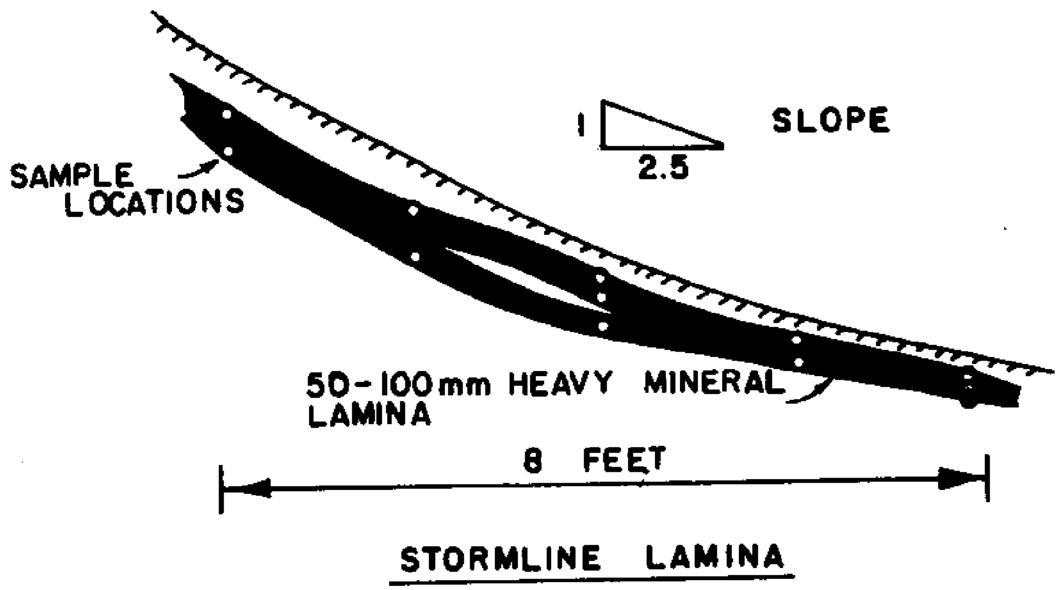
Horizontal textural trends in stormline deposits are the same as those observed on the foreshore (Figure 5-12). It was possible to collect heavy mineral samples of the top, bottom, and central portion of a stormline due to its thickness. This was done to establish textural trends with depth. If present, consistent trends might indicate the scope of changing flow conditions as the deposit formed. Of special interest is the tidal state, flood or ebb, during formation. Center samples were taken from the lamia to authenticate any textural trend in depth and to identify the lamina as a single tidal cycle event. Textural parameters are assumed to be relatively constant in a longshore lamina direction when swash impinges on the dune face in a constant magnitude and direction.

In only two of the six middle lamina textural parameters does the central portion of the lamina not follow the trend established by the upper and lower margins (Figure 5-12). The two anomalous values are representative of samples taken from the contact of a light mineral lense and the upper heavy mineral lamina. Textural parameters from mid-lamina of the seaward station are in all instances midway in value between upper and lower samples. Textural uniformity through the thickness implies a single, continuous, slowly changing sequence of events responsible for forming the lamina.

Figure 5-11

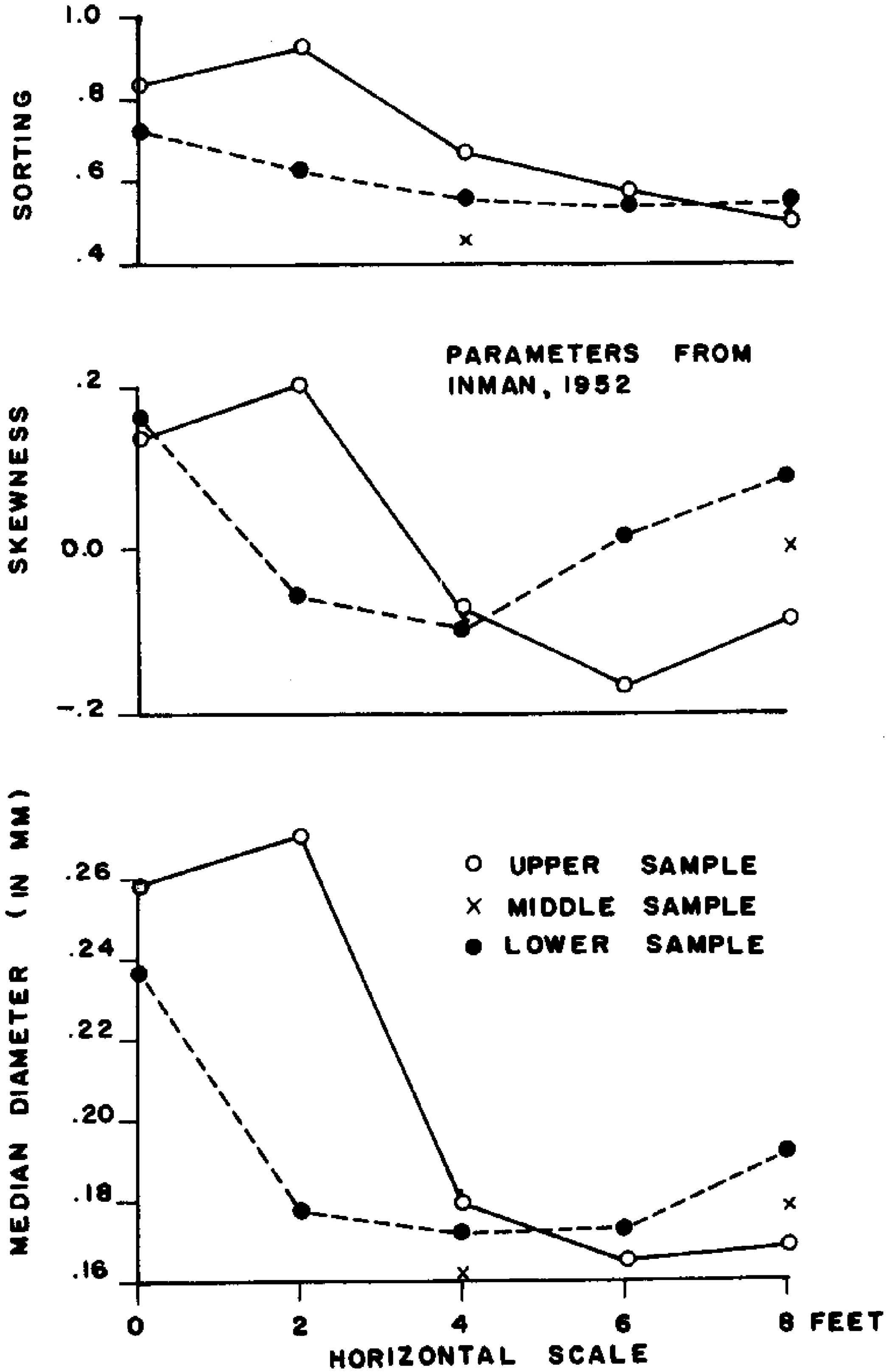


SORTING	.84	.93	.67	.58	.52
			.45		.54
	.71	.63	.56	.54	.55
SKEWNESS	.131	.203	-.075	-.173	-.077
			-.090		0.00
	.155	-.064	-.093	.019	.091
MEDIAN DIAMETER (mm)	.259	.270	.178	.165	.167
			.155		.178
	.237	.178	.173	.173	.191



STORMLINE PLACER - DARE COUNTY, N.C.

Figure 5-12



TEXTURAL PARAMETERS - FROM FIGURE 5-10

The difference between the lower and upper textural profiles appears not to be one of form, but of location. If the textural values for the lower sample profiles were transposed with the station two feet seaward, the values would nearly coincide with those of the upper profiles. This suggests that the placer-forming dynamic processes change in a uniform manner throughout its constructive building stage.

As observed in the heavy mineral textural parameters on the foreshore, the stormline median grain diameter increases in an upslope direction. Skewness also becomes positive landward, but sorting becomes poorer upslope. This is not the case on the foreshore where sorting remained fairly constant at 0.5 to 0.6.

Heavy Mineral Laminae - Mechanism of Formation

Swash, the uprush and backwash of waves on the foreshore, is responsible for constructing heavy mineral laminae on beaches. Light and heavy minerals are moved together, but during lamina formation heavy minerals are selectively deposited near the upper swash terminus, while the light minerals remain in motion. Summer foreshore and stormline laminae originate under similar dynamic conditions near the summer and storm berms respectively. Discussion of one is applicable to the other. Stormline laminae are the only important beach placers, however, because they are often preserved while summer foreshore deposits are constantly modified and destroyed. Summer-formed placers sometimes provide a portion of the heavy mineral source available for stormline concentration.

Heavy minerals move in suspension through the step zone, but tend to remain on the foreshore during periods of longshore transport because they are temporarily deposited between swash cycles rather than returned to the step by the backwash. Under certain conditions of shallow flow depth and an accelerating backwash flow, the largest available heavy minerals begin to fill light mineral interstices on the upper foreshore. The variance in critical bed shear for different sized heavy mineral grains results in deposition of the largest particles first (Figures 2-7 and 2-9). The flow intensity necessary to pass these large heavy mineral particles over a smaller sized heavy mineral bed would be enough to erode the entire bed. This would preclude lamina construction. Flume experiments confirm the field observation that large heavy mineral grains will not bypass any sedimentary bed found on a Dare County beach. Heavy minerals will move only when the entire bed, of which they are a part, is moving. This suggests that when heavy minerals are moving, the sediment surface beneath them is eroding, thus there are definite ranges of bed shear values for which only heavy minerals of various sizes and densities are deposited.

Light minerals bypass a heavy mineral bed when flow conditions move the light minerals, but are not capable of initiating motion in the heavy particles. The lamina becomes covered with light minerals when bed shear is insufficient to transport light grains across it. Conversly, the lamina scours when bed shear conditions are sufficiently intense to initiate heavy mineral motion.

The genesis of a heavy mineral lamina is reflected in the textural parameters obtained from the Dare County beaches (Figures 5-10 to 5-12). Median grain diameters decrease in a seaward direction suggesting a diminution of critical bed shear from the landward margin of the lamina to the step. The downslope decrease in backwash flow intensity was substantiated by flow measurements on the foreshore. Positive upslope skewness values indicate many finer heavy minerals have been entrapped in larger heavy mineral interstices. This is the situation expected at the beginning of a depositional sequence. Poor sorting and positive skewness often reflect the coarse depositional region of a sedimentary sequence. Sorting values decrease in the downslope direction in the stormline lamina reflecting the absence of the coarse end (0.18 to 0.26 mm) of the total heavy mineral range available. On the wider foreshore lamina, the distance over which the swash flow regime acts, is capable of maintaining low sorting values throughout the deposit.

As soon as the supply of heavy minerals moving on the foreshore is sufficient to cover the surface, the shear required to move that surface increases (Figure 2-7). The increase is the difference between critical bed shear for the light mineral surface and that of the heavy mineral covering. The heavy minerals accumulate first as a moving surface because the variance in bed shear necessary to transport heavy and light minerals is such that it causes the heavy grains to move at a slower rate. As flow conditions decrease, a point is reached when the placer particles are no longer moved and deposition of the moving lamina commences. This beginning was observed to be accomplished with a heavy grain thickness of one or two grain diameters. Once the light mineral surface is covered, heavy minerals will continue to be deposited as long as swash flow conditions remain relatively the same over the lamina throughout a swash cycle. Light minerals may be intermittently deposited between uprush and backwash sequences, but will be removed again when flow increases. This mechanism indicates the importance of maximum bed shear conditions on lamina formation. Under suitable bed shear circumstances the lamina will continue to increase in thickness until the placer mineral supply is exhausted. Heavy mineral supply undoubtedly plays a role in terminating lamina growth even when bed shear conditions are compatible for placer expansion.

The increase in median grain diameter, skewness, and sorting (Figure 5-12) from the bottom vertically to the upper surface of a stormline deposit suggests decreasing flow conditions as the lamina builds because the upward limit of highest flow intensity, as evidenced by the sediment texture, has moved seaward. In this case flow conditions prior to deposition must have been erosional because light minerals in the dune areas all have critical shear values much less than those of the heavy minerals. High intensity flow conditions were observed to concentrate placer minerals in a moving surface accumulation on the upper foreshore during erosive cycles so that heavy minerals would be present in quantity when bed shear circumstances were reduced sufficiently for a lamina to form. This lessening of flow intensity could be the result of a receding tide or waning swash oscillations as wave intensity decreases. The uniformity of changing textural parameters in a vertical section (Figure 5-12) indicates the deposit was formed during a continuous, decreasingly less intense sequence of events.

CHAPTER VI

INLET PLACERS

INTRODUCTION

Ultimately heavy minerals drifting along beaches encounter barriers hindering their movement. These obstacles include inlets, estuaries, deltas, coastal projections, submarine canyons, and man-made structures such as jetties and groins. Inlets and estuaries, in particular, are of interest to the prospector as they may prove prospective coastal plain traps for heavy minerals. Although this chapter emphasizes inlet sedimentation, the related sedimentary processes, especially those which concentrate heavy minerals, are often similar to estuarine processes.

Caldwell (1969) defines a tidal inlet as a short, restricted waterway connecting the tidal ocean or sea with a sizable bay or lagoon. The restriction is such that the passage of the ocean tide wave is noticeably hindered by the inlet and the tide range in the bay is appreciably less than that for the adjacent ocean waters.

Purpose

The threefold objective of this chapter is to define: (1) sediment distribution, (2) the inlet flow regime, and (3) the inlet bathymetry as they aid in predicting areas favorable to heavy mineral accumulation. In conjunction with predicting areas for placer formation, it is a further objective of this investigation to place limits on the spatial geometry of the deposits as well as to consider the probabilities of placer preservation. This chapter will also serve as an assist in relating placers in the active sediment wedge (Figure 3-2) to possible burial and retention in the offshore zone following a marine transgression.

Approach

A discussion of the inlet throat region and shoals and channels in the offshore and sound regimes provide a basis on which to define heavy mineral concentrations. Descriptions of individual placers within these locations and a review of the mechanisms responsible for their formation aid in establishing concepts applicable to other inlets.

FIELD INVESTIGATION

Oregon Inlet is located on the coast of North carolina and central within the placer study area (Figure 1-1). It is the first break in the coastal barrier south of Cape Henry, Virginia which is 75 miles north. The south-southeast trending barrier island remains unbreached for 35 miles south to Cape Hatteras, North Carolina.

Oregon Inlet was open in the sixteenth century and closed sometime thereafter (Villianos, 1967). Its re-opening in 1846 was the result of a large storm which piled Pamlico Sound water on the western shores of Bodie and Pea Islands. During the 1846 storm the water level attained an elevation sufficient to breach the low coastal barrier dunes there and the washover and subsequent channel scour formed the present inlet.

Based on sediment distribution patterns in the vicinity of Oregon Inlet, the inlet system can be subdivided into three connecting regions: (1) the offshore regime, (2) the inlet throat, and (3) the Pamlico Sound region. Figure 6-4, illustrating the median diameter dispersal pattern, indicates that the portion of the offshore region influenced by inlet processes extends in a crescent shape 15,000 feet from the inlet throat to the east and 11,000 feet parallel to the coast. The throat is the constricted region bisecting Bodie Island and Pea Island. The Pamlico Sound region expands in an arc to a maximum distance of 24,000 feet from the throat. Processes and flow intensities in each of these regions are significantly different, warranting this division in the following discussion of inlet sedimentation.

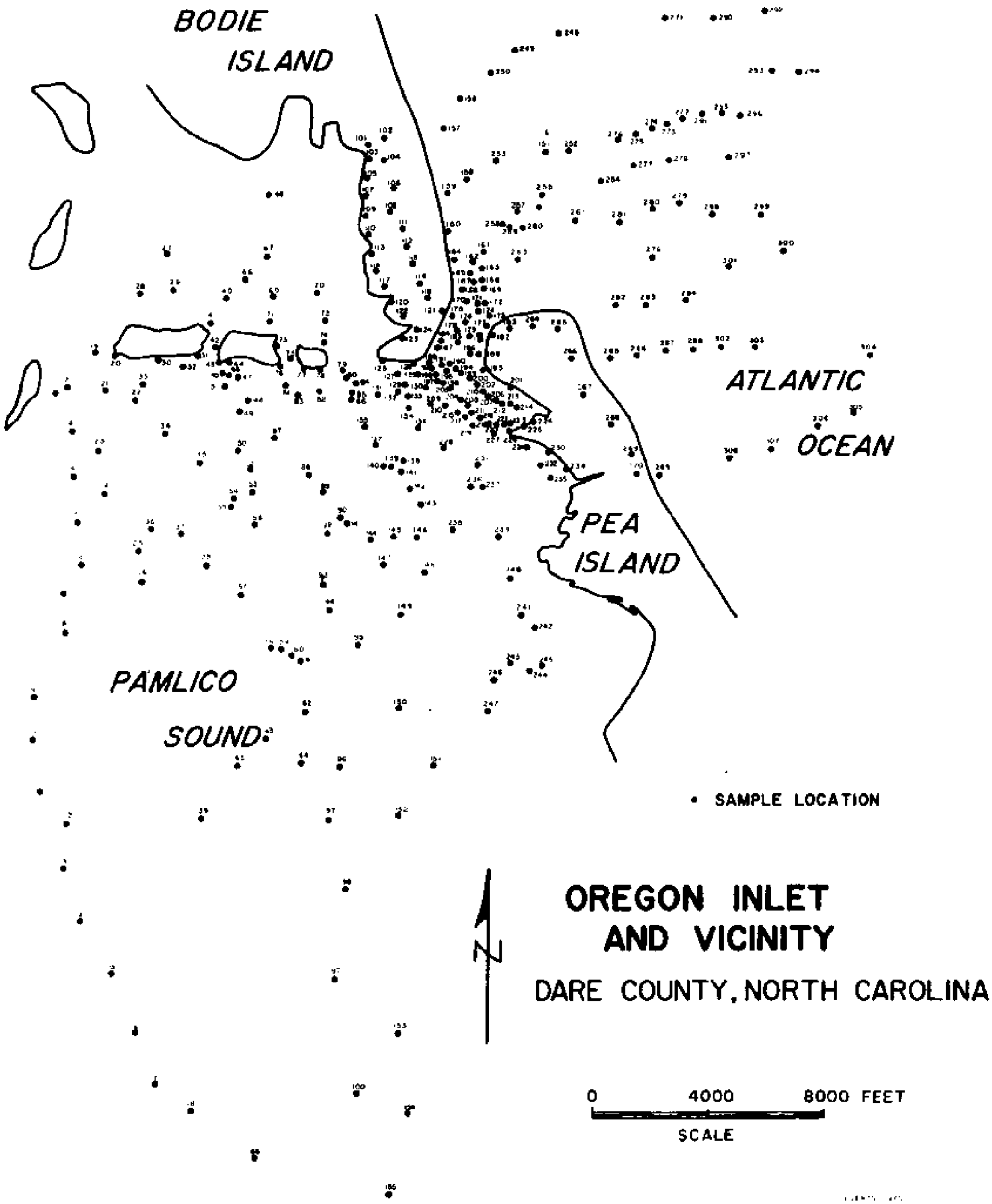
Procedure

Bottom samples were collected near Oregon Inlet during the months of July and August, 1969. Sample stations were occupied over 34 square miles. (Figure 6-1). Thirty-two stations were positioned above mean sea level, and 276 below. The spit on Bodie Island was sampled because it was topographically low and appeared to be periodically inundated. Although not an active sedimentation region during the field survey, the spit occasionally becomes an active erosional and depositional area during storms, especially those capable of raising the local sea level two or three feet. Samples from Pea Island were taken from coastal dunes.

All sample stations and topographic features were located using reference points obtained from an enlarged portion of Coast and Geodetic Chart 1229, revised to 1968. The stations were positioned by three-point triangulation with a sextant held in the horizontal plane. Reference points included fixed navigational structures and the center of the Oregon Inlet bridge.

Collection of samples was accomplished by free-surface diving and by dredging, each using a 14-foot boat as the collection platform. Employing a pipe dredge, samples were taken of approximately the two centimeters of surficial bottom material. To maintain survey uniformity, the diver also collected material to the same two centimeter depth. Dredge samples were collected only in those channels deeper than 15 feet.

Figure 6-1


BODIE
ISLAND
ATLANTIC
OCEAN
PEA
ISLAND
PAMLICO
SOUND
• SAMPLE LOCATION
OREGON INLET
AND VICINITY
DARE COUNTY, NORTH CAROLINA
N
0
4000
8000 FEET
SCALE

Bathymetry

The topographic expression of the complex Oregon Inlet sedimentation system reflects the response of unconsolidated sediment to the flow regime existing in the area. Channel-cut shoals bound the inlet throat on its Atlantic Ocean and Pamlico Sound margins (Figure 6-2). Each shoal and channel is an important depositional and distributory feature for heavy minerals entering the system.

INLET THROAT

The throat region is characterized by the deepest channel in the inlet system, reaching a depth of between 35 and 40 feet. The thalweg is located near the northeast shore of Pea Island, resulting in a skewed bottom profile. Vallianos (1967) reported that the main channel has been located on the south shore since 1957. He also noted that the historical trend of the main channel conformed to the northeast-southwest direction as it is observed today.

OFFSHORE REGION

On the Atlantic Ocean side of the throat the main channel is joined by two smaller troughs entering from the east and north. The smaller channels bound the barrier island beaches and two large offshore shoals. The shoals, bisected by the main channel, form an arc with its center of curvature on the coast of Bodie Island.

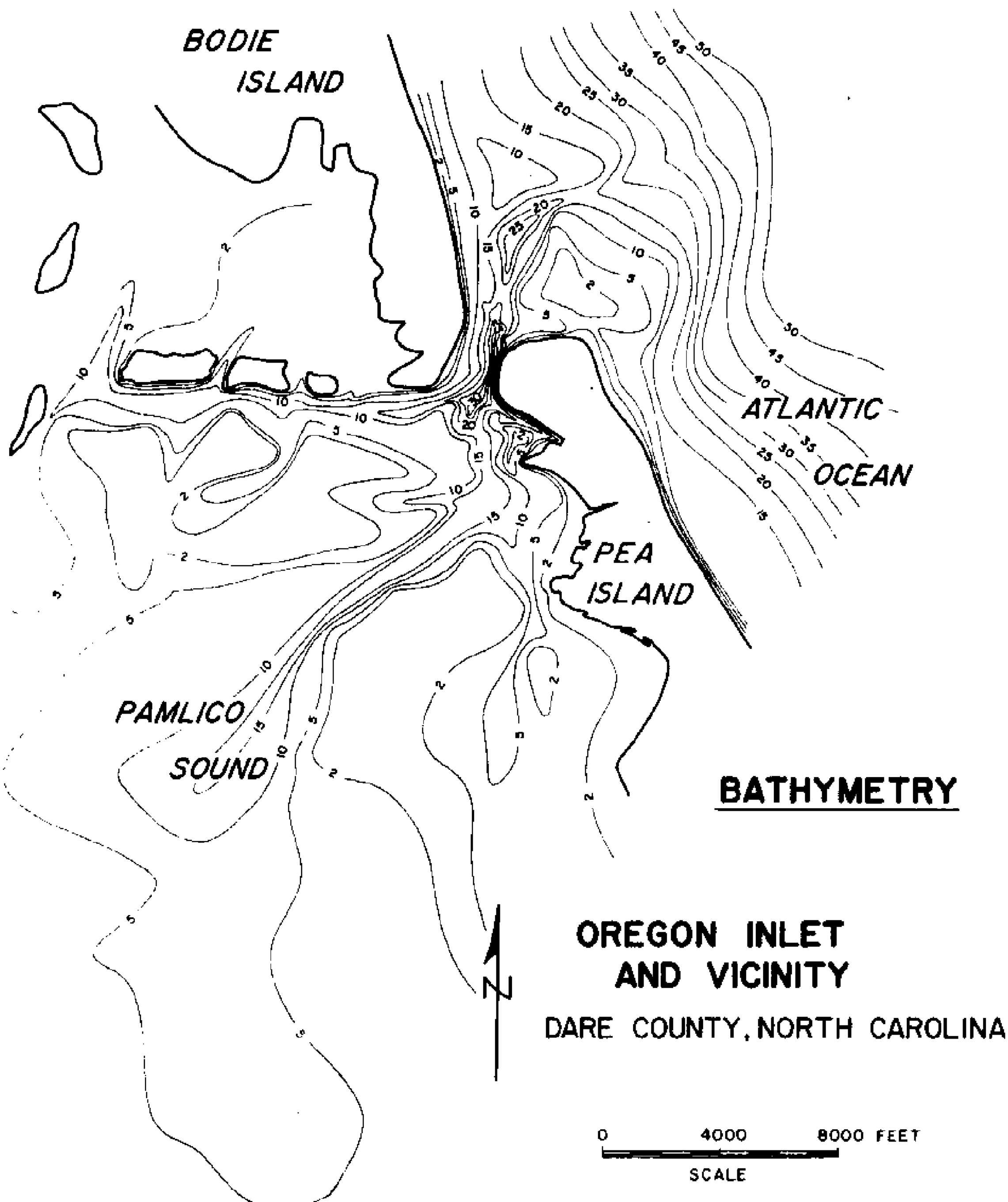
Depths of 25 feet or more in the main channel are interrupted in two locations where the channel shoals to 15 to 20 feet. The first hump occurs 2,000 feet from Pea Island in the north trending section of the channel. The other hump is located on the eastward extension of the main channel near the outer margin of the shoals.

Lying south of the main channel, the largest offshore shoal extends almost to the beaches of Pea Island. The southern shoal was observed to be almost awash at low tide. Breakers were observed on this shoal in response to all sea conditions existing during the summer of 1969. Waves broke on the central portion of the north shoal in response to moderate seas.

SOUND REGION

The pattern of channels and shoals in the Pamlico Sound region is remarkably similar to the offshore pattern. Each region has a main channel flanked by two smaller channels with intermediate shoals. In each region the smaller troughs parallel land and are connected to their counterparts having relatively the same depth. The patterns reflect the tidal influence on bathymetric features.

Figure 6-2


BODIE
ISLAND
ATLANTIC
OCEAN
PEA
ISLAND
PAMLICO
SOUND
BATHYMETRY
OREGON INLET
AND VICINITY
DARE COUNTY, NORTH CAROLINA
N
0
4000
8000 FEET
SCALE
DEPTH IN FEET

Inlet Flow Regimes - General Concepts

Because of the high cost in time and equipment of effectively monitoring an inlet flow system, it is impractical in this study to define and prove a quantitative inlet flow model. It is, however, possible to derive an idealized conceptual model based on the inlet flow regime that may be used as an aid to: (1) describe the channel and shoal system of the inlet, (2) explain the cross-sectional throat profile, (3) delimit the barrier island outline, (4) explain the influence topography and flow play in sediment transport and deposition, and (5) most importantly, describe the behavior of heavy mineral grains in the inlet system.

Three unique flow fields exist in the inlet system: tidal currents, longshore currents, and wave-activated motion. A combination of all three results in the overall flow pattern of the inlet. Each flow regime changes with time, especially tidal currents, so that the overall inlet flow pattern is unsteady. The most significant change is the reversing flow caused by astronomical tidal waves.

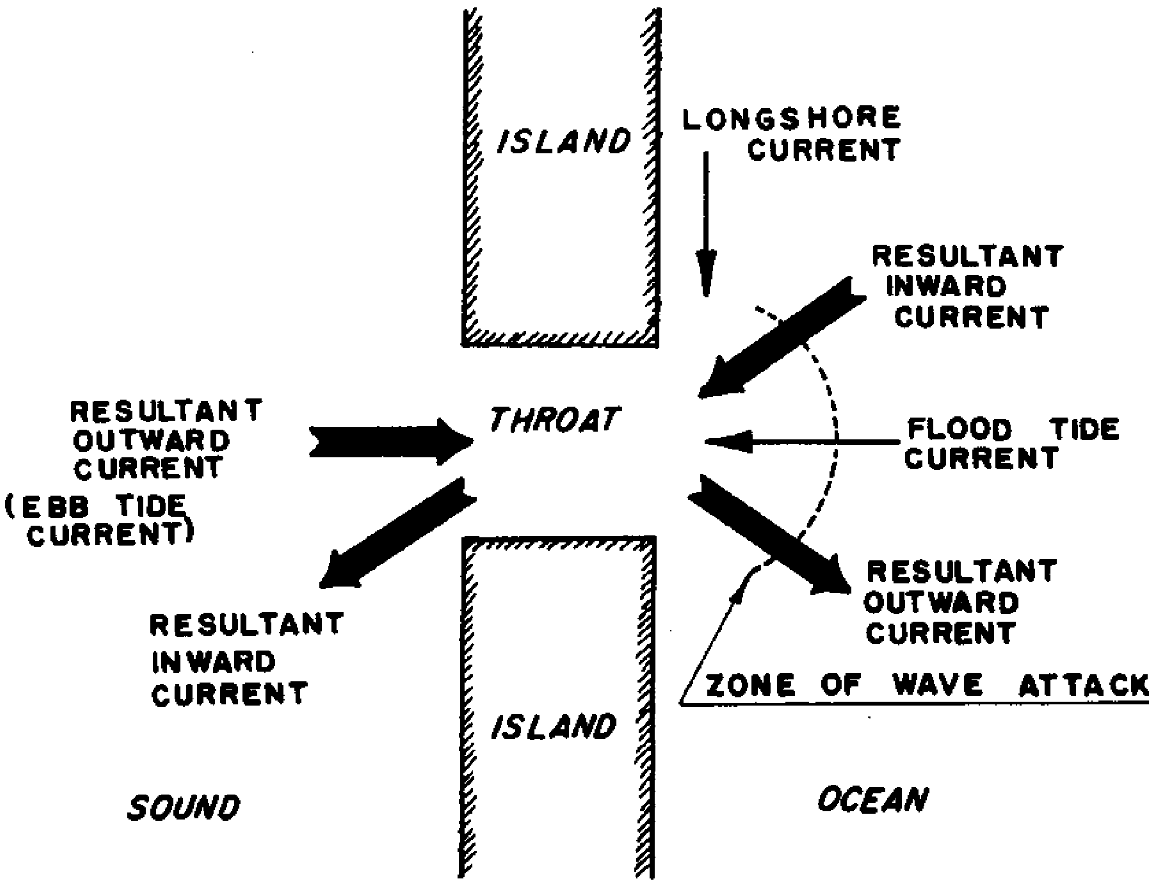
Tidal currents are directional and reversing and include predictable astronomically-induced tidal waves and meteorologically-activated low frequency waves. The latter are the result of water piling on, or removal from, one side of the inlet causing a hydrostatic gradient between the sound and ocean.

The flow pattern from a tidal wave is illustrated in Figure 6-3. Current intensity is greatest in the throat and on the discharge side of the inlet. Discharge may be simulated by a jet (Harrison _et al._, 1964). The least intense current velocities are on the input side of the throat. In this region flow moves towards the throat from all directions of an arc extending from the shore. Flow intensities are a function of conceptual streamline spacing in the lower diagram of Figure 6-3.

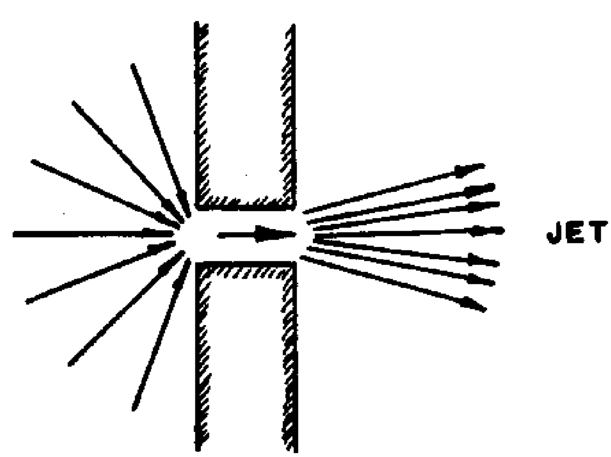
Flow moving into the sound is known as the flood current and flow moving out of the sound as the ebb current. The effect of wind on tidal currents in Oregon Inlet is shown in Table 6-1. When wind-driven currents and tidal currents are in phase, predicted tidal ranges, maximum throat velocities, and flow rates increase. The opposite condition prevails when these currents are in opposition.

Fresh water discharge may also be an important factor in total inlet discharge. This is not the case for Oregon Inlet, however, because fresh water input into Pamlico Sound is 33,700 acre-feet per day (Vallianos, 1967) while the total volume of the sound is 19,400,000 acre-feet. Assuming equal discharge by each of the three inlets draining Pamlico Sound, the discharge per day per inlet would be 0.058 percent of the total sound volume. A stratified flow situation is therefore not likely to exist in the Oregon Inlet area.

Figure 6-3



IDEALIZED INLET FLOW



TIDAL JET

SCHEMATIC OF INLET FLOW SYSTEM

Table 6-1

TIDAL FLOW AT OREGON INLET

Date	Cross-Sectional Area	Winds Direction	Winds Ave. Speed (mph)	Predicted Tidal Range (ft) Flood	Predicted Tidal Range (ft) Ebb	Maximum Velocities (ft/sec) Flood	Maximum Velocities (ft/sec) Ebb	Observed Total Flow (acre-feet) Flood	Observed Total Flow (acre-feet) Ebb
9/9/31	39,000	NE	6.2	2.0	1.5	2.5	2.3	47,769	37,399
8/31/32	-	W	11.4	2.2	1.8	2.4	2.6	42,726	40,054
11/10/32	-	SW	11.9	1.7	1.8	2.4	3.2	34,873	57,208
8/24/37	44,000	S	6.9	2.2	2.1	3.7	3.3	63,500	55,900
8/14/39	56,000	SW	10.5	2.5	2.2	2.8	3.8	37,800	71,500
4/32/50	28,000	SSW	20.5	-	-	-	3.2		38,200
9/27/65	66,580	NE	8.6	2.8	2.5	3.1	2.3	98,200	54,200

(modified from Vallianos, 1967)

No astronomical tides were observed in Pamlico Sound because the extensive shoal area between the inlet throat and deeper sound waters creates rapid energy dissipation of the tidal wave. Therefore, current velocities with respect to time are in phase with tidal heights because a maximum hydraulic gradient exists at times of high and low tides (Caldwell, 1955).

Many coasts, including Dare County, are subject to a variable direction of wave attack. When not normal to the coast, the angle of attack creates one form of longshore current which may affect the inlet flow regime. Figure 6-3 illustrates a longshore current entering the inlet system from the upper quadrant.

Littoral transport, a direct function of the amount of wave energy being expended on a coast plus the angle of wave attack, has been reported for the Oregon Inlet region by Vallianos (1967). His results are recorded for an average year in Table 6-2. Note the net littoral drift is to the south.

The third flow regime incorporated into the inlet flow system is a consequence of shoaling gravity waves. Many inlets with an appreciable longshore littoral drift exhibit arcuate shoals on the seaward margin of the inlet (Bruun, 1967b). Incoming shoaling waves are refracted on the shoals with their crests bending parallel to the bottom contours (Figure 6-3). Waves often break on the offshore shoals, and in the case of tidal flood flow, the translational wave-induced current strengthens the flood current. Ebb currents, on the other hand, are diminished near the shoals by the incoming waves.

The upper diagram of Figure 6-3 illustrates the resultant inward and outward current directions for a superposition of tidal, longshore, and wave-induced currents. The angle of wave attack in this figure is normal to the coast.

The remainder of this chapter is devoted to applying this general flow model to the specific case of Oregon Inlet.

Inlet Sediment Distribution

SEDIMENT AVAILABILITY

Sediment may move into the Oregon Inlet system from nearby beaches, the Atlantic continental shelf, or Pamlico Sound. Once it reaches the inlet it may bypass on the outer shoals or move into the throat and out again in response to the reversing tidal flow regime. Sediment may also move into the inlet system and be deposited.

Vallianos (1967) reports a net southerly littoral drift of 365,000 cu yds/yr near Oregon Inlet and a total sand in transit of 2,389,000 cu yds/yr, suggesting a large sediment availability to inlet transport and depositional mechanisms. Vallianos also reports

Table 6-2

SUMMARY OF LITTORAL TRANSPORT ANALYSIS - VICINITY OF OREGON INLET, N.C.

Direction of Wave Approach	Magnitude of South-bound Drift (cu yds)	Magnitude of North-bound Drift (cu yds)
N	1,690	
NNE	83,580	
NE	1,291,960	
ENE		458,580
E		188,630
ESE		289,920
SE		43,430
SSE		31,550

(from Vallianos, 1967)

that a relatively constant quantity of sediment is retained in shoals over a period of several tens of years; therefore, although the inlet receives the material, it retains only small quantities of it, an interesting observation in as much as that part of this investigation is to determine if heavy minerals moving in the inlet do displace light minerals and thereby accumulate there.

Various authors have reported the net landward transport of bottom sediments into estuaries and inlets from offshore regions (Meade, 1969; Simons, 1965; Hareleman and Ippen, 1967). Harrison and others (1967) recovered seabed drifters in Oregon Inlet that had moved there from offshore placement stations to the northeast, some from as far as 35 miles. These record residual bottom drift on the continental shelf, but one cannot assume the gentle day-to-day currents that move neutrally bouyant seabed drifters will move significant amounts of sand.

TEXTURAL DISTRIBUTION

Inlet Throat

Coarse material, that with a median diameter greater than 0.30 mm, is axially orientated in the throat direction but offset against the northwest shore (Figure 6-4). This poorly sorted deposit (sorting greater than 0.7) is located in a bottom region sloping toward the channel thalweg. Sediments near the thalweg are fine-grained (0.20 to 0.25 mm) and well sorted (Figure 6-5). Skewness within the throat region is not diagnostic of mappable sedimentary trends (Figure 6-6).

Offshore Region

The coarse ribbon-like deposit observed in the throat continues into the offshore region near the beach of Bodie Island although somewhat landward of the main channel. Landward of the first topographic high in the main channel (Figure 6-2), a decrease in the median diameter of the coarse ribbon is apparent. At this location the coarse deposit swings away from shore and becomes coincident with the main channel. The coarse ribbon continues to be confined to the main channel until it reaches a point just prior to its northward peak. The coarse material is everywhere poorly sorted to this location.

The median diameter of the coarse ribbon decreases as the axis of the deposit trends south of the second topographic high in the main channel. A fine-grained sediment lobe appears to be encroaching on the main channel near the hump.

Sorting trends wrap around Pea Island and approach the beach in the south. Sorting north of the main channel is generally good. Skewness for the most part in the offshore inlet region is non-diagnostic.

Figure 6-4

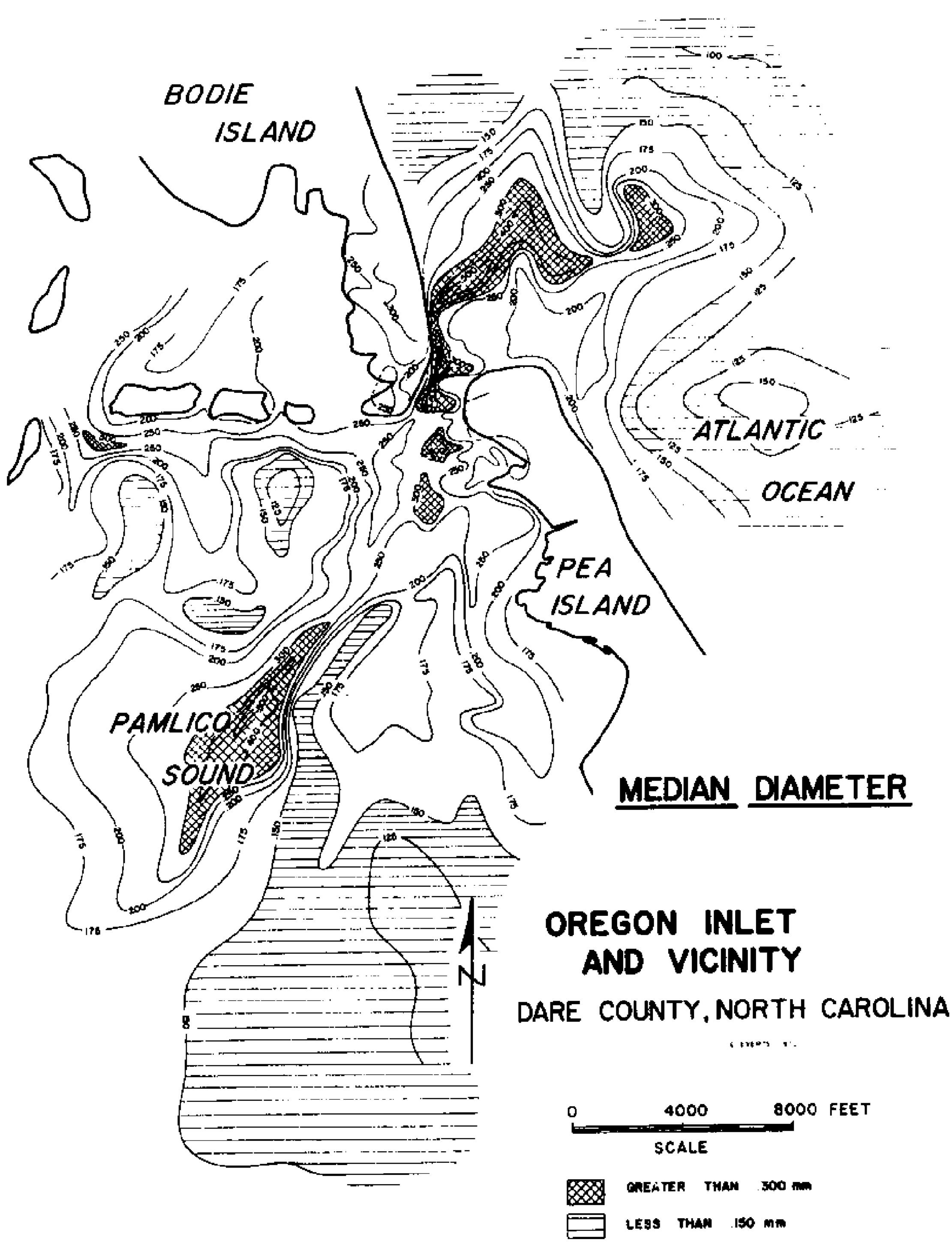

BODIE ISLAND
ATLANTIC OCEAN
PEA ISLAND
PAMLICO SOUND
MEDIAN DIAMETER
OREGON INLET AND VICINITY
DARE COUNTY, NORTH CAROLINA
0 4000 8000 FEET
SCALE
GREATER THAN .300 mm
LESS THAN .150 mm
N

Figure 6-5

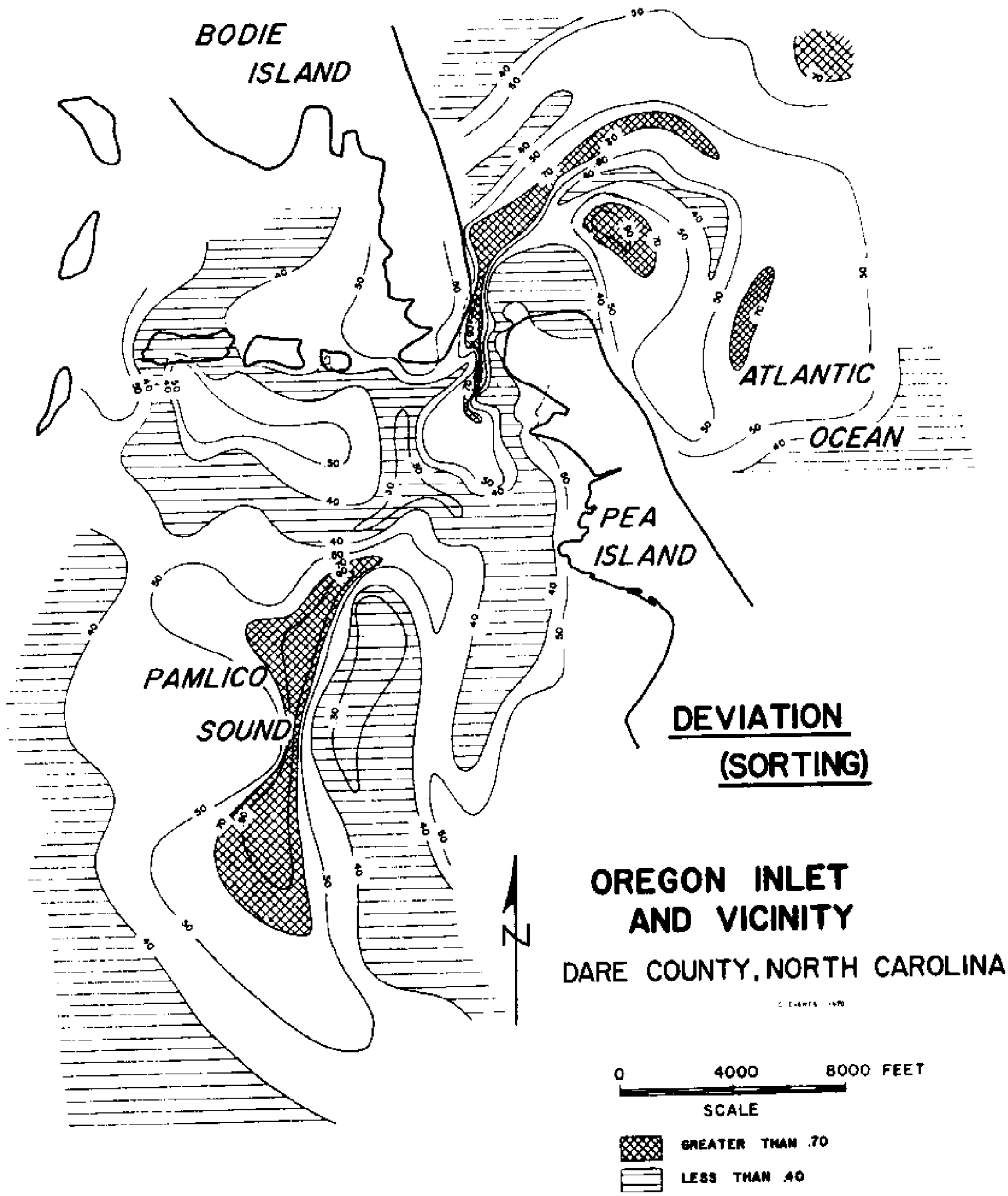
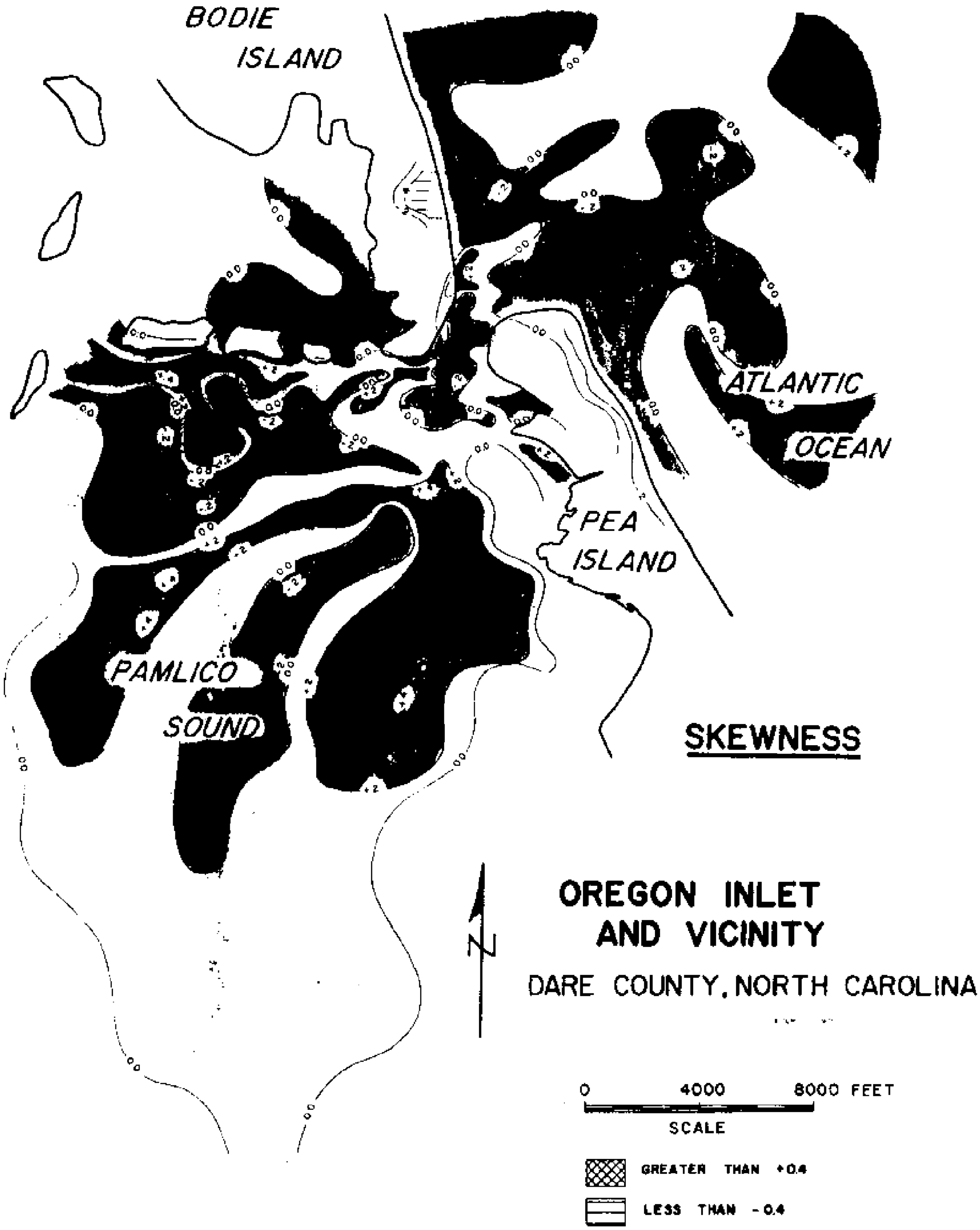

BODIE
ISLAND
ATLANTIC
OCEAN
PEA
ISLAND
PAMLICO
SOUND
DEVIATION
(SORTING)
OREGON INLET
AND VICINITY
DARE COUNTY, NORTH CAROLINA
0
4000
8000 FEET
SCALE
GREATER THAN .70
LESS THAN .40

Figure 6-6


BODIE
ISLAND
ATLANTIC
OCEAN
PEA
ISLAND
PAMLICO
SOUND
SKEWNESS
OREGON INLET
AND VICINITY
DARE COUNTY, NORTH CAROLINA
N
0
4000
8000 FEET
SCALE
GREATER THAN +0.4
LESS THAN -0.4

Sediments on the south shoal are characterized by sorting values of 0.5 or less and a median diameter of less than 0.20 mm. The northern shoal is not obvious on either the median diameter or sorting charts.

The median grain diameter decreases away from the offshore terminus of the main channel. Fine grains of diameter less than 0.125 mm located 8,000 to 15,000 feet from the inlet throat are not of the general sediment size of the shelf material in this area (Figure 4-5). The decrease in particle median diameter away from the inlet and their anomolous position in the offshore region suggest they are of inlet-process origin.

Sound Region

Channel locations in the sound are shown in Figure 6-4 where all coarse grains are located in channels. Lesser grain sizes (smaller than 0.20 mm), however, appear at constrictions and topographic highs in the main and north channels, as well as on shoals. Texturally, the south channel is not well defined.

Sorting is poor for deposits in the main channel, while sediments in the north channel are well sorted. The sorting difference between channels is duplicated in skewness values which are negative in the main channel and positive in the north channel. The west end of the north trough is especially positively skewed. The bathymetric highs in each channel are characterized by positive skewness and good sorting. Skewness and sorting values at the main trough high are especially divergent from values in the remainder of the channel. Sorting improves and skewness approaches zero near the shallowing finer-grained terminus of the main channel.

Shoal sediments in the sound are uniformally fine-grained (median diameter less than 0.175 mm). Sediments on the shoals are also well sorted and reported skewness values are generally positive, except along the rim of the north channel.

Discussion

SEDIMENT SOURCE

Littoral drift supplies much sediment to the inlet system (Table 6-2). With the exception of sizes finer than 0.175 mm, sediments of all grain sizes in the inlet region are also moving on beaches in Dare County (Figure 5-4). Coarse material, that with a median diameter in excess of 0.30 mm, is transported predominantly in the breaker zones of the beach conduit.

The total volume of sediment directly entering the inlet system from offshore sources is probably negligible because the input area parallel to the coast is small. Pierce (1969) balanced a net sediment budget in the Core Bank area of North Carolina by assuming 337,000 cu yds of material entered the barrier region from offshore sources, or about 4,000 cu yds per year per mile of coast. The offshore supply was one percent the volume passing as littoral drift. Core Banks is located 100 miles south of Oregon Inlet in a region of less intense wave activity.

Further evidence suggesting that an appreciable volume of offshore sediment does not move directly into the inlet is the observed decrease in median diameter of sediments away from the offshore inlet region. Sediments of median diameter less than 0.125 mm are not found on the immediate continental shelf nor in the beach conduit, suggesting the fine material is depositional and inlet-derived. Moreover, the arcuate shape of the fine dispersal pattern also indicates an inlet source.

Since sediments of grain size less than 0.175 mm are not found in the littoral zone feeding the inlet and are not moving into the inlet from an offshore source, a source in Pamlico Sound is indicated. Peat and very fine-grained sand deposits were observed as erosional outcrops along channels in the sound and near the southwest shore of Pea Island. The eroding peat deposits near the throat were at sea level while those on the sound were located four or five feet below mean sea level suggesting formation periods during different sea level stands. All peat deposits were from one to two feet thick. One peat crop-out was observed along the entire north bank of the north channel at a depth of four feet.

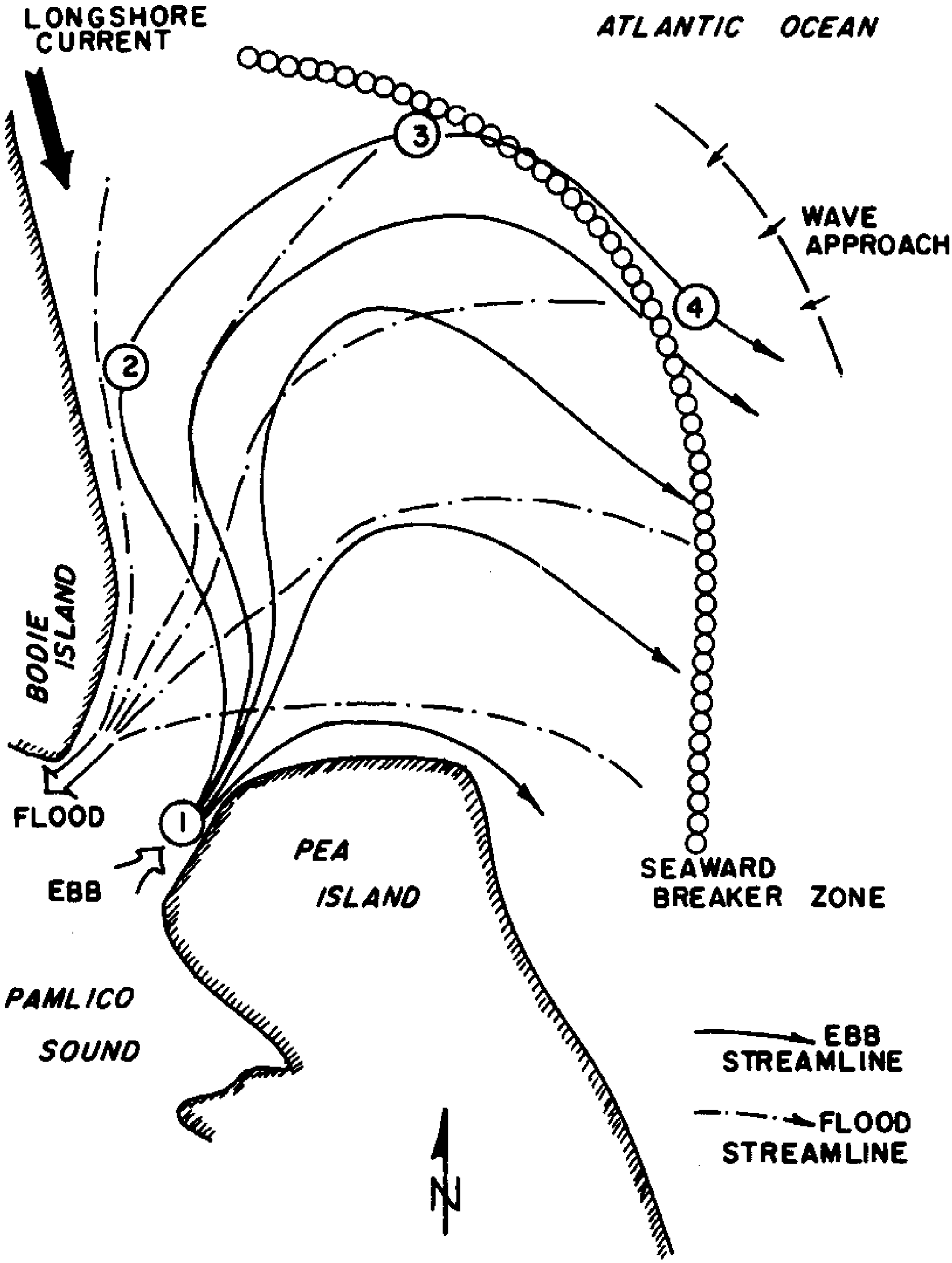
The bathymetry of the sound along the barrier islands is almost the same as the bathymetry near the inlet suggesting that inlet processes do not significantly change the total volume of sediments in the sound (Figure 6-2). Inlet processes may, however, cause erosion and displacement of fine-grained sound material with replacement by littoral sediment moving into the inlet barrier. The small displaced particles may move offshore and form the fine-grained deposits ringing the inlet mouth or move toward the sound into quiescent water.

As further evidence of the displacement hypothesis, the median grain diameter of material in the sound north of Oregon Inlet is considerably coarser (0.20 mm to 0.125 mm) than that located three miles south of the inlet (Figure 6-4). The median sizes of material at the southern margin of the figure correspond to sizes found offshore. Since the inlet is migrating southward (Vallianos, 1967), channeling would have had the opportunity to displace the fine sound-formed material north of the inlet, but not that in the south.

Sediment Dispersal

It is important to describe sediment dispersal patterns with respect to the flow regime in order to define the general processes active in inlet sedimentation. Armed with this knowledge we are better able to describe the formative processes existing during placer genesis, thereby aiding placer location and deposit geometry predictions in similar environments. The velocity and direction of

Figure 6-7



SCHEMATIC REPRESENTATION OF OFFSHORE FLOW - OREGON INLET, N.C.

currents discussed in the following sections are based on semi-quantitative observations made by the author during the field survey.

Inlet Throat

Sediment was observed to enter the throat region of Oregon Inlet from the beaches of Bodie Island when the flood and longshore currents were acting in tandem. The flood tide wave enters the throat from the east while the longshore current is directed southward near the beach. The resultant landward current is from the northeast (Figure 6-7). This complex flow impinges on the northeast shore of Bodie Island and distributes some of the incoming material in that region. Since most coarse sediment is transported in the nearshore beach zone, the larger grains are moved into the inlet throat near the northeast shore. Both the directional nature of the incoming current and the feeder location of incoming sediment result in the coarse nearshore ribbon as shown in Figure 6-4.

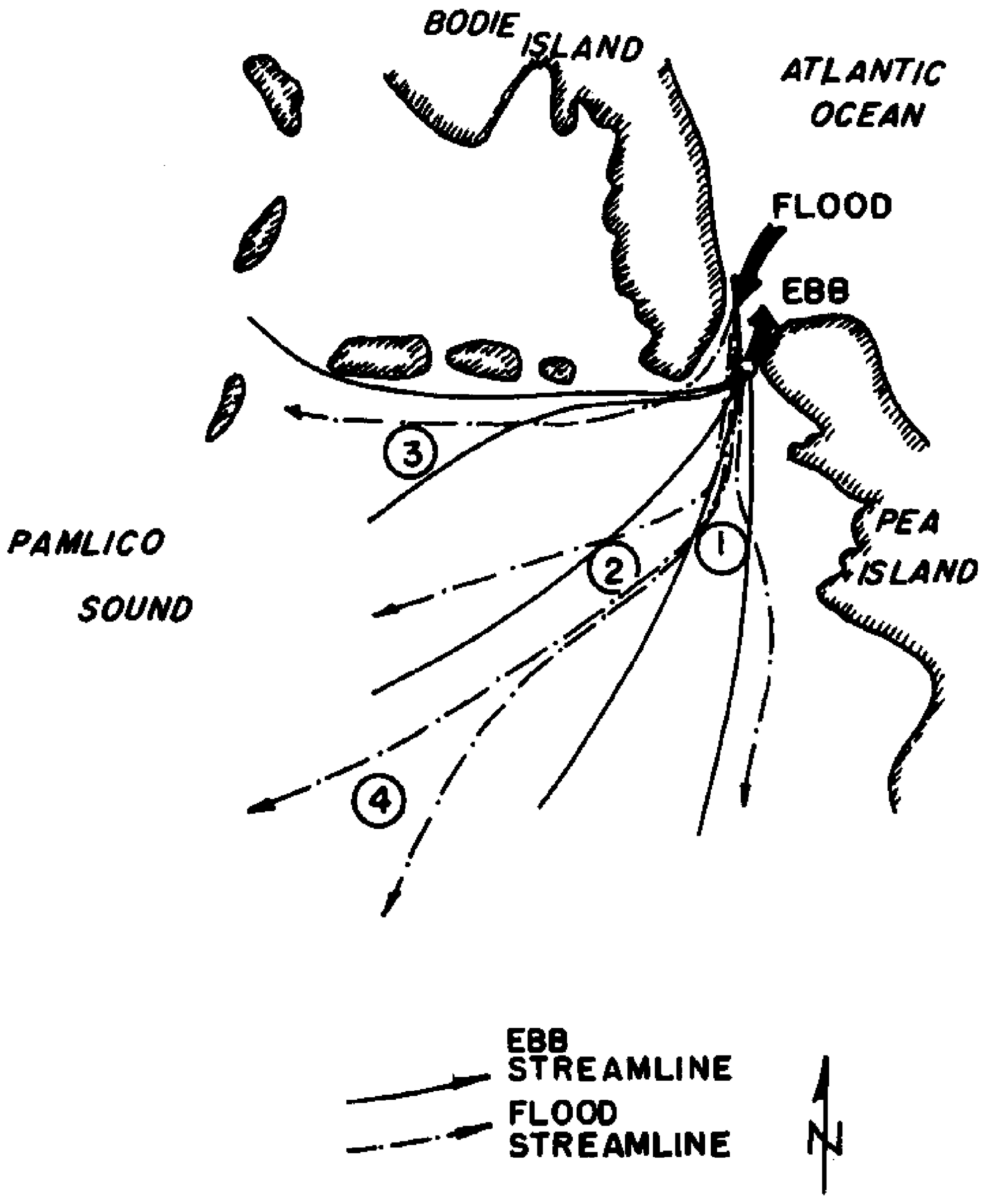
The reversed ebb flow into the throat enters from the east and impinges on the western shore of Pea Island. The ebb current is then deflected northward from the erosional Pea Island shore. The ebb regime does not move the same amount of coarse material in the throat in comparison to flood currents because the coarse sediments are not available near Pea Island. As indicated in Figure 6-8, entering flow from the entire sound becomes channeled only where it nears the throat. Because the ebb regime drains the shoals in the sound it moves considerable fine material seaward, thereby accounting for the finegrained deposits near Pea Island.

The historical record reveals that during periods of high sediment availability from the north, the southernmost tip of the Bodie Island spit progrades southward (Vallianos, 1967). Further, considerable accretion occurs during moderate sea conditions and high longshore drift. Because progradation of the spit decreases the cross-sectional area of the throat, tidal velocities increase and the western shore of Pea Island erodes. This process is responsible for the southward migration of Oregon Inlet.

The coarse updrift throat deposit is buried as the spit migrates, leaving a coarse lenticular deposit at depth. The deposit trends parallel to and beneath the barrier island dunes. Since the deposit forms at depth in the channel and is preserved, a transgressing wave-active face (Figure 3-2) may later exhume it as a relic offshore deposit.

The dominant longshore drift, superimposed on tidal ebb and flood flows, is responsible for the cross-sectional profile of the throat and the distribution of grain sizes. The shape of the barrier islands near the inlet mouth is also a direct result of the cyclic flow regime that causes the updrift island to be displaced landward and the downdrift island to erode on its western margin and prograde on its eastern boundary.

Figure 6-8



Scale 1" = 6,000'

SCHEMATIC REPRESENTATION OF PAMLICO SOUND FLOW - OREGON INLET, N.C.

Offshore Region

Sediment distribution in the offshore region is a function of ebb-tide jet, flood-flow regime longshore currents, wave activity, and the amount of material bypassing the inlet on its offshore shoals, plus the sediment volume moving through the throat. As the ebb current is reflected from the erosional northwest bank of Pea Island (location 1, Figure 6-7), it swings north-northwest and moves some of the incoming coarse material offshore. At location 1, although the ebb current is deflected northward, much of the fine material moving with the ebb flow continues in a northeasterly direction to subsequently accumulate on the south shoal. Sorting values onthe south shoal suggest an absence of coarse sediment either moving or being deposited in this flow region.

The deflected ebb-flow jet is slowed in its northward journey in the region of the nearshore topographic high in the main channel (location 2, Figure 6-7). The high forms where net incoming flood and longshore current intensity is equal to the outgoing ebb current intensity over a tidal cycle. The result is a low shoal and a depositional region capable of concentrating sediment sizes smaller than those maintained in the higher energy main channel depths. Coarse material between the inlet mouth and the topographic high is offset shoreward by the westerly flow direction of the flood current. The southward-moving longshore current dissipates some ebb flow energy near location 2 and deflects the ebb jet to the northeast. Coarse sediments are moved in the main channel away from shore because ebb and flood currents are coincident.

The north shoal (location 3, Figure 6-7), is formed as a consequence of outward-moving ebb-carried material and wave containment. Waves reaching the shoal region slow the outward flow of inlet waters and cause deposition. The shoal, in turn, acts as a barrier to the outgoing current and deflects it eastward.

Out of the influence of the ebb-flow regime, waves, attacking the shoals from a predominantly northeasterly direction, are responsible for moving large quantities of sediment south. This is evidence in the southward median diameter and sorting trends. The topographic high at location 4 (Figure 6-7) is a nodal depositional region of equal net wave and ebb regime intensity. The coarse material seaward of this region was probably moved during an abnormally high ebb flow. Under normal conditions it would be expected to remain relatively immobile because it is effectively out of ebb regime influence.

Bruun (1967a, b) derived an empirical formula for quantitatively computing the relative amount of the total longshore drift which bypasses an inlet on its offshore shoals without entering the throat. When the ratio of net littoral transport to maximum instantaneous tidal flow is greater than 200, he reports most littoral material will cross the offshore bar. The ratio for Oregon Inlet is 261 (Tables 6-1 and 6-2). The bypassing is accomplished by strong ebb currents, a dominant southerly longshore drift, and wave transport of sediments southward along the outer margin of the north and south shoals. Textural trends on the outer bars support the hypothesis that considerable material moves south.

Because of the complexity and varying intensity of flow regimes in the offshore region, skewness, which defines depositional and transportational trends in quasi-steady flow, is non-diagnostic. Sorting trends are alligned with the resultant flow direction. Poor sorting suggests unsteady flow and an admixture of grains carried and deposited by different flow fields. Offshore inlet sediment dispersal is, therefore characterized by a deflected ebb-current jet distribution in the nearshore regions and southerly wave dominated transport in the outer regions.

Sound Region

Flow in the Pamlico Sound region is less complex than that for the area seaward of the inlet mouth because the hydraulic regime in the sound is basically tidal in origin. Ebb and flood currents dominate the flow field and are responsible for the distribution of sediments, the formation of heavy mineral concentrations, and the orientation of tidal channels. Due to the cyclic flow hydraulics and the relative absence of other currents and wave action, sorting and skewness in the sound are diagnostic of flow intensity and direction.

Figure 6-8 is a schematic representation of Pamlico Sound flow near Oregon Inlet. It illustrates the relative magnitude and direction of flood and ebb flow as observed during the normal summer climatic conditions as were extant during the field sampling program of 1969. The distance between streamlines is indicative of relative current velocities; i.e. the closer the streamline spacing, the greater the current velocity in that region.

The incoming flood jet, carrying Bodie Island beach material, is directed into the main channel (Figure 6-2). Near location 1 on Figure 6-8 the jet splits and the main flow body is deflected southwestward while a minor flow continues in the south channel. The reason for the bifurcation is related to the very shallow depths existing south of the throat near Pea Island. Shallowing flow conditions, as observed in the flume, increase flow resistance. The increase southward is so great that the jet deviates and seeks deeper water and less resistance in the sound.

Ebb streamlines are spread over the sound in a somewhat radial pattern which is in accord with Figure 6-3. The central shoal regions were observed to be dominantly influenced by ebb currents directed toward the throat. Ebb flow, however, appeared to be concentrated to some extent in the north channel as considerable water is focused between the islands and shoal regions in the northwest quadrant of the study area.

Channels in the sound region are the principal transportational conduits for the distribution of sediments. Material moves in on the flood cycle and seaward in response to ebb flow. Sediment distribution in Figure 6-4 is either the result of net depositional equilibrium over a tidal cycle or deposition during a unique high intensity event.

When material enters the inlet it moves in response to tidal current-induced bed shear until it is deposited in the sound or until it passes to a seaward depositional site east of the throat. A given particle, we presume, will be deposited during its movement past the inlet in a region of net tidal-cycle shear approximating the maximum shear necessary to move it. This accounts for the greater median diameter of sediments in the channels (0.25 to 0.50 mm) than on shoals (0.15 to 0.20 mm).

A complication in defining distribution is found in accounting for high energy events capable of moving and depositing sediments away from their normal areas of deposition. This is probably the case for the coarse region near the terminus of the main channel (location 4, Figure 6-8). A high intensity flood current (greater than 4.0 ft/sec through the throat) may have brought the sediments in while normal ebb current intensities were, at the same time, insufficient to move them seaward. The negative skewness and above-average median diameter values suggest that this region is non-depositional for particles of the sizes found on the nearby shoals.

Near location 2 (Figure 6-8)the net opposing flood and ebb shear on the bottom is balanced. This results in a build-up of sizes smaller than the average for the main channel because the material is in quasi-equilibrium. Neither flood nor ebb is capable of removing the smaller fractions without near-equal replacement by the opposing cycle.

The coarse zone (location 4, Figure 6-8) in the main channel is probably flood controlled with the net bed shear greater than at location 2 and directed and toward the sound. The decrease in the size of material away from the channel terminus and the pattern of median diameter contours supports the flood-dominated flow hypothesis.

A skewness trend near the north channel (location 3, Figure 6-8) bifurcates away from the sound suggesting deposition during the ebb sequence of the tidal cycle. The good sorting and highly positive skewness indicate little dilution by coarse (greater than 0.40 mm) sediments moving in from the Bodie Island beaches.

Fine material (median diameter less than 0.20 mm) is ubiquitous on shoals. These well-sorted, shallow water deposits suggest steady and uniform depositional processes. Most movement on the shoal margins is of flood current origin while ebb currents modify central shoal regions. Positive skewness, the shallow depths relative to the rest of the sound, and the median grain diameter trends away from the channels indicate shoals are depositional areas. The majority of shoal material is probably brought in during high intensity flood flow.

Heavy Minerals

GENERAL CONCEPTS

Heavy minerals, subject to different hydraulic conditions, form two distinct depositional placer types, disseminated and laminar. The disseminated placers of the sound and offshore shoals are formed in a current ripple regime and will be discussed in detail below. Throat placers are of the laminar type formed above mean sea level by a mechanism analogous to that which constructs beach storm-line deposits except the throat placers were created in a uni-directional rather than reversing flow field. Deep water, disseminated, offshore placers are formed in a wave-dominated oscillatory environment comparable to the rippled bed region as discussed in Chapter IV. It may be noted that, with an understanding of how and where a given particle moves, one might predict the composition of an aggrading layer if such a layer is formed as the bed migrates.

Should the sediment volume entering an area be greater than that leaving the region, accretion occurs. In this case there is a possibility of heavy minerals accumulating during the process of migration.

Figure 6-9 illustrates the pressure and velocity distribution over a typical current ripple where u_o is the local velocity, p is the local pressure, and p_r is a reference pressure (Vanoni and Hwang, 1967). The flow converges on the upstream side of the ripple and diverges downstream. A separation streamline attaches itself at the crest and reattaches where u_o equals zero. This streamline is shown as a dashed line on Figure 6-9.

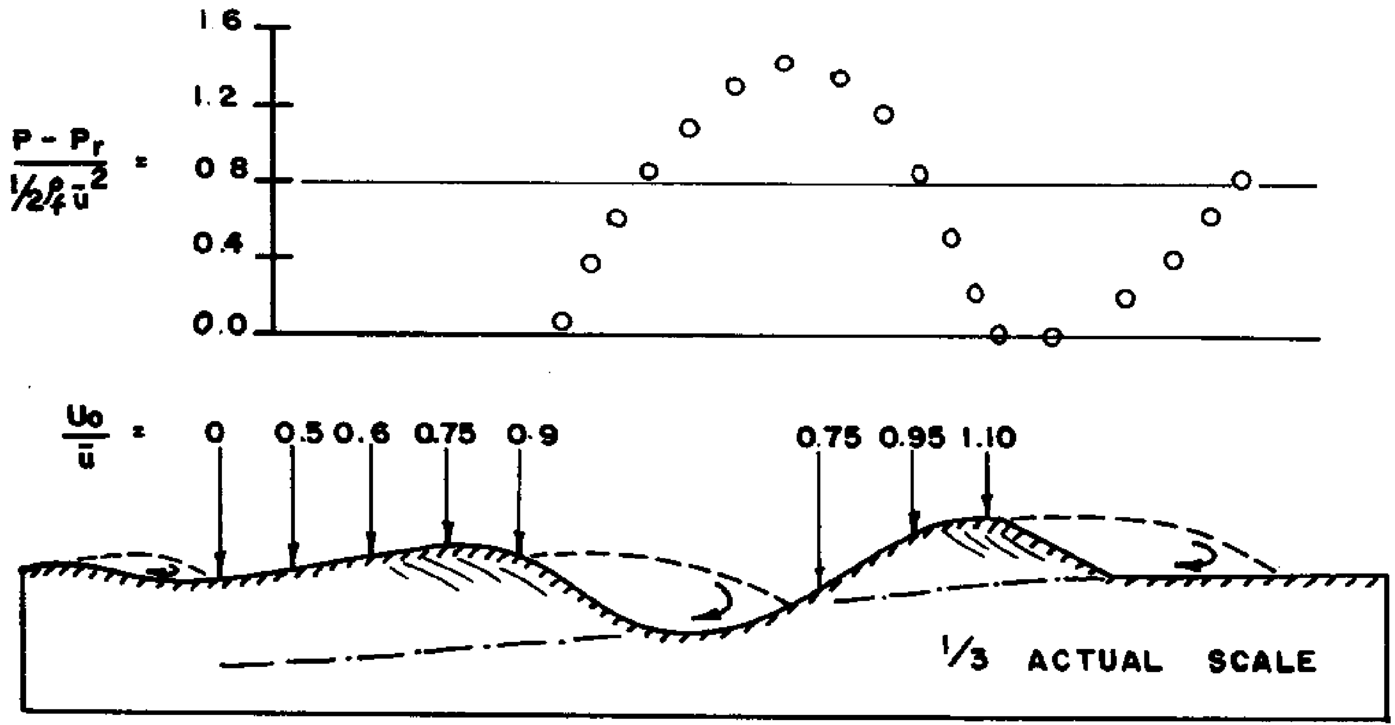
Using a dye and observing fluid motion near a ripple, Sutherland (1966) noted the trough region to be an area of strong eddies which cause the bed grains to be lifted in radial trajectories from the disturbance center. When the flow intensity is strong enough the grains continually move in the trough without preferred orientation except that imposed by flow above the bed. Pronounced scour occurs at the stagnation point.

Flow on the upstream slope is parallel to the boundary and conditions are similar to those on a flat bed. The velocity on the slip face increases toward the crest while the importance of impinging eddies diminishes. Almost all particles move as creep or in high frequency saltation movement. Bed creep is most prominent near the crest.

Grains at the crest are launched into the main flow or tumbled down the lee slope. Flow on the lee face is again influenced by strong eddy impaction because there is little opportunity for a viscous sublayer to develop.

Heavy minerals were observed in the flume to form a surface cover on the upstream face of current ripples and to a lesser extent in the ripple troughs. The amount in the trough varied with flow intensity and the size of light minerals associated with them. Optimum

Figure 6-9



PRESSURE AND VELOCITY DISTRIBUTION OVER A TYPICAL RIPPLE

(FROM VANONI AND HWANG, 1967)

conditions of flow strength (bed shear) and light mineral association allowed the heavy minerals in the trough to be overridden and buried. In an aggrading environment the trough is the lowest region on the ripple surface and material in the trough is most likely to be buried. Only short-term qualitative observations were made of migrating ripples because there was no means of recirculating sediment in the flume. Equilibrium conditions for observed migrating beds were probably never reached.

The vertical dimensions, rate of migration, and ripple index of a bed form are a function of mean velocity, flow depth, and bed consistency (Harms, 1969). Spacing between ripples observed on Pamlico Sound shoals varied from 7 to 12 times the ripple height. The lee face of current ripples usually sloped at their static angle of repose, or about 34 degrees for sand-sized particles, with the coarsest grains located at the base of the lee face, and the finest particles concentrated on the slip face. A sharp angularity was often present at the base of the lee slope marking the junction of contemporary ripple material prograding an erosional, or a depositional surface.

Heavy minerals were observed in ripple troughs and in bed depressions on shoals in Pamlico Sound. Light minerals migrated toward the throat leaving heavy minerals as an accumulating deposit in the troughs when the mean ebb velocity in one or two feet of water was less than 0.6 ft/sec (18 cm/sec) and the median diameter of light minerals was 0.175 mm. When the mean flow velocity reached 1.0 ft/sec (30 cm/sec) the heavy minerals also moved and accumulated on the slip face of the ripple. As the flow velocity increased, the heavy mineral accumulation on the slip face moved toward the crest indicating non-depositional conditions for the placer grains.

When the migrating ripple was composed of fine-grained (less than 0.20 mm in diameter) light minerals and when the heavy minerals were moving, the entire bed appeared to be eroding. However, when there were coarse light mineral sizes (greater than 0.25 mm in diameter) available on the migrating ripples, they were present in the trough and often appeared to be deposited and buried while the heavy minerals and smaller light minerals bypassed them. This bypassing mechanism is not possible on a flat boundary. These observations are important in our explanation of disseminated placers formed on shoals.

Figure 6-10 illustrates the equivalent sizes of light and heavy minerals that have been concentrated and buried on rippled surfaces in the inlet and offshore regions. The densities of heavy minerals shown are given as a function of grain size in Figure 1-5. Average equivalent sizes for the offshore oscillatory ripple zone (dashed line) and the inlet current ripple zone (solid line) are similar, suggesting that depositional processes in the trough are equivalent. The equivalent sizes are not comparable to equivalent settling velocities (Figure 1-2).

THROAT PLACERS

The largest laminar-type deposit observed in the Dare County study area was located on the shore of Pea Island near the southeast side of Oregon Inlet (Figure 6-11). A three-dimensional model and plan view of the deposit is illustrated in Figure 6-12. The length of the deposit is 200 feet with a maximum width of 50 feet and an average total thickness of two feet. Single heavy mineral laminae attained a maximum thickness of five to six inches while intermediate light mineral laminae rarely exceeded three inches. A three-fold light/heavy mineral separation was observed near the base of the upper sequence of laminae. A basal erosional light mineral surface was covered with a black heavy mineral lamina composed of dense opaque minerals. The opaque lamina was in turn followed by a lamina of less dense amphibole, garnet, and epidote with a light mineral lamina on top. All contacts were sharp. Although the thicknesses of the individual lamina varied along the shore, the total thickness of the three-fold sequence remained relatively constant until it pinched out along the placer margins.

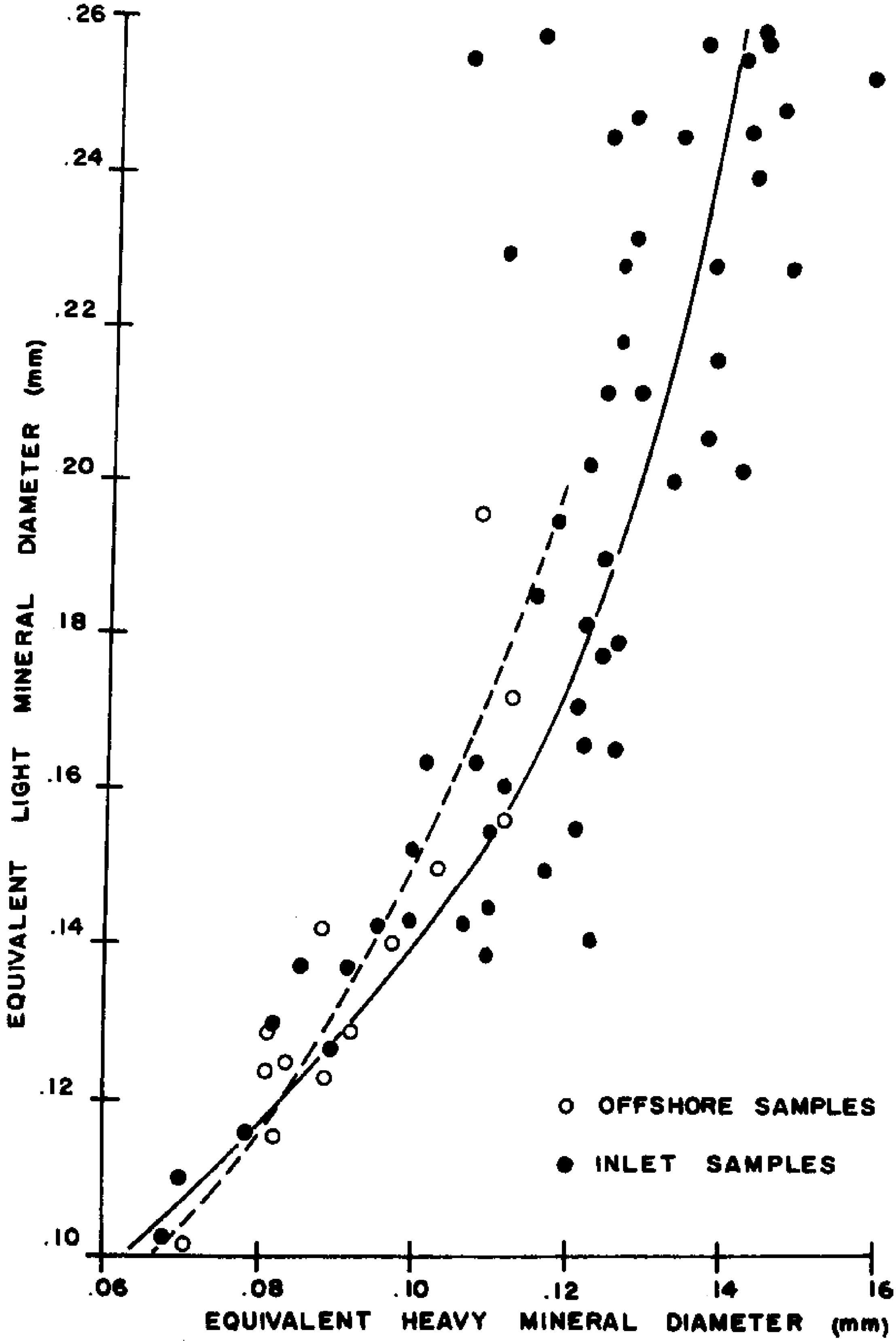
Heavy minerals cropping out from a two foot erosional scarp first drew attention to the deposit. Several channels which were later dug normal to the shoreline established the laminae as inlet derived rather than a relic dune-covered deposit. All laminae terminated at an erosional surface on the dune face. Protruding roots on the upper surface of the dune provided initial evidence of erosion. The material covering the upper laminae sequence appeared to be derived from slippage of dune sands.

Evidence of two events in the creation of the deposit was provided by a discontinuity separating the upper relatively level laminae (2-4^{o} dip toward the inlet) from the lower laminae (10-15^{o} dip toward the inlet). The lower sequence, whose basal lamina was found near the normal high water mark, cuts the upper sequence and was probably responsible for the erosional scarp.

The upper and lower laminar-placer sequences appear to have been formed during increasing high water conditions following an erosive event. Winds approaching the North Carolina coast from an easterly direction, depending on their duration and intensity, often create storm surges. Large waves coupled with an exceptionally high water condition serve as an ideal erosional agent against the southern margin of the inlet. Resultant flood currents moving into the sound also aid scour by removing eroded material. Incoming flood-flow deposits most of its beach-derived sediment on the northern spit and is not erosional in that region. The erosional surface shown in Figure 6-12 suggests a bench cut during an exceptionally high water event.

As a storm center passes inland and the wind direction changes, water piled on the western shore of Pamlico Sound moves east and increases the water level at Oregon Inlet. Heavy and light minerals from the sound are then moved to the throat as the high water ebb current impinges on Pea Island. When depth and

Figure 6-10



EQUIVALENT SIZES – RIPPLE ENVIRONMENTS

Figure 6-11

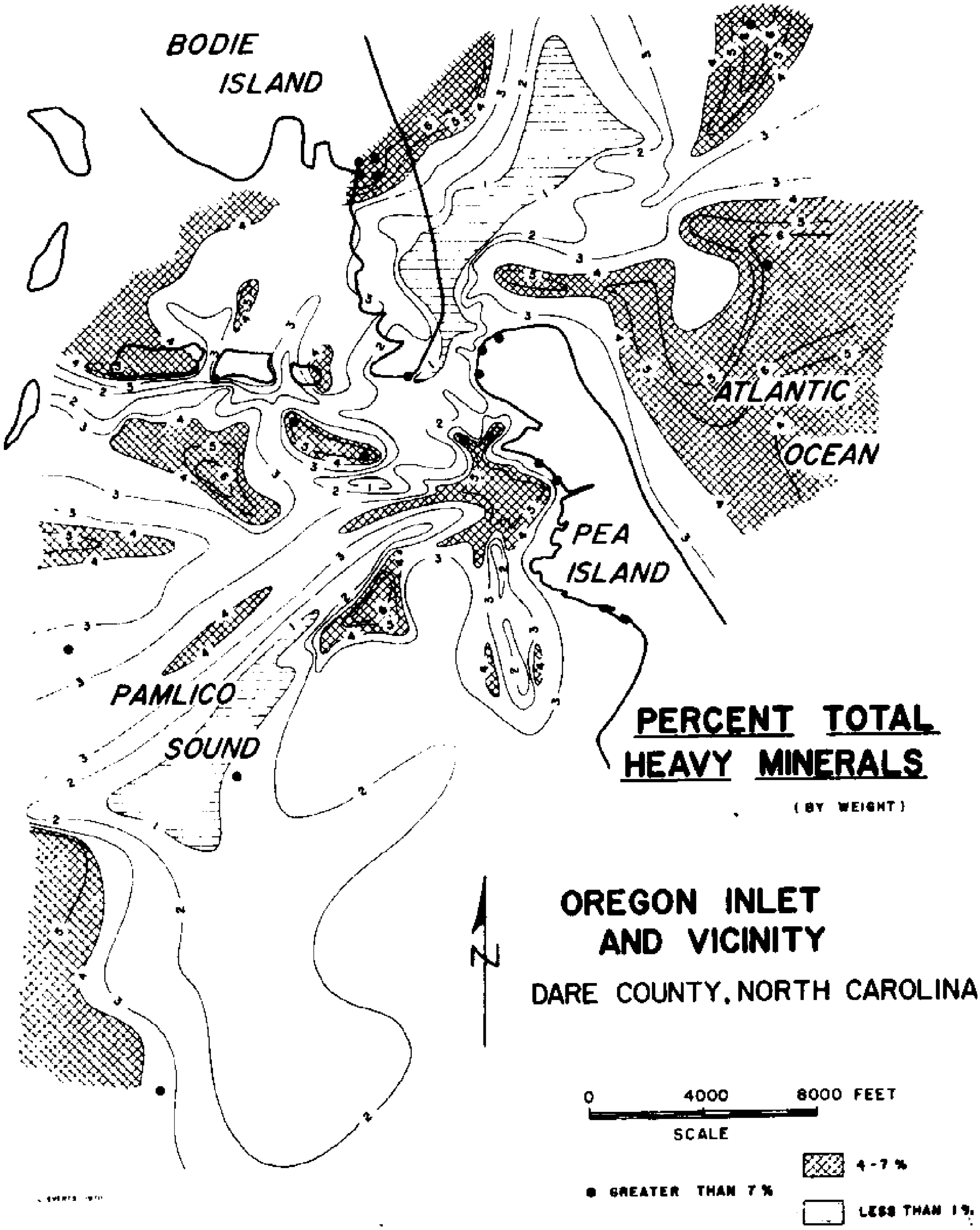
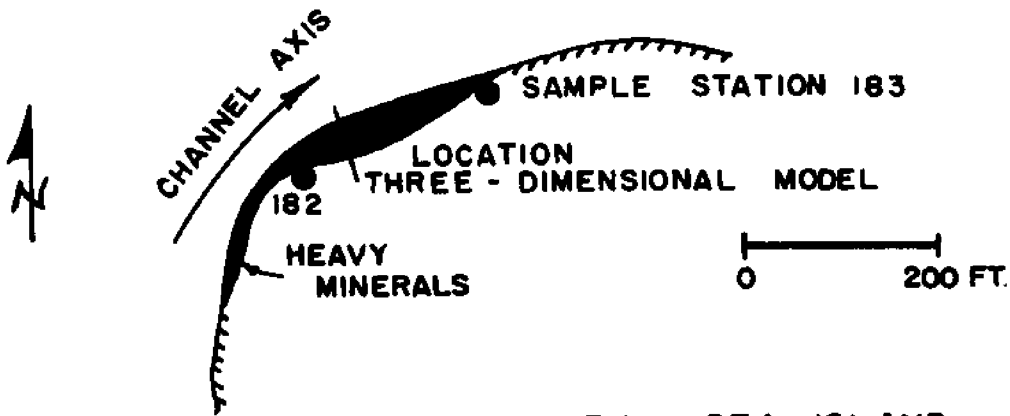
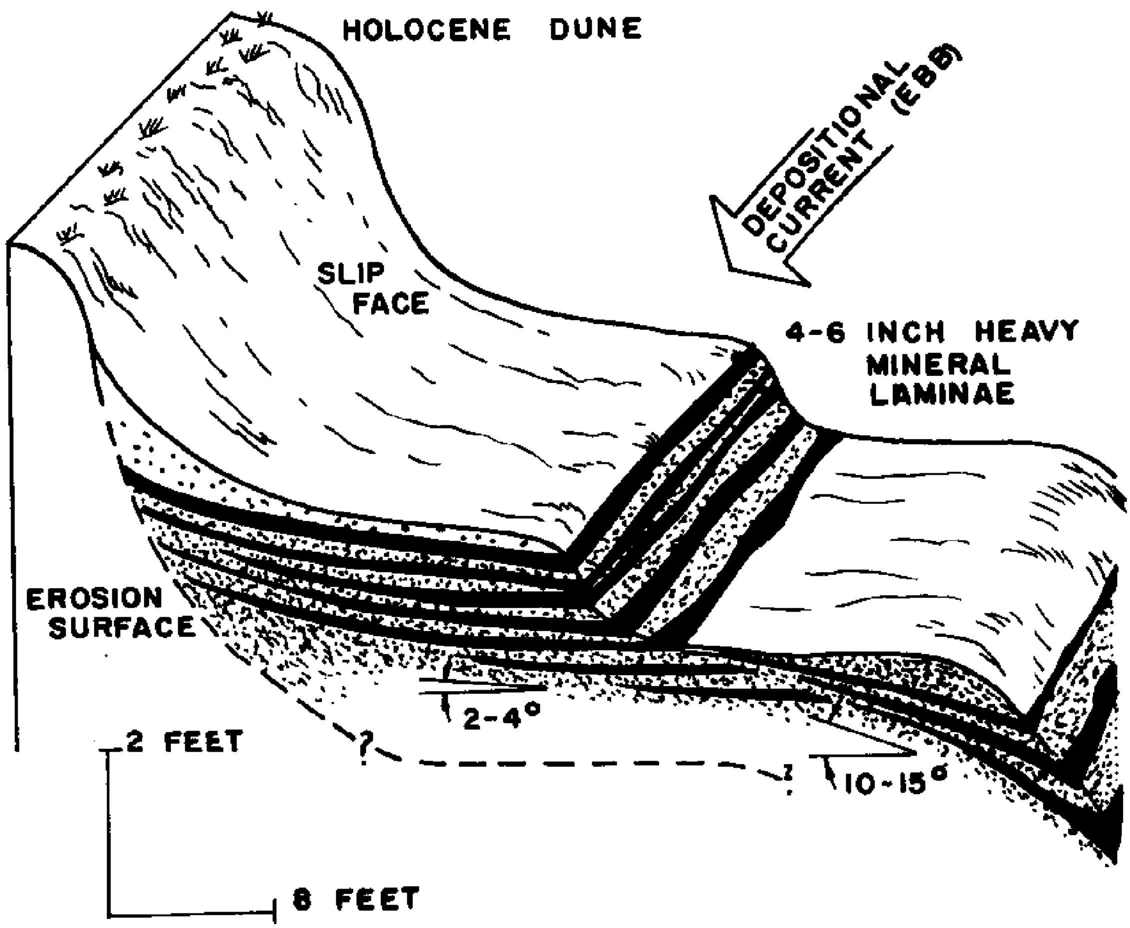

BODIE
ISLAND
ATLANTIC
OCEAN
PEA
ISLAND
PAMLICO
SOUND
PERCENT TOTAL
HEAVY MINERALS
(BY WEIGHT)
OREGON INLET
AND VICINITY
DARE COUNTY, NORTH CAROLINA
N
0 4000 8000 FEET
SCALE
4-7 %
GREATER THAN 7 %
LESS THAN 1 %

Figure 6-12



PLAN VIEW - PEA ISLAND



THREE - DIMENSIONAL MODEL OF PLACER

HEAVY MINERAL CONCENTRATION,
OREGON INLET, N.C.

velocity conditions over the previously eroded bench are optimum, heavy minerals are deposited (Figure 2-7 and 2-9), light minerals are bypassed (Figure 2-8), and a placer lamina is formed. Since flow depth increases as Pamlico Sound water is piled against the western shore of the barrier islands, the deposit will increase upward. Sediment availability is a significant factor in forming a lamina, and along with cyclic changes in bed shear due to changing velocity and depth over the bench, it is responsible for the vertical sequence of laminae shown in Figure 6-12. The boundary sharpness between various laminae indicate the close cutoff between heavy and light mineral depositional conditions.

Conditions favoring the formation of significant placer laminae in the throat are: (1) a wide depositional environment such as an erosional bench, (2) suitable avilability of heavy minerals passing the region, (3) a shallow depositional site which encourages heavy mineral loss and light mineral bypassing, (4) relatively steady flow at an optimum bed shear capable of concentrating placer grains, and (5) formation above sea level in an area favoring long-term preservation. The last criterion is lacking at Oregon Inlet because of its southward migration.

Ingram (1968) discussed a similar narrow placer containing 5 to 30 percent heavy minerals located between a tidal channel and a beach on the sound side of a barrier island. The location is near Beaufort Inlet, North Carolina, on the eastern end of narrow Bogue Sound. Beaufort Inlet is a distributory system closely resembling Oregon Inlet except that Bogue Sound is much narrower. The Beaufort Inlet placer probably formed at lower water level conditions than that at Oregon Inlet, but under similar circumstances.

That section of the Oregon Inlet throat that lies below mean sea level is barren in heavy minerals on the surface (Figure 6-11). The total percentage lies well below the two percent average for the inlet region. The implication is that the deepest channels, those containing the sedimentary deposits most likely to be preserved during a transgression, lack placer minerals and will be unlikely prospects in a relic environment.

The northern flood-dominated shore of Oregon Inlet is an optimum preservational site due to the southward migration of the throat. A small laminar-type placer is located at the extreme southern tip of Bodie Island. It lies on a coarse (greater than 0.3 mm) depositional bench near the mean high tide level, and landward of a steep submerged channel bank. Total thickness of the deposit is one foot while individual laminae vary from 0.5 to 3 inches. The placer apparently is forming and prograding with the spit in response to incoming beach-derived heavy minerals carried by the flood current. The spit tip is a region of decreasing current velocity and shallow depth, and is therefore ideal for placer deposition. Unfortunately, optimum depositional conditions prevail over a limited area and the surface extent of the deposit is small.

Neiheisel (1962) observed multiple-event heavy mineral deposits on the updrift shore of Jekyll Island, Georgia, that appear to have been formed under conditions similar to those which produced the prograding spit deposit at Oregon Inlet. At Jekyll Island the deposits are parallel to the shore, one to three inches thick, and 10 to 30 feet wide. They are formed on a bench near or above high water level. The surface seams are reported as now being overridden by coastal dunes.

OFFSHORE PLACERS

Offshore placers are located in three regions: (1) immediately seaward of the main channel, (2) south of the main channel and on the south shoal, and (3) near a washover region on the Bodie Island spit (Figure 6-11). Each represents a slightly different depositional mechanism although all but the last were formed in a ripple environment.

The northeasterly trending heavy mineral concentration near the seaward terminus of the main channel was formed in response to a storm surge in Pamlico Sound. The directional nature of the deposit suggests an ebb-current mechanism and the 40 foot depth of the deposit and its distance from the throat suggest formation following an intense outward flow. Figure 4-8 indicates little likelihood of an offshore heavy mineral source.

The northern concentration of heavy minerals appears to be areally connected with the laminar Bodie Island washover deposit. Several discontinuous laminae were observed with depth on the spit, each was less than one inch in thickness. The spit deposit was formed during a washover from the northeast as evidenced by the skewness gradient across the island (Figure 6-6). The location of the most concentrated laminae, at the southern terminus of the coastal dune complex,suggests a storm surge piled water on the beaches to the north with its only exit along the beach and across the island at a topographic low. It should be noted the northern beaches contain significant quantities of heavy minerals (Figure 5-9). The dunes on Bodie Island would be expected to migrate south thereby burying and preserving the washover deposit.

The concentrating mechanism of the placer seaward of the spit is not apparent; however, it is located near the updrift margin of the north shoal suggesting longshore drift and entrapment at the proximal end of the shoal.

The largest of the inlet placer regions is located east and south of Oregon Inlet (Figure 6-11). This disseminated-type placer locality, containing four to six percent heavy minerals, appears to have been formed as the placer minerals moved out from the throat. Two lobes of the deposit trend east from bathymetric highs in the main channel suggesting the opposing flow fields observed near the highs sorted and moved the heavy grains and associated fine-grained light particles seaward. Under the action of shoaling waves, the deposit is turned south near the outer margin of the south shoal.

A broad linear sedimentary feature of low heavy mineral concentration (Figure 6-11), and coarse light mineral composition (Figure 6-4), bisects the offshore inlet region. The continuity, directional trend, and outer depth of this feature suggests the amount of placer grains bypassing the inlet on its outer bars is insignificant. This means heavy minerals must enter the throat from the beach zone in order to bypass Oregon Inlet, thus allowing for concentration and deposition in the sound or throat regions.

SOUND PLACERS

Several relationships appear between disseminated placer occurrences in Pamlico Sound and their location with respect to tidal channels (Figure 6-11). Shoals adjacent to, and generally on the south or southeast margins of, or the terminus of, the main and north channels often contain anomalous heavy mineral values. The shallowest regions of the shoals and those near steep channel banks are likely locations for heavy mineral concentrations while channels contain low percentages.

Placer accumulations are elongate deposits in regions of good sorting and positive skewness indicating depositional conditions characteristic of fine-grained material in the absence of coarse (greater than 0.25 mm) grains. All deposits are formed on rippled surfaces. High flood current intensity may have transported the heavy grains to the depositional site while ebb currents return light minerals to the channels. Ripples moving only light grains were observed during ebbing conditions on most shoals suggesting that ebb currents winnow the initial flood deposit. The upper one-half to one inch of surface sediment on shoals was observed to be poor in placer grains which further supports the preceding winnowed deposit hypothesis. The depth of ripple reworking was one to four inches suggesting new flood-derived placer particles must be deposited on the shoals for each few inches of enriched placer formed.

Since Pamlico Sound channels serving Oregon Inlet migrate as the throat migrates, all shoals and heavy mineral accumulation south of the channels will also be modified with little chance of preservation. Inasmuch as the placer deposits in the sound regions are shallow, preservation during a transgression will not be as likely as it would be if the placers occurred in the deeper channels. In short, heavy mineral accumulations farthest from the throat, i.e. at the landward terminus of the tidal channels, have the best opportunity for preservation during a transgression.

CHAPTER VII

SUMMARY AND CONCLUSIONS

This study has attempted to clarify the processes responsible for the formation of certain specified marine placers. The main conclusion to be drawn is that there are two types of loose boundary marine placers: 1) the laminar deposit formed on a flat bed, and 2) the disseminated deposit formed in a ripple environment.

LAMINAR-TYPE PLACER

1. A laminar deposit is the most important of the marine placer types because it often approaches a one-hundred percent concentration of heavy minerals. It also contains a high ratio of economically significant minerals such as rutile, ilmenite, monazite, and zircon and a paucity of near-worthless garnet, amphibole, epidote, and other low density non-opaque minerals.

2. Concentrated heavy mineral laminae are formed in either a uni-directional or reversing flow regime.

3. Heavy minerals in placer laminae and bounding light mineral laminae are well sorted indicating relatively steady flow conditions at the time of deposition.

4. Skewness values of laminar placers are neutral or slightly positive suggesting steady, slightly decreasing flow intensity during deposition.

5. Heavy minerals (density of 3.3 to 5.2 gm/cc), on a flat bed composed of light mineral grains in the 0.1 to 4.0 mm size range, move only when the entire bed moves. In order to deposit placer particles the light mineral bed must be erosional or in mass equilibrium with the transported material.

6. Mean flow velocity is partially responsible for segregating heavy and light minerals, however, it is not sufficient to completely describe bed shear. Flow depth is the other critical factor.

7. Placer grains are deposited when the bed shear is below critical. The thickness of a lamina is dependent on heavy mineral availability at the depositional site and the uniformity of the depositional boundary. The longer optimum flow conditions prevail, the greater will be the lamina thickness.

8. Light minerals bypass a heavy mineral bed when bed shear is less than that required to initiate motion in the placer boundary, but greater than the minimum necessary to commence deposition of light minerals. Bypassing is greatly aided in very shallow flow situations. Shear on large light mineral particles increases at a higher rate than shear on a fine-grained bed as the flow depth decreases. Flat laminae preferentially form in shallow water when the Froude Number is approximately one.

9. The preservation of a heavy mineral boundary is assisted when the bed has a low sorting value and positive skewness. Each results in a smooth boundary easily bypassed by light mineral particles.

10. Flow intensity must slowly decrease or remain steady, and/or heavy mineral availability must increase, for a lamina to thicken and be preserved.

11. In order to preserve a laminar deposit over a long time period, conditions of formation dictate creation above mean sea level. Deposition below mean sea level usually results in subsequent modification of the deposit by near-normal flow fluctuations.

DISSEMINATED-TYPE PLACER

1. A disseminated-type deposit may contain two to six times the heavy mineral concentration existing in the average sediments of a region. The depositional ripple environment cannot distinguish between low and high density heavy minerals and therefore concentrates minerals according to their equivalent sizes.

2. Heavy minerals and the equivalent light mineral sizes are deposited in the trough of a bed form as it migrates over its boundary. Preservation is only possible when the boundary is aggrading. Material is transported on the ripple crests.

3. The critical shear necessary to deposit heavy minerals (concentrate them in an aggrading trough) was observed in the field to be approximately 0.6 times the critical shear necessary to maintain a uniform flat bed in the flume. When the bed shear exceeded this limit the heavy minerals moved on the ripple face and bypassed the region.

4. Heavy minerals in a ripple environment may bypass a depositional surface composed of material coarser than their equivalent light mineral size. Coarse light minerals may therefore be deposited in the ripple trough while heavy grains bypass them.

5. The highest heavy mineral accumulations in disseminated deposits are associated with well sorted material of positive skewness suggesting a slightly decreasing flow intensity with time.

6. Disseminated deposits may occasionally be reworked and enriched as some light minerals are winnowed to the depth of ripple action. The amount of light minerals remaining in the deposit is dependent on the ratio of heavy and light minerals of equivalent size in the depositional system.

OFFSHORE PLACERS

1. The most important offshore placer variable is heavy mineral availability. This should be the first criterion considered in planning an exploration program.

2. Heavy minerals move and are deposited in the offshore oscillatory wave zone in a ripple environment. Disseminated deposits form when placer grains are concentrated in ripple troughs on an aggrading bed.

3. Heavy minerals move offshore through the influence of river currents, inlet flushing, rip currents, and storm induced undertow. In the absence of seaward trending currents, placer minerals do not move seaward through the breaker zone. Sediment movement in response to shoaling gravity waves is onshore.

4. Tidal currents may be an important mechanism in moving and concentrating placer minerals in constricted areas where they may be deposited in bathymetric or hydraulic traps. On open shelves, currents, except off the mouths of rivers or inlets, are not likely to be the singular factor in placer formation.

5. Offshore shoals when nourished with placer minerals from rivers or inlets may retain their heavy minerals during a transgression when the seaward-impinging waves are insufficient to move the entire shoal offshore. The shoal would be expected to remain as an offshore bathymetric high.

6. An important offshore placer possibility is a relic lower beach, lagoon, sound, or coastal plain stream placer sequence that was not eroded during a transgressive cycle. Optimum conditions for such preservation include a rapid rise in sea level, an abundant sediment input onto the wave-active face, or a low intensity wave climate on the coast. The bathymetric expression of a relic offshore deposit may be a submerged ridge or terrace.

7. Placers forming under the influence of waves alone will be rare because the wide range of wave state and large bottom area precludes the steady or quasi-steady flow conditions in a limited area necessary for heavy mineral accumulation.

8. Surface sampling for offshore heavy minerals is effective only in locating contemporary heavy mineral accumulations and will likely miss the more important non-eroded relic deposits. Geophysical techniques will be required to locate buried placers, which must be confirmed by coring or relatively deep dredging.

BEACH PLACERS

1. Beaches are optimum regions for concentrating heavy minerals because they provide a mechanism for creating and preserving laminar deposits. The present stillstand provides a maximum opportunity for wave action to transport heavy minerals from offshore relic deposits and to move placer grains in the beach conduit.

2. Placer minerals move preferentially on the foreshore and to a lesser extent in the surf zone. Few heavy minerals move in the breaker zone.

3. Heavy minerals on the foreshore move slower than associated light minerals in the longshore conduit.

4. Stormline placers are located in the most landward storm berm, generally at the base of coastal dunes. This region provides the greatest opportunity for long-term preservation.

5. Preservation is unlikely on the foreshore and berm because of fluctuations in the beach profile in response to varying wave and tidal conditions.

6. Stormline laminae form at the end or near the end of a light mineral erosive cycle on the backshore. As water level or wave intensity lessens, heavy minerals are the first to be deposited. As long as bed shear conditions remain conducive to placer grain deposition, and light mineral bypassing, the laminae will thicken. Heavy minerals are deposited on the backwash cycle while light grains return to the step.

7. Long-term uniform conditions of heavy mineral deposition build the thickest laminae. Beaches where storms are of relatively the same intensity and periodic in nature may concentrate significant quantities of placer minerals over a long time period.

8. Heavy minerals concentrated in "summer" berms formed under moderate sea conditions may supply placer grains to stormline deposits formed during high intensity storm conditions.

9. The terminus of a littoral drift system, i.e. a topographic barrier to sediment transport, should have a sediment buildup. Heavy minerals may, however, be concentrated because storm erosion removes the seaward-deposited light minerals. Light minerals may also be lost along the littoral conduit to features such as submarine canyons.

10. There is little likelihood of a wide placer being formed on a transgressive beach even though the heavy minerals move up the beach as it transgresses the coast. The opportunity for a stormline deposit to be preserved during a transgression is slight and will be the result of a prograding beach as the sea level rises, however, such a beach would provide optimum conditions for placer formation.

The rise in the location of stormline deposition as water level rises will increase the total depth of the deposit.

11. Regressive beaches are optimum sites for heavy mineral preservation.

INLET PLACERS

1. Heavy minerals move into the inlet system in the littoral transport conduit. There is little possibility of significant quantities of placer minerals entering the region from offshore sources.

Placer minerals are transported past the inlet in a path that takes them through the throat and into the sound. Ebb currents then flush them seaward on the downdrift shore of the inlet. Although large quantities of light minerals appear to pass the inlet on its offshore bars, few heavy minerals follow that route.

2. Tidal currents cut channels in original sound-formed sediments. This fine-grained material is displaced by littoral material as the channels migrate. The fine-grained sediments move offshore or landward into quiescent regions in the sound. Moreover, heavy minerals are not concentrated in channels of the inlet system. They, therefore, will not be found preserved as relic channel deposits formed as the result of longitudinally migrating barriers.

3. Shoreline regions near the inlet throat provide conditions favorable for heavy mineral laminae formation provided: (a) a wide depositional environment such as an erosional bench exists, (b) suitable quantities of heavy minerals pass the depositional site during high flow conditions, (c) the depositional site encourages shallow flow, (d) relatively steady flow conditions prevail for a sufficient period of time to form thick laminae, and (e) formation is above sea level in an area favorable for longterm preservation. The latter case is encouraged by updrift deposition. A deposit formed on the downdrift shore will be eroded as the inlet migrates.

A favorable location for a laminar-type spit placer is immediately downdrift of a coastal dune complex. During a storm surge, water piles on the beach and moves in a downdrift direction. The flow breaches the barrier at the first topographic low it encounters and moves toward the sound as a shallow sheet flow ideal for the deposition of heavy minerals if any are present.

4. Disseminated-type placers form at the terminus of the main tidal channels both offshore and in Pamlico Sound. Disseminated placers also occur on shallow shoals adjacent to steep margins of major tidal distributory channels in Pamlico Sound.

SUGGESTIONS FOR FURTHER STUDY

There follow suggestions for future studies of marine placers. The results of these studies would not necessarily be of equal value, nor is the list meant to be comprehensive.

Still needed is a third-dimension look at the placer types described in this report. Although the surface geometry of the heavy mineral accumulations is known, little can be inferred as to their depth dimension.

Individual deposits should be systematically studied on a fine scale in order to pick subtle changes in the texture of the heavy and light minerals. Data obtained would assist the investigator in defining the actual genesis of the deposit.

The hydraulics responsible for the creation of disseminated ripple environment deposits should be investigated to provide dynamic limits on flow conditions during deposition.

Flume experiments concerned with a wider range of heavy mineral densities should be initiated. Experiments dealing with various sorting and skewness values and with undulating boundaries would greatly aid in defining natural placer behavior.

Inferences of natural sediment transport rate and direction based on neutrally bouyant seabed drifter observations are questionable. An investigation of the behavior of each in a controlled wave environment and under field conditions is required.

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