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MAPS SHOWING DIRECTIONS OF LONGSHORE SEDIMENT TRANSPORT ALONG THE ALASKAN BERING SEA COAST

by

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INTRODUCTION

The interpreted directions of net longshore sediment transport along the Alaskan mainland coast of the Bering Sea, from Bering Strait in the north to Unimak Pass in the south, are shown on sheets 1 to 5 of this report. Longshore sediment transport is the movement of sediment, mostly sand and gravel, in alongshore directions on beaches and in the adjacent shallow-water zone of breaking waves. On most coasts, waves and wave-driven currents are the main transporting agents, but tidal and other currents may be locally important causes of transport. We do not attempt to show the directions of transport in deeper water, where tidal and other currents are more likely to be important.

The rate of longshore transport increases with increasing wave size. Where wave size is constant, the transport rate is greatest when the waves approach shore at an angle of about 45° (Zenkovitch, 1967; Komar, 1976).

Longshore transport at any given point along a coast can be in either of the two shore-parallel directions. Along many coasts the transport direction alternates many times a year as wave conditions change. Thus, the direction of net transport over a period of a year may differ from the direction at any instant, and the direction of net transport over many years may differ from the direction over a single year. The symbols shown on the summary map (sheet 1) are the interpreted directions of long-term net transport, where long-term means at least several years and perhaps as many as a few thousand years. The locations of the specific features used in making these interpretations are shown on sheets 2 to 5, and the symbols used on these sheets are explained on sheet 1.

Directions of longshore transport can be determined by several methods, such as monitoring the movement of artificially introduced tracer grains (Ingle, 1966), calculating the transport from known wave conditions by the use of a formula (Komar and Inman, 1970), making interpretations from longshore variations in the grain size or composition of the beach sediments, or making interpretations from coastal landforms. All these methods have advantages and disadvantages. Monitoring the movement of tracer grains, for example, is the most reliable method for determining short-term transport but would be a very time-consuming method of determining the long-term net transport. The use of a formula is probably fairly reliable but requires extensive wave data that are not now available for the Bering Sea. The methods based on grain size, sediment composition, and landform interpretation are moderately reliable; of these, the interpretation of landforms is the quickest and most inexpensive method.

The directions of longshore transport shown in this report were interpreted from coastal landforms. The interpretations were made from aerial observations during two summers and from maps and aerial photographs. Only certain kinds of coastal landforms, which are described later in this report and sketched on sheet 1, indicate the direction of longshore transport.

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GENERAL PROBLEMS OF INTERPRETING TRANSPORT DIRECTIONS FROM LANDFORMS

One of the more serious problems in interpreting the directions of longshore transport from coastal landforms is that some landforms mimic those produced by longshore transport but in actuality owe their form to other causes, such as the geologic structure of the bedrock. A problem leading to similar uncertainties is that some coastal landforms can be interpreted as indicating transport in either of the two longshore directions. On sheets 2 to 5 of this report we indicate by question marks those interpretations that we judge to be questionable.

Another problem is to estimate how much time was required for a landform to develop. The largest landforms take many centuries to develop and thus indicate the long-term transport direction, whereas the smallest landforms, such as oblique bars, may be produced during a single storm and thus indicate only short-term transport. Depositional landforms, which are composed of unconsolidated sediment, may develop more quickly than erosional landforms cut in bedrock. We distinguish by the symbol "m" those minor landforms that we judge could have formed in less than a year. The significance of minor landforms is increased, however, if they are numerous along a section of coast at any one time or if they were observed during several years; we note multiple occurrences by the symbol "s". We also note by a special symbol (an arrow marked by an X) those short-term drift directions that we consider to be opposite from the directions of long-term net drift.

Still another problem is that a landform may cause a change in the pattern of transport as it develops. Many coastal landforms, such as hooked bays, develop in ways that reduce the net transport rate (see, for example, Zenkovitch, 1967); these may eventually reach equilibrium states, in which there is no net transport. Some coastal landforms, such as spits, sometimes produce reversals of transport direction as they grow and shelter downdrift parts of the coast from the dominant waves. A few of the landforms in the Bering Sea may have reached equilibrium; if so, the transport directions shown on the maps indicate the directions during the development of these landforms.

The absence of landforms that indicate transport direction along a stretch of coast does not necessarily mean that transport is negligible. If landforms indicate transport in the same direction at two points along the coast and if the intervening shoreline is straight or curves gradually from one point to the other, it can be assumed that the transport is in the same direction along that stretch of coastline. Considerations of this kind were used in preparing the generalized map of transport directions (sheet 1) from the specific features shown on sheets 2-5. On some sections of coast, such as the active Yukon River delta, landforms that indicate transport direction are too rare for interpretations to be made even with these considerations.

The grain size of the available sediment affects the interpretation of a coastal landform. Coarse gravel can be moved only by the largest waves, whereas fine sand or silt can be moved by relatively small waves. It is conceivable, then, that fine-grained sediment might be transported in a different direction from coarse-grained sediment if the smaller waves moved in a different direction from the larger waves and if the smaller waves persisted long enough to transport more fine-grained sediment than was transported by the infrequent larger waves.

TYPES OF LANDFORMS THAT INDICATE TRANSPORT DIRECTION

As a general rule, if the direction of longshore transport is to be inferred from a coastal landform, the landform must be asymmetric with respect to a line normal to the generalized shoreline. In most coastal landforms, the asymmetry of the shoreline itself provides a sufficient basis for interpretation. Coastal landforms have been described by many other workers, including Zenkovitch (1967), King (1972), Silvester (1974), and Komar (1976).

Shoreline Offsets

Offsets or steps in the map view of a coastline were recognized as an indicator of transport direction by Gulliver (1896, 1899). Geometrically, offsets are of two types. In one type, a downdrift section of the shoreline is offset seaward from an updrift section (sheet 1, symbols G_H, G_G, G_J, G_R, and G_I). In the second type, a downdrift section of the shoreline is offset landward from an updrift section (sheet 1, symbol I). Offsets can be produced entirely by erosion, entirely by deposition, or by a combination of the two effects. They occur at resistant headlands, at manmade structures, and at channel mouths.

Offsets at Groins

Groins are manmade structures placed across a beach to slow down or halt coastal erosion. They produce a downdrift-landward offset (sheet 1, symbol G_G) by causing deposition or lessening erosion updrift of the structure. Because most groins trap relatively small amounts of sediment, the offsets visible at any one time do not necessarily indicate the direction of longterm net transport. The only examples of what might be called groins in the Bering Sea are some accidental collections of metal debris on the beaches near Nome.

Offsets at Headlands

Headlands of resistant bedrock tend to act like groins, producing downdrift-landward offsets (sheet 1, symbol G_H), and in fact they have been called "natural groins" (Inman and Frautschy, 1965). Most offsets at headlands are produced by differential erosion, but deposition updrift of the headland may be involved in the formation of some offsets. Offsets at most headlands are large-scale features that indicate the direction of long-term net transport.

When inferring transport directions from offsets at headlands, one must take care to rule out the possibility that the offset was produced directly by faulting or folding, by differing erosional resistances of the bedrock on either side of the headland, or by differing amounts of sediment supplied to the shoreline on either side of the headland. Such possibilities can most easily be ruled out where the offsets occur in a series, forming what might be called a "stepped shoreline."

Offsets at Jetties

Jetties are manmade seaward-extending structures built at the sides of stream mouths, tidal inlets, or harbors to prevent shoaling or migration of the channel. They tend to act like groins, producing a downdrift-landward offset of the shoreline across the channel mouth (sheet 1, symbol G_J). However, the complicated patterns of water flow and sediment transport in the vicinity of the channel mouth can produce exceptions to this tendency; these possible complications are discussed in the next section. The only jetties in the Bering Sea are the pair at the mouth of the Snake River in Nome.

Offsets at Tidal Inlets

Offsets at tidal inlets can be either downdrift-landward or downdrift-seaward (sheet 1, symbols G_I and I). The causes of these offsets have been studied by Todd (1968), Hayes and others (1970), Goldsmith and others (1975), and Lynch-Blosse and Kumar (1976), but are still not completely understood. Among the factors responsible for offsets at inlets are tidal currents and wave refraction, both of which can be very complex. Another factor probably important in controlling the kind of offset is whether the inlet is a net sink (that is, an area in which sediment is lost from the transport system) or a net source for sediment moving along the coast. If the inlet traps sediment coming from the updrift direction and does not supply an equal amount to the downdrift coast, it is a net sink, just as a groin is, and will tend to produce a downdrift-landward offset. If the inlet is a net source of sediment, it will tend to produce a downdrift-seaward offset.

It may not be easy to determine if a tidal inlet is a net source or sink of sediment. One indication is the size ratio of the ebb-tidal and flood-tidal deltas; a flood-tidal delta that is larger than the ebb-tidal delta suggests that the inlet is a net sink, and conversely. An inlet that migrates alongshore and cuts into a large body of sand or gravel is likely to be a net source of sediment.

Offsets at Stream Mouths

If the interaction of stream mouths with the longshore transport follows the rule hypothesized for tidal inlets, all stream mouths should be characterized by downdrift-seaward offsets, for all streams are net suppliers of sediment to the sea. In actuality, no offsets of any kind occur at the mouths of many small streams, evidently because they are too small to have any significant effect.

Surprisingly, many small streams in the upper part of Bristol Bay and in southern Norton Bay have downdrift-landward offsets at their mouths (sheet 1, symbol G_R), the direction of net drift being confirmed by other evidence. All offsets of this kind occur only on low, eroding coasts vegetated by tundra and underlain by unconsolidated clay, silt, or peat. Evidently the streams draining the low tundra carry very little sediment and somehow act as groins, perhaps by their water discharge stopping part of the littoral drift, as suggested by the common occurrence of small spits at the updrift sides of the stream mouths.

Hooked Bays

Certain bays have a distinctively asymmetric plan form (sheet 1, symbol H) as noted in 1906 by Halligan (quoted by Davies, 1973, p. 136). The significance of such bays as indicators of transport direction has been stressed by Silvester (1960). These bays, which have variously been called zetaform bays, spiral bays, headland bays, half-heart bays, crenulate bays, and hooked bays (Davies, 1973; LeBlond, 1972; Rea and Komar, 1975; Silvester, 1960, 1970, 1974; and Yasso, 1965), are produced by coastal erosion downdrift of headlands. Although the reasons for their form are not fully understood, they seem to be one of the more reliable indicators of the direction of long-term net transport. They are commonly associated with offsets at headlands.

Channel-Mouth Deflections

The fact that stream mouths tend to become deflected in a downdrift direction (sheet 1, symbol D_R) has been known for many years (Gulliver, 1896, 1899). Such deflection is caused by (1) the halting of longshore transport as the sediment reaches the deeper water of the channel, (2) the consequent tendency for the channel to be filled on the updrift side, (3) the diversion of the stream flow to the downdrift side, and finally, (4) the erosion of the downdrift side of the channel. Channel mouths can migrate many miles downdrift by the continued operation of this process. Tidal inlets can be deflected in the same way (sheet 1, symbol D_T) and therefore commonly migrate to the downdrift ends of the lagoons that they drain (Price, 1952).

Channel-mouth deflections caused by longshore transport can be confused with those due to other causes. Some deflections are caused by longshore variations in wave height, these variations being caused most commonly by wave refraction (Bascom, 1954). Deflections of this kind are produced when storm waves build a berm or beach ridge that dams a stream mouth, whereupon the impounded water breaks through the lowest part of the berm. Since the height of the berm is roughly proportional to wave height, the point of breakthrough is controlled by longshore variations in wave height. As deflections of this kind are produced suddenly, any evidence of a gradually increasing deflection rules out this process and indicates an origin by longshore transport.

Stream-mouth deflections take different forms on stable coasts, on straight prograding coasts, and on deltas. A deflected stream on a stable coast flows parallel to the shoreline in back of a berm ridge for some distance, whether the deflection is produced by longshore transport or by longshore variations in wave height. Streams that flow obliquely to the shoreline across a straight prograded strand plain can be interpreted as having been gradually deflected by longshore transport (Allen, 1965); some well-developed deflections of this kind occur on the north side of Bristol Bay between Cape Constantine and Kulukak Point. Deflections of delta-building streams can also be ascribed to the gradual effect of longshore transport during delta growth; some well-developed though small-scale examples occur on the east sides of Kuskokwim Bay and Norton Bay. A delta itself is likely to be made asymmetric by longshore transport (Komar, 1973).

Depositional Bodies

Oblique bars. Bars are ridges of sand or gravel in the intertidal or shallow subtidal zones of a beach (sheet 1, symbol O). Only bars that are oblique to shore can be used to interpret transport directions; oblique bars trend seaward in a downdrift direction (Guilcher, 1974). If the bars are submerged, they may be visible from aircraft if the water is clear; otherwise, their form may be indicated by the pattern of breaking waves or stranded ice on the bar crests. Also, oblique bars can often be recognized from protuberances of the shoreline where the bars are attached to shore; these protuberances are called giant cusps, though not all giant cusps are associated with oblique bars. Most oblique bars are relatively small depositional bodies and thus do not necessarily indicate the long-term net transport direction.

Spits. Spits differ from oblique bars in being at least partly above the high-water line. Some spits protrude from a shoreline that was relatively straight before the spit grew, whereas others continue the line of an updrift shoreline where the shoreline originally curved landward into an embayment; in either type, the free end of the spit points in the downdrift direction (sheet 1, symbol S). Most spits are large enough to indicate the direction of long-term net transport. They are one of the most reliable indicators of transport direction.

A spit may produce a local reversal in transport direction in the area behind and downdrift of the spit by protecting that part of the shoreline from the dominant waves while leaving it exposed to waves from other directions. The diffraction of the dominant waves around the tip of the spit can also cause a reversed transport direction in the area behind the spit.

Cuspate forelands and spits. A cuspate foreland is a roughly triangular depositional plain that protrudes from the generalized shoreline (sheet 1, symbol F). The seaward tip of a cuspate foreland may be either sharply pointed or rounded. Smaller features of this type have been called cuspate spits. Most cuspate forelands and spits are composed of sediment that was carried to the depositional site by longshore transport, but some have been interpreted as having formed by other processes, such as modification of a previously formed delta (Hoyt and Henry, 1971). The origin of cuspate forelands and spits has been a subject of much discussion (Gulliver, 1896; Zenkovitch, 1959; Tanner, 1962; Price, 1964; Voskoboynikov, 1966; White, 1966; Hoyt and Henry, 1971; Swift and others, 1972).

Cuspate forelands composed of material carried by longshore transport can occur in two different settings, either where the net transport approaches the foreland from both sides or where the net transport is unidirectional but decreases in a downdrift direction. The second kind, which is typically asymmetric in form, commonly occurs at pre-existing shoreline irregularities. The few cuspate forelands in the Bering Sea seem to be of the asymmetric type.

Downdrift-tapering barriers. Barrier islands that have one wide, bulbous, rounded end and one narrow, pointed end (sheet 1, symbol T) are very suggestive of net transport in the direction of narrowing (Price, 1954; Hayes and others, 1973). This interpretation is based on the similarity of a tapering island to a spit.

Depositional Ridges Marking Former Shorelines

Many coastal depositional features are covered by ridges that mark former shorelines and thereby record the stages of growth of the depositional bodies. These ridges include (1) beach ridges, most of which are probably formed by wave swash during storm surges, (2) foredune ridges, which are formed by onshore winds blowing sand into vegetated areas behind the beach, where the sand is trapped, and (3) cheniers, which are beach ridges separated from one another by tidal flat or marsh deposits.

The pattern of ridges may permit the long-term transport direction to be interpreted even where the gross form of the depositional body itself does not furnish sufficient evidence (sheet 1, symbol B). The historical type of evidence furnished by beach ridges is of the same type furnished by repeated maps or aerial photographs of the coast. The principle on which interpretation is based is that a depositional coastal landform generally grows or migrates in a downdrift direction. However, some coastal landforms, such as cuspate forelands, can perhaps migrate updrift under certain conditions (R. C. H. Russell, in Williams, 1956), and this possibility should be considered when transport directions are being interpreted from the pattern of ridges.

GENERAL PROPERTIES OF LONGSHORE TRANSPORT SYSTEMS

The patterns of longshore transport along shorelines can best be analyzed in terms of coastal cells (Inman and Frautschy, 1965; Longinov, 1965; Zenkovitch, 1967; King, 1972; Davies, 1973; Inman and Brush, 1973; Tanner, 1973; and Silvester, 1974). A coastal cell is defined here as a section of shoreline bounded by points of net divergence of the longshore transport; other authors use somewhat different definitions. Because a cell is defined as having divergence points at either end, it must have a convergence point somewhere within the cell.

Coastal cells are commonly marked by coastal erosion at and near the divergence points that bound the cells and by accretion at and near the convergence points. The divergence points tend to be located at headlands and the convergence points within embayments, but exceptions can occur.

Every coastal cell has a sediment budget consisting of the various sediment sources and sinks, which may be located onshore, offshore, or alongshore. The most common onshore sources are rivers, while the most common onshore sinks are coastal dunes. The most common offshore sources are eroding shelf deposits, while the most common offshore sinks are growing shelf deposits, shelf edges, and submarine canyons; canyons do not occur near the Bering Sea coast. The most common alongshore sources are eroding coastal bedrock and unconsolidated deposits, while the most common alongshore sinks are growing coastal depositional bodies.

INTERPRETED DIRECTIONS OF NET LONGSHORE TRANSPORT

General Discussion

Wave-induced sediment transport along the eastern Bering Sea coast is seasonal. During winter it is curtailed by ice cover, which begins to advance southward in October, covers most of the coast by February, and remains on northern parts of the coast until June (Arctic Environmental Information and Data Center, 1974). Offshore ice inhibits the generation of large waves by limiting the wind fetch, and ice attached to shore or sheathing the beach gives protection against wave-induced sediment movement. In general, the period during which ice does not directly affect the coast varies from three months (July-September) in the northern part of the study area (Bering Strait to Norton Sound) to seven months (May-November) along the Alaska Peninsula coast of Bristol Bay.

Coastal weather stations in the southern Bering Sea (Cape Newenham and Port Moller) report an essentially bimodal distribution of wind directions for the ice-free season, with the stronger component from the south and the other from the northwest (Arctic Environment Information and Data Center, 1974). Net littoral drift directions, which are summarized on sheet 1, typically reflect transport by waves from one of these two directions. The Alaska Peninsula coast of Bristol Bay is protected from the south, and transport along the coast is to the northeast in response to waves generated by winds from the northwest or west. Between the Alaska Peninsula and Norton Sound the transport is generally to the north in response to waves generated by southerly winds. Wind directions recorded at Nome, in the northern part of the study area, are more nearly uniformly distributed. However, the north-trending sections of coast on the eastern side of Norton Sound and between Bering Strait and Norton Sound still reflect transport predominantly to the north. The east-trending sections of coast on the north and south sides of Norton Sound indicate transport to the east in response to the considerably longer fetch to the west.

During storms, much more sediment can be transported by waves than under normal conditions, and many of the coastal features used to interpret transport direction may have been formed largely during storms. A major storm track crosses the eastern Bering Sea from southwest to northeast during the latter part of the ice-free season (late July to early September) (Arctic Environmental Information and Data Center, 1974). The cyclonic air flow about the low-pressure storm systems following this track causes a northward air flow along the coast, which correlates well with the interpreted dominantly northward transport along the coast.

Discussion of Sheets 2-5

For this report the nearly 3,000 km of open coastline of the eastern Bering Sea has been divided into four sections. Coastal strip maps for each section have been compiled on sheets 2-5. Longshore transport directions and the geomorphic criteria used in interpreting the directions are indicated on these sheets. Sheet 2 covers the Alaska Peninsula coast of Bristol Bay and Unimak Island. Sheet 3 includes the coastal area between the Alaska Peninsula and the Yukon-Kuskokwim Delta complex and Sheet 4 includes the Yukon-Kuskokwim Delta. Sheet 5 covers the northern portion of the study area from Norton Sound to the Bering Strait.

Alaska Peninsula Coast of Bristol Bay (sheet 2)

This coast is relatively straight but is segmented by six large bays that are more or less protected from the Bering Sea by barrier spits or islands. Between the bays are subtle headlands, most of which are bluffs eroded in unconsolidated sediment. In general, the net transport direction along this coast is to the northeast with local reversals near the bays.

The coast is divided into a series of coastal cells in which the convergences of longshore transport occur at the bays and the divergences are a short distance northeast of each bay. The positions of divergence occur most commonly about 10 km to the northeast of each bay (Bechevin Bay, Izembek Lagoon, Port Heiden, Ugashik Bay, and Egegik Bay). The Port Moller area forms a much larger cell with a divergence occurring at Cape Kutuzof, 37 km to the northeast of Port Moller. The long-term effect of wave action in this system of coastal cells is to straighten the coast by eroding the headlands and filling the bays. One former cell, whose embayment was at the mouth of Cinder River, has been incorporated into the Ugashik Bay cell by the filling of the embayment, and other cells will be combined as coast straightening continues.

High-Tidal Estuaries (sheet 3)

Between the Alaska Peninsula and the Yukon-Kuskokwim Delta complex are a series of funnel-shaped embayments. As the tides move into these bays they are amplified, creating very large tidal ranges. For example, the diurnal range at Clarks Point in Nushagak Bay is 5.9 m and at Kvichak at the head of Kvichak Bay is 5.0 m. Within Kvichak, Nushagak and Kuskokwim Bays are found linear shoals that are alined parallel to the tidal currents. These shoals, approximately 10 m in height and spaced 4 km apart, are characteristic of high-tidal environments.

In general, the net longshore transport directions range from northeastward to northwestward depending on shoreline orientation. The bays, most of which are kept open to the Bering Sea by strong tidal currents, form coastal cells with convergence points at the heads of bays and divergence points at the headlands between bays. The transport along the shorelines of the bays is probably only a small fraction of the sediment transport caused by tidal currents in the deeper parts of the bays.

Yukon-Kuskokwim Delta Complex (sheet 4)

The Yukon-Kuskokwim delta complex encompasses the region between Kuskokwim Bay to the south and Norton Sound to the north. It is an area of relatively low relief, interrupted only by the Askinuk Mountains south of Scammon Bay and Nelson Island south of Hazen Bay. The coastline is extremely varied, in part because of the heterogeneity of source material and the variability in tidal range. Extremely broad tidal flats, locally bordered by short barrier islands, flank the macrotidal Kuskokwim delta, whereas the microtidal Yukon delta is fringed by prograding distributary-mouth bars and interdistributary tidal flats. Relatively coarse-grained sandy beaches occur in the vicinity of Hooper Bay, where eroding Pleistocene sand dunes supply the sediment. Steep gravel beaches occur along the cliffed shorelines in the vicinity of Nelson Island, Cape Romanzof and Point Romanof. Elsewhere, the coast consists mainly of eroding low bluffs cut into unconsolidated peaty silt and sands.

The general trend of nearshore sediment transport is to the north due to the combined effects of local winds and storms, southwesterly swell tidal currents, and north-flowing coastal waters. Where examined in detail, however, the patterns are more complex.

Most of the sediment along the northern margin of Kuskokwim Bay is transported perpendicular to the shoreline by tidal currents; such transport is not shown on the map sheets. Tidal flats are locally 10 km wide, with the result that wave action at the shoreline is negligible except during major storms; sediment transport by waves is important only on the sandy shoals along the outer fringe of the tidal flats.

The western margin of the delta complex between Cape Avinof and Cape Romanzof is dominated by northward sediment transport, as evidenced by oblique bars, prograding spits, and longshore variations in grain size. There are, however, some significant exceptions to northward transport. The shoreline of Hazen Bay is essentially erosional, with few depositional features to define a dominant direction of longshore transport. Most of the sediment appears to be transported perpendicular to the shoreline by tidal currents and so is beyond the scope of this report. There is a significant divergence of longshore sediment transport near Dall Point, with southward transport south of the divergence point as evidenced by Nuok Spit. This divergence appears to be caused by the refraction of southwesterly waves due to offshore bathymetry.

The Sand Islands north of Cape Romanzof show some evidence of northward transport; however, the direction of sediment transport along the adjacent mainland shoreline, where broad mudflats tend to dissipate the effect of waves, is less clear. A prograding spit at the mouth of the Black River may indicate some northward transport, but lobes of river-derived sand extending to the south over mud flats suggest that most of the sediment is transported to the south, perhaps in part by river-induced currents.

There are virtually no geomorphic indicators of longshore transport along the margins of the presently active Yukon delta. Much of the sediment is trapped in distributary-mouth bars or adjacent tidal flats. Much of the remaining sediment remains in suspension to be transported across the shelf (Nelson and Creager, 1977). The shelf is extremely shallow in this area, with breakers having been reported as much as 15 km offshore (Natl. Ocean Survey Nautical Chart 16240). The breadth of the shallow shelf tends to reduce wave height by frictional attenuation and to decrease the angle of wave incidence by refraction, thereby decreasing longshore currents (see Wright and Coleman, 1973).

Norton Sound to Bering Strait (sheet 5)

Norton Sound is a roughly rectangular embayment approximately 110 km wide and 220 km long. It has a remarkably uniform depth averaging about 20 m. Along the southern side of Norton Sound between the Yukon delta and Stuart Island, longshore transport is to the east. East of Stuart Island is a lava-flow coast composed of boulders and bedrock with little indication of net transport direction. Probably little longshore transport occurs along the lava-flow section of coast, due to the scarcity of material fine enough to be transported. Along the north-trending coast forming the east side of Norton Sound, the transport is to the north. Inlet migration to the north at Unalakleet threatens several houses at the south end of the village. The transport in Norton Bay, a relatively shallow embayment at the northeast corner of Norton Sound, is generally to the northeast, with convergence at Koyuk Inlet at the head of the bay. Golovnin Bay to the west of Norton Bay forms a closed coastal cell with northward transport into the bay. Between Golovnin Bay and the northwest corner of Norton Sound the drift is inferred to be more variable through time than on most parts of the Bering Sea coast. Stream-mouth deflections, a small sediment accumulation behind a jetty, beach ridges on the barrier spits enclosing Safety Lagoon, and spit growth at Safety Lagoon inlet all indicate transport generally to the east. One set of vertical aerial photographs of the region between Topkok Head and Sledge Island, however, shows longshore bars which are slightly oblique to the shoreline and attached to shore at their eastern ends, indicating drift to the west at the time of the photographs.

Coastal features between Norton Sound and the Bering Strait include a classic hooked bay, a recurved spit over 30 km in length encompassing Port Clarence, and rocky headlands west of Port Clarence. Between Norton Sound and Port Clarence the long-term drift is generally to the north. The morphology of the hooked bay may, however, have reached equilibrium with waves coming from the south, so that there may now be little net transport within the bay. Two cells operate within Port Clarence, one with convergence at the mouth of Grantly Harbor and the other with convergence at the southern margin of Port Clarence. Between Port Clarence and Bering Strait there is a divergence point near King River. To the east of the divergence, sediment is transported into Port Clarence, and to the west transport is toward the Bering Strait.

REFERENCES

Allen, J. R. L., 1965, Coastal geomorphology of eastern Nigeria: beach-ridge barrier islands and vegetated tidal flats: Geol. en Mijnbouw, v. 44, p. 1-21.

Arctic Environmental Information and Data Center, 1974, The Bristol Bay environment, a background study of available knowledge: Arctic Environmental Information and Data Center, Univ. Alaska, 858 p.

Bascom, W. N., 1954, The control of stream outlets by wave refraction: Jour. Geology, v. 62, p. 600-605.

Davies, J. L., 1973, Geographical variation in coastal development: Hafner, Riverside, N. J., 211 p.

Goldsmith, V., Byrne, R. J., Sallenger, A. H., and Drucker, D. M., 1975, The influence of waves on the origin and development of the offset coastal inlets of the southern Delmarva Peninsula, Virginia, in Cronin, L. E., ed., Estuarine research, v. 2: Geology and engineering: Academic Press, New York, p. 183-200.

Guilcher, Andre, 1974, Studies in coastal geomorphology contributing to coastal engineering: Conf. Coastal Engineering, 14th Proc., v. 1, p. 1-20.

Gulliver, F. P., 1896, Cuspate forelands: Geol. Soc. America Bull., v.7, p. 399-422.

________, 1899, Shoreline topography: Am. Acad. Arts Sci. Proc., v. 34, p. 151-258.

Hayes, M. O., Goldsmith, V., and Hobbs, C. H., III, 1970, Offset coastal inlets: Conf. Coastal Engineering, 12th Proc., v. 2, p. 1187-1200.

________, Owens, E. H., Hubbard, D. K., and Abele, R. W., 1973, The investigation of form and process in the coastal zone, in Coates, D. R., ed., Coastal geomorphology: Binghampton, State Univ. New York Pubs. Geomorph., p.11-41.

Hoyt, J. H., and Henry, V. J., Jr., 1971, Origin of capes and shoals along the southeastern coast of the United States: Geol. Soc. America Bull., v. 82, p. 59-66.

Ingle, J. C., Jr., 1966, The movement of beach sand: Developments in sedimentology 5: Elsevier, Amsterdam, 221 p.

Inman, D. L., and Brush, B. M., 1973, The coastal challenge: Science, v. 181, p. 20-32.

________, and Frautschy, J. D., 1965, Littoral processes and the development of shorelines: Santa Barbara Specialty Conf. on Coastal Engineering, Am. Soc. Civil Engr., p. 511-536.

King, C. A. M., 1972, Beaches and coasts (second edition): St. Martins Press, New York, 570 p.

Komar, P. D., 1973, Computer models of delta growth due to sediment input from rivers and longshore transport: Geol. Soc. America Bull., v. 84, p. 2217-2226.

________, 1976, Beach processes and sedimentation: Prentice-Hall, Inc., Englewood Cliffs, New Jersey, 429 p.

________, and Inman, D. L., 1970, Longshore sand transport on beaches: Jour. Geophys. Res., v. 75, no. 30, p. 5914-5927.

LeBlond, P. H., 1972, On the formation of spiral beaches: Conf. Coastal Engineering, 13th, Proc., v. 2, p. 1331-1345.

Longinov, V. V., 1965, The possibility of calculating the sand drift along a shallow-water coast. II. Definition of the basic concepts and terms and formulation of the problem of determining the drift: Oceanology, v. 5, p. 69-76.

Lynch-Blosse, M. A., and Kumar, Naresh, 1976, Evolution of downdrift-offset tidal inlets: a model based on the Brigantine Inlet system of New Jersey: Jour. Geology, v. 84, no. 2, p. 165-178.

Nelson, C. H., and Creager, J. S., 1977, Displacement of Yukon-derived sediment from Bering Sea to Chukchi Sea during Holocene times: Geology, v. 5, p. 141-146.

Price, W. A., 1952, Reduction of maintenance by proper orientation of ship channel through tidal inlets: Conf. Coastal Engineering, 2nd, Proc., p. 243-255.

______, 1954, Shorelines and coasts of the Gulf of Mexico, in Gulf of Mexico, its origin, waters, and marine life: U. S. Fish and Wildlife Service Fishery Bull. 89, v. 55, p. 39-65.

______, 1964, Cyclic cuspate sand spits and sediment transport efficiency: Jour. Geology, v. 72, p. 876-879.

Rea, C. C., and Komar, P. D., 1975, Computer simulation models of a hooked beach shoreline configuration: Jour. Sed. Petrology, v. 45, p. 866-872.

Silvester, R., 1960, Stabilization of sedimentary coastlines: Nature, v. 188, p. 467-469.

______, 1970, Growth of crenulate shaped bays to equilibrium: Am. Soc. Civil Engrs. Proc., v. 96, Jour. Waterways Harbors Div., no. WW2, p. 275-287.

______, 1974, Coastal engineering, II: Developments in geotechnical engineering, 4B: Elsevier, Amsterdam, 338 p.

Swift, D. J. P., Kofoed, J. W., Saulsbury, F. P., and Sears, Phillip, 1972, Holocene evolution of the shelf surface, central and southern Atlantic shelf of North America, in Swift, D. J. P., Duane, D. B., and Pilkey, O. H., eds., Shelf sediment transport: Process and pattern: Dowden, Hutchinson, and Ross, Inc., Stroudsburg, Pennsylvania, p. 499-574.

Tanner, W. F., 1962, Reorientation of convex shores: Am. Jour Sci., v. 260, p. 37-43.

______, 1973, Advances in near-shore physical sedimentology: a selective review: Shore and Beach, v. 41, p. 22-27.

Todd, T. W., 1968, Dynamic diversion: influence of longshore current-tidal flow interaction on chenier and barrier island plains: Jour. Sed. Petrology, v. 38, p. 734-746.

Voskoboynikov, V. M., 1966, Shore dynamics of elongated bodies of water: Oceanology, v. 6, p. 541-545.

White, A., 1966, Drainage asymmetry and the Carolina capes: Geol. Soc. America Bull., v. 77, p 223-240.

Williams, W. W., 1956, An east coast survey: some recent changes in the coast of East Anglia: Geogr. Jour., v. 122, p. 317-327.

Wright, L. D., and Coleman, J. M., 1973, Variations in morphology of major river deltas as functions of ocean wave and river discharge regimes: Am. Assoc. Petroleum Geologists Bull., v. 57, p. 370-398.

Yasso, W. E., 1965, Plan geometry of headland-bay beaches: Jour. Geology, v. 73, p. 702-714.

Zenkovitch, V. P., 1959, On the genesis of cuspate spits along lagoon shores: Jour. Geology, v. 67, p. 269-277.

______, 1967, Processes of coastal development: Interscience Publishers, New York, 738 p.

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