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SOME CONSIDERATIONS ON COASTAL PROCESSES RELEVANT TO SEA LEVEL RISE

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16. Abstract

The effects of potential sea level rise on the shoreline and shore environment have been briefly examined by considering the interactions between sea level rise and relevant coastal processes. These interactions have been reviewed beginning with a discussion of the need to reanalyze previous estimates of eustatic sea level rise and compaction effects in water level measurement. This is followed by considerations on sea level effects on coastal and estuarine tidal ranges, storm surge and water level response, and interaction with natural and constructed shoreline features. The desirability to reevaluate the well known Bruun Rule for estimating shoreline recession has been noted. The mechanics of ground and surface water intrusion with reference to sea level rise are then reviewed. This is followed by sedimentary processes in the estuaries including wetland response. Finally comments are included on some probable effects of sea level rise on coastal ecosystems.

These interactions are complex and lead to shoreline evolution (under a sea level rise) which is highly site-specific. Models which determine shoreline change on the basis of inundation of terrestrial topography without considering relevant coastal processes are likely to lead to erroneous shoreline scenarios, particularly where the shoreline is composed of erodible sedimentary material.

With some exceptions, present day knowledge of shoreline response to hydrodynamic forcing is inadequate for long-term quantitative predictions. A series of inter-related basic and applied research issues must be addressed in the coming decades to determine shoreline response to sea level change with an acceptable degree of confidence.

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ABSTRACT

The effects of potential sea level rise on the shoreline and shore environment have been briefly examined by considering the interactions between sea level rise and relevant coastal processes. These interactions have been reviewed beginning with a discussion of the need to reanalyze previous estimates of eustatic sea level rise and compaction effects in water level measurement. This is followed by considerations on sea level effects on coastal and estuarine tidal ranges, storm surge and water level response, and interaction with natural and constructed shoreline features. The desirability to reevaluate the well known Bruun Rule for estimating shoreline recession has been noted. The mechanics of ground and surface water intrusion with reference to sea level rise are then reviewed. This is followed by sedimentary processes in the estuaries including wetland response. Finally comments are included on some probable effects of sea level rise on coastal ecosystems.

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With some exceptions, present day knowledge of shoreline response to hydrodynamic forcing is inadequate for long-term quantitative predictions. A series of inter-related basic and applied research issues must be addressed in the coming decades to determine shoreline response to sea level change with an acceptable degree of confidence.

1. INTRODUCTION

The complexities of shoreline response to sea level rise are contingent upon a very wide range of inter-relationships between physical/ecological factors. The focus of resource analysis for the present purpose must ultimately be on predictive capability, since we are principally dealing with the question of how shorelines and shore environment will change with future sea level rise. Prediction in turn requires an understanding of process fundamentals and adequate data. Therefore, much of what follows pertains to these aspects, which in many cases have more to do with the basics of resource response to hydrodynamic and meteorologic forcing than to sea level rise. If this can be elucidated, then imposing and evaluating the effect of sea level rise becomes a far less difficult task.

Organization of basic knowledge is intertwined with the question of resolution of spatial and temporal scales. The desired resolution for the evaluation of a resource is set by criteria which are dependent upon many non-technical factors. At a built-up shoreline, a 10 m recession could severely damage a structure, while at a natural shoreline the concerns will be less stringent. Then again, in low lying areas such as the Florida Everglades, just a few centimeter rise in sea level would prove to be disastrous to water management, and would cause extensive ecological changes associated with salinity intrusion. A rapidly rising sea level can generate a materially different response than a slow one, an example being the fragile barrier island shoreline. Finally, there is the question of absolute sea level rise and the associated shoreline scenarios. By keeping the issues focussed on the coastal processes themselves, we have in the most part stayed clear of centering on specific temporal and spatial scales explicitly, even though such considerations are inherent in evaluating the degree of uncertainty in the state-of-the-art knowledge and in future research needs.

The interactive nature of coastal processes renders it difficult to isolate resource issues and place them under well-defined "umbrellas" for descriptive purposes. We have selected ten headings (sections 2 through 11) within which a range of topics has been referenced. The first of these - Estimates of Eustatic Sea Level Rise - does not deal with process description in a general way, but highlights a fundamental issue, namely the quality of the data base that has been used to calculate past secular trends in sea level change, and what needs to be done to improve this base. Following this is the section Compaction Effects, which is directly associated with problems in water level measurement.

Sections 4 though 11 deal with coastal processes. In section 4 the effect of sea level rise on tidal ranges is discussed, and section 5 deals with non-astronomical factors including storm surge and waves. The next two sections are concerned with shoreline response. While section 6 deals with physical processes in shoreline response in broad categories, section 7 focusses on specific issues relative to the scope and limitations of the well known Bruun Rule for estimating shoreline recession rate. Physical considerations upon which this rule must be re-examined have been noted.

Section 8 describes problems with saltwater intrusion in groundwater as a result of sea level rise or analogous effects, while the same problem in surface waters is highlighted in section 9. Sedimentation problems in tidal entrances, estuarine mixing zone and wetlands is described in section 10. Finally, ecological changes, including research needed to quantify these better, have been noted in section 11.

Some overlap between the various sections is inevitable. This extends to both the physical description and research needs. Also, by and large, the coastal processes have been reviewed from an engineering perspective, and evaluation of present day knowledge has been made from the viewpoint of the availability of quantitative (as opposed to qualitative) criteria.

In general it appears that with the possible exception of tidal hydrodynamics and salinity intrusion, considerable further research is required for assessing shoreline and shore environmental response in a confident manner. Strides made during the past decade have been impressive, but for example where sediment transport is a key factor, we are significantly limited in long-term predictive capability. This is partly due to the lack of good quality synoptic hydrodynamic/meteorologic data. This problem in turn has an impact on ecological modeling, which is contingent upon a knowledge of flows and sediment movement.

Section 12 is essentially a summary of future research needs. There is a table for each of the ten broad research issues described in sections 2 through 11. A special issue ranking procedure has been used for the ultimate purpose of a numerical ranking of research areas in terms of their importance to the sea level rise problem.

Bibliography is contained in section 13. Division is by sections. In some cases, additional references not cited in the text, but considered to be of potential interest to the reader, have been included.

2. ESTIMATES OF EUSTATIC SEA LEVEL RISE

2.1 INTRODUCTION

Eustatic sea level rise is the global average sea level rise primarily due to: 1) additional water mass in the oceans through release of water contained in polar ice caps and alpine glaciers, and 2) steric expansion of water presently in the oceans due to increased temperature, thereby increasing the volume of an existing water mass. Sea level change data from 20,000 years before present (BP) to 1,000 years BP have been obtained from radiometric dating of plants and animals that lived only in intertidal or shallow marine waters. Data from the last 100 or so years are based on measurements from long-term tide gages. Both of these sources include not only the "signal" of eustatic sea level change, but the "noise" or contamination by local vertical movement of the land where the measurements are made. Additionally, local and temporal oceanographic and meteorological factors may contribute to anomalously high or low water levels for periods of many years. The degree of contamination in any one tide gage record may be severe with the annual contamination exceeding up to 40 years of eustatic trend. Much of the contamination is spatially and temporally coherent over fairly long distance and time scales and the physics of this contamination is poorly understood. If the available tide gage data provided a representative distribution over the world's oceans, the noise could be eliminated by simply averaging over these gages. However, the available tide gage data are heavily concentrated in the northern hemisphere and along continental margins.

Tide gages measure the local relative sea level which is important and is the water level relevant to that area. However, an understanding of recent eustatic sea level rise is critical, because models developed for predicting future sea level rise are calibrated based on estimates of recent rise. Most of these estimates suggest a rate of 10-15 cm/century (1 to 1.5 mm/yr) with some investigators inferring an increase in the rate of rise over the past 40 or so years. Most of the studies leading to the above estimates have been based on gages located in reasonably stable low- to mid-latitude areas. Clearly the most significant neotectonic contribution to relative sea level rise is the earth's rebound from the ice loading in the polar regions during the last (Wisconsin) ice age. This rebound is causing uplift in the high latitudes on the order of 1 meter per century and land subsidence at the lower latitudes on the order of 5 cm per century. There have been suggestions that most of the studies of eustatic rates, in excluding the high latitudes of relatively rapid uplift, have yielded overestimates. A very preliminary analysis presented here based on United States data tends to support this contention.

Areas in which future studies appear warranted include: 1) understanding the physics of the noise in tide gage records with the objective of extracting this portion of the record, 2) revisiting the question of extracting recent eustatic sea level rise rates from the tide gage records with an emphasis on proper recognition of the contribution from glacial rebound at all latitudes, and 3) if the changes resulting from 2 are significant, recalibrating the models employed for predicting future sea level rise based on scenarios of future changes in CO_2, other trace gases and a gradual warming trend.

2.2 LITERATURE REVIEW

There has been a wide range of techniques and degree of sophistication applied in an attempt to extract eustatic sea level (ESLR) rise from tide gage records. One of the first comprehensive published studies on ESLR based on tide gages was by Gutenberg (1941). A total of 69 gages was analyzed encompassing the period 1807 to 1937. Gutenberg excluded tide gages known to be in areas of crustal uplift, yet gages were included in areas known to be sinking, some at fairly high rates. Gutenberg concluded that ESLR was approximately 1 mm per year.

Many investigations following those of Gutenberg have tended to adopt his data selection procedures with similar results, i.e. rates of 1 to 1.5 mm/yr, see Table 2.1. Emery (1980) concluded that ESLR has been accelerating with a rate up to 3 mm/yr over the past 40 years. Subsequent studies by Aubrey and Emery (1983) and Barnett (1983) conducted specifically to examine the change in rate concluded there was no convincing evidence for such a conclusion.

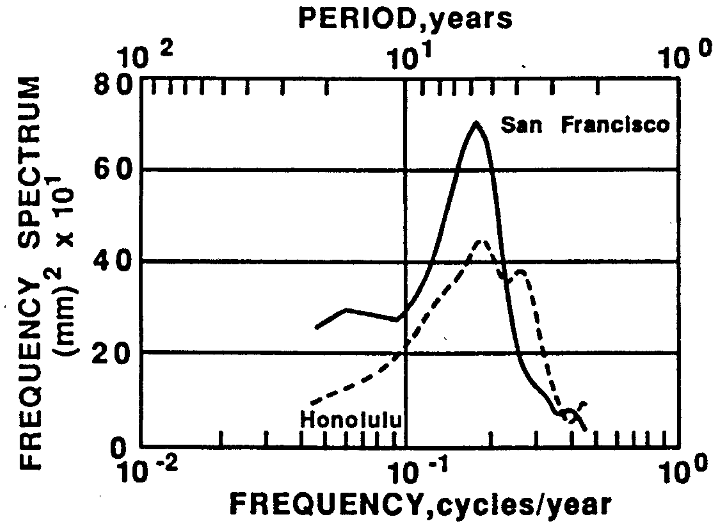
The difficulties of extracting the sea level rise (SLR) "signal" from a record containing substantial noise has been studied carefully by Sturges (1987). The coherency of spatially separated tide gage records was investigated with the hypothesis that coherent signals with no lag could be interpreted as global sea level rise whereas lags with a certain character

Table 2.1. Estimates of Eustatic Sea Level Rise Based on Tide Gage Data (adapted from Barnett, 1983; and Hicks, 1978)

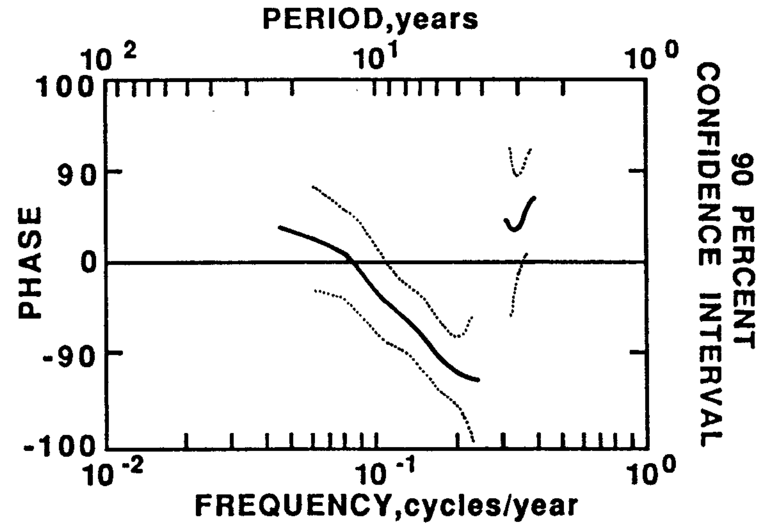
Author(s)	Estimate (cm/100 yr)
Thorarinsson (1940)	> 5
Gutenberg (1941)	11 ± 8
Kuenen (1950)	12 to 14
Lisitzin (1958)	11.2 ± 3.6
Fairbridge and Krebs (1962)	12
Hicks (1978)	15 (U.S. only)
Emery (1980)	30
Gornitz et al. (1982)	12 (10 cm excluding long-term trend)
Barnett (1983)	15

could be interpreted as due to atmospheric forcing or long water wave (Rossby wave) motions. As an example, the records at San Francisco and Honolulu were found to be coherent at periods of 5 to 10 years and longer, although with a phase lag. A comparison of the energy spectra obtained from these two stations is presented as Fig. 2.1a and other spectral information is presented in Figs. 2.1b,c,d. The amplitudes of these coherent components are 5-15 cm. Similar coherence results were found for tide gage records located on both sides of the Atlantic. Sturges concluded that the available records are contaminated by substantial energy with periods up to 40 to 50 years, thus exacerbating the problem of identifying any change in the rate of SLR. The ability to extract the SLR signal may possibly be enhanced through an analysis which recognizes the probable cause of the noise components, thereby guiding their removal from the record.

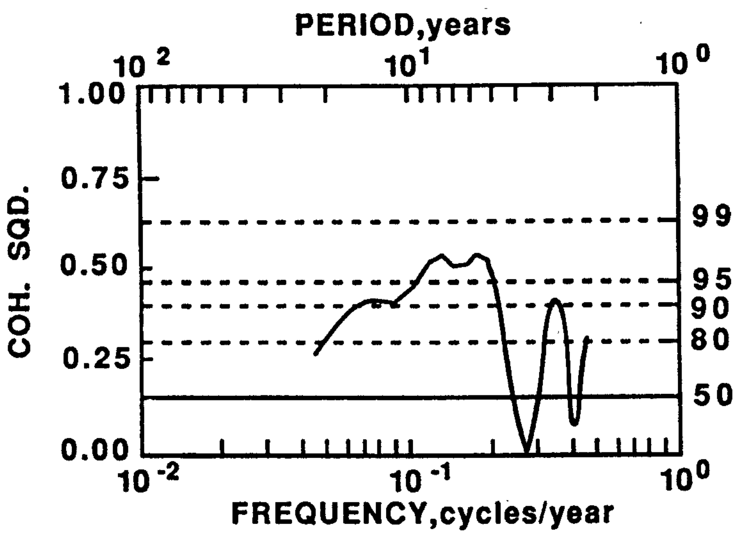
Aubrey and Emery (1983) applied the method of eigenanalysis to United States tide gage data in an attempt to identify fluctuations that were spatially and temporally coherent. This method, among the most sophisticated applied to date, has the potential advantage of retaining in the first few temporal eigenfunctions, those fluctuations that have the same form and that are either exactly in or exactly out of phase. The principal disadvantage is that the method is purely statistical and does not recognize the physics of the phenomenon, although it may isolate features that will assist in identifying physical components. A particular drawback is that the method



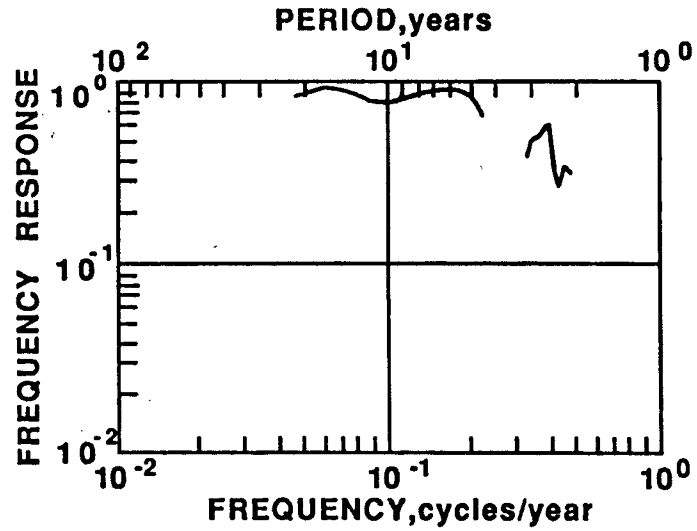
a) Energy Spectra



b) Phase Spectra with 90% Confidence Intervals



c) Coherence Squared



d) Frequency Response Function, Honolulu to San Francisco

Fig. 2.1. Cross-Spectral Characteristics between Sea Level at San Francisco and Honolulu: Yearly Data, 1905 through 1971 at San Francisco and Beginning 1907 at Honolulu (after Sturges, 1987).

only recognizes correlations which are either in phase or exactly out of phase as "signal". Thus a very long and slowly propagating wave would be rejected as noise whereas a pure standing wave would be recognized as "signal". Aubrey and Emery first applied the technique to 12 U.S. gages each of which encompassed 61 years of data and secondly to 41 tide gages with a common time base of 40 years of data. Different rates of rise were found for the East and West coasts. From the longer term data set of 12 stations, the eustatic values on the West and East coasts were found to be rising by averages of 1.4 mm/year and 1.3 mm/year, respectively. For the shorter term (40 years) of 41 stations, the rates of change for West and East coasts were -0.3 mm/yr and +2.5 mm/yr, respectively. It was found that the long-term rates of sea level rise are increasing from Cedar Key on the Florida west coast to Cape Hatteras, decreasing from Cape Hatteras to Cape Cod and increasing from Cape Cod to Eastport, Maine. These results are presented in Fig. 2.2. Finally, it was concluded that there is no evidence from this analysis that rates of SLR are increasing over the past 10 years.

Pirazzoli (1986) has analyzed the results from 1,178 tide gage stations provided primarily by the Permanent Service for Mean Sea Level. This appears to be the largest data set considered in an individual analysis. The analysis method was straightforward, first taking averages for each station over five

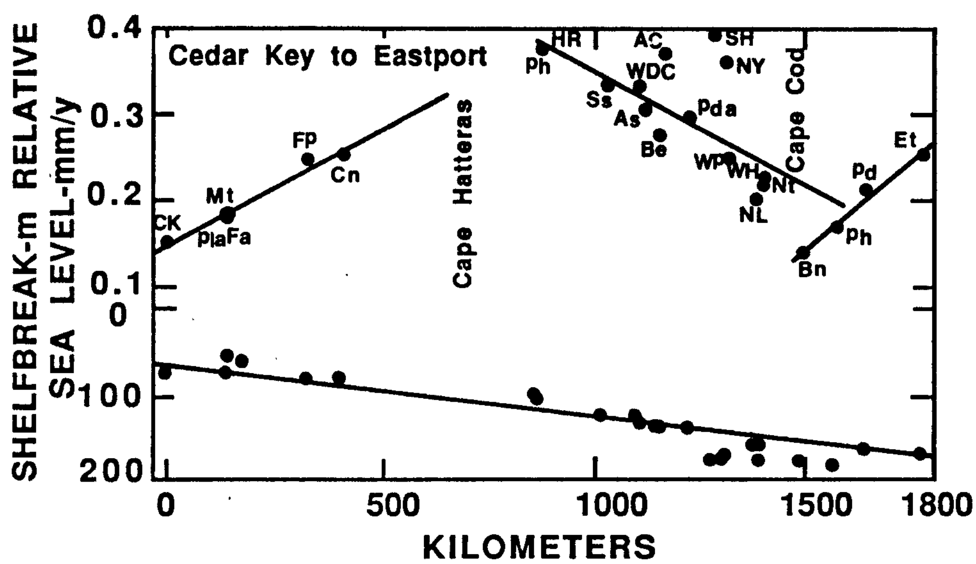


Fig. 2.2. Mean Annual Relative Sea Level Changes During 40 Year Record. Lines Define Three Main Segments of East Coast with Differing Sea Level Trends (after Aubrey and Emery, 1983).

year periods, then averaging over the two ends of the resulting data to obtain a change in sea level from which the rate is determined. The results are presented regionally and on a global basis. The effects of glacio-eustatic adjustment to the last ice age are very apparent in the data with relative sea level (RSL) rising and lowering in most low and high latitudes, respectively. The possible effects of earthquakes in causing sudden displacements and altering the trend after the earthquake are illustrated. As an example, the tide gage at Messina, Italy recorded an abrupt increase in RSL of 57 cm during the earthquake of 1908. Anthropogenic effects, primarily the extraction of water and hydrocarbons, causing compaction are noted with Venice, Italy particularly evident as a consequence of ground water pumping. In attempting to infer global rates from the available data, it is noted that if the earth is divided into 30° latitude and longitude sectors, a total of 72 compartments result of which 71 have marine coasts. The data distribution in these compartments is very non-uniform. Most of the tide gages (70%) are situated in only 4 compartments whereas there are no data in 70% of the compartments. Long-term tide gage data in the southern hemisphere are particularly sparse with over 97% of the stations examined by Pirazzoli in the northern hemisphere. Without the assumption that the results from the northern hemisphere are globally representative, the available data are clearly inadequate. Fig. 2.3 presents a distribution of the tide gage locations according to the longitude-latitude compartments noted earlier. Fig. 2.4, also from Pirazzoli, presents the distribution of tide gages and median trend of RSL by 5° increments of latitude. The earlier noted effect of relative rises in the mid-latitudes and lowering RSL in the higher latitudes is evident.

Pirazzoli concludes that the results presented by most investigators ($\ll 1$ mm/yr) probably are an overestimation of the ESLR. Local and regional factors including tectonic movements and oceanic factors are generally larger than eustatic factors. The bias due to downwarping as a result of loading of the continental shelves by sediment transport and deposition is noted. Finally, when centimeter accuracy is attainable from satellite altimetry, the potential to contour the open ocean is regarded as a major advance in our general knowledge of eustatic sea level rise rates which have both good geographic coverage and are free from much of the contamination which attends measurements of tide gages located along the coastline.

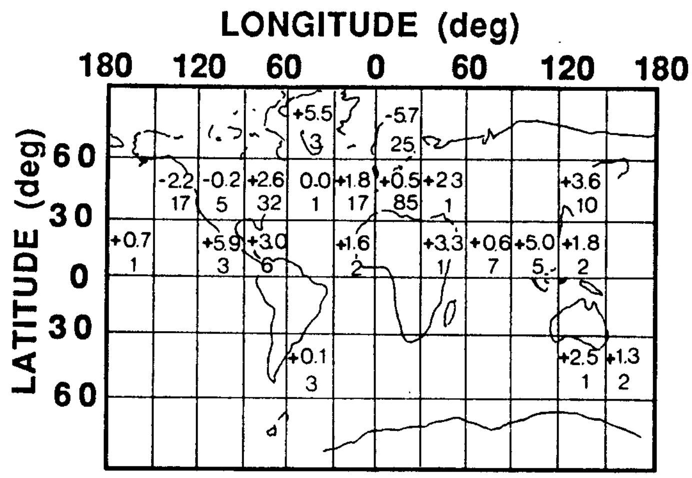


Fig. 2.3. Characteristics of Tide Gage Data by 30° Longitude and Latitude Sectors. The Lower Values Represent the Number of Tide Gages in Each Sector. The Upper (Signed) Numbers Represent the Linear Long-Term Relative Sea Level Change Resulting from those Gages (after Pirazzoli, 1986).

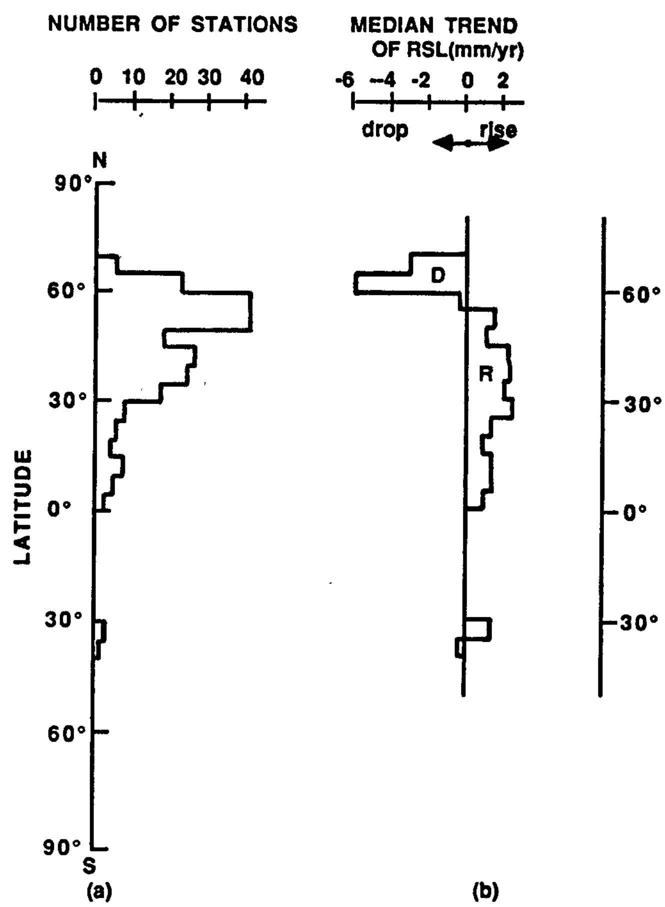


Fig. 2.4. Distribution by 5° Latitude Belts of a) Tide Gage Stations, and b) Median Values of Linear Long-Term Trends of Relative Sea Level. Note the Tendency for a Relative Drop in Sea Level for the Higher Latitudes (after Pirazzoli, 1986).

Lambeck and Nakiboglu (1984) have carried out an analysis of the effect of post-glacial adjustment on estimates of ESLR. For this purpose, a viscous model of the earth was adopted with the assumption of a uniform mantle viscosity. To quantify the effect of rebound on estimates of ELSR as determined from tide gage records, the apparent or RSL rises predicted by the model without any additional water mass or steric changes were computed for the same eight long-term tide gage stations selected by Barnett (1983). Two values of viscosity, μ, were used: Model 1, $\mu = 5 \times 10^{21}$p and Model 2, $\mu = 10^{22}$p. for the eight stations, Models 1 and 2 predicted apparent (relative) sea level rises of 0.5 and 0.8 mm/yr, respectively whereas Barnett found 1.5 mm/yr. Based on this comparison, Lambeck and Nakiboglu conclude that the post-glacial rebound contribution may be as high as 30% to 50% of published estimates of ESLR.

A limited analysis has been carried out here to attempt to determine the effects of employing only the lower latitude tide gate data. The U.S. data for the East and West coasts and Gulf of Mexico as published by Hicks *et al.* (1983) were used. The trend estimates in Hicks *et al.* were simply plotted against latitude as presented in Fig. 2.5. A problem is that the data only encompass latitudes from approximately 25° to 58° and thus it is necessary to extrapolate liberally. At the lower latitudes, the data were extrapolated uniformly at approximately 3.2 mm/yr and at the higher latitudes, due to the uncertainties, two extrapolations were adopted to determine sensitivity as presented in Fig. 2.5. Based on the latitudinal variation, $\dot{\eta}(\phi)$, estimates of the ESLR, $\dot{\eta}_E$, were based on the following

$$\dot{\eta}_{E_j} \approx \int_0^{\pi/2} \dot{\eta}_j(\phi) \cos\phi \, d\phi \qquad (2.1)$$

where j = I,II represents the different high latitude extrapolations. The resulting values were

$$\dot{\eta}_{E_I} = 0.32 \text{ mm/yr, Extrapolation I}$$
$$\dot{\eta}_{E_{II}} = 0.67 \text{ mm/yr, Extrapolation II}$$

These results are qualitatively in agreement with those of Lambeck and Nakiboglu.

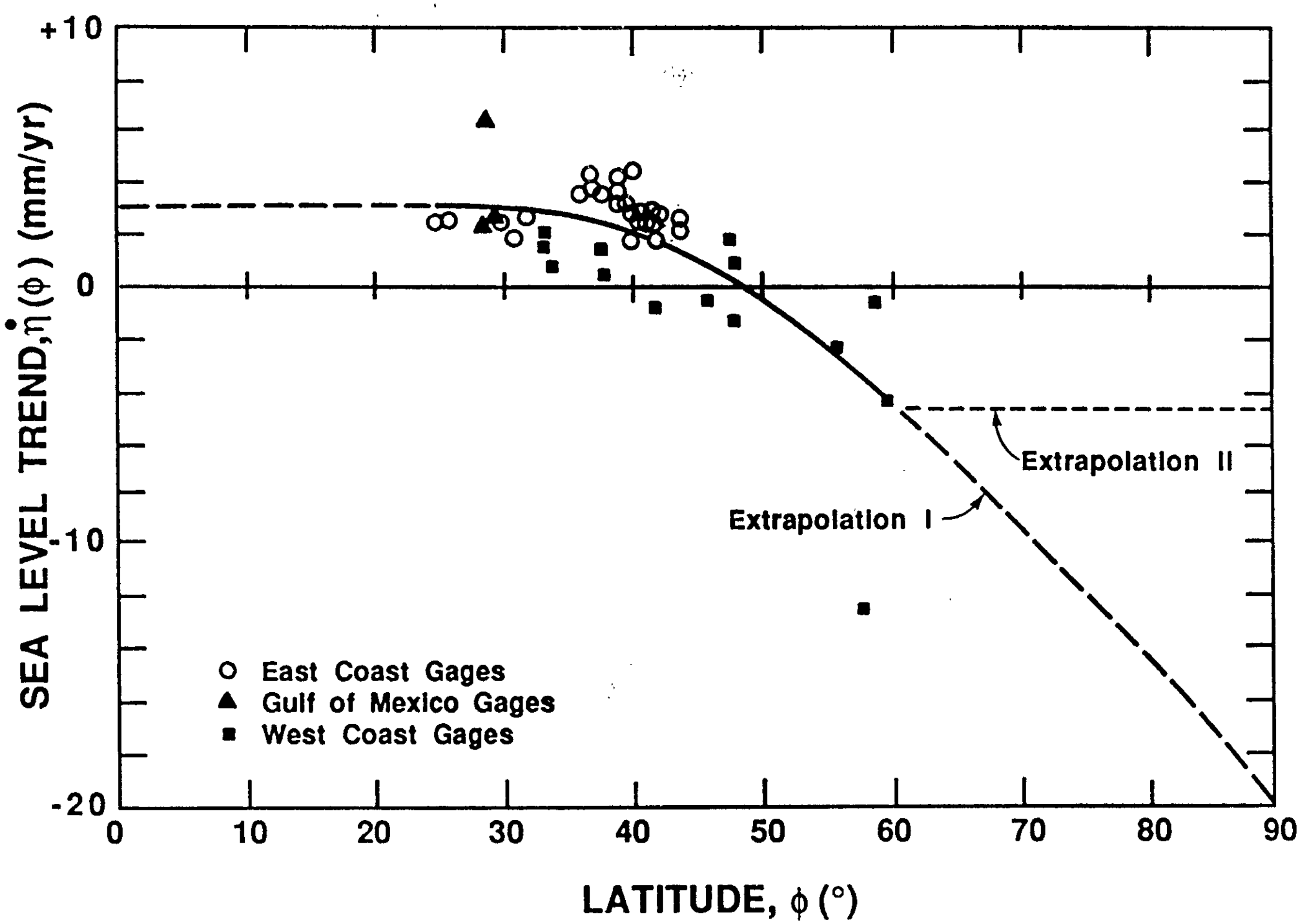


Fig. 2.5. Long-Term Tide Gage Trend Results, $\dot{\eta}$, versus Latitude, ϕ. Continental United States and Alaska. Based on Hicks et al. (1983).

2.3 THE NATURE AND ANALYSIS OF SEA LEVEL DATA

From the standpoint of extracting eustatic sea level change, it is useful to represent the total RSL, $\eta_i(t)$, as measured by the i^{th} tide gage as

$$\eta_i(t) = \eta_E(t) + \eta_{N_i}(t) \tag{2.2}$$

in which $\eta_E(t)$ is the eustatic sea level at time t and $\eta_{N_i}(t)$ is the total "noise" at the i^{th} tide gage. The noise can contain many components including vertical ground motion, effects of freshwater in the vicinity of the gage, coastal currents, long waves, barometric pressure anomalies, wave effects, etc. Several obvious results follow from Eq. 2.2. First, if there were a uniform coverage of tide gages on the oceans, an average of the elevations from all such tide gages would yield the eustatic sea level. Additionally, the eustatic sea level change rate need not be constant, but could vary substantially year-to-year with temperature, etc. Considering two or more tide gages, the noise may be correlated in space and time positively, negatively, with an arbitrary phase or uncorrelated. The more widely separated the gages, the greater the likelihood that the noise will be uncorrelated. Thus, there are advantages to averaging many records along a coast, possibly with an appropriate coastal length weighting factor. Finally, the best estimate of eustatic sea level (and thus eustatic sea level rise) and one which yields the most understanding as to the stability of the results is a progressive averaging in which larger and larger data bases are averaged, i.e.

$$\bar{\eta}_{IK}(t) = \frac{\sum_{i=1}^{IK} \eta_i(t) w_i}{\sum_{i=1}^{IK} w_i} \tag{2.3}$$

where w_i is a distance weighting factor and IK is the total number of gages along a selected coastal segment, perhaps a continent. The worldwide estimate of eustatic sea level, $\tilde{\eta}_E(t)$ could then be obtained by averaging over all available coastal segments

$$\tilde{\eta}_E(t) = \frac{1}{IKTOTAL} \sum_{IK=1}^{IKTOTAL} \bar{\eta}_{IK}(t) \tag{2.4}$$

Other ways of extracting meaningful information relating to post-glacial rebound could include averaging first over longitude for certain increments of latitude.

2.4 RESEARCH NEEDS

In general, improvements in our understanding of eustatic sea level change can come about through use of the existing data base or development of new data. Extraction of more meaningful results from the existing data base will require either more powerful analysis procedures or an improved understanding and application of the physics of relative sea level change, including the noise present in the records. Enhancement of the existing data base through new measurements will most likely occur through satellite altimetry once this is proven to centimeter accuracy over the open ocean. Additionally, in some cases much can be learned locally about anthropogenically generated compaction in areas of tide gages through the installation of rather simple compaction measurement devices. One feature of new data is the length of time that will be required for such data to "mature" to yield significant meaningful information.

2.4.1 Use of Existing Data

Analysis in light of the physics of RSL change appears to be the most effective and productive use of existing data. In particular, accounting for the contribution of long period waves as explored by Sturges (1987) would allow interpretation and removal of a major portion of the noise in the RSL measurements.

A second productive area is a more thorough analysis than presented previously of the contribution of post-glacial adjustment of the earth following the last ice age. As noted previously, Lambeck and Nakiboglu (1984) have inferred from viscous models of the earth that the actual eustatic rise is roughly one-half to two-thirds the value determined from analysis of records based only on areas of relative stability. Improved estimates of eustatic sea level rise could be based on either a more inclusive data set with or without the use of a viscous earth model. Obviously more meaningful results could be obtained with the combined approaches simultaneously. The approach envisioned here is in general the same as applied in "physical principles" with the addition that the global viscous model would be employed for interpretation, guidance and confirmation of the results obtained.

Most approaches of direct analysis attempt to reduce the noise in a record on a station-by-station basis through determining some sort of RSL estimate through fitting to the data. Unfortunately, the noise in individual records is such that at least 20 to 40 years of data must be available at the individual gages before these results can be considered meaningful. An approach that would make these results meaningful early after their availability is the weighted averaging of many stations along a coastline to establish a more stable value. This averaging length could encompass, for example, the North American or North and South American shoreline(s). Thus, if a wave with length exceeding the expanse of the stations encompassed were contributing to the "noise", this process would tend to reduce or (in the very fortuitous cases) eliminate its contribution. By first averaging over long segments of the shoreline, weighting each station by its alongshore influence length, then combining appropriately the results for various such shoreline segments, a much more stable year by year value could be obtained, i.e. Eqs. 2.3 and 2.4. This would allow effective use of such data as are available for the east coast of South America where eight of the twelve available gages are less than 30 years in duration. As is evident from Fig. 2.6 which presents the mean annual sea level variation of Pensacola, Florida, 30 years is not adequate to obtain a stable estimate from an individual gage.

2.4.2 Need for New Data

There are two types of new data that would contribute to improved estimates of ESLR: those that contribute immediately and those that would require a data base of at least several years before meaningful results could be obtained. It is anticipated that even with the potential benefits of satellite altimetry, at least one decade and possibly two decades will be required before adequate confidence will be placed in these data to yield accepted reliable estimates of eustatic sea level rise. Three research needs in the category of "new data" are described below.

Compaction Gages - As is well-documented by a number of studies, withdrawal of ground water and hydrocarbons can contribute to substantial subsidence and thereby a "relative sea level rise" (see also section 2 for a

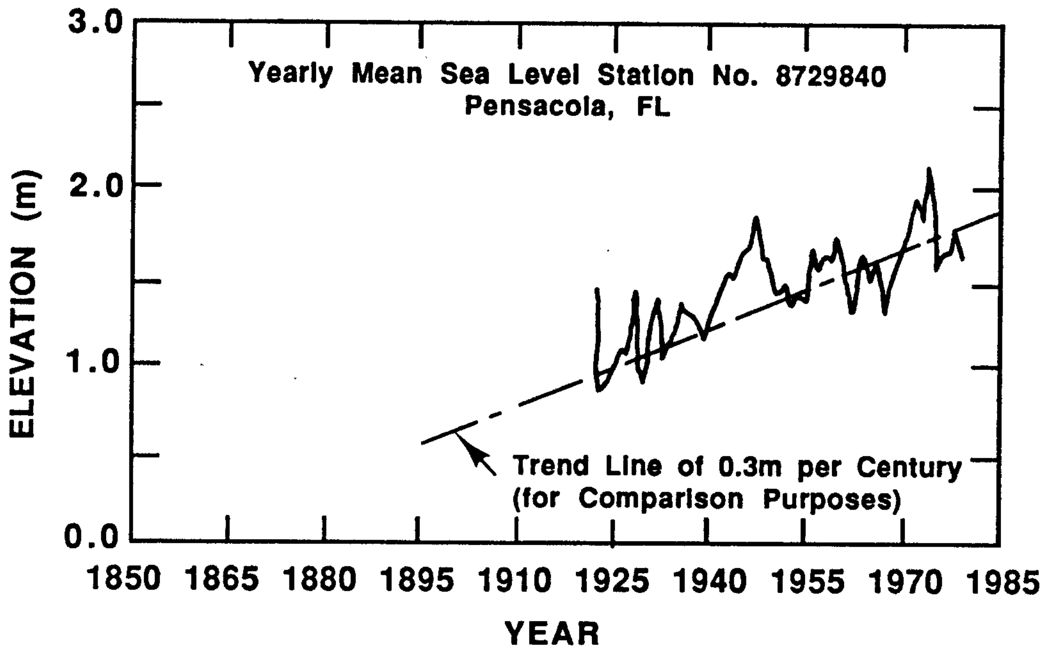


Fig. 2.6. Average Annual Sea Level Variations for Pensacola, Florida (adapted from Hicks et al., 1983).

discussion of compaction effects). It is worth noting that this is probably the only component that realistically can be controlled by humans. The obvious general but not universal correlation of areas of tide gage locations and ground fluid extraction near population concentrations justifies a possible concern over this activity. Also the fact that these are the areas that continued RSL rise may contribute most to the ultimate response cost (relocation, defense, repair, etc.) makes it important that the significance of anthropogenically induced subsidence be quantified and possibly controlled as early as possible.

Very simple and sensitive compaction meters have been utilized in quantifying this effect in the vicinity of Osaka and Niigata, Japan among other locations. A schematic of two such gages is presented in Fig. 2.7. Each installation consists of an outer casing lining a hole drilled to some depth, h. The inner pipe of slightly smaller diameter is founded on the stratum at depth h. Thus the relative vertical movement between the top of the inner pipe and the general ground level represents the total compaction over the upper sediment column of thickness, h. To establish differential compaction, several such devices would be required at each location of interest. Ideally installations would be made near tide gages and also remote from cities but say inland and in the same geological formations as those near the tide gages. These gages would commence yielding valuable data immediately, and it may be possible to supplement the compaction data collected with models using data representing the geological formations and the history of past ground fluids extraction to estimate earlier compaction. Such results would be invaluable in providing more reliable estimates of past and future eustatic sea level rise.

New Tide Gage Data - Referring to Figs. 2.3 and 2.4a, it is clear that the southern hemisphere is especially deficient in long-term tide gage data. A number of relative short-term tide gage records are available along the east and west coasts of South America; however, there needs to be an effort on an international basis to install and maintain additional gages to provide a representative distribution. In addition to the southern hemisphere, more insular tide gages and tide gages along the open coast are needed. A first phase effort could be a survey to identify such sites.

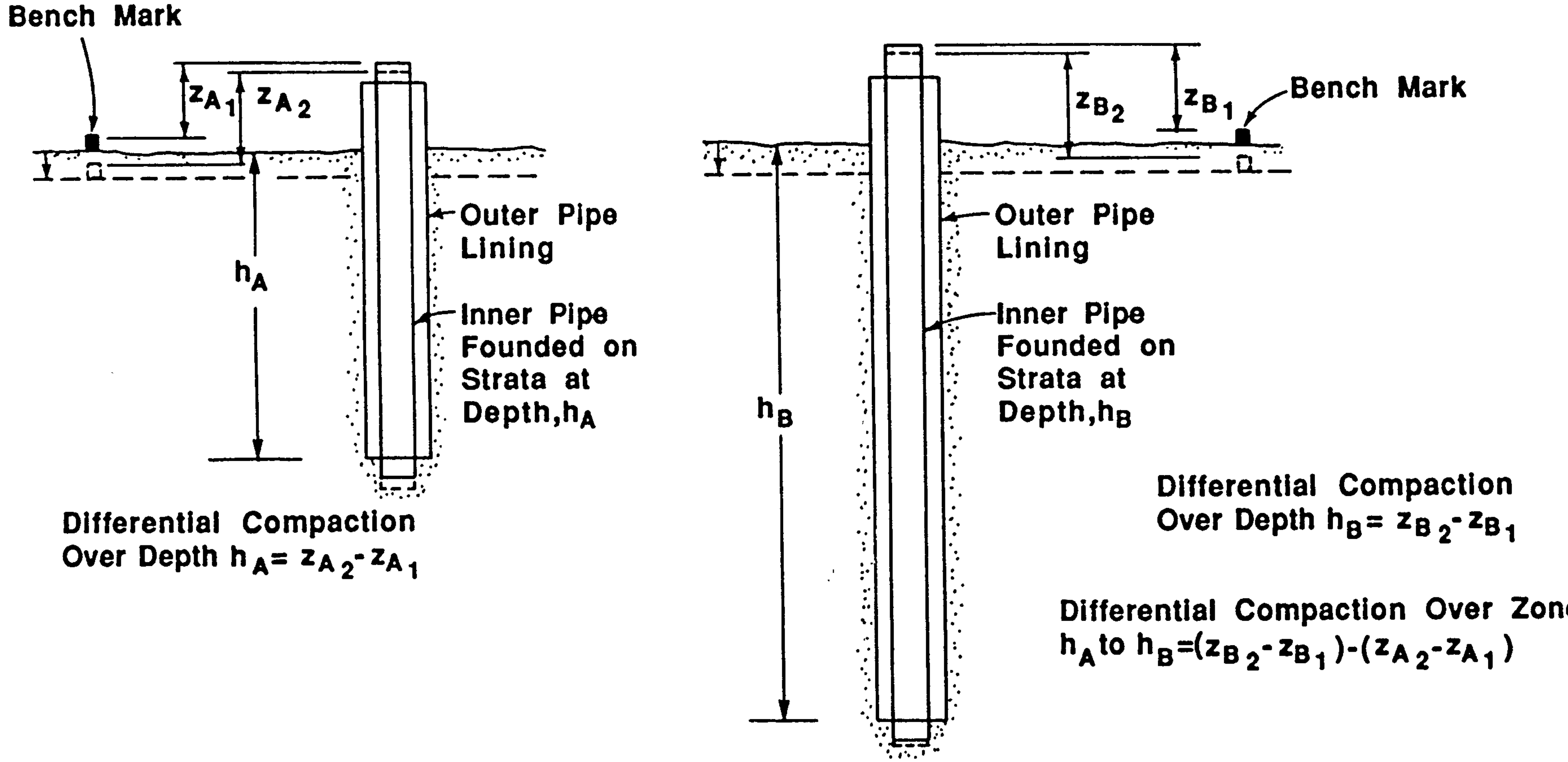


Fig. 2.7. Use of Two Compacting Gages to Obtain Compaction Distribution over Depth Zones h_A, h_B, and $h_B - h_A$.

Satellite Altimetry - This new technology should soon yield absolute vertical accuracies of centimeter accuracy. Thus, sounding much of the ocean surface would allow much broader coverage and very importantly does not require reliance on coastal measurements. It would appear appropriate to continue a dense network of tide gages for sea level rise purposes for several decades after such accuracy is claimed to assure that future needs will be met, and also to allow comparison of the broader satellite coverage and the long-term tide gage results.

3. COMPACTION EFFECTS

3.1 INTRODUCTION

Compaction results in the subsidence of ground level due to reduction in the void ratio of the underlying soil, and in coastal areas contributes to a local relative rise in sea level. Reduction in void ratio is often the natural response of a soil to an increase in loading, because an increase in the interstitial stresses between solids is required. An increase in the loading of a soil stratum can be the result of an increase in loading on the ground surface (e.g. building construction or additional sediment deposition), or due to removal of ground fluid (e.g. water, oil, or natural gas). Compaction occurs in nature as mud is deposited on the beds of rivers and estuaries, and especially in river deltas. Another example is the increase in loading as a barrier island migrates over a stratum of peat, causing the peat to compact and ground level to subside. Because compaction is a time-dependent process, the relative rate between deposition and compaction will determine whether bed elevation increases or decreases. Compaction of a region can also be induced by man, due to 1) loading by the weight of structures, 2) the extraction of oil and natural gas, and 3) depletion of the groundwater table due to active pumping or by preventing recharge of aquifers.

The literature in soil mechanics and foundation design is too replete with articles on the general topic of compaction to review in detail. The proceedings of a symposium "Land Subsidence" held in Tokyo in 1969 (in reference list in section 13) provides a thorough treatment of the causes of compaction, its theoretical description, field measurement techniques and analysis, physical consequences and remedial measures. Much of the subsequent material is gleaned from this collection of studies. However, no investigations have been found which identify any specific effects of the inverse problem, i.e. the effect of sea level rise on compaction and subsidence.

Shiffman _et al._ (1985) review the available theories regarding consolidation (compaction). The simplest is Terzaghi's "Conventional Theory" governed by

$$c_v \frac{\partial^2 u}{\partial z^2} = \frac{\partial u}{\partial t} + \frac{\partial u_o}{\partial t} - \frac{\partial \sigma}{\partial t} \tag{3.1a}$$

$$c_v = \frac{k(1 + e_o)}{\rho_w a_v} \tag{3.1b}$$

where u is the excess pore water pressure, u_o is the hydrostatic pressure, σ is the total stress applied to the system, k is the hydraulic conductivity, e_o is the initial void ratio, ρ_w is the mass density of the fluid (water), and a_v is the compressibility of the soil skeleton. Solving Eq. 3.1 for u and applying the continuity equation for conventional theory

$$\frac{\partial}{\partial z} \left(\frac{k}{\rho_w} \frac{\partial u}{\partial z}\right) = \frac{\partial n}{\partial t} \tag{3.2}$$

soil porosity n is determined. Knowing the porosity as a function of time and the initial thickness of the soil layer, the time history of ground level subsidence can be calculated. Except for very idealized cases, this problem must be solved numerically. Shiffman et al. (1985) also describe a nonlinear finite strain theory, which removes several assumptions of conventional theory but requires difficult numerical solution. Fig. 3.1 displays comparison of the two theories to centrifuge experiments, with the finite strain theory providing good results.

3.2 MEASURING COMPACTION

As noted in section 2, a simple yet effective device for measuring compaction rates has been developed in Japan and has been widely used there for at least the past 30 years, see Murayama (1970). This device, shown in Fig. 3.2 (see also Fig. 2.7), consists of two concentric pipes that penetrate to a desired non-compactable stratum. The outer pipe is perforated to allow the groundwater table to move freely up and down in the casing. A float-type gage monitors the water level. A strip chart and pen displacement gage, mounted on a foundation that "rides" the ground surface, records the subsidence as the pipes appear to protrude from the ground. Several of these gages located in the same area, but penetrating to different strata, provide information about the vertical distribution of compaction. A single gage which penetrates to bed-rock will record the total subsidence.

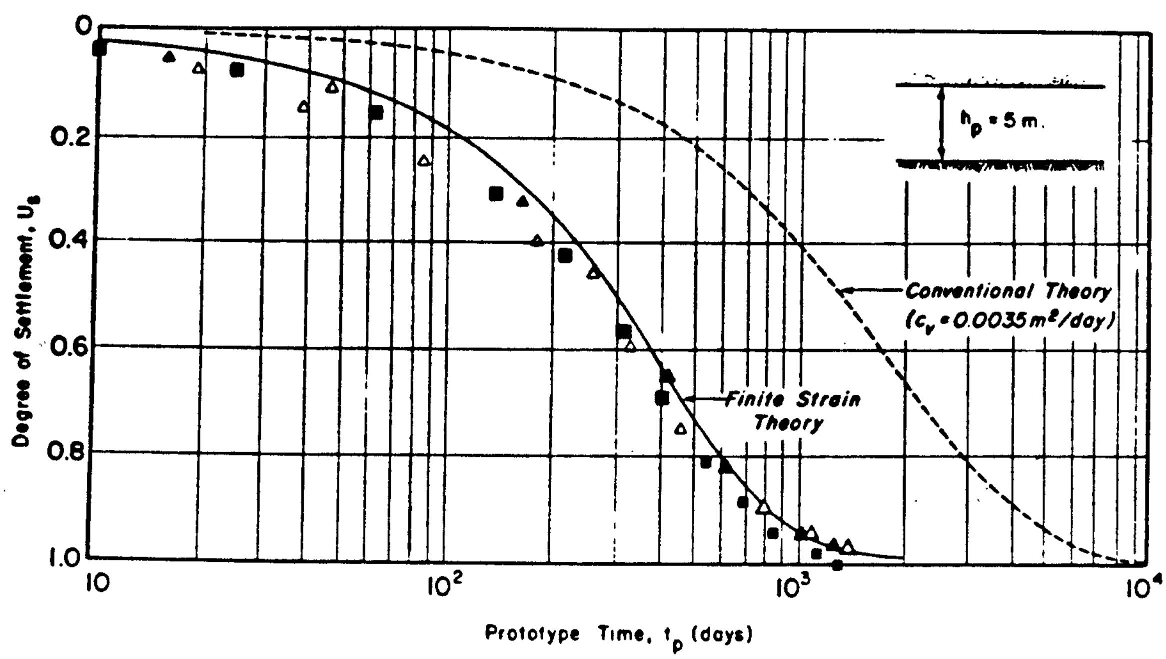


Fig. 3.1. Results of Centrifuge-aided Compaction in Comparison to Two Theories (after Schiffman et al., 1985).

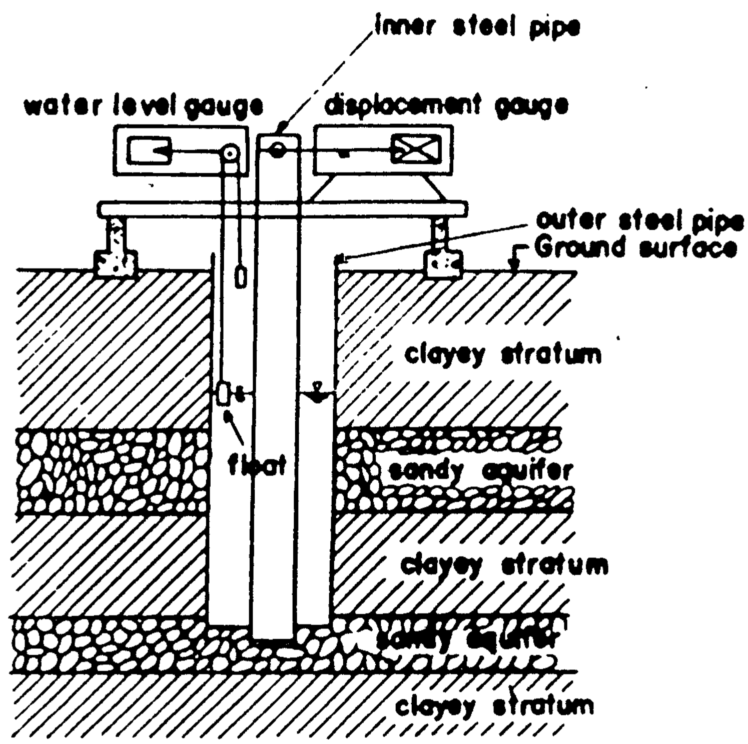


Fig. 3.2. Device for Monitoring Compaction and Groundwater Elevation (after Murayama, 1970).

3.3 IMPLICATIONS OF COMPACTION

Compaction enters the discussion of sea level rise in two distinct places. First is the obvious effect that relative sea level will rise as ground or bed level subsides, resulting in deeper water in rivers and estuaries, and increasing the likelihood of erosion and flooding in coastal communities. This will occur even without global sea level changes and seismic activity. Second is the possible contamination of estimates of eustatic rise due to compacting of regions where tide gages are located. Although most such estimates as detailed in section 2 have avoided using records from areas subject to "obvious" compaction, compaction rates comparable to estimates of eustatic sea level rise (~ 1 mm/yr) are not obvious without detailed measurements using devices such as that described. Because tide gages are usually located near coastal cities where both loading by structures and groundwater extraction/depletion are to be expected, the potential for compaction contamination of the measurements exists.

3.4 REMEDIAL MEASURES

Of all types of subsidence, only that which is man-induced can be prevented, arrested, and perhaps partially reversed. Extraction of oil and gas can be accompanied by recharge of the soil stratum with water, as was the case at Terminal Island, California to be discussed. Protection of the surface recharge areas of aquifers, and water use management to avoid extreme draw-down of the water table also can prevent or reduce compaction.

3.5 EXAMPLES

Mississippi River Delta - A striking example of subsidence due to natural compaction is the delta of the Mississippi River. According to May *et al.* (1983), the Louisiana coast is retreating at an average rate of 4.2 m/yr, most of which is attributed to erosion and inundation in response to relative sea level rise induced by natural compaction. The levees built along the river have cut off the source of sediment to the mud flats, and their natural rate of compaction is causing some areas to sink at rates of 1 cm/yr or more (see also Table 10.1). Only in a small area of delta formation is the rate of deposition greater than the rate of compaction. This high rate of rise in relative sea level is drowning salt marshes and causing existing small sandy barriers to migrate over the backbarrier muds, further exacerbating the compaction. Penland et al. (1985) predict that at present rates of sea level rise, the Chandelieur Islands and Isles Dernieres will be lost during the next 100 years. Because the loading in this region is naturally-induced and the affected area so large, the only functional remedial measure would be to remove the levees in the delta region in hopes of restoring the sediment supply and deposition rate. Although proven successful on a local scale, this is not a cost-effective nor practical solution on a regional basis.

Terminal Island, California - This classic example of the increase in relative sea level due to man-induced subsidence demonstrates many of the possible consequences of natural sea level rise. Due to withdrawal of oil and gas from the Wilmington Oil Field, an area 5 km wide and 6.5 km long subsided an average of about 1.5 m, and encompassed Terminal Island and a portion of Long Beach, California. In some areas the overall subsidence reached 7 m and resulted in considerable damage to harbor facilities as relative sea level rose. This damage required substantial remedial efforts including diking in areas of extreme subsidence, reconstruction of damaged facilities, and bridge repair. The compaction was arrested by injecting water into several of the existing wells in order to maintain pore pressure as the production wells continued operation. It should be stressed that the rate of increase in relative sea level in this instance was much greater than any expected rates due to eustatic or neotectonic changes.

Japan - Several regions of Japan have experienced large rates of subsidence due to compaction, generally caused by overpumping of groundwater. Ground elevations in Niigata Prefecture and the cities of Osaka and Tokyo have dropped as much as 4 m in the past 40 years, sometimes reaching rates as high as 16 cm/yr (Takeuchi et al., 1970). Fig. 3.3 displays the isolines of the total amount of land subsidence in Osaka from 1935 to 1968. The subsidence is greatest near the coast (280 cm) and small (40 cm) in the hilly region in the center of the city where the compactible stratum is thin. Fig. 3.4a displays monthly measurements of groundwater elevation and Fig. 3.4b shows the corresponding monthly rates of compaction. The two are clearly correlated. The period where subsidence stopped is due to destruction of the city during the bombing of World War II when pumping of groundwater ceased. The installation of an industrial water system and the reduction in pumping

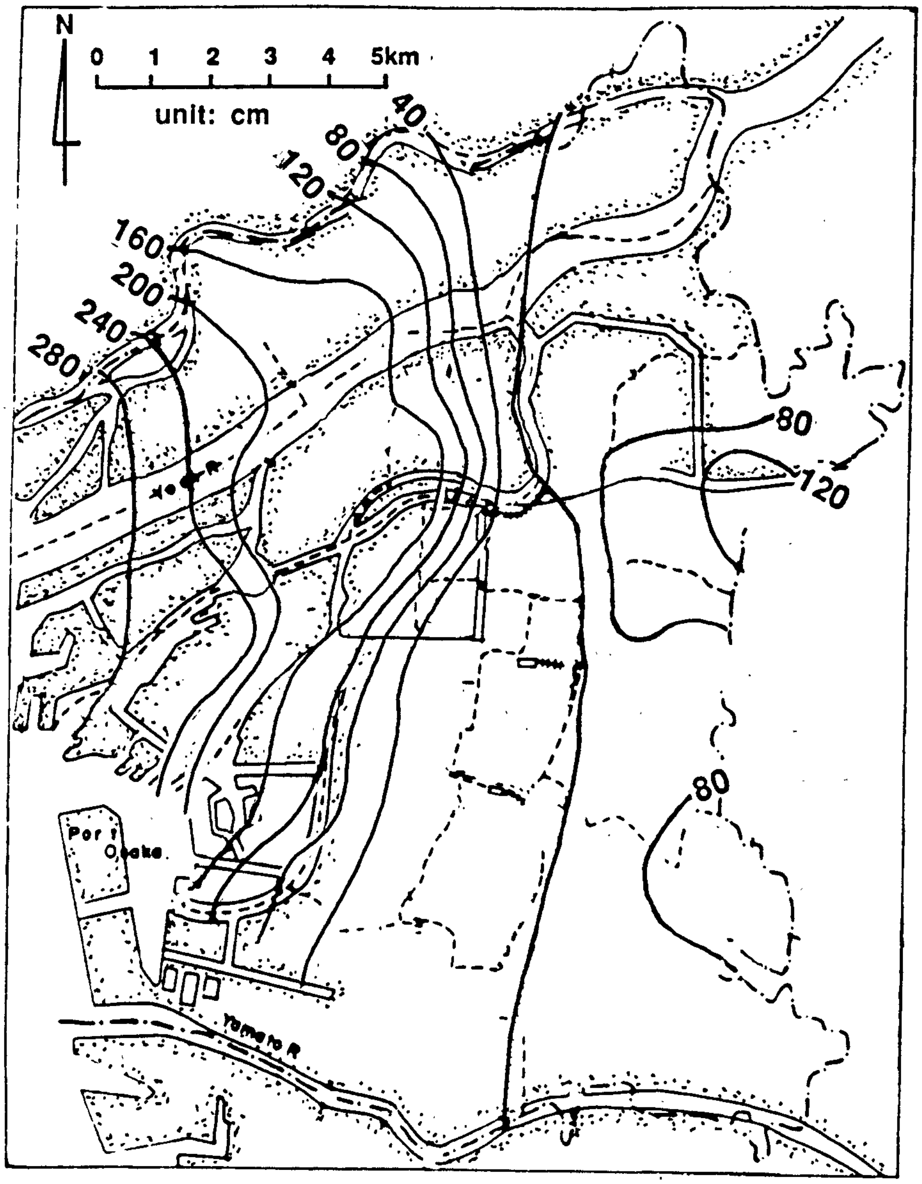


Fig. 3.3. Isolines of Total Subsidence (in cm) from 1935-1968 in Osaka, Japan (after Murayama, 1970).

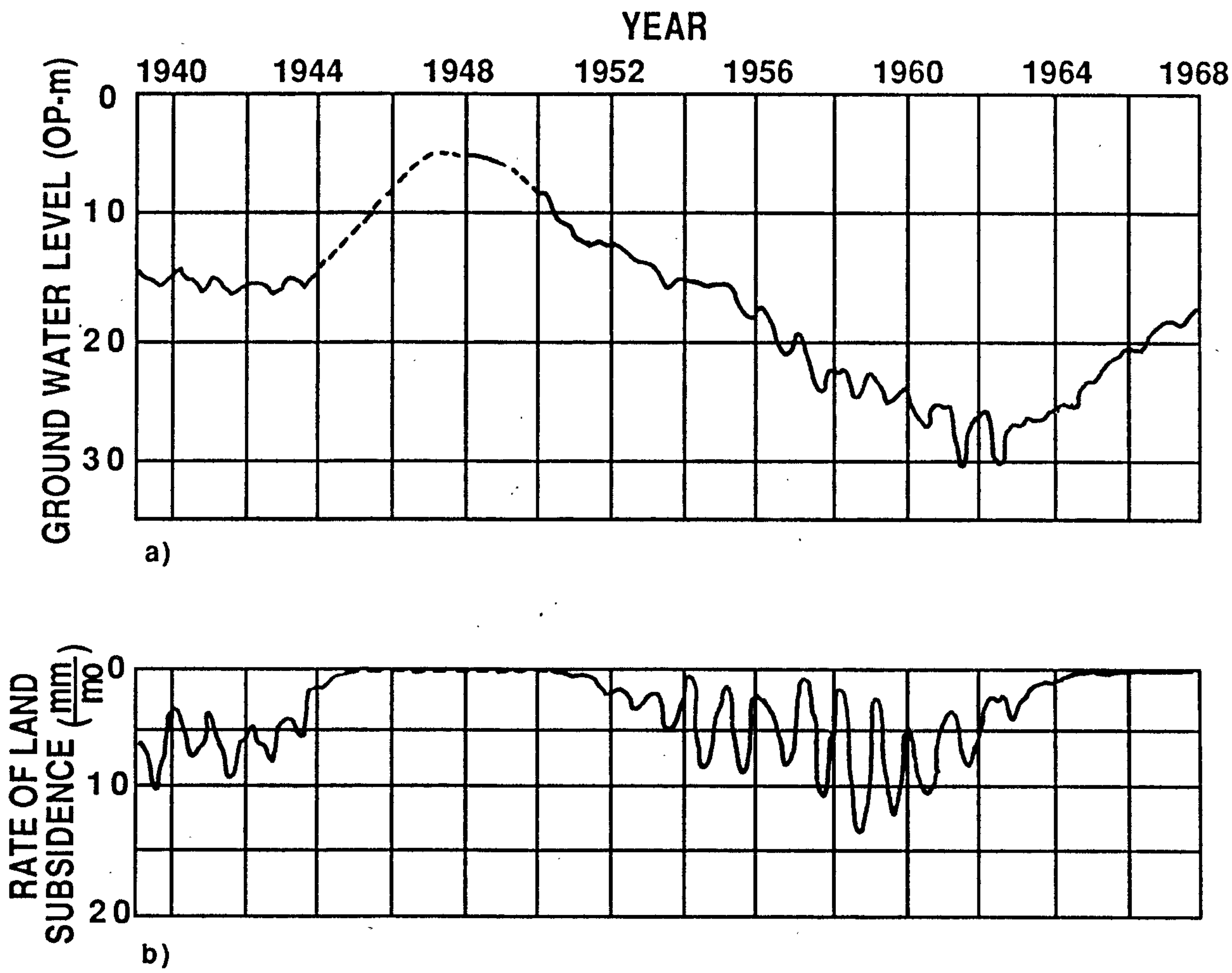


Fig. 3.4. Monthly Record of a) Groundwater Level and b) Rate of Subsidence in Osaka, Japan (after Murayama, 1970).

started in 1961 have since raised the groundwater table and arrested the subsidence. In Niigata the most severe subsidence has also occurred right on the coast. In all of these cases, regulations controlling groundwater pumping have since been enacted, plus recharge has been practiced in several of the regions where the subsidence is particularly acute. These measures have always proved successful in at least slowing the rate of compaction. In the Tokyo region however, 253 km of embankments, 41 sluice units, and 9 pumping stations were required to protect against typhoon flooding and extreme tides, and to provide drainage for rainwater (Ukena et al., 1970; Tagami et al., 1970). These are precisely the types of measures that may be required in many coastal cities within the next century.

3.6 RESEARCH NEEDS

One important aspect of compaction that requires investigation is its effect on the tide gage measurements used to determine sea level rise, as noted in section 2. Although gage elevations are often surveyed in relation to bench marks that are anchored to bedrock, the error inherent in leveling over possibly long distances would favor a more direct indication of any local compaction. It is recommended that a few experimental groundwater table/compaction devices be installed near selected tide gages. These would be located in communities where demand for the local groundwater is high, and compressibility of the underlying strata significant. If these devices prove useful, more should be added until, ideally, every tide gage used in making sea level rise estimates has at least one accompanying compaction device.

Another task necessary to resolve the eustatic component of sea level rise is to survey each tide gage in relation to orbiting satellites. Once each gage is tied-in to an absolute datum and compaction devices installed, the relative rise (or fall) in sea level at any particular site can be separated into its compaction, tectonic, and eustatic components.

A program is also needed to document compaction rates in those coastal areas currently experiencing high rates of erosion and shoreline retreat to see if compaction is playing a role, and to determine if remedial measures can be implemented. Installing arrays of compaction measuring devices will also permit study of the long-term behavior of the local subsidence as well as its relation to fluctuations and secular depletion of the water table. By sinking nearby devices to different depths, the vertical distribution of compaction can be determined, and by placing arrays farther inland along a transect the spatial behavior of compaction rates can be studied.

4. TIDAL RANGE EFFECTS

4.1 INTRODUCTION

The effect of sea level rise on the open coast and estuarine tidal ranges is a matter of significance as far as the dynamics of shoreline response is concerned, including such processes as coastal flooding, salinity intrusion and sediment transport. An obvious question is whether a rise in the range, should it occur, would overshadow the effect of the mean sea level rise itself. The phenomenon is strongly site-specific, depending upon local morphological and meteorological conditions, and also on remote forcing due to macro-scale oceanographic phenomena.

Astronomical tides are shallow water waves even in the deepest ocean, and therefore "feel the bottom". Conversely, therefore, the bottom topography and frictional resistance influence tide propagation in the sea. Since shorelines define the boundaries of the offshore shelf which is usually quite "shallow," nearshore tides are strongly influenced by the shelf topography. The distinction between tide measured "along the open coast" and, for example, at a bay entrance therefore becomes somewhat blurred. With reference to tide measured inside a bay as opposed to outside, Mehta and Philip (1986) noted that "the definition of 'outside' remains somewhat obscure in physiographic terms...". However, they added that "restrictive dimensions of bays compared to the sea impose water level oscillations whose range and frequency may be partially unrelated to oscillations outside." Furthermore, from the point of view of organizing data, the distinction between open coast tide and bay or estuarine tide may be retained, as in the following description.

4.2 LITERATURE REVIEW

The principal tide-generating forces arise from the gravitational pull exerted on the earth's surficial water mass (and to a much smaller extent on the entire mass of earth; see, e.g. Hendershott, 1972) by the moon and the sun. Darwin (1898) presented an "Equilibrium Theory of the Tides," which provides a useful qualitative description of some of the main features of the tide phenomenon based on a force balance involving gravitational attraction and centrifugal reaction for the system comprising the earth, the moon and the sun. This theory has been summarized by Dean (1966); it highlights the role of the basic forces in generating periodic oscillations of the water surface, and their dependence on such factors as the latitude, the declination of moon and the relative effects of the moon and the sun.

During the 1920's, Proudman (see e.g. Proudman, 1925) published a series of articles in which he investigated various aspects of tidal motion including the Coriolis effect due to earth's rotation. The significant advance made relative to the equilibrium theory was accounting for the actual motion of water particles on the rotating earth. Computer technology has now made it feasible to simulate tidal motion over entire oceanic masses. Early computations were based on solutions of Laplace's tidal equations (LTE). A review of numerical models of the sixties and the seventies has been provided by Hendershott (1977). Subsequently, more general forms of the Navier-Stokes equations of motion have been solved. A recent review of solutions of these ocean tidal equations (OTE) has been provided by Schwiderski (1986).

Tides in the nearshore environment are considerably influenced by winds, waves, bottom topography as well as temperature- and salinity-induced stratification. Where astronomical tides are small, e.g. along U.S. Gulf coast, non-tidal forcing often assumes overwhelming significance and modeling of a purely deterministic nature becomes difficult. Physical considerations along these lines have been reviewed by Csanady (1984).

Proudman's contributions also included considerations for tidal motions in channels of various cross-sectional shapes, and the effect of coastal configuration on offshore tidal features. A good review of simple analytic approaches for tidal propagation in estuaries, without and with bottom frictional effects, has been presented by Ippen and Harleman (1966). For the fundamentals on numerical methods for estuarine hydrodynamics, the works of Dronkers (1964) and Abbott (1979) may be cited. Nihoul and Jamart (1987) have edited a series of contributions on the state-of-the-art modeling techniques of marine and estuarine hydrodynamics using three-dimensional numerical approaches.

A special class of tidal hydraulics pertains to the hydraulics of tidal inlets or entrances connecting the sea to relatively small and deep bays. A simple, coherent theory for predicting water level variation in the bay for a given, sinusoidally forced, sea tide has been presented by Keulegan (1967). Mehta and Özsoy (1978) have reviewed various approaches including developments previous and subsequent to Keulegan's contribution.

4.3 PHYSICAL PRINCIPLES

4.3.1 Tidal Propagation

According to the equilibrium theory of tides, the tidal amplitude can be shown to be proportional (to leading order) to the fourth power of earth's radius, considering the moon-earth system. Since this number (6,378 km) is so large compared to any expected effect of sea level rise (i.e. increase in earth's radius), the corresponding change in the tidal range on this account would be negligible. In order to evaluate the effect of sea level rise on the tidal range, the nature of propagation of tide in very shallow waters must be considered.

The simplest description of tide in the dynamic sense is that of a shallow water wave moving along the x-direction with a speed or celerity, C_o. If a frictionless bottom is assumed, the wave equation is

$$\frac{\partial^2 \eta}{\partial t^2} = C_o^{\,2} \frac{\partial^2 \eta}{\partial x^2} \qquad (4.1)$$

where $\eta(x,t)$ is the instantaneous water surface elevation. The celerity, $C_o = (gh)^{1/2}$ where g is acceleration due to gravity and h is water depth.

The effect of friction can be accounted for by including an additional term on the right hand side of Eq. 4.1. Thus, for example, this term under the assumption of linearized friction is $-gM\partial\eta/\partial t$, where M is an empirical coefficient accounting for the magnitude of bottom friction. Friction slows down the speed of propagation (celerity), decreases the current speed and reduces the tidal range compared with frictionless tide. The effect is depth-dependent, and it can be shown that in fact it varies with $h^{-1/3}$, which means that increasing the water depth would decrease frictional damping, thereby increasing the tidal range. Observations in the German Bight (southern North Sea) suggest this type of a trend, as will be noted later.

Within the estuary itself, increasing the water depth can have a drastic effect on the tidal range. The majority of present day estuaries are of holocene origin, having been formed since the last ice age and accompanying sea level rise. In some, sea level rise has caused the depths to increase while in others, sedimentation rates have been high enough for the depths to have "kept pace" with sea level rise. In a few cases, e.g. some estuaries in China (Qitang, for example), sedimentation rates have essentially exerted an overwhelming control, causing the depths to decrease inspite of sea level rise, and thereby pushing the mouth seaward.

While, in general, increased water depth would increase the estuarine tidal range, the opposite effect could occur, for example, in cases where tidal resonance is a significant factor. This can be illustrated in a simple way by considering the case of a tidal wave entering a frictionless channel closed at the upstream end. In this case, considering complete wave reflection at the closed end, the incident and reflected progressive waves combine to form a standing wave, as shown in Fig. 4.1. The estuary is of length ℓ, with the closed end at x=0 and the mouth at x=$-\ell$. If the range of the progressive wave is H, the range of the standing wave at the closed end will be 2H. The standing wave envelope is thus defined by an antinode at the closed end and a node in the sea. It can be shown (Ippen and Harleman, 1966) that the ratio, R, of the amplitude, η_{om}, at the closed end to the amplitude, $\eta_{-\ell m}$, at the mouth will be (ignoring bottom friction)

$$R = \frac{\eta_{om}}{\eta_{-\ell m}} = \frac{1}{\left|\cos\left(\frac{2\pi\ell}{L}\right)\right|} \tag{4.2}$$

Since $|\cos(2\pi\ell/L)| \leq 1$, in general, the tide at the closed end will be higher than that at the mouth. This type of a resonance effect is well known, and occurs in such estuaries at the Bay of Fundy, Canada, and at Cambay in India. Given such a behavior, a situation can arise whereby an increase in water depth would in fact decrease the difference between the tide at the closed end and that at the mouth.

Consider first the case of an estuary of mean water depth, h = 15 m. Given an estuary length, ℓ = 108 km, from Eq. 4.2 R = 3.7, for a semi-diurnal tide. Now if h is increased, for example, by 2 m, R is reduced to 2.60 (assuming no change in the estuary length). Further suppose that as a result of the 2 m sea level rise, the tidal range at the mouth increases by 10%, say from 1 m to 1.10 m. Then, by virtue of Eq. 4.2, the range at the closed end will <u>decrease</u>, from 3.2 m to 2.9 m.

A bay-like water body connected to the sea via an entrance will experience range amplification as the frequency of tidal forcing approaches the natural period of oscillation of the water body. The situation is analogous to the response of a damped harmonic oscillation (Mehta and Özsoy,

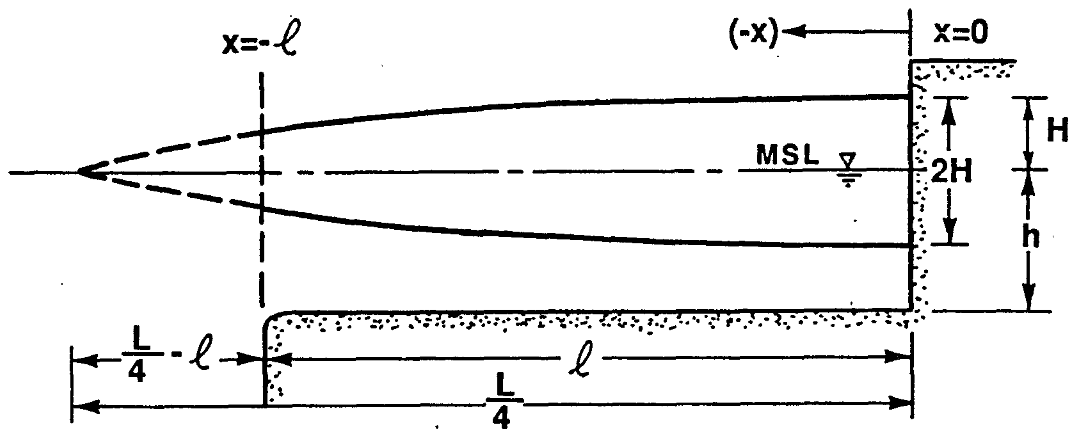


Fig. 4.1. Tidal Wave Envelope in an Estuary in which the Wave is Reflected at the Upstream Closed End.

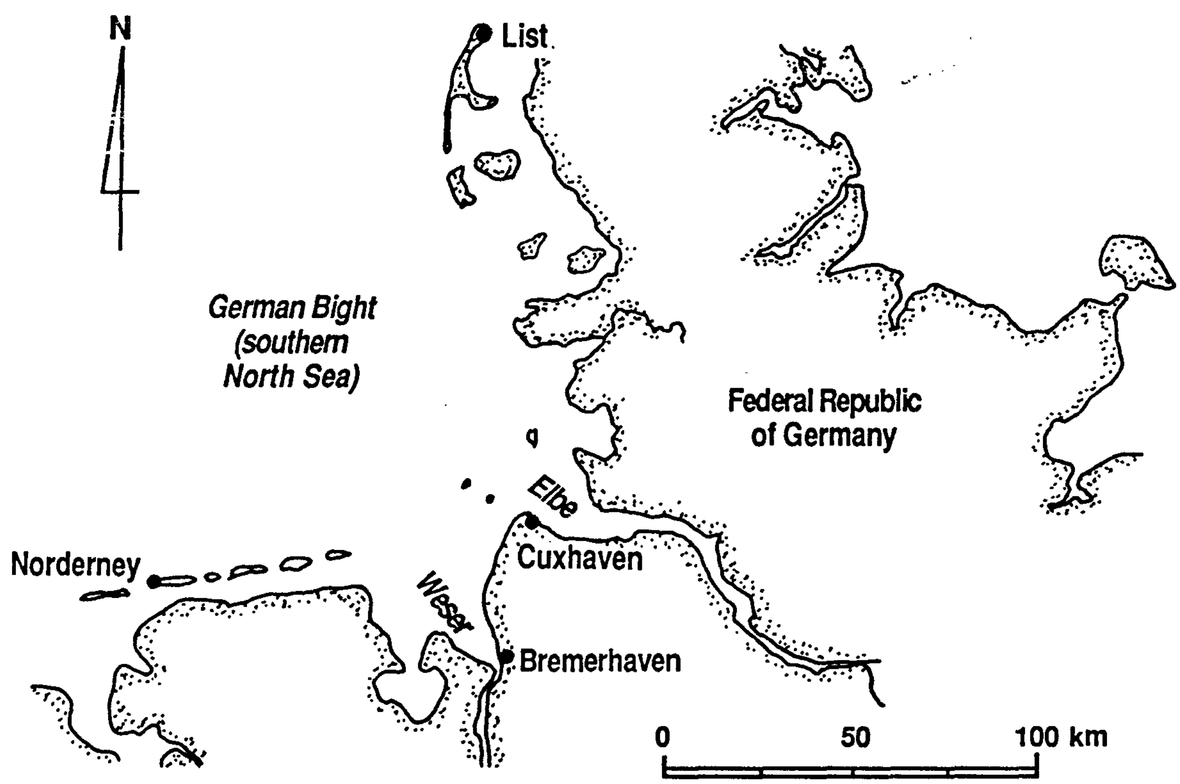


Fig. 4.2. Locations of Four Tide Gages in the German Bight.

1978). In a number of bays along the U.S. coastlines, for example, the tidal range in the bay is greater than that outside (O'Brien and Clark, 1974). Amplification becomes most pronounced when the forced and natural frequencies are equal. If therefore an increase in water depth due to sea level rise were such as to shift bay response away from resonance, the tidal range relative to that at the mouth could, as illustrated previously, decrease in spite of the opposing trend caused by decreasing bottom friction and increasing tidal admittance with increasing water depth. In a great many inlet/bay systems, however, bottom friction in the inlet channel controls the bay tide; hence in these cases sea level rise will increase the bay range, as will be illustrated later.

4.3.2 Superelevation Effect

In most bays, the tidal mean water level is usually different, often higher than mean sea level. The difference, referred to as bay superelevation, results from a number of physical factors. Mehta and Philip (1986) reviewed these factors, and the physical mechanisms by which they generate superelevation. Representative maximum superelevation corresponding to each cause, as might be found from measurements, were suggested; Table 4.1 gives a summary of the findings. Among the listed causes, sea level rise will directly or indirectly influence inlet/bay geometry, sea tide, salinity, wave penetration and some other factors. Since these in turn influence the mean bay level, in the evaluation, for instance, of the change in tidal range due to sea level change, the associated change in superelevation must be additionally considered in calculating the net water change.

Mann (1987) examined the superelevation effect resulting from inlet/bay response to tidal forcing. Tide-averaged hydrodynamic equations were developed and it was shown that bottom friction in the inlet channel is the primary cause of superelevation. Stokes drift, tidal current asymmetry and river runoff were identified (in the absence of such effects as those arising from salinity, wind waves, etc.) as the major governing physical processes. Mann considered the case of a small, deep bay connected to the sea via a long inlet channel. The combined effects of tide and superelevation resulting from sea level rise were evaluated, as will be noted in the next section.

Table 4.1. Representative Bay Superelevations (after Mehta and Philip, 1986)

Cause	Superelevation[a] (cm)	Comment
Inlet/Bay Geometry	5–30	Effect of shallow bar is more important than changing geometry with tide; hence seasonal and episodic response
Sea Tide	10	Theoretical estimate; no verification; believed to be a small contribution compared to others
Runoff	50	Major factor; strong seasonal variation
Salinity	15	Important in estuaries rather than bays (no runoff); seasonal variation
Wind	10–15	Local forcing and remote forcing can both be equally important; seasonal and episodic response
Waves	5–10	Induced pileup behind reefs may be important; seasonal and episodic response
Other Factors	1–30	Modification of tide during upland propagation and Coriolis effect are significant

[a]Only positive values are indicated. Superelevation can also be negative, i.e. lower bay level than sea, e.g. due to offshore wind.

4.4 EXAMPLES

Führböter and Jensen (1985) evaluated long-term sea level trends at ten gages in the German Bight. The evaluation was based on records obtained over a 100 year period from 1884 to 1983. Trends relevant to the present purpose may be illustrated by considering four gages, at Norderney, List, Cuxhaven and Bremerhaven, shown in Fig. 4.2. Of these, the gages at Norderney and List may be considered as "open coast" gages, while Bremerhaven is decidedly up estuary (Weser). Cuxhaven is at the mouth of estuary (Elbe). Table 4.2 gives relevant results.

Table 4.2. Secular Trends in Mean Tidal Range in the German Bight (after Führböter and Jensen, 1985)

Location	Rate of Change of Mean Range (m/100 yr)	
	N = 100	N = 25
Norderney	–[a]	0.431
List	–[a]	0.369
Bremerhaven	0.380	1.293
Cuxhaven	0.065	0.949

[a]Insufficient data

The rate of change of mean tidal range has been calculated in two different ways for each location. The first is the average rate based on the entire 100 year period (no values were computed for Norderney and List due to insufficient data). The second is based on the last 25 year (=N) record, converted to an equivalent 100 year rate. Comparing Norderney and List to Bremerhaven, it is observed that the tidal rise (N=25) has been far more significant (three-fold) within the estuary than on the open coast. The rise at the estuary mouth is intermediate in magnitude. One likely reason is the effect of reduced bottom friction due to sea level rise. This effect is more pronounced in the shallow estuary than in the deeper sea.

It is also interesting to observe from Table 4.2 (for Bremerhaven and Cuxhaven) that the increase in tidal range has been considerably more significant in recent years (N=25) than what is obtained based on a 100 year record (N=100). At Bremerhaven, the mean tidal range 100 years ago was ~ 3.30 m. Thus the range increased there by ~ 9% during the subsequent 75 years. During the next 25 years the range increased again by about the same percentage.

Führböter and Jensen noted a trend of rising tidal range approximately over the past century at all ten locations examined. They concluded that this trend is not due to any long-term changes in meteorological conditions, but is possibly due to the morphology of the North Sea, a very shallow water body in which the global rise of the mean water level effect is amplified via a standing wave effect. This possibly suggests a situation in which the natural frequency of the water body approaches the tidal forcing frequency with increasing water depth and changing boundaries.

Mann (1987) theoretically simulated the response of inlet/bay systems of assumed geometries to a total sea level rise of 1.3 m, corresponding to a 0.3 m rise over the past century and a 1.0 m projected rise. The bay was assumed to be relatively small and deep, with a surface area of $5 \times 10^6 m^2$. The inlet channel was 1,800 m long and 150 m wide. It is illustrative to consider here the case of an initially 1.5 m deep channel. For this shallow system, the ratio of the (semi-diurnal) tidal frequency to the natural frequency is 0.16, which is $<< 1$, thus signifying a friction-dominated (as opposed to resonance-dominated) system.

In Fig. 4.3, the resulting changes in the mean bay level and bay tidal range are shown. A 1.3 m rise in sea level decreased bay superelevation (head above mean sea level), from 0.27 m to 0.11 m. On the other hand, reduced friction resulted in an increased tidal range. Initially, the high water (HW) and low water (LW) amplitudes of tide relative to mean bay water level were 0.28 m and 0.25 m, respectively. The tidal range was thus 0.53 m. After a 1.3 m sea level rise, the amplitudes became 0.66 m and 0.56 m, i.e., range 1.22 m.

These data on the effect of sea level rise enable the determination of the high water level within the bay initially, and following sea level rise. Let S = sea level rise, a_{HW} = HW tidal amplitude in the bay relative to mean bay level and B = bay superelevation. Let Δa_{HW} and ΔB represent changes in a_{HW} and B, respectively. Then, initially, the HW level with respect to the initial mean sea level will be $a_{HW} + B$. After sea level rise, it will be $S + a_{HW} + \Delta a_{HW} + B + \Delta B$. Note that in the example considered, ΔB is a negative quantity. Relevant quantities in the present case are: S = 1.3 m, a_{HW} = 0.28 m, Δa_{HW} = 0.38 m, B = 0.27 m and ΔB = - 0.16 m. Thus the initial HW level relative to initial sea level was 0.55 m, which rose to 2.07 m subsequent to sea level rise.

The significance of the above result is self-evident; sea level rise could, in addition, increase the tidal range so that, in spite of a decrease in bay superelevation, high water level rise within the bay would become greater than that corresponding to sea level rise alone.

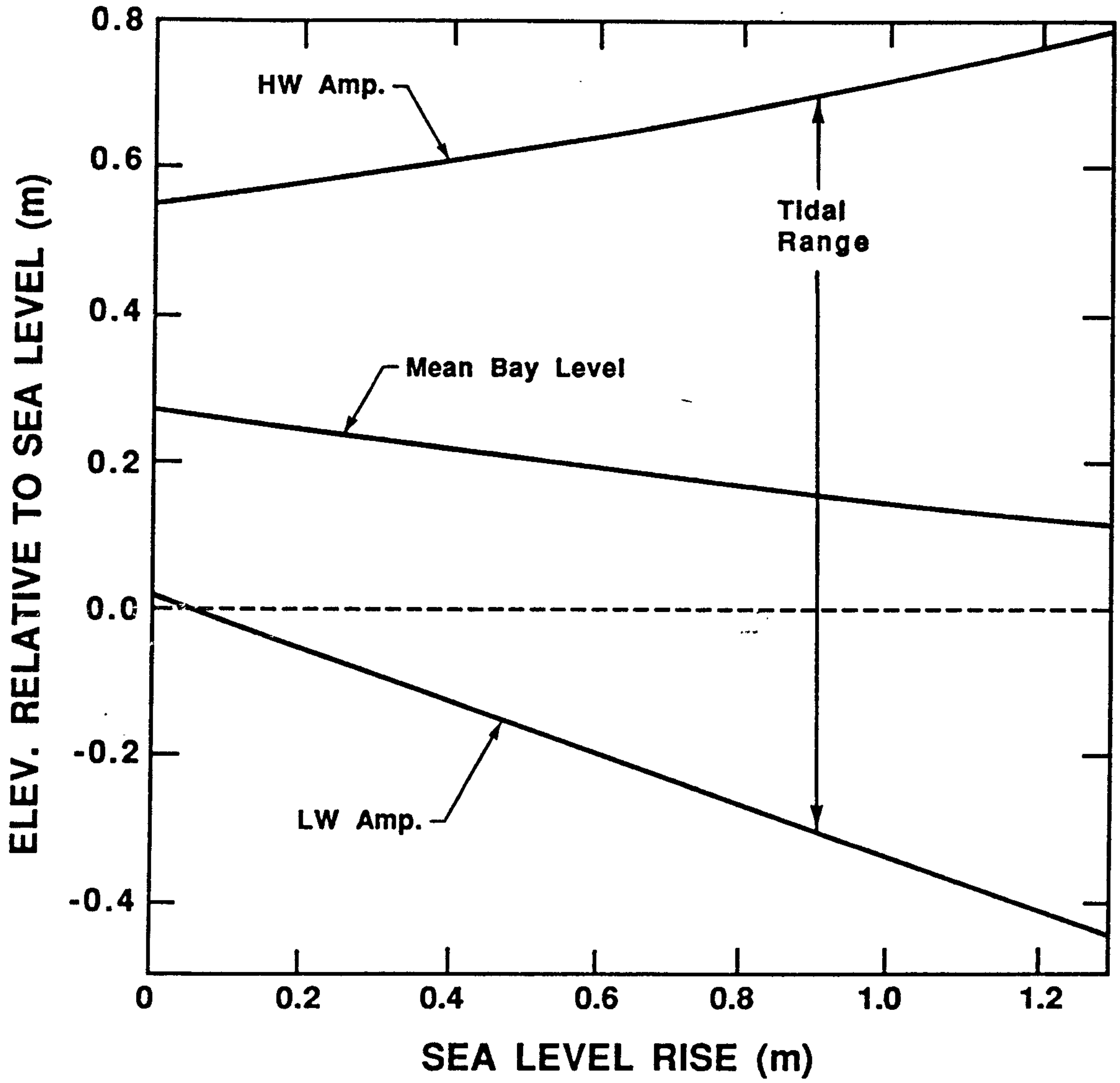


Fig. 4.3. Response of a Shallow Inlet/Deep Bay System to Sea Level Rise: Changes in Mean Bay Level and Tidal Amplitudes (based on computations by Mann, 1987).

A noteworthy conclusion based on the result of Fig. 4.3 is that the secular rate of water level rise would be lower in the bay than in the sea, on account of the decrease of bay superelevation. Hicks (1984) selected 19 pairs of gages, one inside the bay and the other at the closest location outside the entrance, for which long-term data were available. For each pair, the difference (outside minus inside) in the secular rate of change of mean water level (mm/yr) was calculated. In 12 cases, this difference was positive, which means a greater water level rise outside than inside the bay. With the exceptions of the Long Branch (NJ)/New York (NY) and Springmaid Pier/ Charleston (SC) pairs, where the differences were large (13.1 and 13.6 mm/yr, respectively), the mean of the remaining 10 pairs was 2.6 mm/yr. If bay superelevation changes were the sole effect involved (which is not by any means certain, since the gage data were probably contaminated by any number of physical phenomena), this 2.6 mm/yr change would be indicative of the rate of decrease of superelevation.

Mann (1987) showed that the changes in bay response are greater in shallow inlets than in deep ones. He also found that considering, for example, the bay to have a gentle boundary slope as opposed to a vertical wall-like perimeter would reduce the changes in superelevation and tidal range compared with the vertical wall case (Fig. 4.3). In general, however, it was concluded that due to an increase in sea level, "additional coastal flooding may occur beyond that due merely to the changes in sea level." Observations by Führböter (1986) in the German Bight estuaries seem to corroborate such a trend.

4.5 RESEARCH NEEDS

Fast computers with large memory storage have made numerical modeling of tides rather sophisticated. In many cases, it seems, modeling capabilities have "outstripped" data quality such that inaccuracies in collected data limit the accuracy of mathematical prediction. Data limitations arise from many causes; it suffices to note two factors.

One pertains to a lack of physical understanding, on a micro-scale, of phenomena which ultimately affect water level prediction. An example is our understanding of bed forms, the manner in which they change with flow, and the precise relationship between their occurrence and the flow resistance they generate. Such forms may be as small as ripples to large, migratory sand waves found in estuaries and in nearshore waters.

The second factor is related to historic tide records. Many records are highly contaminated by such unaccounted for effects as arising from land subsidence, poor leveling between gages, shifting gage locations, and a general lack of knowledge of the physical surroundings and variations in parameters characterizing these surroundings over the duration of tidal record. Thus, an accurate, quantitative evaluation of superelevation effects would require the deployment of better monitored gages. In addition, Mehta and Philip (1986) noted that our understanding of bay response and its relation to response outside would be considerably enhanced by: 1) establishment of additional primary stations along the open coast, 2) collection of long-term records at several presently designated secondary stations in bays, 3) accurate geodetic leveling connecting additional outside and inside stations, and 4) publication of relevant data in a user-oriented format. National Ocean Service initiated marine boundary programs and tidal datum survey programs appear to be directed towards this type of effort, particularly with respect to the first three items.

5. STORM SURGE AND WIND-WAVE RESPONSE

5.1 INTRODUCTION

Storm surge is the response of mean water level to the high winds, pressure differential, and rainfall associated with tropical (hurricane) and extratropical (northeaster) storms. The forces which appear to elicit the greatest responses are wind-induced shear, which tends to push water onto the beach, and the inverse barometer effect, which elevates the water level under the eye of a hurricane. For example, Fig. 5.1 displays the observed tides and storm surge associated with Hurricane Carla in the Galveston, Texas area. A complete discussion of all the relevant forces and the equations governing flows induced by storms can be found in the Shore Protection Manual, U.S. Army Corps of Engineers (1984). Solutions to idealized cases are given by Bretschneider (1966a) and Dean and Dalrymple (1984). The dependence of these solutions on nominal water depth will be examined in order to postulate some of the possible effects of long-term sea level rise.

The stress applied to the water by the high winds associated with storms is also responsible for wave generation. Bretschneider (1959) developed a family of curves from non-dimensional significant wave height induced by a hurricane, shown in Fig. 5.2. Wind waves are affected by sea level (water depth) both in their generation and as they propagate over the continental shelf. Shallow water limits the height a growing wave can attain due to steepness-induced breaking and bottom friction, while bottom friction continues to drain energy from the waves as they propagate out of the generation region. Wave generation in shallow water and losses due to bottom friction will be briefly examined in order to identify effects of depth, and hence the consequences of long-term sea level rise. Reference can be made again to the Shore Protection Manual and Bretschneider (1966b) for information on these topics.

5.2 STORM SURGE

For the idealized situation shown in Fig. 5.3a where the continental shelf is uniform in depth, we consider a spatially and temporally uniform surface shear stress due to the wind associated with the storm. According to Dean and Dalrymple (1984) the set-up, η, for steady-state conditions is given by

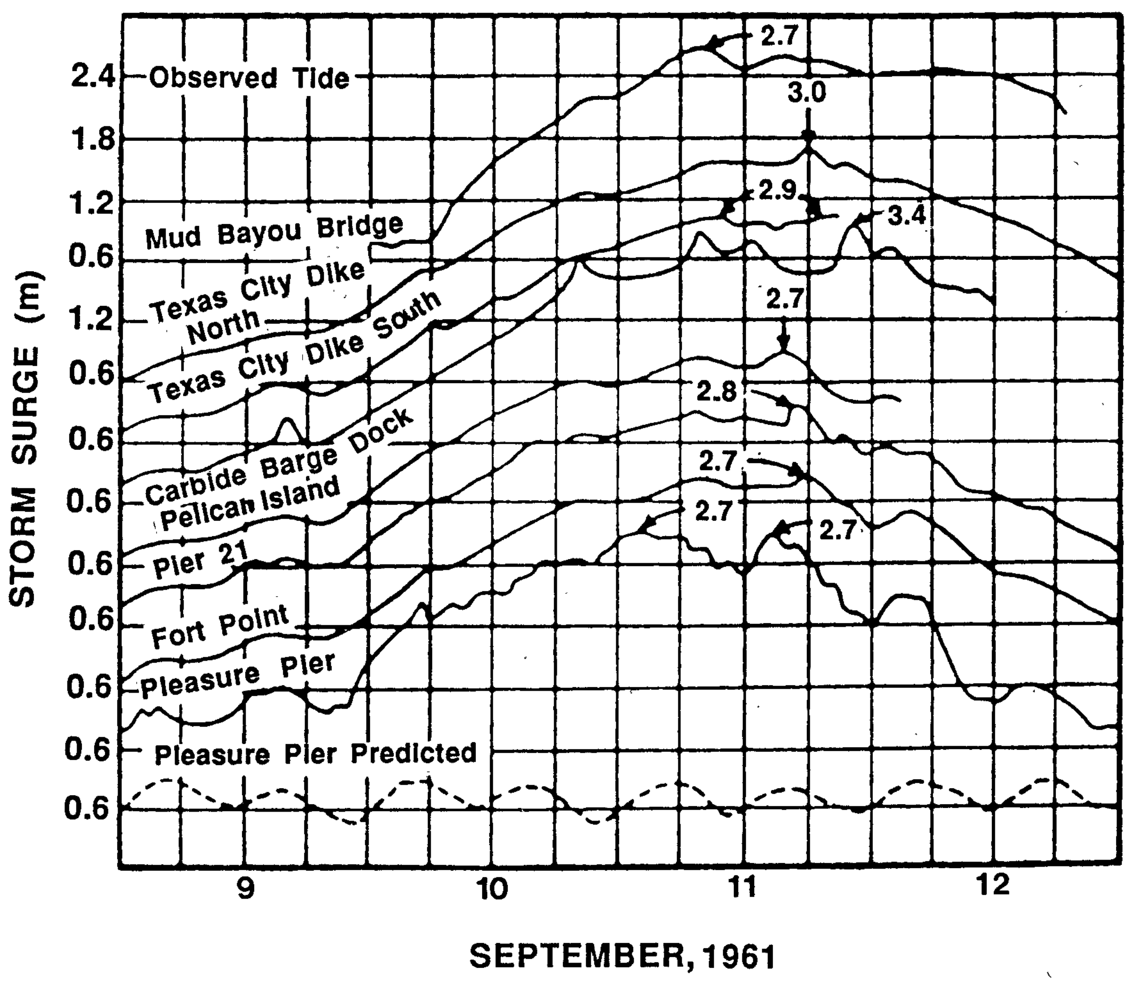


Fig. 5.1. Measured Storm Surge in Galveston, Texas Area during Hurricane Carla (adapted from Army Corps of Engineers, 1984).

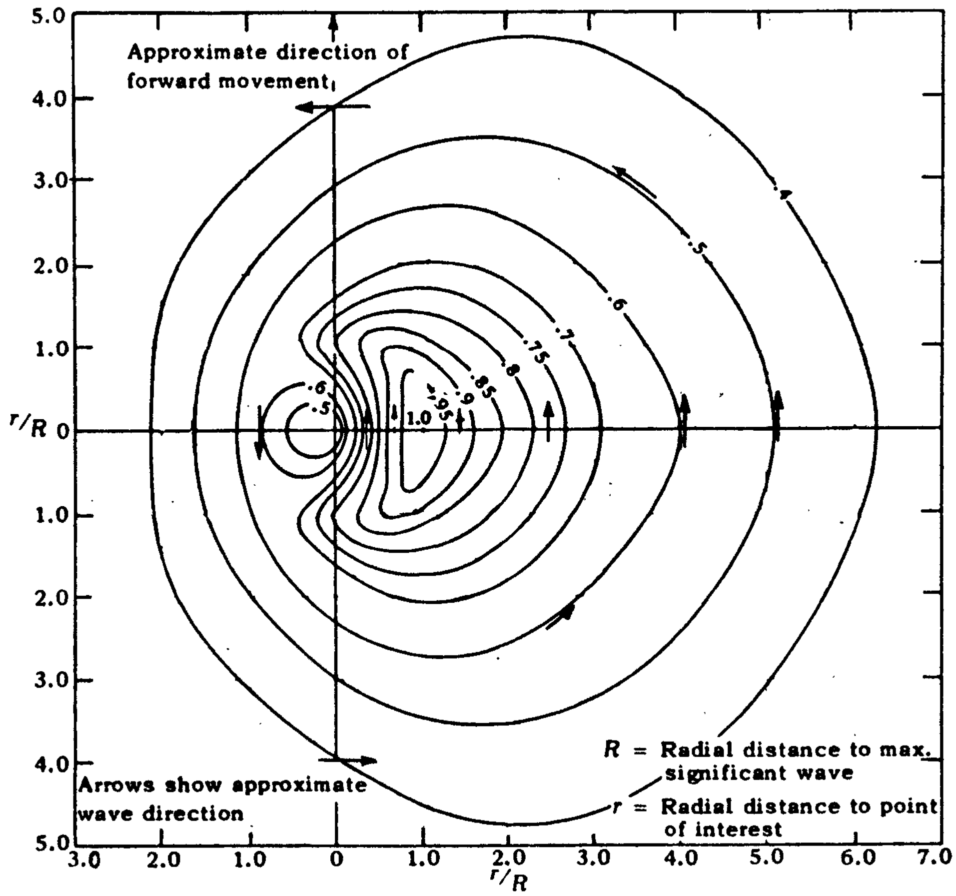


Fig. 5.2. Isolines of Non-Dimensional Significant Wave Height for Hurricane-generated Wind-waves (after Bretschneider, 1959).

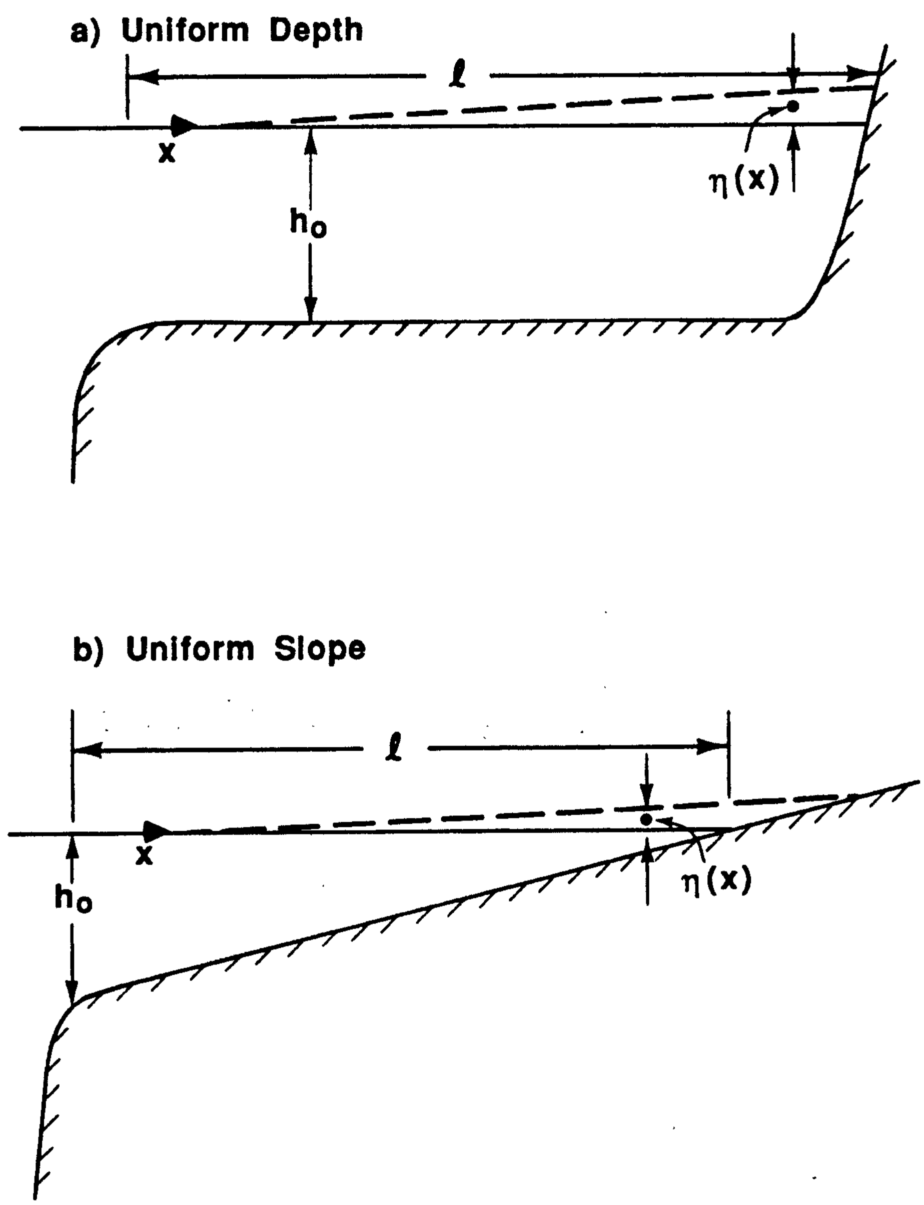


Fig. 5.3. Idealized Geometries for the Continental Shelf: a) Uniform Depth, b) Uniform Slope.

$$\frac{\eta(x)}{h_o} = \sqrt{1 + \frac{2B\ell}{h_o^{\,2}} \frac{x}{\ell}} - 1 \tag{5.1}$$

where B includes the wind induced shear stress and h_o is the original water depth. Note that at the shoreline ($x=\ell$), the set-up increases with the expanse of the shelf. Rearranging slightly and evaluating this expression at the shoreline yields

$$\eta(\ell) = \left(h_o^{\,2} + 2B\ell\right)^{1/2} - h_o \tag{5.2}$$

Taking the derivative with respect to h_o, the dependence of $\eta(\ell)$ on h_o can be examined:

$$\frac{\partial\eta(\ell)}{\partial h_o} = \left[h_o\left(h_o^{\,2} + 2B\ell\right)^{-1/2} - 1\right] \tag{5.3}$$

$$\frac{\partial\eta(\ell)}{\partial h_o} = \frac{1}{\left(1 + \frac{2B\ell}{h_o^{\,2}}\right)^{1/2}} - 1 \tag{5.4}$$

Because Eq. 5.4 is always negative, η decreases as h_o increases. This indicates that according to this simple model, as long-term sea level rises (h_o increases) the set-up induced by a given wind shear will decrease.

Consider the following situation:

average depth h_o = 10 m

shelf width ℓ = 150 m

average wind shear "head" $B = 3.3 \times 10^{-5}$m (wind speed ≈ 12.5 m/s)

and using Eq. 5.2 the wind-induced set-up at the shoreline is calculated: $\eta(\ell)$ = 0.49 m. For the same shear and shelf width, but including a 1 m rise in sea level yields $\eta(\ell)$ = 0.45 m. There is 4 cm less set-up with sea level rise, but of course the total water level would be 96 cm higher than without the 1 m rise in sea level.

Another relevant idealized geometry for the continental shelf and nearshore region is that of a uniformly sloping bottom, as shown in Fig. 5.3b. Dean and Dalrymple (1984) present an implicit solution for wind-induced set-up

$$\frac{x}{\ell} = \left(1 - \frac{h+\eta}{h_o}\right) - \frac{B\ell}{h_o^2} \ell n \left[\frac{\frac{h+\eta}{h_o} - \frac{B\ell}{h_o^2}}{1 - \frac{B\ell}{h_o^2}} \right] \tag{5.5}$$

where

$$h = h_o \left(1 - \frac{x}{\ell}\right) = h_o\left(1 - \frac{xm}{h_o}\right) \tag{5.6}$$

where m is the shelf slope. If we attempt to follow the same procedure as before, $\eta(\ell)$ is given by the transcendental expression

$$0 = \eta(\ell) + \frac{B}{m} \ell n \left(\frac{\eta(\ell) - \frac{B}{m}}{h_o - \frac{B}{m}}\right) \tag{5.7}$$

Taking the derivative of Eq. 5.7 with respect to h_o yields

$$\frac{\partial \eta(\ell)}{\partial h_o} = \frac{1 - \frac{B}{m\eta(\ell)}}{\frac{h_o m}{B} - 1} \tag{5.8}$$

from which it is difficult to immediately draw firm conclusions. However, for positive $\eta(\ell)$, from Eq. 5.7 we know that

$$0 < \frac{1 - \frac{B}{m\eta(\ell)}}{\frac{B}{m\eta(\ell)} \left(\frac{h_o m}{B} - 1\right)} \tag{5.9}$$

Quite sophisticated 2-D (planform) and multiple layer numerical models have been developed since the mid-1950's, which treat more realistic bottom topography and boundary geometry, see e.g. Jelesnianski (1965), Reid and Bodine (1968), Heaps and Jones (1975), Wang and Connor (1975), Wanstrath et al. (1976), Forristall et al. (1977), Chen et al. (1978) and Thacker (1979). The caliber of these models has outrun the quantity and quality of available field data with which to verify them. Complicated numerical models also do not lend themselves easily to examining the general interaction of the forcing mechanisms and surge response, and it may be necessary to look for ways to parameterize and scale the models to extract the effects of sea level rise on storm surge.

5.4 WAVE CHARACTERISTICS

The characteristics of waves generated in deep water should not change in response to sea level rise. However, for the same wind speeds and fetch lengths, waves generated over the continental shelf and shallower water will be higher and longer due to the reduced effects of bottom friction and steepness-limited breaking. From the Shore Protection Manual, U.S. Army Corps of Engineers (1984) the wave height H generated by wind speed U blowing over a fetch length F in water depth h is given by the expression

$$\frac{gH}{U^2} = 0.238 \tanh\left[0.53 \left(\frac{gh}{U^2}\right)^{0.75}\right] \tanh\left\{\frac{0.025\left(\frac{gF}{U^2}\right)^{0.42}}{\tanh\left[0.53\left(\frac{gh}{U^2}\right)^{0.75}\right]}\right\} \tag{5.10}$$

In shallow water this reduces to

$$\frac{gH}{U^2} = 0.126 \left(\frac{gh}{U^2}\right)^{0.75} \tag{5.11}$$

It follows that

$$\frac{\partial H}{\partial h} = (0.126)(0.75) \left(\frac{gh}{U^2}\right)^{-0.25} = 0.75 \frac{H}{h} \tag{5.12}$$

and it is clear that wave height will increase with water depth. Wave period follows an expression similar to Eq. 5.10

$$\frac{gT}{2\pi U} = 1.20 \tanh\left[0.833 \left(\frac{gh}{U^2}\right)^{0.375}\right] \tanh\left\{\frac{0.077 \left(\frac{gF}{U^2}\right)^{0.25}}{\tanh\left[0.833 \left(\frac{gh}{U^2}\right)^{0.375}\right]}\right\} \tag{5.13}$$

which in shallow water becomes

$$\frac{gT}{2\pi U} = \left(\frac{gh}{U^2}\right)^{0.375} \tag{5.14}$$

so that

$$\frac{\partial T}{\partial h} = \frac{(2\pi)(0.375)}{U} \left(\frac{gh}{U^2}\right)^{-0.625} = 0.375\,\frac{T}{h} \tag{5.15}$$

and it is apparent that waves become longer as sea level rises.

After a wave leaves the storm area where it was generated, bottom friction will drain energy and reduce its height, but should not alter wave period. The losses due to friction can be expressed by the equation

$$\frac{\partial ECg}{\partial x} = -\,\overline{\tau_b u_b} \tag{5.16}$$

where E is energy density, Cg is group velocity, τ_b is the bottom shear stress and u_b the water particle velocity. The overbar denotes time-averaging over one wave period. Defining the instantaneous shear stress as

$$\tau_b = \rho\,\frac{f}{2}\,u_b|u_b| \tag{5.17}$$

performing the time average and integrating Eq. 5.16 yields

$$H(x) = \frac{H(o)}{1+\Gamma} \tag{5.18a}$$

where

$$\Gamma = \frac{f\sigma^3 H(o)x}{3\pi g\ Cg\ \sinh^3 kh} \tag{5.18b}$$

The effect of rising sea level will depend on the geometry of the continental shelf. For a uniform depth where the rise in sea level does not affect the shelf length, Γ will decrease and H at the shoreline will increase. However, if the geometry is such that a rise in sea level results in a wider shelf, Γ may tend to increase and H at the shoreline will then decrease, because bottom friction has had a longer time to drain energy.

Consider a storm on the continental shelf that results in the conditions:

wind speed $U = 30$ m/s ($g/U^2 = 0.01089$ m^{-1})
fetch length $F = 50$ km
average water depth $h_o = 10$ m

so that $gF/U^2 = 544.4$ and $gh/U^2 = 0.1089$. According to Eq. 5.10 waves will be generated whose heights are 2.06 m, and according to Eq. 5.13 the period will be 6.3 s.

Next consider the same storm after a 1 m rise in sea level. Following the same procedure as before yields $H = 2.18$ m and $T = 6.4$ s, or an increase in wave height and period of 5.8% and 1.6% respectively. If the continental shelf is 150 km wide and has a friction coefficient $f=0.01$, the loss in wave height due to bottom friction is calculated using Eq. 5.18 and the wave height in the nearshore is found to be $H = 0.82$ m for the case without sea level rise. With the initial wave conditions for the 1 m sea level rise, the wave height on the inner shelf is found to be 0.96 m, or a 16.6% increase in wave height due to the combined effects of sea level rise during generation (slight) and reduced bottom friction on the shelf (marked).

More detailed numerical models for wind-wave generation have been developed, e.g. Cardone et al. (1976) and Resio (1981). Several models have been intercompared by the Sea Wave Modeling Project (SWAMP, 1985) but without definite conclusions due to lack of data. Cardone (1986) concludes that the level of error in wave height, period, and direction is on the order of 10% if high quality wind data are available. However, such data seldom are, and for predictive purposes the use of less accurate models for representing winds is often necessary.

5.5 RESEARCH NEEDS

The aspects of storm surge and wind-wave generation that require research have less to do with long-term sea level rise, than with the basic phenomena themselves. Storm surge has received intensive theoretical and numerical study over the past three decades, and several sophisticated numerical models exist. However, there is a conspicuous lack of field measurements of hurricane and extratropical storm surge with which to calibrate and verify these models. Required are concurrent time series from devices placed along the coast at intervals small enough to resolve the behavior of the surge as a storm moves out of the open ocean and makes landfall. The ability to model and predict storm surge cannot improve significantly without such data. Also, several phenomena associated with storms such as the superelevation of water level before arrival of the storm (often referred to as a forerunner) are still a mystery.

Research on wind-wave generation in deep and shallow water has progressed well. However, as noted there is a lack of detailed, high quality wind and wave data with which to verify these models. The basic process of damping of wind-waves as they cross the continental shelf due to bottom friction and breaking induced by wave-wave interaction are other areas in which research is needed. Theoretical work has progressed, but accurate field measurements are lacking. It is also necessary to stress the spectral approach to damping, as most methods available to date are limited to the assumption of monochromatic waves. Basic research on the directionality of wave spectra, in both deep and shallow water, is also necessary before a better understanding of the effects of sea level rise on ocean waves can be assessed accurately.

6. INTERACTION WITH NATURAL FEATURES AND CONSTRUCTED WORKS

6.1 INTRODUCTION

Assuming that sea level will rise a significant amount over the next century, and that shorelines will generally respond in some manner, the question arises as to by what means can (or will) this response be modified or prevented. Natural features such as shoals, headlands, inlets and even barrier islands themselves will cause the neighboring shorelines to respond in a manner different from that of the typical "open" coast. Man-made engineering works, e.g. breakwaters, jetties, and beach fills, by their very purpose alter shoreline response from that of nature, and so can modify shoreline response to sea level rise. Alternatively, the design, construction, and cost of coastal projects is highly dependent on local water depth. Relative sea level rise must therefore be addressed for a project having a long design-life.

On a sandy coast, sea level rise generally invokes shoreline response by two mechanisms. First is simply the retreat due to flooding or inundation, which is often small because natural beach profiles are usually concave upwards in shape. However, the rise in sea level builds a large potential for additional erosion and shoreline retreat induced by wave action, which can be quite severe. The only means of preventing shoreline retreat due to inundation is by constructing dikes and seawalls. All other features which modify shoreline response, both natural and man-made, do so by altering or reducing the wave climate and have little effect on the inundation component. These features/structures are now discussed individually.

6.2 NATURAL FEATURES

Barrier islands - are the elongated natural islands composed of sandy material, which front a substantial portion of the mainlands of the world. These islands block out the wave activity to which the mainland shoreline would otherwise be subjected, essentially acting like large breakwaters. Although the mainland shorelines are still vulnerable to flooding due to sea level rise and wind-waves generated locally in the bays, barrier islands are the paramount safeguard against realization of the full erosional potential of sea level rise in the back-bay region. This potential is especially strong because the sediments found here are often fine sands, silts, clay and peat, all highly erodible.

If sea level rise causes local barrier islands to deteriorate and "drown in place" rather than migrate landward, progressively more wave energy will penetrate through the chain and attack the mainland shoreline. This can result in enlargement of the bay area as the mainland erodes. An example is the Isles Dernieres on the coast of Louisiana, see Penland, et al. (1985). As shown in Fig. 6.1, since 1853 the large barrier has deteriorated drastically to become a series of five small islands, which have retreated about 2 km in 125 yrs. Most of this is due to inundation and erosion accompanying the rapid subsidence of the delta region. Concurrently, Lake Pelto has been greatly enlarged by erosion of both the isles and the mainland. As the isles continue to disappear, erosion of the mainland should accelerate.

Shoals - are large deposits of sediment, usually associated with relict barrier islands, inlets, and large headlands. They serve to naturally limit the wave energy that impacts a shoreline, as a result of dissipation due to bottom friction and breaking, as well as partial reflection. As sea level rises, shoals become less effective unless their natural response is to grow, as is the case at inlets as described subsequently. Such growth of course requires sediment and may demand it from neighboring shorelines or inlets. An example of the effects of offshore shoals on the neighboring shoreline is Cape Canaveral, Florida, shown in Fig. 6.2. This cape has an extensive system of offshore and shore-connected shoals, which generally protect the cape from storm wave activity out of the northeast. Little protection is afforded from the southeast. The regional direction of net longshore drift is from north to south, and the offshore shoals A, B, C and the Hetzel Shoal have afforded enough protection for the Chester Shoal and False Cape to form utilizing this supply of sediment. The entire shoal system is responsible for the formation and protection of Cape Canaveral and Southeast Shoal. Farther to the south the shoreline assumes a crenulate shape as is common for such features. Field and Duane (1974) report that since 1878 Chester and Southeast Shoals have become broader and thicker, and the offshore shoals have migrated slightly to the southeast. Since 1898 accretion has occurred on the southern sides of Chester and Southeast, while the shoreline between these shoals and to the south of the cape have experienced erosion. This seems to indicate that as

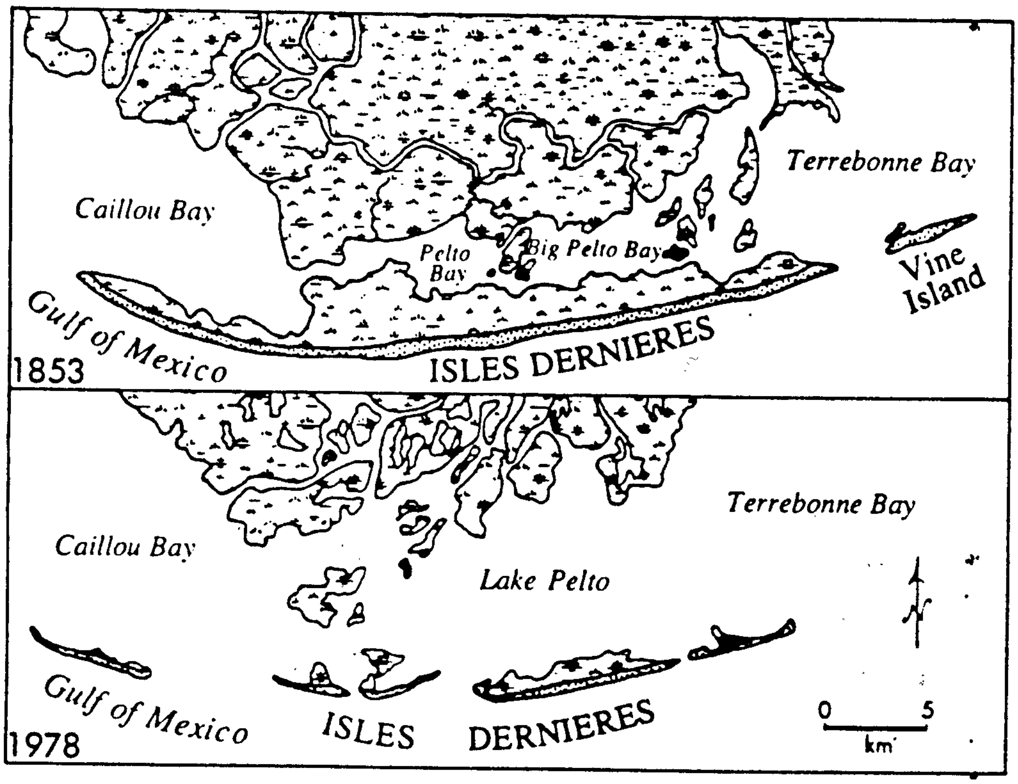


Fig. 6.1. Historical Shoreline Changes at the Isles Dernieres, Mississippi (after Penland <u>et al.</u>, 1985).

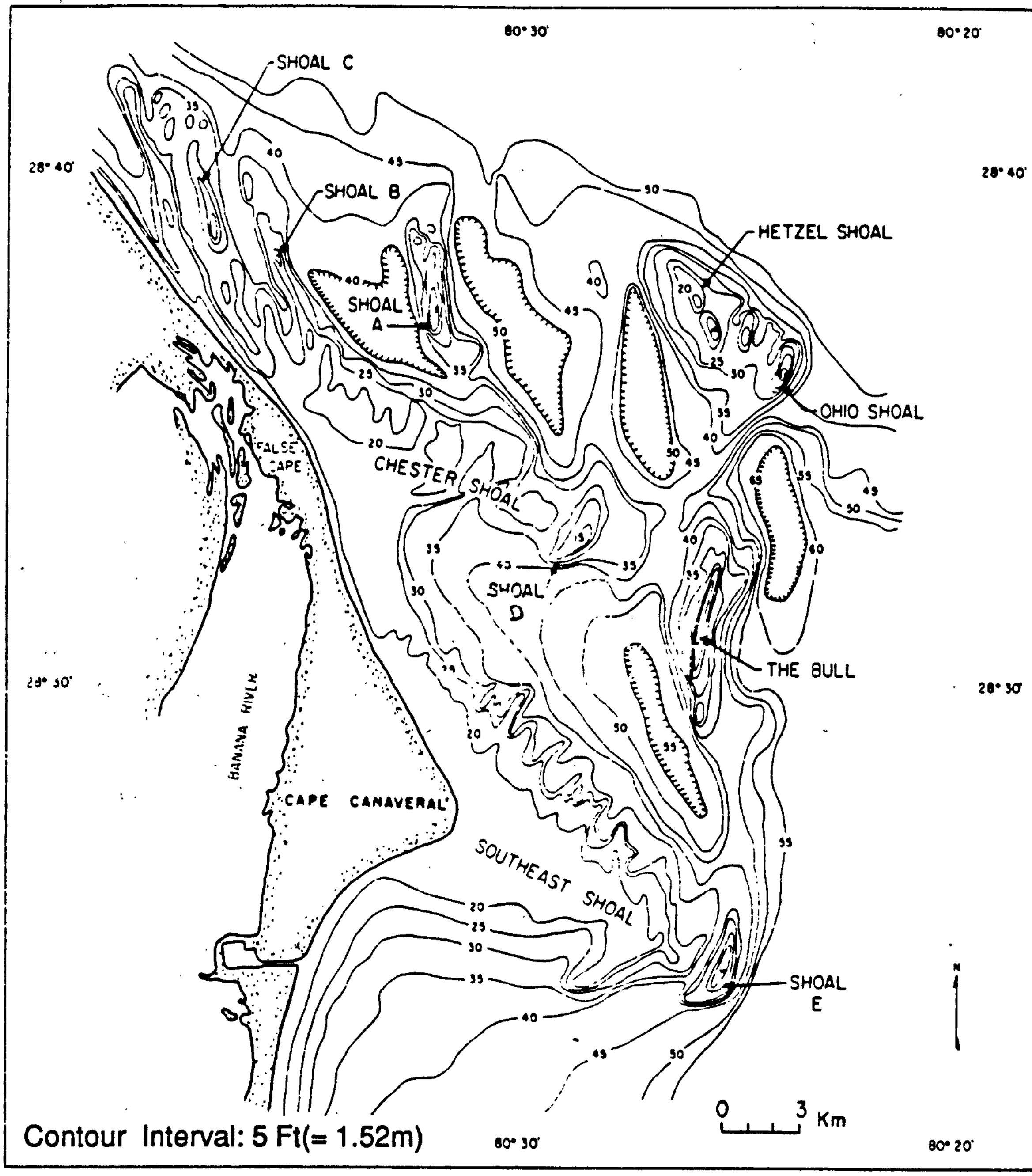


Fig. 6.2. The Shoal System at Cape Canaveral, Florida (after Field and Duane, 1974).

sea level has risen the protected areas at False Cape-Chester Shoal and Cape Canaveral-Southeast Shoal have continued to be maintained, apparently at the expense of the region to the south which has become more crenulate in shape due to blockage of longshore drift from the north.

Natural inlets - are the breaches between barrier islands, usually cut during storms. They generally affect neighboring shorelines as they migrate alongshore, which causes the updrift side to accrete and the downdrift side to erode. After a natural or man-made inlet is cut, ebb shoals along the mouth of the inlet grow and shunt sediment across the inlet and partially maintain the supply of sand to the downdrift beach. Sea level rise will tend to trigger a chain of events which could result in larger shoreline retreat than would occur if the inlet were not present. A rise in mean water level increases the depth of a bay and increases the hydraulic efficiency of an existing inlet, thereby increasing the tidal prism. An increase in tidal prism increases the velocities in an inlet which in turn may deepen or enlarge the throat. The ebb shoal then demands more sediment from the neighboring shorelines as the inlet grows in size.

Natural tidal inlets also trap sediment in shoals on the bay side of the inlet (called flood shoals). Because of the reduced wave climate in the bay, these shoals are usually left behind as the inlet migrates and thus become a sink for sediment. As is the case with ebb shoals, the size of these shoals generally increases with the size of the inlet, and an increase in inlet size due to sea level rise will tend to remove more sediment from the beaches and store it in flood shoals.

A good example of the effect of long-term sea level rise on a natural inlet is Nassau Sound, Florida, shown in Fig. 6.3. Calculations of the volume of sediment contained in the ebb shoals by Marino and Mehta (1986) indicate addition of $6.3 \times 10^6 m^3$ of material from 1871-1970. The shoal volume is currently about $40.5 \times 10^6 m^3$. During this 99 year span, relative sea level rose 0.3 m. Long-term sea level rise will also promote creation of additional inlets, each with their own demand for sediment to maintain shoals.

Headlands - are natural intrusions of hard material on an otherwise sandy shoreline. These less-erodible features act as natural groins or breakwaters and compartmentalize a shoreline. A large single isolated headland usually causes a crenulate embayment to form on its downdrift shoreline, as is the

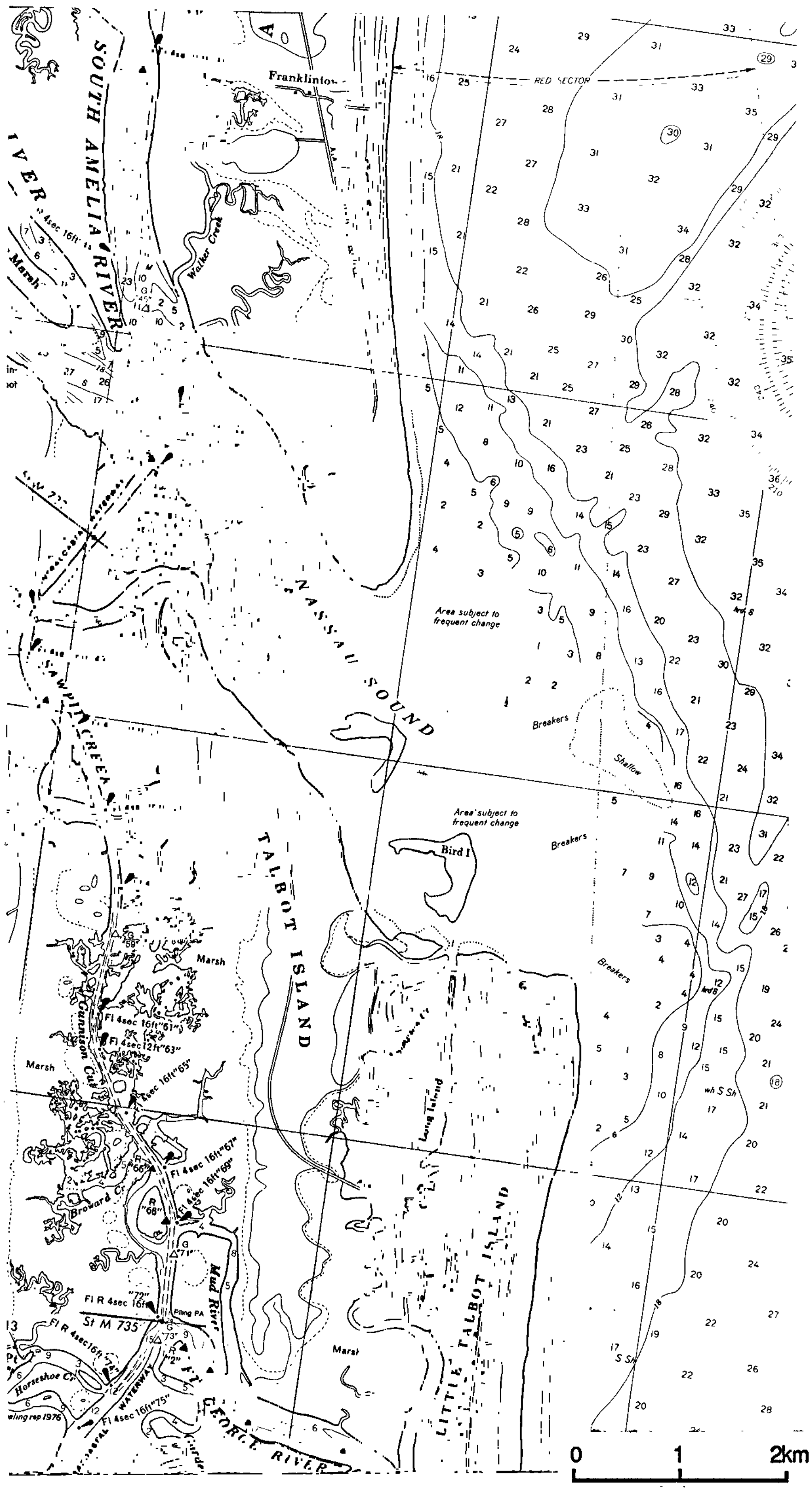


Fig. 6.3. Bathymetric Chart of Nassau Sound, Florida Showing Ebb Shoals. Depths are in Feet (from NOS Nautical Chart 11489).

case at Cape Canaveral. A series of two or more headlands spaced closely enough to act as a system will cause formation of embayments that are more semi-circular in shape.

The role that a headland or series of headlands will play in modifying shoreline response to sea level rise will depend on the amount of incident wave energy dissipated or reflected by the headland(s), and the aspect ratio (ratio of width to length) of the embayment(s). Those with broad faces parallel to the coast block significant amounts of energy, and their embayments have larger aspect ratios. This means they significantly increase the length of shoreline available to "resist" a given amount of wave energy - the amount being controlled by the fixed distance between headlands. Because of the reduced energy density at the shoreline, less of the potential erosion takes place as sea level rises. This situation is analogous to the performance of offshore breakwaters. However, narrow headlands do not block significant amounts of wave energy and although the shoreline in between may be reoriented, it is not lengthened substantially. Because the energy density at the shoreline is not reduced, little is done to affect on/offshore transport and therefore the full potential for erosion associated with a rise in sea level can be realized. This situation is analogous to a groin field, to be discussed subsequently along with offshore breakwaters.

An example of the effect of headlands on shoreline evolution is Wreck Bay on the west coast of Vancouver Island, British Columbia, shown in Fig. 6.4. Quisitis Point and Wya Point are two natural headlands responsible for the large embayment inbetween. Although historical shoreline changes for the bay are not readily available, the general behavior in response to future sea level rise is expected to be as described. Part of the erosive potential of the rise will be spent on lengthening the shoreline of the bay as it enlarges. So, the average retreat of the shoreline will be less than on the open coast.

6.3 CONSTRUCTED WORKS

Dikes and levees - are free-standing, elongated mound-like structures used to prevent coastal and riverine flooding and to create usable land from low-lying, previously inundated wetlands. They are usually constructed of earth or sand (armored by clay, asphalt, rubble or vegetation), masonry, and

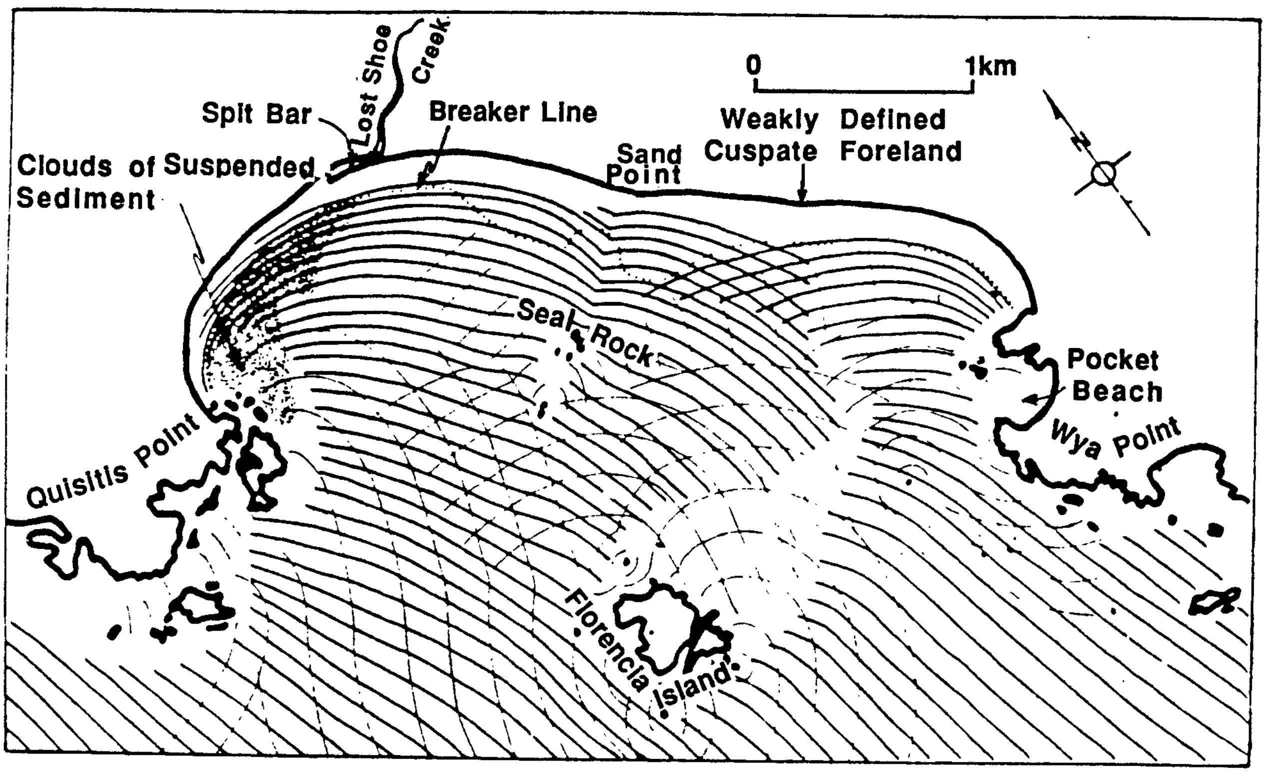


Fig. 6.4. Shoreline Between Two Headlands at Wreck Bay, Vancouver Island, with Observed Wave Patterns (after Bremmer and LeBlond, 1974).

concrete, and are often assisted by pumps to remove seepage. A few typical design cross-sections are shown in Fig. 6.5. Although costly to construct and maintain, dikes are the only means of totally preventing shoreline retreat (both inundation and wave-induced) due to a rise in relative sea level. They "modify" shoreline response essentially by creating a new shoreline at the structure location, and have been successful in many places throughout the world, more notably the Netherlands. If long-term sea level rise is significant, dikes may be the only workable means of protecting coastal cities.

The effect of sea level rise on existing dikes and the design of new ones is manifested predominatly in the required crest elevation of the structure. This is the elevation that prevents significant overtopping during the design storm. Crest elevation in turn determines the cross-sectional area of the structure and the volume requirements for material. A crude relationship is that the height increases directly with the rise in sea level and area increases with the square of the increase in sea level. More precise estimates depend on the actual design cross-section. The major sea level related question confronted in the design of new dikes and levees is whether to include projected long-term estimates or not. Answers depend on the site specific estimate of the rate of sea level rise, the type and method of construction to be used, the expected lifespan of the structure, and the expected frequency of maintenance. In regions where relative sea level is rising rapidly due to ground subsidence or tectonics, a projected estimate of suitable length (on the order of 10-50 years) may be most appropriate. "Hard" structures such as dikes built with masonry, concrete, and rubble are usually very expensive to maintain or improve, and should be designed using the maximum long-term sea level estimate projected during the design life of the structure. "Soft" structures built with sand or earth and armored with vegetation usually require frequent (but relatively inexpensive) maintenance and are more easily altered and improved. This permits raising of the crest elevation in response to actual sea level rise, rather than designing for a perhaps uncertain projection of sea level.

The best example of the use of dikes and levees to prevent coastal flooding, and their interaction with long-term sea level rise, is the Netherlands. Dikes have existed in the Netherlands since pre-Roman times and

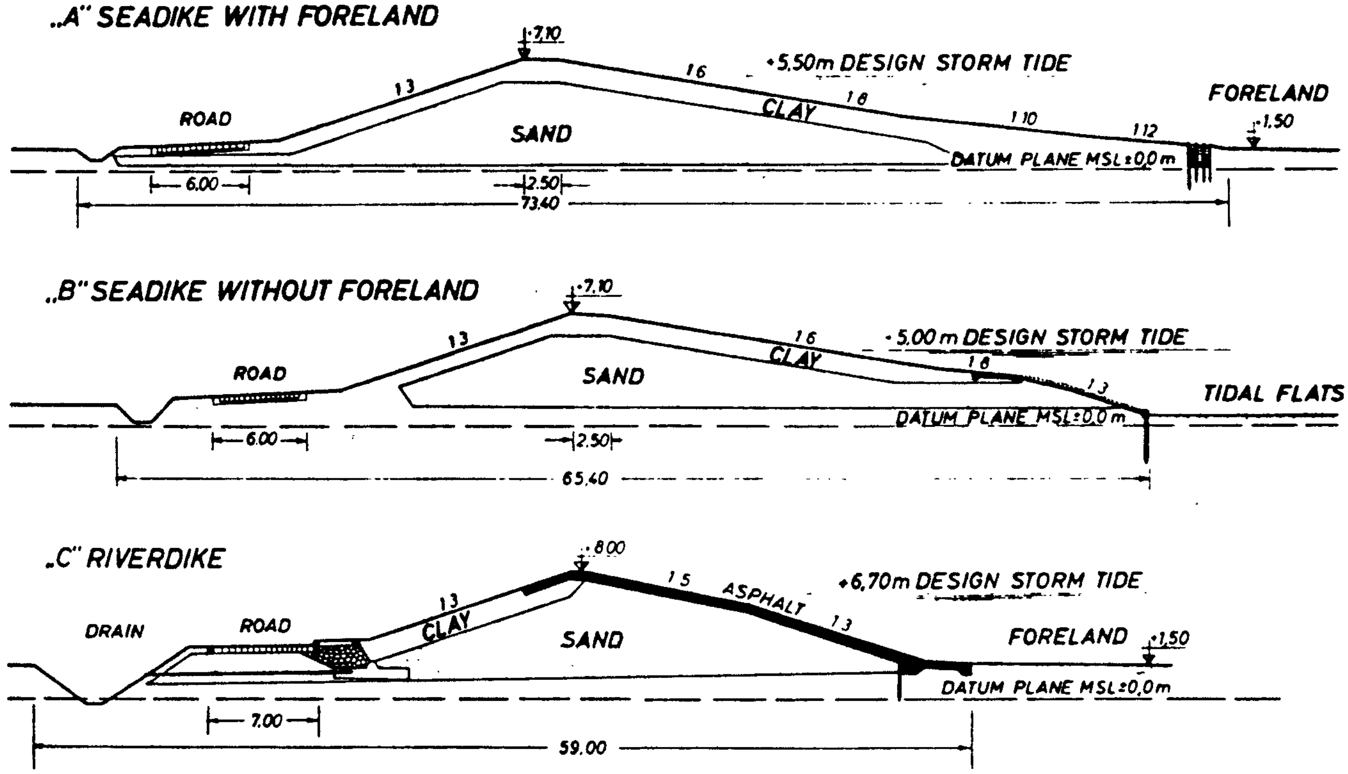


Fig. 6.5. Examples of Design Cross-sections for Sea Dikes (after Kramer, 1971).

over 1,000 km now exist (Lingsma, 1966). However, several catastrophic failures during storms have served to periodically demand a review of their use and design, the most recent being the flood of February 1, 1953 where 1,783 people were killed and total damage was estimated at 250 million dollars (Wemelsfelder, 1953). This disaster prompted construction of the massive Delta Project, whose large storm surge barriers were just recently completed (see Kohl, 1986). As shown in Fig. 6.6, almost half of the Netherlands is below mean sea level and protected by dikes. The situation here is a clear microcosm of the future of many regions around the globe if the greatest estimates of sea level rise prove accurate.

Seawalls bulkheads, and revetments - are structures of concrete, masonry, steel sheet pile, or rubble used to armor the shoreline and prevent retreat due to the combination of wave activity and sea level rise. Although performing much like dikes, they usually are not free-standing and are always "hard" features, with vertical or steeply sloping faces. They generally are used on a local rather than regional basis and are built to protect the upland along a limited section of beach. Besides their cost, the major drawback to seawalls is that as sea level rises progressively less sandy beach is available for recreation and additional storm protection. Periodic beach nourishment is often required as mitigation. Typical cross-sections for a seawall, bulkhead and revetments are shown in Fig. 6.7.

Sea level rise affects the design, construction, and maintenance of these structures in the same general manner as with dikes. As sea level rises, higher crest elevations are required, but because the structures are not free-standing the required cross-sectional area increases more linearly than quadratically with sea level. As with dikes, a seawall, bulkhead or revetment can either be designed with enough crest elevation to account for projected sea level rise, or else the crest elevation can be periodically raised in response to sea level. Because these are hard structures, it is usually difficult and expensive to exercise the second option.

Galveston, Texas is fronted by a seawall constucted after the city was demolished during a major hurricane in 1900, in which more than 6,000 people were killed. The wall, whose cross-section was shown in Fig. 6.7a and planform is displayed in Fig. 6.8, is 4.9 m high and over 16 km long. Nine million cubic meters of fill were placed behind the wall and much of the city

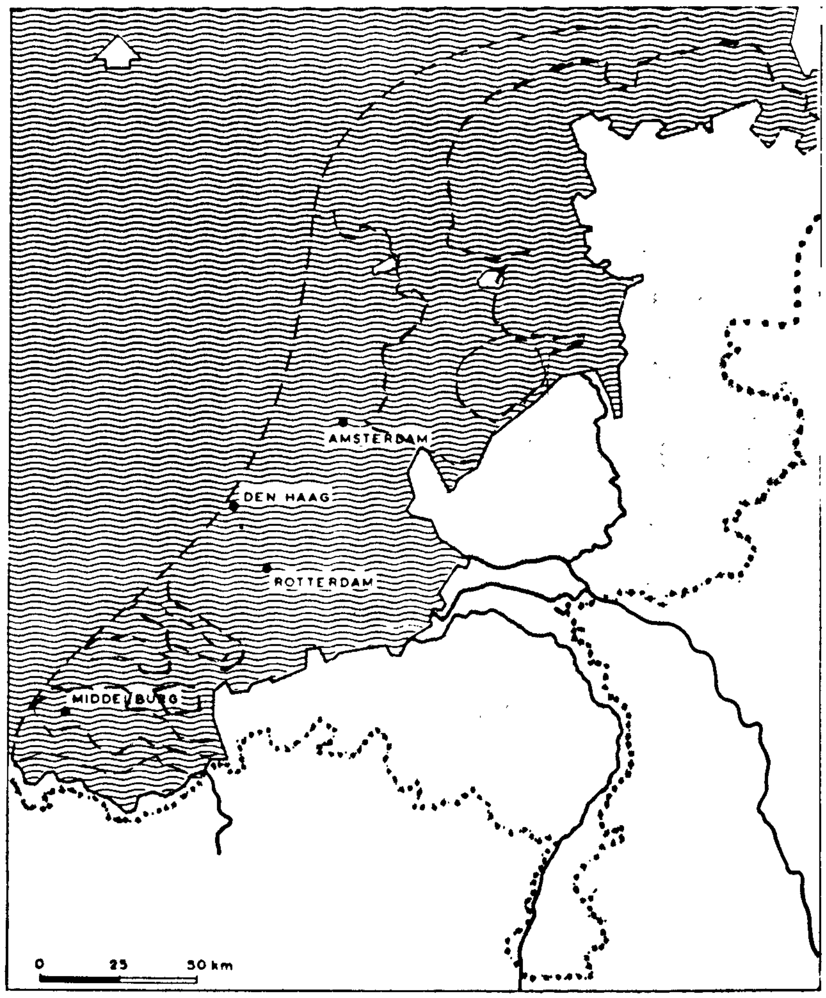


Fig. 6.6. Shoreline of Holland if There Were No Dikes, Showing a 50% Loss in Land Area (after Lingsma, 1966).

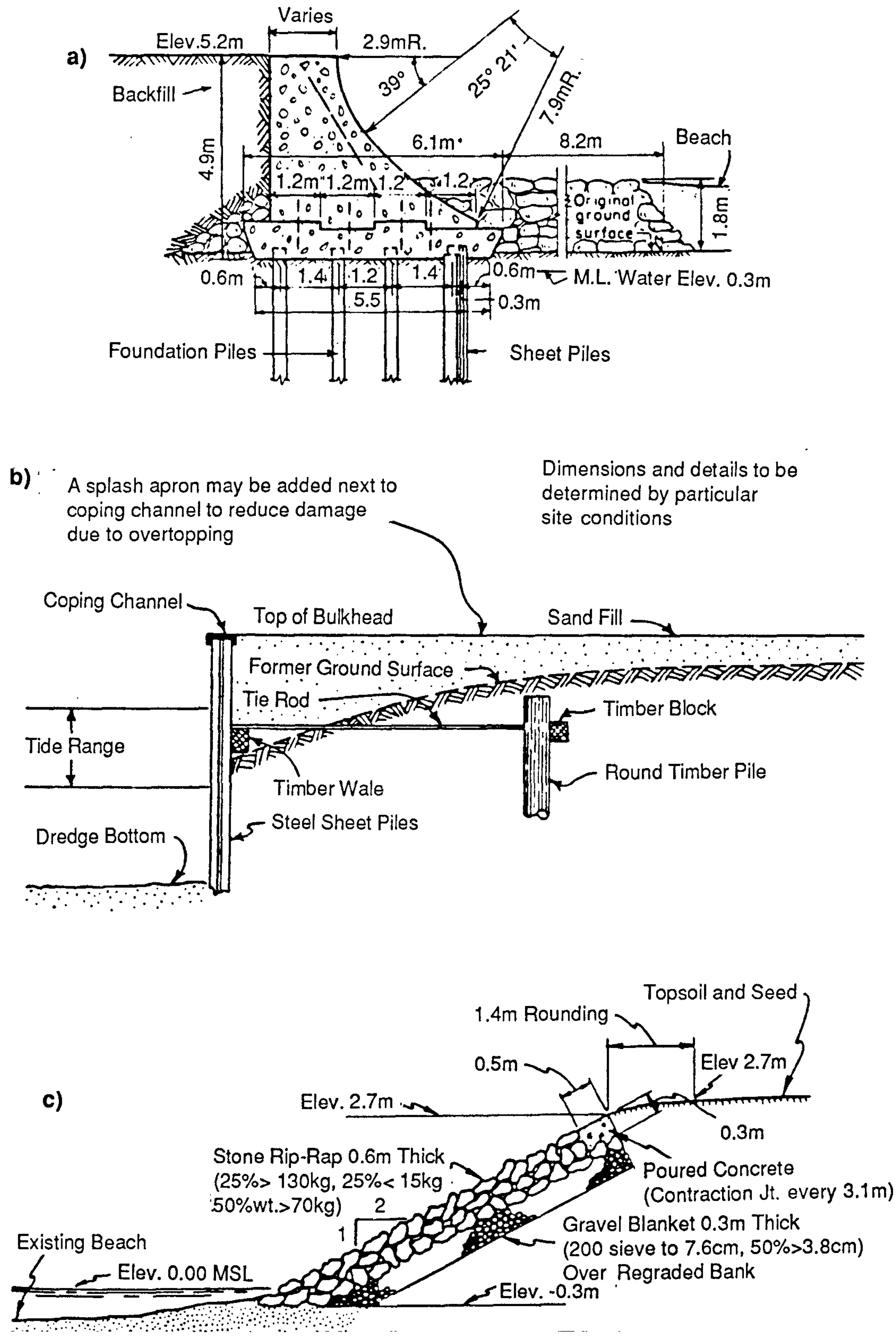


Fig. 6.7. Typical Cross-sections of a) Seawall, b) Bulkhead and c) Revetment (adapted from Shore Protection Manual, U.S. Army Corps of Engineers, 1984).

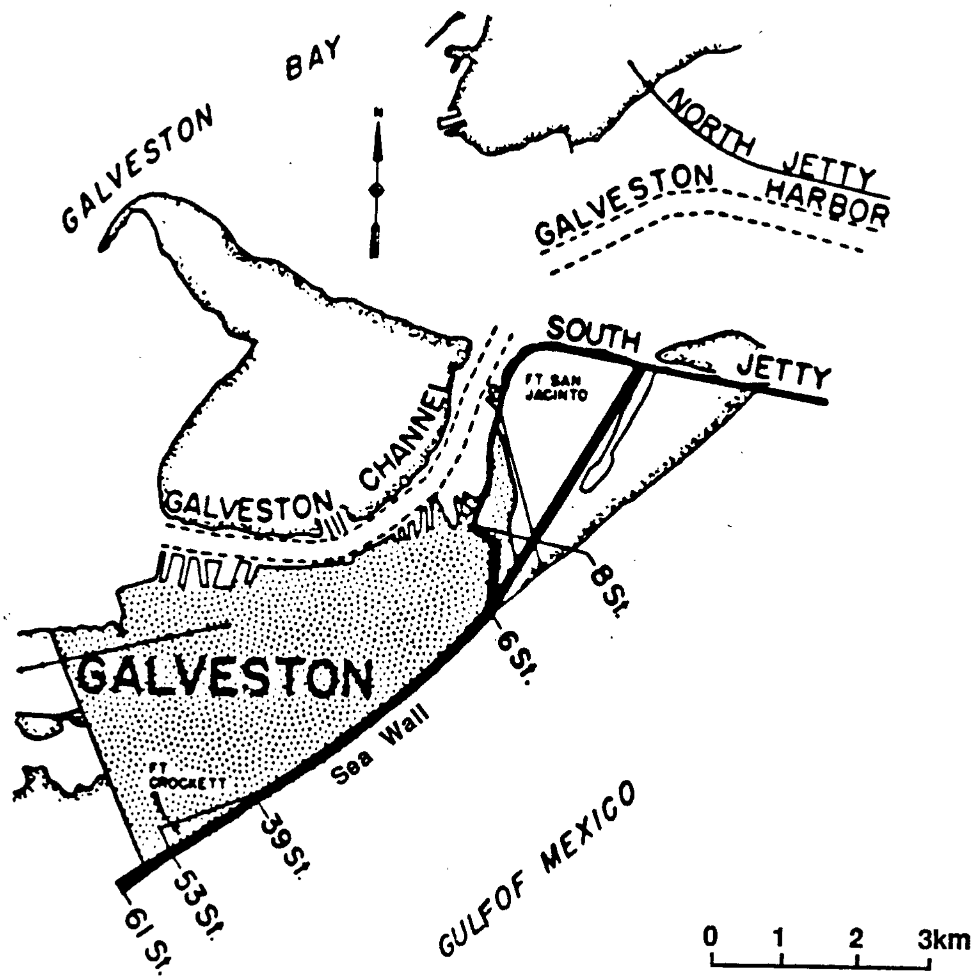


Fig. 6.8. Planview of the Galveston Seawall (after Davis, 1952).

was raised in elevation. The seawall has been subjected to seven major storms since 1915, during which overtopping and toe scour have required additional fill and rubble toe protection. Subsidence of the wall has also been a problem, especially in places where it is located over a soft clay stratum. Relative sea level at Galveston has risen approximately 24 cm since 1904 (Leatherman, 1984), and during that time most of the original beach fronting the wall (up to 90 m wide) has been lost. Leatherman also indicates that diking will be necessary in the future to preserve the city.

Breakwaters - are free-standing structures, usually of rubble mound construction, attached to the shoreline or seaward (detached) of the shoreline. Breakwaters cannot prevent inundation by sea level rise, but can modify shoreline response by blocking some of the incident wave energy. The resulting shoreline (for detached breakwaters) has a bulge associated with each structure, and holds the mean shoreline at a more seaward position. Effective in preventing beach erosion due to both longshore and on/offshore transport, offshore breakwaters have been used for shore protection in the U.S., Canada, Europe, and quite extensively (over 2,500) in Japan. Although initial construction costs can be high, proper design usually ensures low maintenance. The shoreline response and functional design of offshore breakwaters is extensively discussed in Dally and Pope (1986).

As sea level rises, an existing breakwater project will lose sediment from its salient(s) as its relative position moves offshore and overtopping becomes more frequent. In order to maintain shoreline position and a prescribed level of protection, the structure will need to be lengthened and its crest elevated. Otherwise, projected sea level must be used in both structural and functional design, with the margin of safety diminishing as sea level rises during the life of the project.

An example of a segmented breakwater project installed to provide shoreline protection and a recreational beach is found at Presque Isle, Pennsylvania. The project, shown in planform in Fig. 6.9, consists of three segments, each 38 m long and placed 46 m offshore of a beach fill. There is a substantial longshore drift (from left to right) from which the structures have entrapped additional sand to form a series of salients which progressively diminish in size in the drift direction. These salients erode during storms and accrete in calm weather, but the placed fill has remained relatively unscathed.

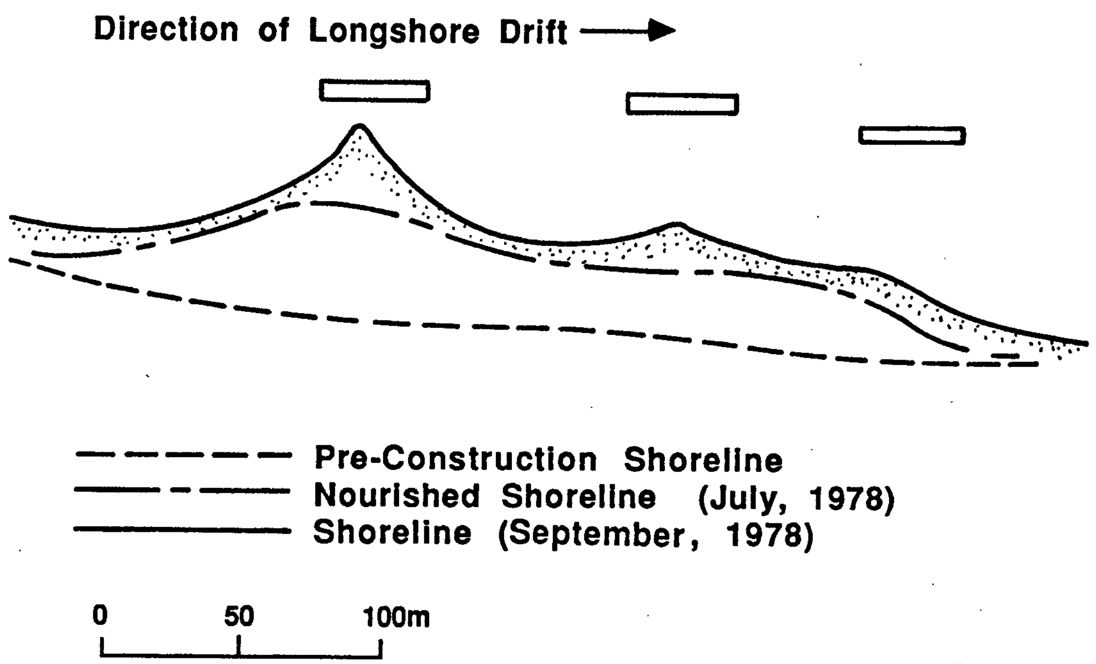


Fig. 6.9. Breakwater Project and Shoreline Response at Presque Isle, Pennsylvania.

Groins - are shore-perpendicular structures made of timber, steel or concrete sheet pile, or rubble, whose purpose is to entrap sediment moving alongshore. The shoreline accretes on the updrift side and erodes on the downdrift until sand is able to pass around the end of the structure and restore the longshore drift. If the fillet is placed artifically during groin construction, much of the downdrift erosion can be prevented. For long stretches of beach, a groin "field" of many structures is used, examples of which are found at Rehoboth Beach, Delaware; West Hampton Beach, Long Island, New York; and Madeira Beach, Florida. It is stressed that groins are only useful if local erosion is due to spatial variation in the longshore drift, and have little positive impact if erosion is due to on/offshore sediment transport. Consequently, the use of groins will do little to modify shoreline response to sea level rise.

Sea level rise will generally result in a loss of efficacy of existing groin projects. Increased water level will allow more overtopping by waves, and eventually flanking could occur at the landward end of the structure as the shoreline retreats. Groins with long useful lives may require lengthening and raising, while those with shorter lifespans should be replaced with redesigned structures.

Fig. 6.10 shows the groin field at Long Branch, New Jersey, where the longshore drift is from south to north. This project has succeeded in trapping sand and building a beach, but apparently at the expense of the North Long Branch shoreline.

Jetties - are shore-perpendicular structures, usually of rubble mound construction, placed at tidal inlets in order to stabilize their position and maintain a navigable channel. Shoreline response to construction of jetties is similar to that of groins, but on a larger scale as jetties are usually very long. A large fillet is formed on the updrift side of the inlet, with the downdrift shoreline often subject to severe erosion. Jetties serve to increase the velocities in a tidal inlet, which deepens the cross-section and pushes the ebb shoals offshore, entrapping even larger amounts of sediment. Without mechanical bypassing of sand from the updrift to downdrift side of an inlet, the downdrift beach will erode until the updrift fillet and ebb shoals are large enough to shunt sand across the inlet.

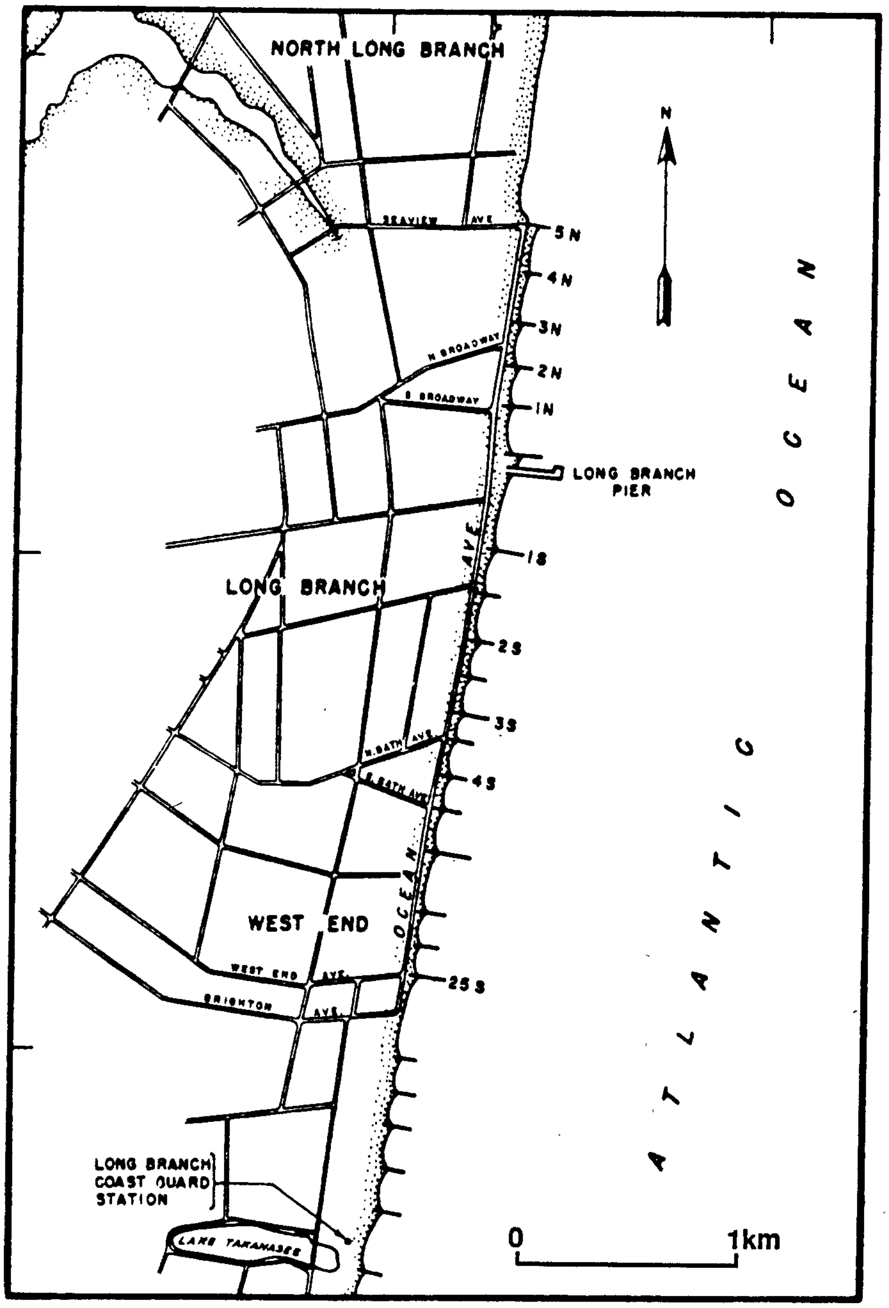


Fig. 6.10. Groin Field at Long Branch, New Jersey (after Army Corps of Engineers, 1964).

Jetties may exacerbate shoreline retreat as sea level rises in the following manner. A rise in mean water level increases the depth of a bay and increases the hydraulic efficiency of an inlet, thereby increasing the tidal prism. An increase in tidal prism increases the velocities through the inlet which in turn pushes the ebb shoal offshore into deeper water. The shoal then demands more sediment from the neighboring shorelines until the regional longshore transport rate is restored (if possible). As noted previously, long-term sea level rise will also promote creation of additional inlets, which if stabilized will each demand sediment to maintain ebb shoals.

As with groins, sea level rise will tend to reduce the efficacy of jetties due to overtopping and possible breaching at the shoreward end. Because jetties usually have long lifespans, they may require lengthening and raising of the crest of the structure.

A typical example of the shoreline response to jetties is the inlet at Ocean City, Maryland shown in Fig. 6.11. A hurricane cut the inlet in August, 1933 and jetties were constructed shortly thereafter. By 1976 the updrift shoreline had advanced 245 m while the downdrift had retreated 335 m. The shoreward end of the south jetty has had to be rebuilt and extended several times. It is doubtful that in the 50 years since "stabilization", the shorelines have regained a state of dynamic equilibrium, so the effects of 50 years of sea level rise cannot be deduced accurately.

Beach nourishment - is the mechanical placement of sand on a beach to advance the shoreline. It is a "soft" protective and remedial measure that leaves a beach in a more natural state than hard structures, and preserves its recreational value. Beach fills cannot "modify" shoreline response to sea level rise because the natural littoral processes remain unaltered, and thus fills can only be regarded as a temporary measure. Although requiring maintenance at regular intervals and after severe storms, beach fills have been successful in many instances such as Miami Beach, Florida; Virginia Beach, Virginia; and Wrightsville Beach, North Carolina.

The greatest effect of long-term sea level rise on beach fill design is to increase the volumetric requirements of the fill and so increase costs. Attempting to hold the shoreline in one location will necessarily require a steeper beach profile as sea level rises. This means increased volumes of placed sand are necessary to satisfy the offshore transport demand, or else placing material of coarser grain size than the native sediment.

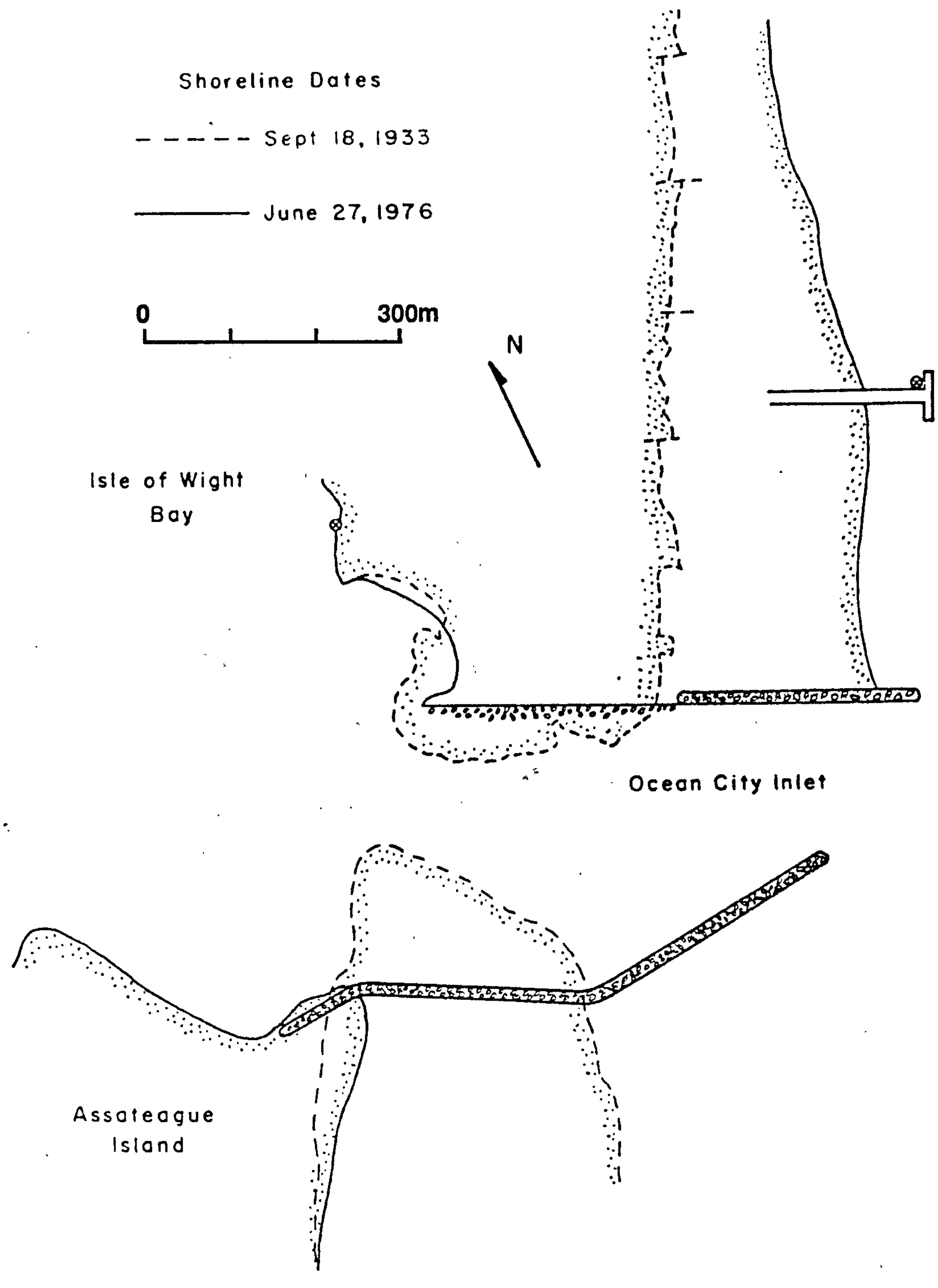


Fig. 6.11. Shoreline Response to Jetty Construction at Ocean City, Maryland (after Dean *et al.*, 1979).

An example of a successful beach nourishment project is that at Harrison County, Mississippi, shown in Fig. 6.12. Constructed in 1951-1952 of 4.6 million m^3 of fill, the project provided 280 hectares of new beach which was 90 m wide with a berm height of 1.5 m, and fronted the seawall constructed in 1925-1928. The project has performed well, with annual losses on the order of 76,500 m^3, and has provided upland protection during several major hurricanes. Several islands provide some shelter to the project from the Gulf of Mexico and may be partially responsible for its longevity. It was renourished with 1.5 million m^3 of fill in 1972-1973, following the effects of hurricane Camille (1969), which caused storm tides locally in excess of 6 m. Relative sea level is estimated to have risen only 8 cm during the life of the project (Hicks _et al._, 1983), forestalling conclusions of the fill's stability in response to sea level rise.

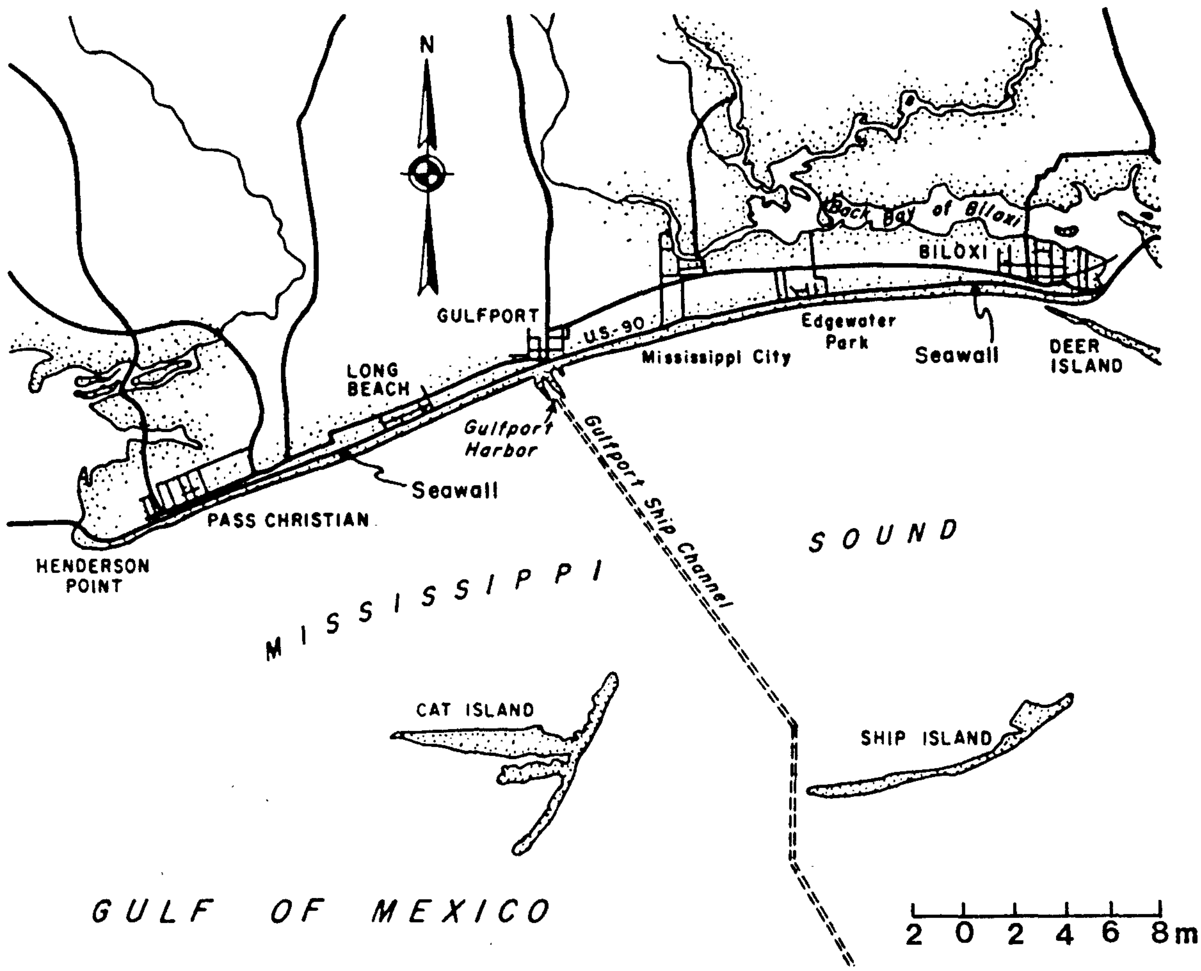
6.4 COST OF COASTAL WORKS

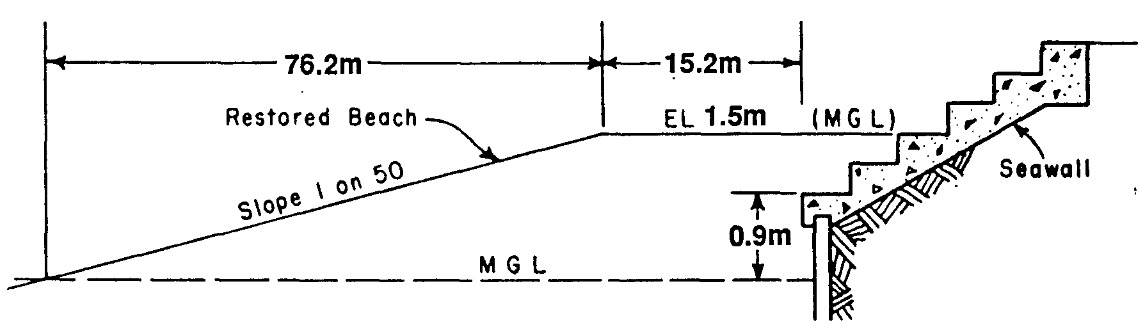
Although the effect of a rise in relative sea level on the cost of a coastal structure or beach nourishment project can only be accurately determined on a case-by-case basis, several crude indicators are available. For rubble mound structures, the cost increases with the required individual weight of the armor stone. Using the well known Hudson formula, found in the Shore Protection Manual (Army Corps of Engineers, 1984), the weight (W) increases with the cube of wave height. From section 5, expression 5.10 for the generated wave height and 5.18 for the height after bottom friction can be used to determine the relative increase in stone weight. For the example presented (sea level rise of 1 m, wind speed of 30 m/s, fetch length of 50 km and shelf depth of 10 m), the ratio of weights is

$$\frac{\text{W (after s.l.r.)}}{\text{W (before s.l.r.)}} = \frac{(0.96)^3}{(0.82)^3} = 1.60 \qquad (6.1)$$

or a 60% increase in stone size. We see that sea level rise may have a significant impact on the design and cost of rubble mound structures.

For beach nourishment projects, the increase in the rate of losses can be examined by assuming the transport rate (Q_s) to be proportional to wave height to the 2.5 power (Dean, 1976). For the same example of section 5 reiterated above, this means that





Profile of Beach

Fig. 6.12. Beach Nourishment Project at Harrison County, Mississippi (after Army Corps of Engineers, 1984).

$$\frac{Q_s(\text{after})}{Q_s(\text{before})} = \frac{(0.96)^{2.5}}{(0.82)^{2.5}} = 1.48 \qquad (6.2)$$

or a 50% increase in the rate of losses from a beach fill.

Using two different methods, the approximate cost of maintaining the existing shoreline of Florida with beach nourishment was examined in the NRC report "Understanding and Responding to Sea Level Rise" (National Research Council, 1987). For the three different scenarios of sea level rise examined, the annual costs range from \$33 to \$204 per linear meter of shoreline, or between 0.1 and 3.4% of the present day value of beachfront property in Florida. The amount varied by a factor of 2.5 between the two methods - highlighting the need for research in this area.

6.5 RESEARCH NEEDS

Research needs in the area of modifying shoreline response to sea level rise and the effects of sea level rise on the design of protective works lie in the realm of ongoing basic studies of natural shoreline processes, and have little requirement for specific treatment of sea level rise. If engineers had a surf zone sediment transport model capable of reproducing and predicting beach response to storms and structures, including the effects of sea level rise would mean an almost trivial matter of increasing the mean water depth in the model. However, until the knowledge of basic processes has grown considerably and such models are developed, there is little reason to expect accurate prediction of the response of beaches to sea level rise to be possible, and that cost-effective techniques for modifying the response will be available.

There are four major areas requiring research in basic physical processes: 1) wave refraction/diffraction, 2) wave breaking, 3) undertow and longshore currents (nearshore circulation) and 4) sediment entrainment under shoaling and breaking waves. The knowledge gained from research in these areas would then be used as input to beach profile and planform response models.

Once a reasonable expertise in shoreline modeling has been reached, the greatest research need is for the engineering community to analyze and quantify the performance and costs of the available alternatives for dealing with sea level rise, and to then determine their cost-effectiveness. Studies should be implemented that are specifically devoted to dikes and artificial dunes, offshore breakwaters, and beach fill design. These measures appear to be the most promising for confronting sea level rise.

7. SHORELINE RESPONSE MODELING

7.1 INTRODUCTION

A potential dominant effect of relative sea level rise is shoreline erosion. An erosional trend on a developed coastline always requires a decision to: 1) retreat, 2) stabilize through coastal structures, or 3) stabilize through nourishment. Each of the above can be costly; accepting that under a given scenario of relative shoreline stability, sea level rise, etc. there is an "optimal" choice, it follows that an inappropriate choice could be inordinantly expensive. Given that eustatic sea level rise affects shorelines on a global basis, that the human rate of shoreline development is increasing and that some projections of future sea level rise are much greater than in the past, it becomes important to attempt to predict the shoreline response to such a rise.

Shoreline response to sea level change depends not only on the rate of change, but also on antecedent conditions and the degree and type of disequilibrium of the shoreface. The dominant engineering approach to predicting shoreline response is the so-called "Bruun Rule" which considers only cross-shore conditions and an offshore "closure depth" seaward of which there is no sediment exchange. The Bruun Rule yields a simple relationship resulting in horizontal shoreline retreat of approximately 50-100 times the rise of sea level. This chapter presents a more complete consideration of the sediment budget on the shoreface and attempts to remove some of the limitations of the Bruun Rule. Specific cross-shore components not included by Bruun but which could be of significance are: 1) shoreward transport of sediment across the shoreface, 2) deposition of suspended sediment, and 3) biogenetic production of sediment. An important factor relating to shoreward sediment transport is the history of sea level change over the past ~ 20,000 years, with the last 6,000 years or so representing a relative still stand.

7.2 LITERATURE REVIEW

Prior to discussing the models for shoreline response, it is instructive to review estimates of sea level rise over the last 20,000 years or so, shown in Fig. 7.1. Sea level rose rapidly (about 0.8 m/century) from 20,000 years before present (BP) to about 6,000 years BP. Over the last 6,000 years, sea

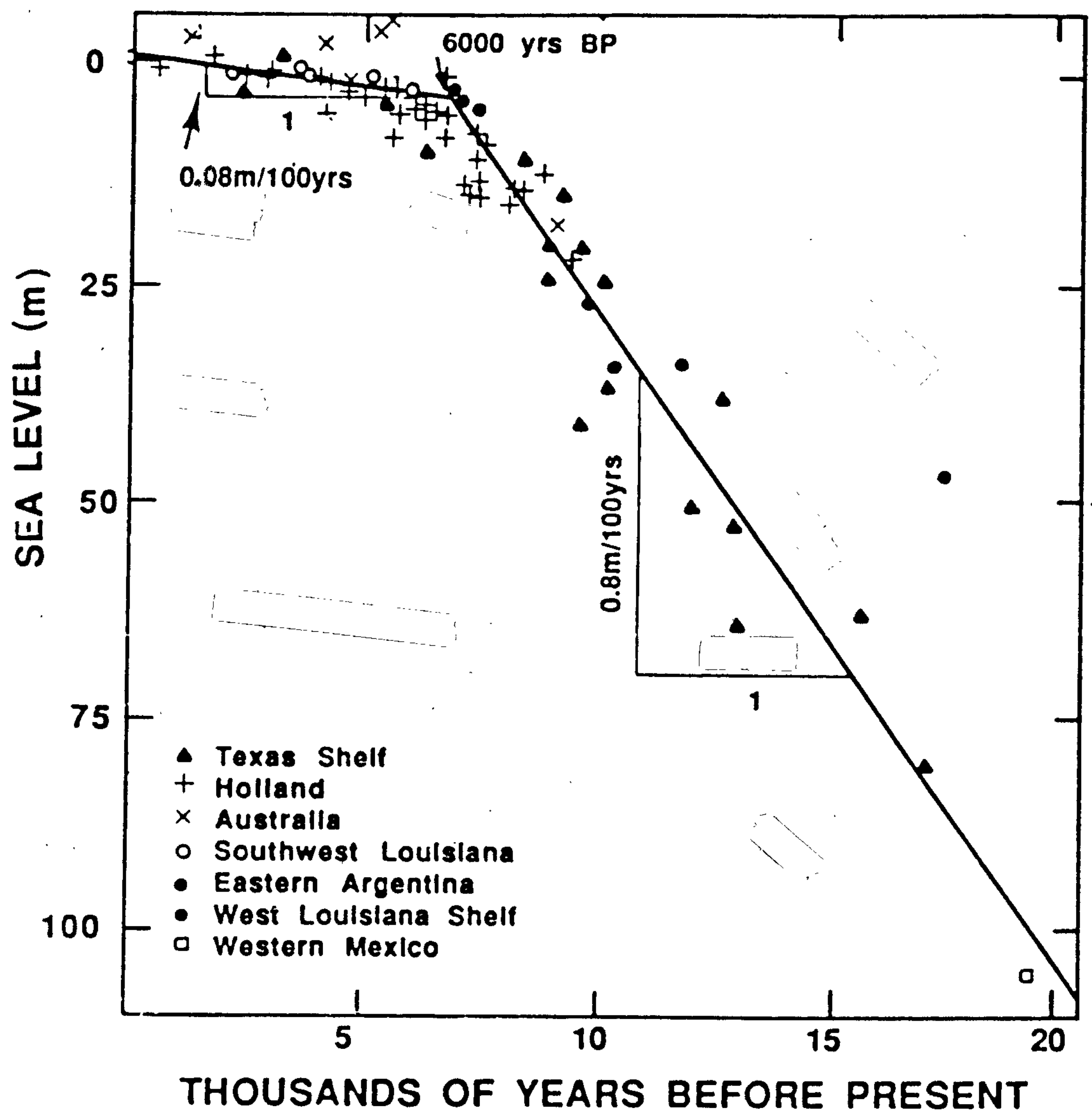


Fig. 7.1. The Rise of Sea Level as Obtained from Carbon 14 Dates in Relatively Stable Areas (after Shepard, 1963). Break in Slope some 6000 Years BP may have Provided Basis for Barrier Island Stability.

level has risen at the greatly reduced rate of 0.08 m/century, which is roughly consistent with estimates of 0.11 m/century based on tide gage data over the last century. As will be discussed later, the earlier much more rapid rise of sea level may still be having an effect.

The most widely applied engineering approach to predicting shoreline response to sea level rise is the so-called Bruun Rule. This rule considers: a) the active profile to always be in equilibrium, and to retain its relative position to sea level, and b) the active portion of the profile to be limited by the "depth of effective motion" seaward of which no sediment exchange occurs. With the above assumptions, when sea level rises a vertical distance, S, the entire active profile must rise also by S, requiring a volume $\forall_R$, of sand per unit beach length

$$\Delta\forall_- = SL \tag{7.1}$$

in which L is the offshore length of active profile. This required sand is provided by a profile retreat, R, over a vertical distance, h_*+B, (see Fig. 7.2). The volume generated by this retreat is

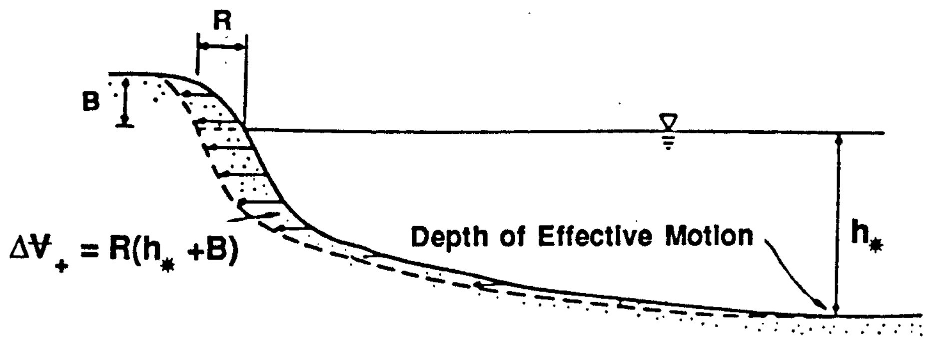
$$\Delta\forall_+ = (h_*+B)R \tag{7.2}$$

and equating the two volumes, the retreat R can be shown to be

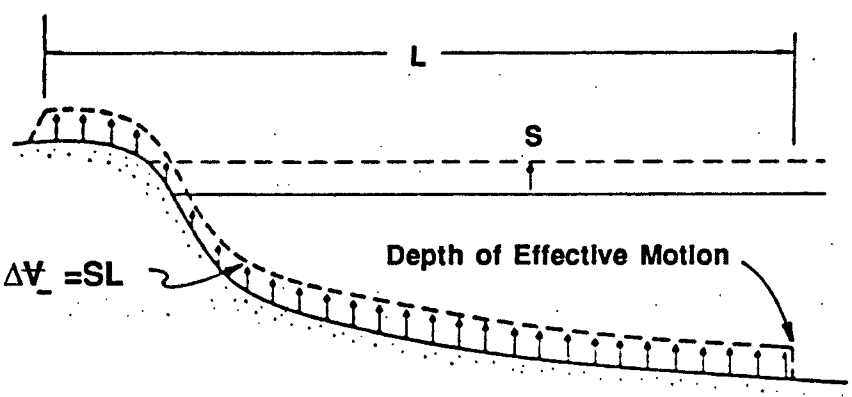
$$R = S\frac{L}{(h_*+B)} = \frac{S}{\tan\theta} \tag{7.3}$$

in which θ is the average slope of the active profile out to its limit of active motion, Fig. 7.3. From Eq. 7.3, it is clear that beach profiles with mild slopes would experience greater recessions due to a given sea level rise than would steeply sloping profiles.

Several laboratory and field studies have been carried out to evaluate the Bruun Rule, usually with confirmation claimed. Schwartz (1965) conducted small-scale laboratory model studies to determine whether an increase in water level caused an offshore deposition equal to the rise in water level as predicted by the Bruun Rule. The wave basin was quite small using medium sized sand of 0.2 mm. Following the development of an equilibrium profile,



a) Volume of Sand "Generated" by Horizontal Retreat, R, of Equilibrium Profile Over Vertical Distance (h_{*} +B)



b) Volume of Sand Required to Maintain an Equilibrium Profile of Active Width, L, Due to a Rise, S, in Mean Water Level

Fig. 7.2. Components of Sand Volume Balance Due to Sea Level Rise and Associated Profile Retreat According to Bruun Rule.

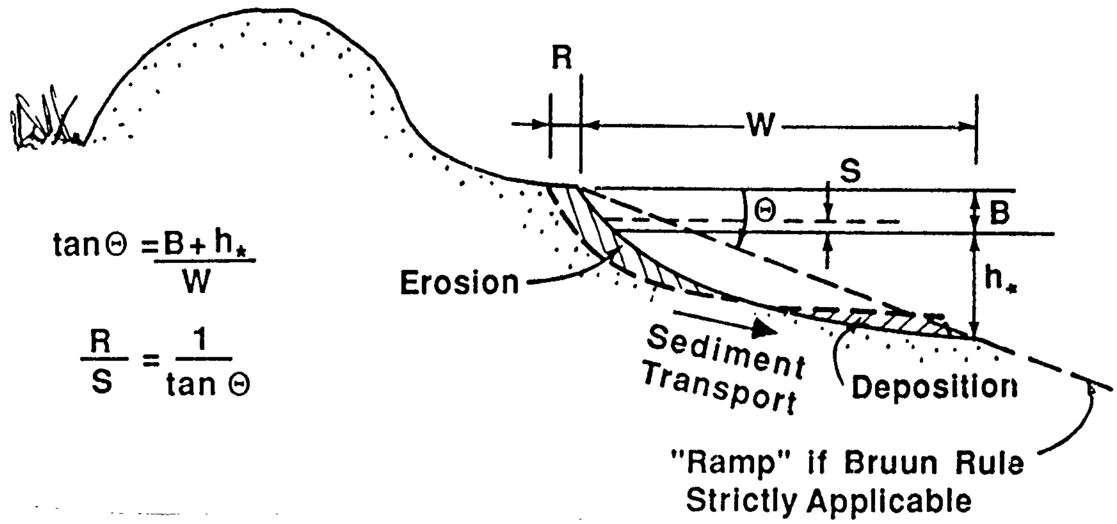


Fig. 7.3. The Bruun Rule with Only Seaward Transport of Sediment and Trailing Ramp Seaward of Active Profile.

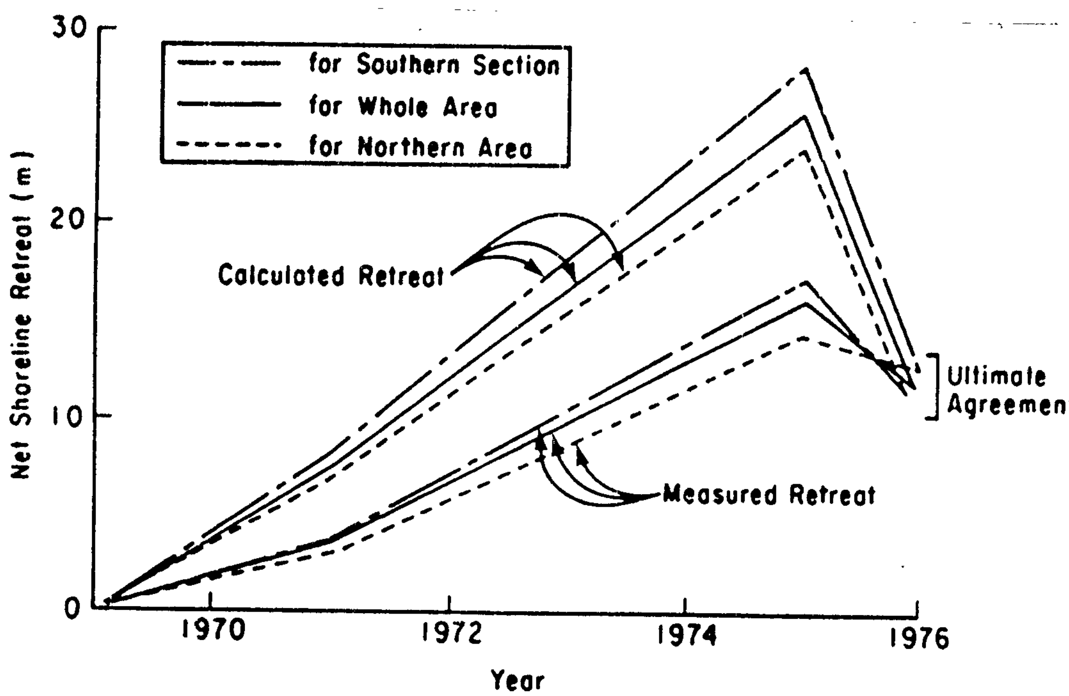


Fig. 7.4. Comparison of Predicted and Measured Shoreline Changes Due to Water Level Increases, Eastern Shore of Lake Michigan (after Hands, 1983).

the water level was increased by 1.0 cm and the test program resumed with the same wave conditions. Following profile equilibration, it was found that the offshore profile had increased in elevation by 0.9 cm which Schwartz considered as confirmation of the Bruun Rule. Schwartz (1967) also conducted a second series of tests with slightly larger facilities, but following the same general experimental framework. Again good agreement with the Bruun Rule was reported. Field measurements were also carried out by Schwartz at Cape Cod, MA in which the shoreline response between spring and neap tides was evaluated in terms of the Bruun Rule. Although "a recognizable upward and landward translation of the profile was noted in the interval between neap and spring tides" was reported and the results were generally regarded as confirmatory, examination of the results is not convincing as to their significance to and agreement with the Bruun Rule. Also, it is not clear that spring tides, which of course have water levels both higher and lower than the average, should be equated to a sea level rise since the average water level is unchanged. Moreover, it is not clear that the Bruun Rule was meant to apply on such a short-term basis especially recognizing that short-term changes in wave climate and convergences of longshore sediment transport can play an important role in beach profile changes.

Dubois (1975, 1976, 1977) has reported on shoreline changes in Lake Michigan in association with a 30 cm rise over a 35 week period. The shoreline recession of 7 m was regarded as substantiation of the Bruun Rule.

Rosen (1978) has evaluated the Bruun Rule on the Virginia shoreline of Chesapeake Bay. Using 14 beach profiles, Rosen found that the errors in predicted erosion rates on the eastern and western shores were +58% and -7% with the positive percentages indicating that the predicted erosion exceeds the measured. As expected, considering smaller groups of profiles, the errors were larger.

Hands (1983) has evaluated the Bruun Rule employing a series of 25 profiles along 50 km of the Lake Michigan eastern shore over a 7 year period. During this period, the water level rose by 0.51 m and then fell by 0.31 m. Fig. 7.4 from Hands shows that the shoreline responded to the changes in water level, although with a lag. Hands recommends that in the absence of other information the "depth of limiting motion" be taken as twice the significant wave height.

Everts (1985) presented a sediment budget approach which encompassed and extended beyond the Bruun Rule. The method was applied to Smith Island, VA and a 75 km segment of the Outer Banks of North Carolina to determine the portion of the shoreline retreat explainable by sea level rise. It was found that 55% and 88% of the measured shoreline retreat was attributable to sea level rise at Smith Island and the Outer Banks, respectively. The remaining component was interpreted to be due to gradients in longshore sediment transport. The sediment budget approach applied by Everts recognizes the limitations of the Bruun Rule and the need to consider a more complete framework for representing and interpreting shoreline response to sea level rise.

Dean and Maurmeyer (1983) have generalized the Bruun Rule to barrier island systems that retreat as a unit filling in on the bay side to maintain their width as they erode on the ocean side. Employing the notation of Fig. 7.5, the shoreline recession, R, due to a sea level rise, S, is

$$R = S \frac{(L_L + W + L_o)}{h_{b_o} - h_{b_L}} \tag{7.4}$$

It is clear from Eq. 7.4 that the recession will always predict a greater erosion than the Bruun Rule because: a) a greater horizontal dimension is being elevated with sea level rise (the entire active barrier island width), and b) the portion of the profile now being "mined" to yield compatible sediment is the difference between ocean and bay depths, $h_{b_o} - h_{b_L}$, i.e. smaller. This equation simplifies to the Bruun Rule if only the ocean side of the barrier system is active. Finally it is noted that as the bay depth h_{b_L} approaches the active ocean depth, h_{b_o}, Eq. 7.4 predicts an infinite retreat rate. This may explain in part the phenomenon of "overstepping" in which barrier islands, rather than migrating landward retaining their identity in the process, are overwashed and left in place as a linear shoal, see, e.g. Sanders and Kumar (1975).

It is noted that Eqs. 7.3 and 7.4 both consider the portion of the profile being "mined" for sand as containing 100% compatible material. If a portion of the profile contains peat or fine fraction that will not remain in the active system, a rather straightforward modification of the equations is required.

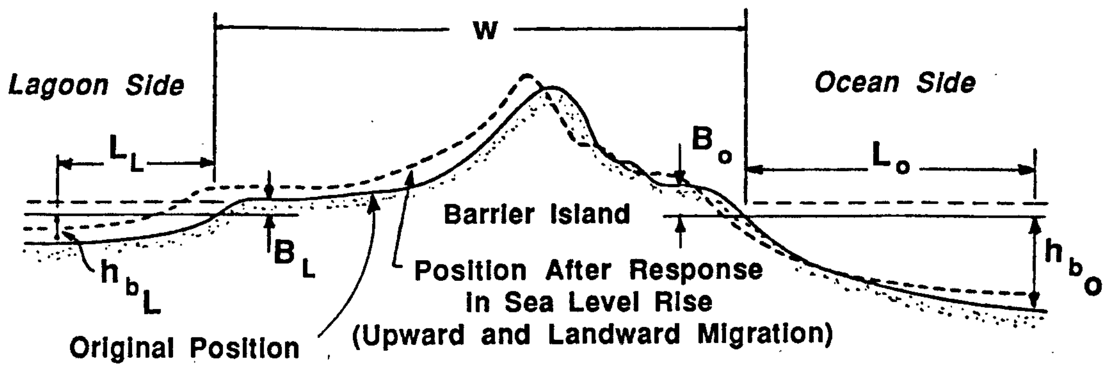


Fig. 7.5. Generalized Shoreline Response Model Due to Sea Level Rise. Applicable for a Barrier Island System which Maintains its Form Relative to the Adjacent Ocean and Lagoon (after Dean and Maurmeyer, 1983).

Kriebel and Dean (1985) have described a dynamic cross-shore transport model in which the input includes the time-varying water level and wave height. In addition to predicting long-term responses, this model accounts for profile response to very short-term events such as hurricanes. Thus an equilibrium profile is not assumed and, in addition to a sand budget "volumetric equation", a "dynamic equation", is required which was hypothesized as

$$Q_s = K(\mathcal{D} - \mathcal{D}_*) \tag{7.5}$$

in which Q_s represents the offshore sediment transport per unit length of beach, K is a universal constant ($K = 2.2 \times 10^{-6} m^4/N$ in the metric system) and $\mathcal{D}$ and $\mathcal{D}_*$ represent the actual and equilibrium wave energy dissipation per unit water volume. Eq. 7.5 is suggested following the determination by Bruun (1954) and later by Dean (1977) that most equilibrium beach profiles are of the form

$$h = Ax^{2/3} \tag{7.6}$$

in which A is a dimensional profile constant depending primarily on sediment size but secondarily on wave climate. Dean (1977) found that Eq. 7.6 is consistent with uniform wave energy dissipation per unit volume. The quantities A and $\mathcal{D}_*$ are related by

$$A = \left[\frac{24}{5} \frac{\mathcal{D}_*}{\rho g \sqrt{g}\ \kappa^2}\right]^{2/3} \tag{7.7}$$

in which ρ is the mass density of water, g is the gravitional constant and κ is the ratio of spilling breaking wave height to water depth ($\kappa \approx 0.8$).

All models of beach profile response described earlier require the identification of a limiting depth of motion h_* in Eq. 7.3 and h_{b_o} and h_{b_L} in Eq. 7.4. Hallermeier (1981) has proposed an approximate method for predicting this depth, h_*, based on average annual significant deep water wave height, $\overline{H}_s$, and period $\overline{T}_s$ and sediment size D,

$$h_* = (\overline{H}_s - 0.3\sigma)\ \overline{T}_s\ (g/5000D)^{0.5} \tag{7.8}$$

in which σ is the standard deviation of the significant wave height.

The models presented heretofore invoke the concept of a limiting depth of motion, a depth seaward of which conditions are static or at least there is no substantial exchange of sediment with the more active shoreface. This assumption seems innocent and quite natural, yet the consequences are very substantial. If no interchange with the shelf profile occurs, erosion is the only possible shoreline response to sea level rise, i.e. there can be no shoreward transport contributions from the continental shelf. There is evidence that shoreward sediment transport is a major contributor to shoreline stability in many areas. The erosion along the south shore of Long Island and at Montauk Point is clearly too small to provide the well-documented westward net transport at Fire Island (Dean, 1986; Williams and Meisburger, 1987).

Dean (1987) has suggested that during the more rapid rate of sea level rise up to 6,000 year BP, the shoreward shelf transport was not sufficient to maintain a stable shoreline. However, with the relative sudden sea level rise reduction by an order of magnitude, the same rates of shoreward sediment transport generally led to reduced erosion rates and in some cases to stable or accreting shorelines; Fig. 7.6 illustrates the concept. The equilibrium mechanics associated with this concept are much different than those employed by Bruun. Recognized are the natural variability of waves and sediment sizes with sorting resulting in coarser sediment close to shore. It is hypothesized that a particle of a given size is in equilibrium when it is in a certain water depth at a particular distance from shore. With sea level rise and

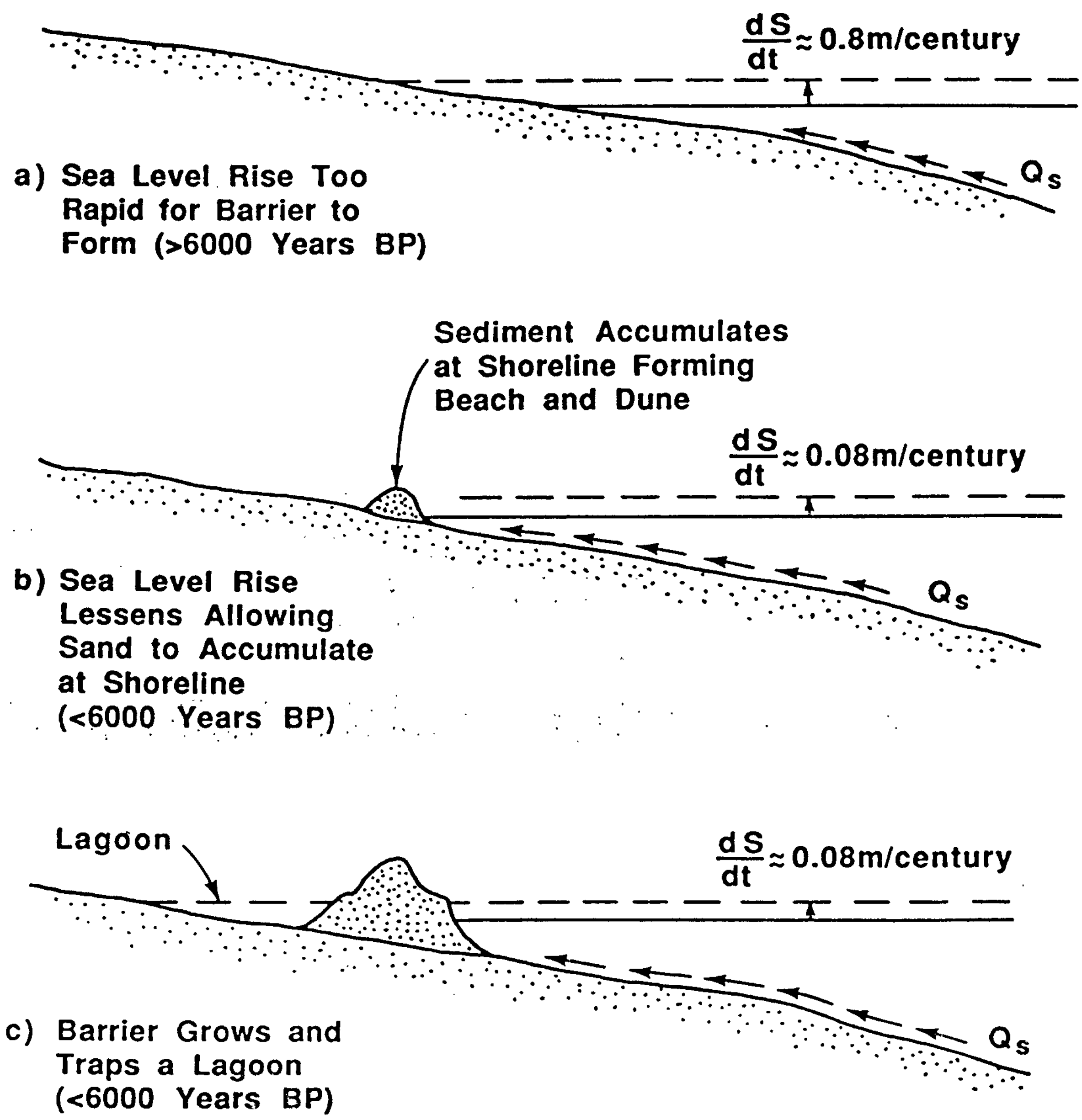


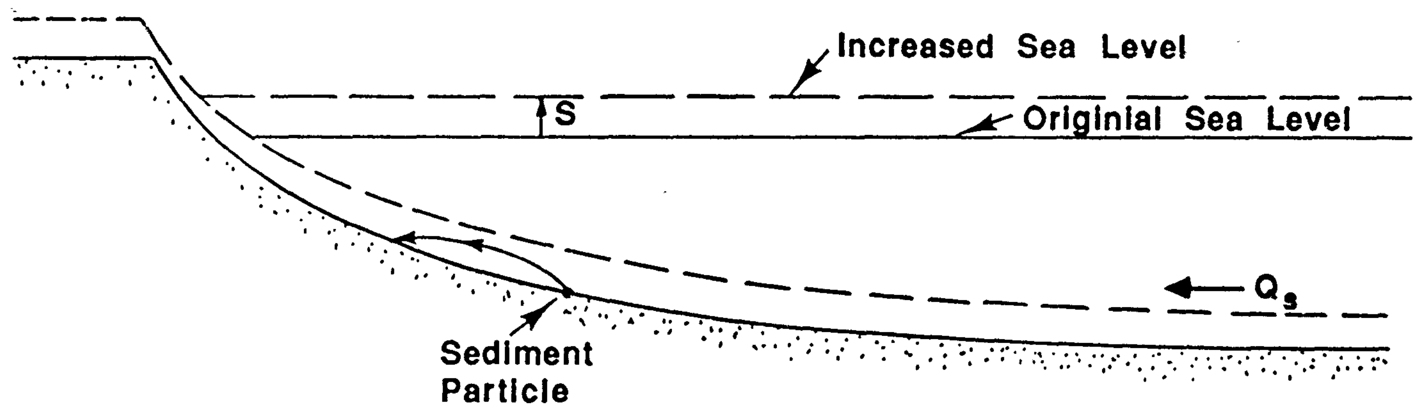
Fig. 7.6. The Role of Shoreward Sediment Transport, Q_s, Across the Shelf and Rate of Sea Level Rise in Causing Barrier Island Formation (after Dean, 1987).

assuming that the wave climate remains the same, the sediment particle would tend to move landward rather than seaward as required by the Bruun Rule. Fig. 7.7 illustrates this mechanism of sedimentary equilibrium.

With greater and greater sea level rise, the general situation will shift toward erosion. Of primary importance is that to predict the response to sea level change, each shoreline segment must be considered on a case-by-case basis with due consideration of the sediment budget. The components of the sediment budget are difficult to quantify. The best basis for developing an appropriate response model for a shoreline segment is an analysis of past response, including a focus on possible anthropogenic effects.

In discussing shoreline response models to sea level change and their development and calibration, it is important to recognize and respect the amount of "noise" in the system including that introduced anthropogenically. Coastal structures and sand management practices at navigational channel entrances are undoubtedly the main contributors to shoreline perturbation by humans. The special attention to documentation following storm activity should also be noted. Along the east coast of Florida, in excess of 38 million cubic meters of beach compatible material has been dredged from channel entrances and disposed at sea. Based on the Bruun Rule, this amount is enough to offset 70 years of shoreline retreat using a eustatic sea level rise of 1.2 mm/year and a retreat/rise multiplier of 100. Data provided by the Jacksonville District of the U.S. Army Corps of Engineers for the period 1980 to 1985 indicates that approximately 50% of the east Florida coast material dredged was still being disposed at sea during this period. This amount (38,000 m^3), again using the Bruun Rule, is sufficient to more than offset their retreat due to the eustatic sea level rise rate employed in the preceding example.

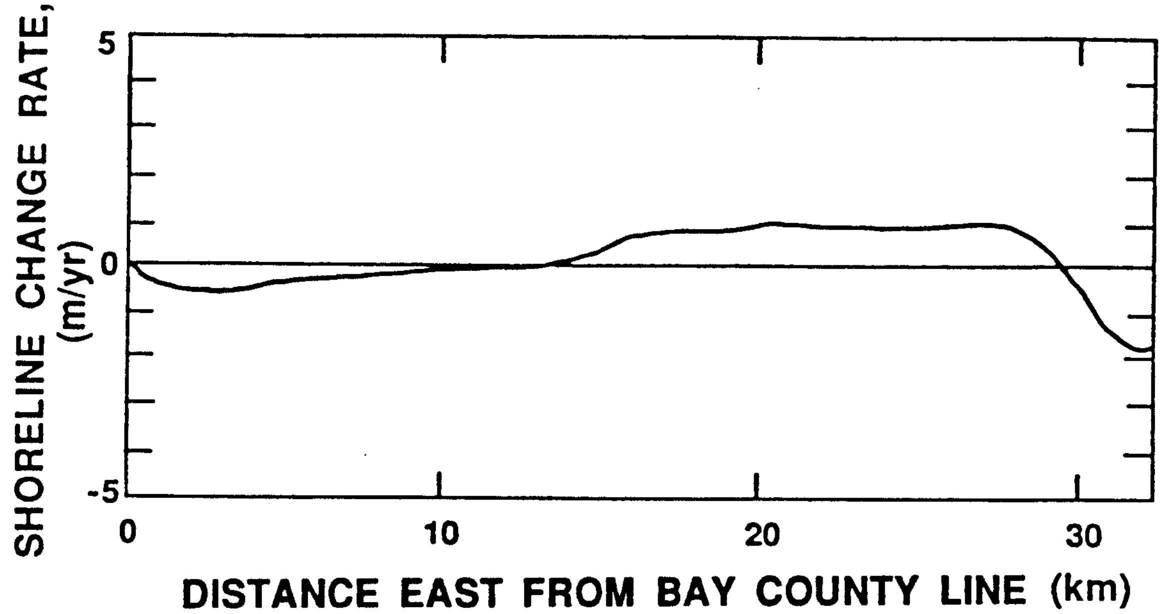
The role of inlets in Florida has been well documented in two cases. The entrance to St. Andrews Bay was cut in 1934 on a previously stable beach. Over the next 50 years, the beach receded at a maximum rate in excess of 2 m/yr where accretion of 1 m/yr had occurred prior to cutting the inlet, Fig. 7.8. The second example illustrates both the adverse effect of cutting the entrance to Port Canaveral in 1951 and the beneficial effects of a beach restoration project carried out in 1974. Again as shown in Fig. 7.9, a



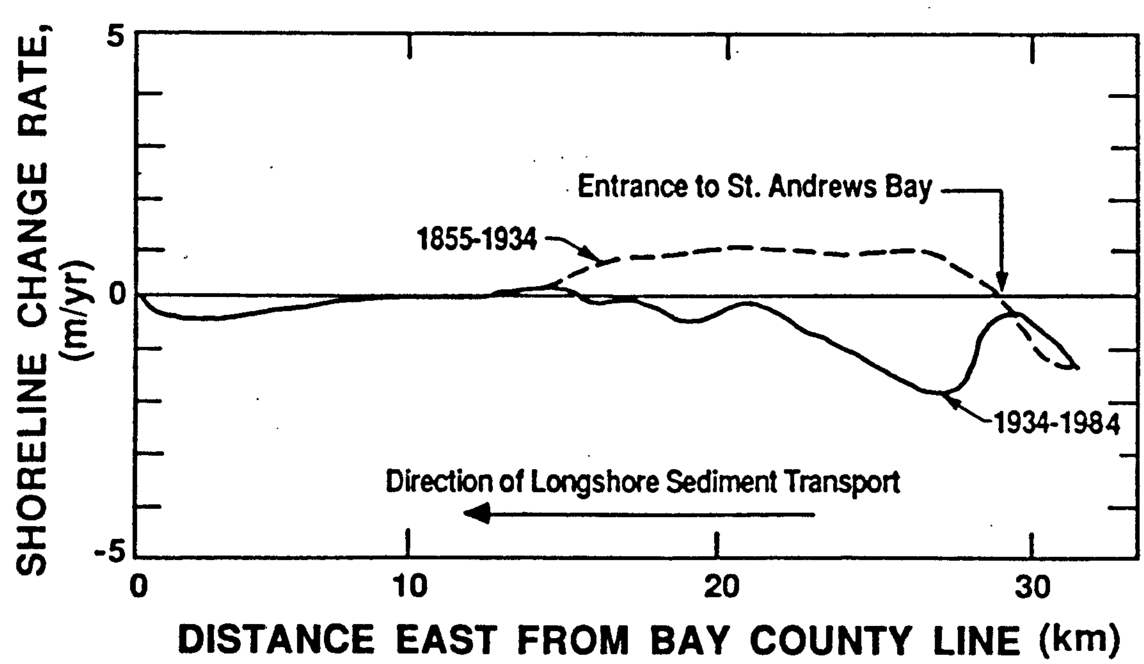
"Subjected to a Given Statistical Wave Climate, A Sediment Particle of a Particular Diameter is in Statistical Equilibrium When in a Given Water Depth"

Thus When Sea Level Increases, Particle Moves Landward

Fig. 7.7. Possible Mechanism of Sedimentary Equilibrium (after Dean, 1987).

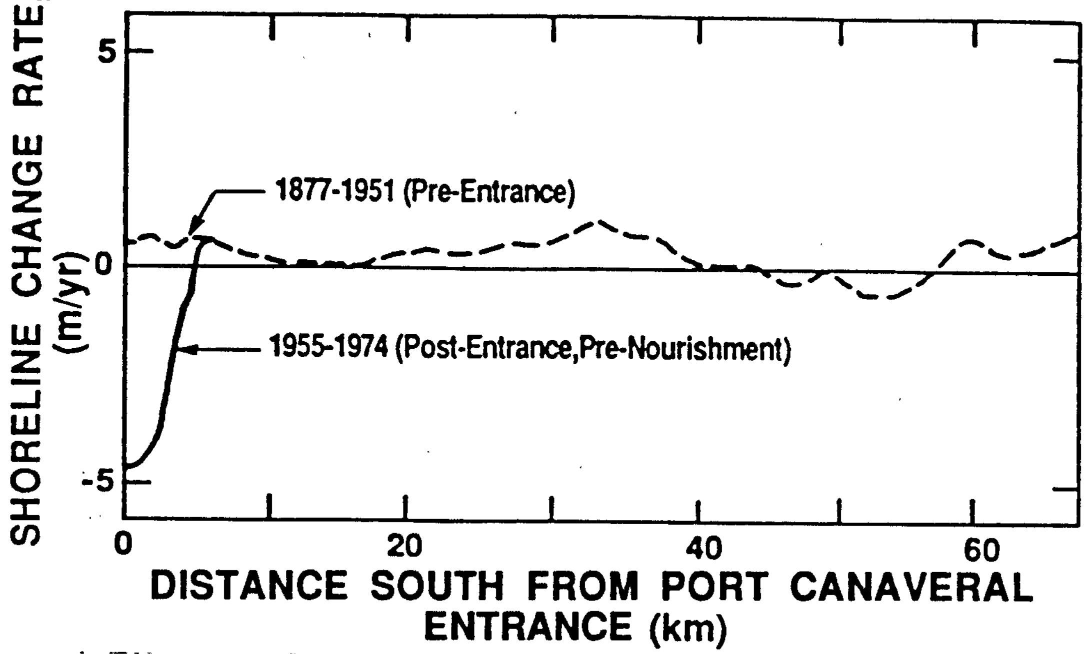


a) Shoreline Change Rates Prior to Cutting Entrance to St. Andrews Bay, 1855-1934 (79 Years).

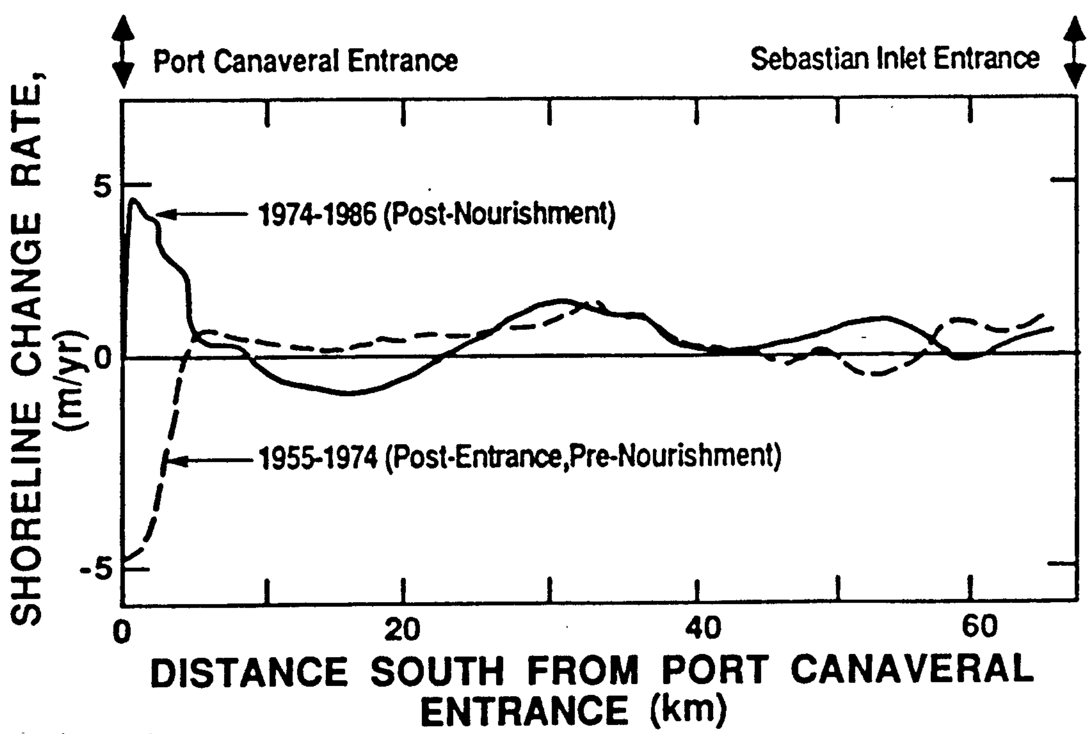


b) Comparison of Shoreline Change Rates Prior to Cutting Entrance to St. Andrews Bay, 1855-1934 (79 Years) and Subsequent to Cutting Entrance, 1934-1984 (50 Years).

Fig. 7.8. Effect of Cutting Entrance to St. Andrews Bay in 1934 on Downdrift Shoreline (after Dean, 1987).



a) Effects of Channel Entrance on Downdrift Beach Stability, Compared to Pre-Entrance Condition.



b) Shoreline Changes Following 1974 Nourishment Project

Fig. 7.9. Effects of Establishment of Cape Canaveral Entrance and Subsequent Nourishment Project on Downdrift Beaches (after Dean, 1987).

beach that had been stable previously underwent dramatic erosion immediately downdrift (south) of the inlet.

Weggel (1986) has examined the economics of beach nourishment under the scenario of a rising sea level. Methods were presented for computing the present worth costs of perpetual renourishment for sea level rising at a uniform rate and projects of limited life (e.g. 50 and 100 years) for increasing sea level rise rates as predicted by Hoffman (1984). The tradeoff between renourishment (repeated costs) and stabilizing structures (initial cost only) was examined and, based on the reduced required frequency of renourishment due to the structures, the justified cost of structures is presented. It is concluded that perpetual beach nourishment is not economically justified under the sea level rise rates predicted by Hoffman.

7.3 PHYSICAL PRINCIPLES

There are two general types of considerations that can be applied to beach profile response. Kinematic considerations relate to sand budget components regardless of the causes of the transports and associated forces. Dynamic considerations relate to the forcing mechanisms. Each of these will be discussed briefly in the paragraphs below.

7.3.1 Kinematic (Sediment Budget) Considerations

The change in absolute elevation, z, of the sediment-water interface at a horizontal point (x,y) can be expressed in terms of the sediment transport components, q_x and q_y and any source terms, $\mathcal{S}$

$$\frac{\partial z}{\partial t} = -\left(\frac{\partial q_x}{\partial x} + \frac{\partial q_y}{\partial y}\right) + \mathcal{S} \tag{7.9}$$

For reference purposes, the x and y coordinates are selected in the offshore and alongshore directions, respectively such that q_y represents the local longshore component of sediment transport. The formal origin of Bruun Rule can be demonstrated by assuming that there are no sources ($\mathcal{S}=0$), that there are no gradients in longshore sediment transport $\partial q_y/\partial y = 0$, and integrating the resulting equation from x_1 to x_2,

$$\int_{x_1}^{x_2} \frac{\partial z}{\partial t}\, dx = - \int_{x_1}^{x_2} \frac{\partial q_x}{\partial x}\, dx = q_{x_1} - q_{x_2} \tag{7.10}$$

and if q_{x_1} and q_{x_2} are sufficiently landward and seaward that no motion occurs, $q_{x_1} = q_{x_2} = 0$, resulting in

$$\int_{x_1}^{x_2} \frac{\partial z}{\partial t}\, dx = 0 \tag{7.11}$$

We now decompose z into two components, one due to the uniform vertical profile displacement, z_1, and the second to the uniform horizontal displacement $z_2(x)$, i.e.

$$z = z_1 + z_2(x) \tag{7.12}$$

and Eq. 7.11 becomes

$$\int_{x_1}^{x_2} \frac{dz}{dt}\, dx = \int_{x_1}^{x_2} \frac{dz_1}{dt}\, dx + \int_{x_1}^{x_2} \frac{dz_2}{\partial t}\, dx = 0 \tag{7.13}$$

Since $z_1 \neq f(x)$ and in fact $\partial z/\partial t = \partial S/\partial t$ over the active profile length, L, and by the chain rule,

$$\frac{\partial z_2}{\partial t} = \frac{\partial z_2}{\partial x} \frac{\partial x}{\partial t} \tag{7.14}$$

and $\partial x/\partial t$ is uniform over depth and equal to $- dR/dt$, Eq. 7.13 becomes

$$\frac{\partial S}{\partial t} L - \frac{\partial R}{\partial t} \int_{z_1(x_1)}^{z(x_2)} \frac{\partial z}{\partial x}\, dx \tag{7.15}$$

and the active depth is h_*+B, we finally obtain

$$\frac{\partial R}{\partial t} = \frac{\partial S}{\partial t} \frac{L}{(h_*+B)} \tag{7.16}$$

which is the Bruun Rule.

We now return to a somewhat more conceptual version of the cross-shore sediment budget equation, expanding the source term. Considering the case of shoreline stability for illustration purposes

$$\frac{dz}{dt} = \frac{dS}{dt} = -\frac{\partial q_x}{\partial x} + SS + B \qquad (7.17)$$

in which the first term on the right hand side is the local convergence of cross-shore bottom sediment transport, the second term, SS, represents deposition of suspended sediment and the last term is the local biogenetic production. The required magnitudes of each of these in order to compensate for a eustatic sea level rise of 1.2 mm/yr alone is presented below.

Convergence of Cross-shore Sediment Transport - Considering a shelf width of 40 km

$$q_{x_2} = 40{,}000 \text{ m}(.0012 \text{ m/yr}) = 480 \text{ m}^3\text{/m/yr} \qquad (7.17a)$$

Suspended Sediment Deposition

$$SS = 0.0012 \text{ m/yr} \qquad (7.17b)$$

Biogenetic Production

$$b = .0012 \text{ m}^3\text{/m}^2\text{/yr} \qquad (7.17c)$$

At present, knowledge is extremely poor concerning the rates of the three components addressed above. However, until improved quantification is available, all should be regarded as potentially substantial contributors to the shoreface sediment budget.

7.3.2 Dynamical Considerations

It is clear that there are net shoreward directed hydrodynamic forces acting on a sediment particle; otherwise an upward sloping beach could not be in equilibrium against the forces of gravity. The near-bottom flow field due to a storm system is complex and consists of net forces and destabilization forces. The dominant destabilization forces are due to the combined effects of gravity and due either to offshore directed flows in water depths greater than breaking or due to wave breaking in the shallower depths. Fig. 7.10 presents the general situation.

The constructive effects include the shoreward predominance in bottom shear stress due to nonlinear waves. The oscillatory bottom water particle velocity, u, associated with a nonlinear wave can be expressed as

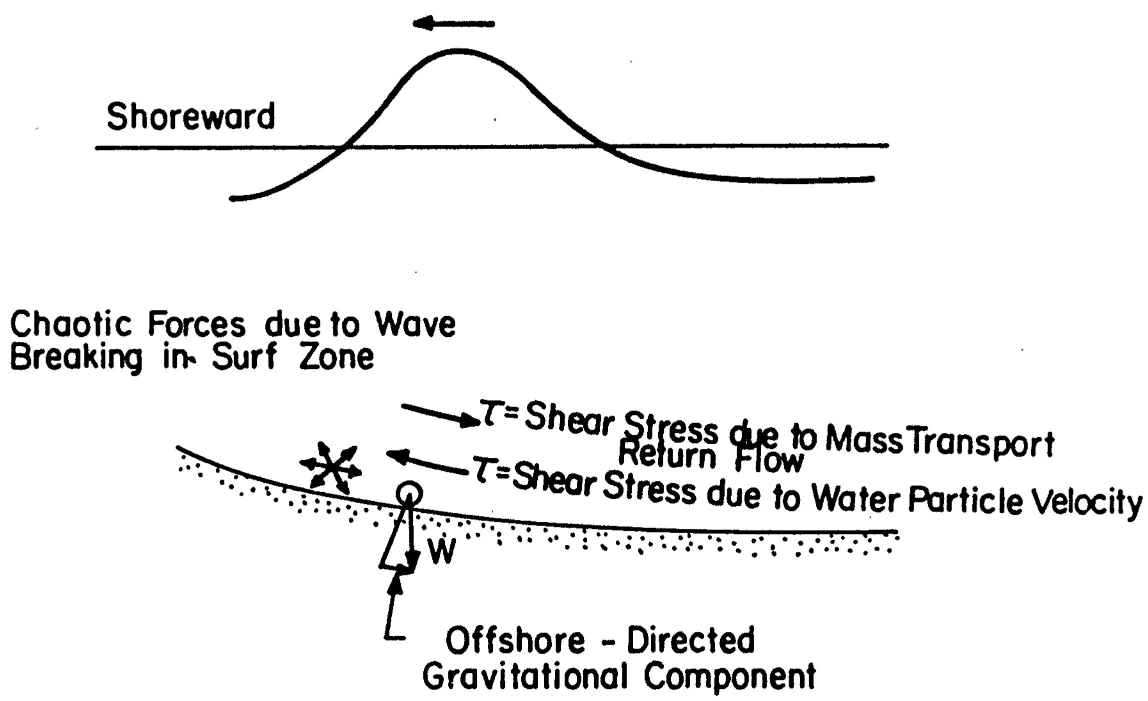


Fig. 7.10. Dominant Forces Acting on a Sediment Particle Resting on the Bottom.

$$u = a_1\cos\alpha + a_2\cos 2\alpha + \ldots \tag{7.18}$$

in which α is the phase angle and the a_n are velocity amplitude coefficients. Even though the time averaged bottom velocity is zero ($\overline{u} \equiv 0$), the net onshore shear stress, $\overline{\tau}$, is positive (i.e. shoreward) since

$$\tau = \frac{\rho f}{8} |u|u \tag{7.19}$$

where ρ is the mass density of water and f is the Darcy-Weisbach fraction factor. Fig. 7.11 is based on stream function (nonlinear) wave theory and presents the average non-dimensional shoreward shear stress, $\overline{\tau'}$,

$$\overline{\tau'} = \frac{\overline{|u|u}}{(H/T^2)} \tag{7.20}$$

as a function of relative water depth, h/L_o and wave steepness, H/L_o.

7.4 RESEARCH NEEDS

It is clear that the development of an adequate capability to predict shoreline response to future sea level rise rates will require a consideration of cross-shore sediment transport fundamentals and applications, and a quantitative understanding of the transport components. The Bruun Rule, while a good first model, is deficient in not allowing for the onshore transport of sand that is clearly occurring at some locations and undoubtedly occurring at many less evident locations. The three types of research needs identified fall in the categories of analysis of existing data, new data, and new technology.

7.4.1 Analysis of Existing Data

Isolation of Anthropogenic Effects - The substantial effects that navigational structures and sand management practices at entrances can have on shoreline stability have been noted and illustrated by the examples in Figs. 7.8 and 7.9. Similar effects are known to occur and be substantial at many other locations, i.e. Folly Beach, SC, Tybee Island, GA, Santa Barbara, CA, and Assateague Island, MD. In addition to the effects at entrances, the effects of groins, seawalls, etc. should be considered. A straightforward methodology could be applied; however, it is believed that development of new, more effective methodology would be worthwhile.

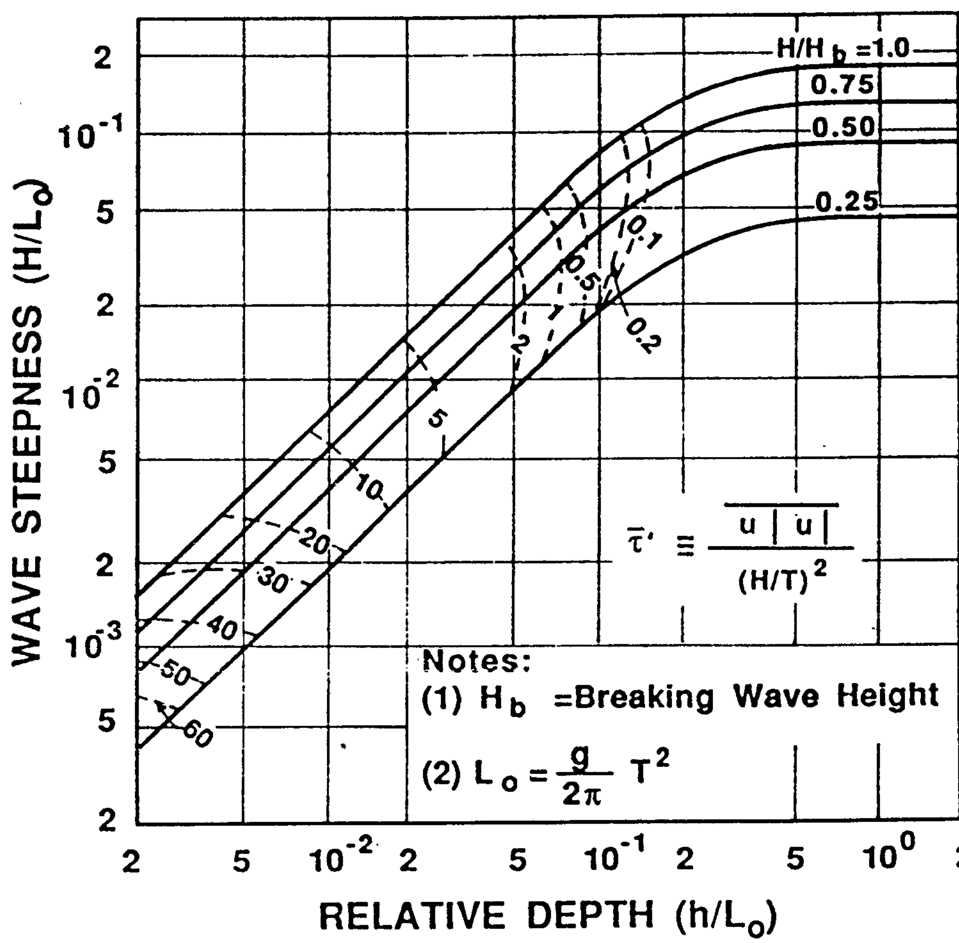


Fig. 7.11. Isolines of Non-dimensional Average Bottom Shear Stress $\overline{\tau'}$ vs Relative Depth h/L_o, and Wave Steepness, H/L_o (after Dean, 1987).

Regional Correlation of Shoreline and Sea Level Change Rates - Previous estimates of long-term shoreline change have been developed (U.S. Army Corps Engineers (1973), Dolan et al. (1983)). These estimates are available on a state wide basis and regionally (such as the entire Atlantic coast). It would be a useful and instructive first broad-brush effort to correlate these estimates with local estimates of sea level rise over the past 50 years or so.

7.4.2 New Data

Quantification of Cross-shore Sediment Budget Components - Referring to Eq. 7.17, the focus of this research element would be to quantify the three terms on the right hand side. The complete methodology has not been developed as yet, but would probably consist of long-term observations of offshore stakes to determine total vertical change, studies of biogenetic production and attrition and deposition rate by suspended sediment traps. It would be useful to conduct this element in conjunction with the experimental element of "Evidence from the Continental Shelf" to be described below.

Shoreline Monument System - The state of Florida maintains a monumented baseline around 1,030 km of sandy shoreline. Since the early to mid 1970's, comprehensive surveys have been conducted on approximately a decadal basis and post-storm studies carried out when appropriate. This type of system provides the only basis for obtaining quality information of shoreline change. It would be very worthwhile, in anticipation of the rising concern over shoreline response to sea level rise, to encourage other states to install, monitor and maintain a monumented system similar to that of Florida.

Evidence from the Continental Shelf - The seafloor of the continental shelf contains information relating to past shoreline response to sea level rise and potential future response. Specifically, Swift (1975) has shown that along much of the Mid-Atlantic Region, there is a "lagoonal carpet" of muds that could not contribute significantly to the sediment budget of the active shoreface. Additionally, the shape of the offshore profiles, along with the availability of sand-sized material contains information (although as yet not completely understood) whether the offshore profile will serve as a source or sink of sand.

In addition to the above, it would be worthwhile to conduct measurements of long-term sediment movement on the continental shelf. These measurements would be conducted along a representative profile; they would document the forcing function (waves, currents, tides, stresses, etc.) and sediment transport (response function). The sediment transport would best be documented through passive means, such as the use of sediment tracing techniques.

7.4.3 New Technology

Laser Profiling - A laser profiling system has been developed in which the airborne laser oscillates in a conical pattern thereby sweeping out a swath of one-half the airplane height, with the ground level pattern being nearly circular overlapping trajectories. The laser return establishes the dry beach elevation, the water surface and the below water profile to approximately two "secchi" disk depths. The potential of this technology to systematically and periodically conduct regional and tidally controlled surveys of shoreline change is extremely encouraging. It is recommended that a pilot program be conducted in Florida in the coming year.

8. SALTWATER INTRUSION

8.1 INTRODUCTION

Many coastal cities rely on local groundwater to meet domestic and industrial needs. With the increasing demand due to greater population and industrial concentrations along the coastline coupled with sea level rise, the potential for saltwater contamination of the aquifer will increase. As in other cases, there are two general approaches to responding to this problem: 1) abandonment of use of the resource, or 2) adoption of management and prevention measures to reduce salinity intrusion. Considerable experience has been gained in coping with saltwater intrusion not principally due to sea level rise, but due to excessive use of the groundwater resource. Yet sea level rise and excessive groundwater usage both decrease the seaward directed piezometric gradient and are, in some respects, comparable. This chapter reviews the various types of saltwater problems that can occur due to sea level rise and the capability to predict and respond to such intrusion.

8.2 LITERATURE REVIEW

The subject of saltwater intrusion into coastal aquifers has been of interest for several centuries. Recently the interest has increased due to intense use of coastal groundwater resources. Methods have been developed to predict the effects of differing usage and management procedures.

Todd (1980) has presented a review of the theory and management practices related to saltwater intrusion in coastal aquifers. In addition to the excessive pumping and sea level rise, saltwater intrusion can result from surface drainage canals which both lower the freshwater head, and if salt water flows into the canals, surface contamination. In recognition of this problem, the State of Florida has enacted legislation to establish a saltwater barrier line in areas prone to saltwater intrusion through canals (Hughes, 1979). There are a number of approaches for controlling saltwater intrusion, as summarized in Table 8.1. Several of these techniques will be illustrated later by examples.

Methods of predicting saltwater intrusion are based on Darcy's law (e.g. Todd, 1980) and include analytical (Henry, 1959; Henry, 1964; van der Meer, 1978; Kozeny, 1953; Strack, 1976; Hunt, 1983) and numerical (McDonald and

Table 8.1. Methods for Controlling Saline Water Intrusion (after Todd, 1980)

Source or Cause of Intrusion	Control Methods
Seawater in coastal aquifer	Modification of pumping pattern Artificial recharge Extraction barrier Injection barrier Subsurface barrier
Upconing	Modification of pumping pattern Saline scavenger wells
Oil field brine	Elimination of surface disposal Injection wells Plugging of abandoned wells
Defective well casings	Plugging of faulty wells
Surface infiltration	Elimination of source
Saline water zones in freshwater aquifers	Relocation and redesign of wells

Harbaugh, 1984; Lennon et al., 1987; Atkinson et al., 1986; Shrivastava, 1978) models. The analytical models are useful for obtaining insight into the interrelationship of various parameters, whereas the numerical models are valuable for predicting detailed response for a particular situation including the benefits of various management strategies. The capabilities of numerical modeling appear to be limited only by knowledge of the transmissive and porosity characteristics of the medium, and the availability of the forcing function, and boundary conditions such as inflow at the boundaries. Most of the efforts conducted to date have not been directed toward evaluating the effects of sea level changes; however, the capability appears to exist. The status of saltwater intrusion in aquifers in all 48 coastal and inland states is presented by Atkinson et al. (1986) as a figure for each state.

In discussing the effects of sea level rise on saltwater intrusion, the distinction between unconfined and confined aquifers is important, see Fig. 8.1. Unconfined aquifers that discharge at or near the shoreline are much more vulnerable to saltwater intrusion than are confined aquifers. The idealized landward displacement of the salt-freshwater interface can be predicted based on the sea level rise, the form of the interface at the level

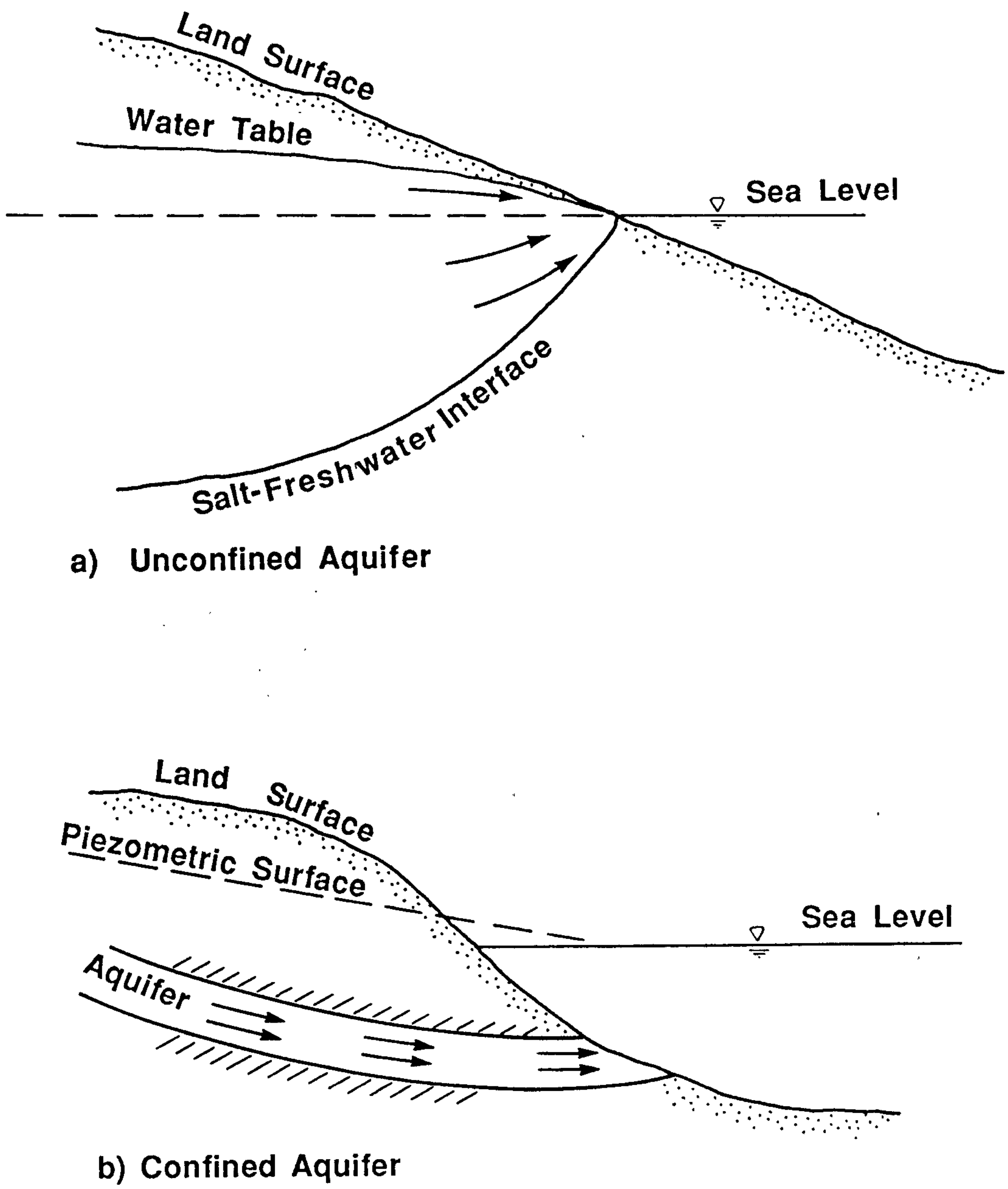


Fig. 8.1. Example of Unconfined and Confined Aquifers.

of interest and the slope of the land. For confined aquifers, the effect of sea level rise is to cause a feedback which tends to offset the rise. The increased head due to the rise at the point of discharge causes a decrease in the piezometric gradient, a reduction in the discharge rate and a resultant transient that, for the same recharge rate, causes an increase in the inland head. The net effect is to reestablish the same discharge rate and same relative (to sea level) piezometric head as before sea level rise.

8.3 PHYSICAL PRINCIPLES AND SOLUTIONS TO IDEALIZED PROBLEMS

8.3.1 General

A simple hydrostatic analysis of the balance between a freshwater aquifier meeting a saline water body yields the Ghyben-Herzberg principle

$$z = \frac{\rho_f}{\rho_s - \rho_f} h \tag{8.1}$$

as illustrated in Fig. 8.2. The above relation applies for a distinct fresh-saltwater interface. However, field measurements have proven definitively that the transition between salt and fresh water is quite gradual with mixing occurring over this transition zone, see Fig. 8.3. The principal cause of the gradual transition, which can be interpreted as dispersion, is the movement of this interface back and forth due to the relatively short period astronomical tide components and also the longer term oscillations due to seasonal variations in replenishment by rainfall and still the more infrequent droughts. During the saltwater advancement and retreat, some salt water is left in the interstices. This is the so-called convective mode. The mixing with the retained fresh water occurs by more slowly occurring molecular diffusion. Experiments have been conducted which demonstrate that the dispersion due to ocean tides can be expressed as

$$D = 4MA/t_o \tag{8.2}$$

where M is a parameter with dimensions of length and A and t_o are the horizontal amplitude of tidal motion and period, respectively. Laboratory studies have been shown that the parameter, M, increases with the uniformity of the stratum and can range from 0.063 cm to possibly as high as 2.8 cm for very uniform sand. Methods employing numerical models combined with Taylor's hypothesis relating the dispersion coefficient tensor, the pore water velocity

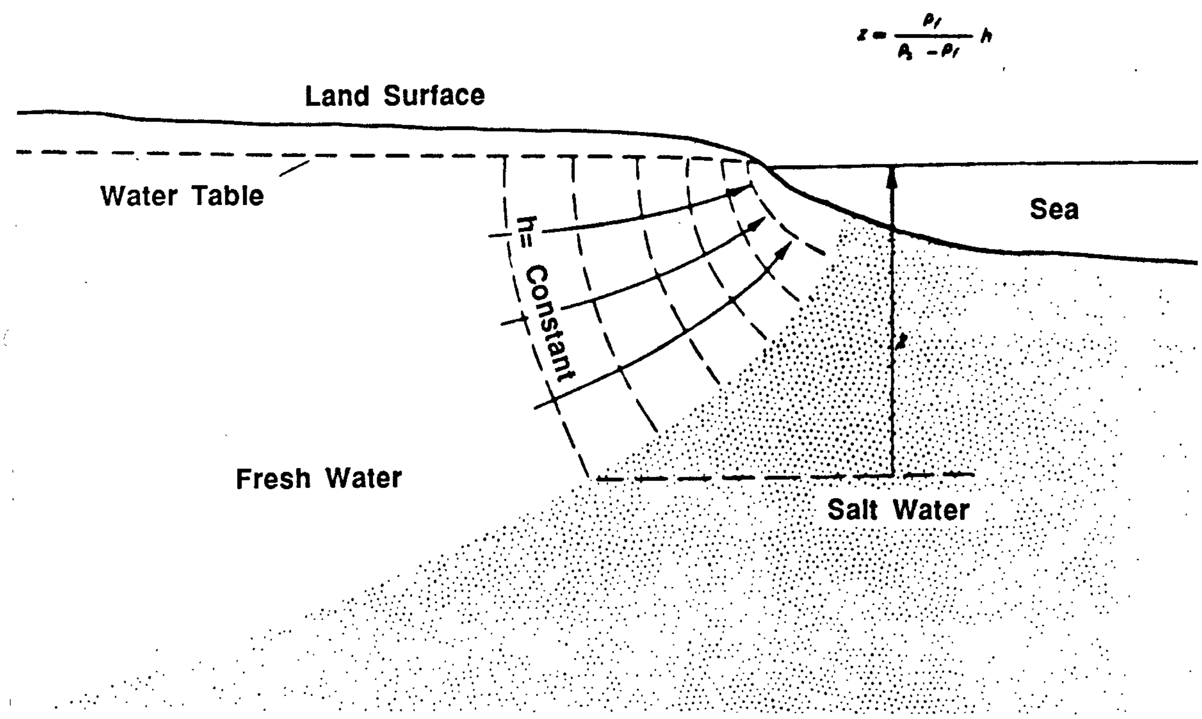


Fig. 8.2. Balance between Fresh Water and Salt Water in a Coastal Aquifer in which the Salt Water is Static (after Cooper, 1964).

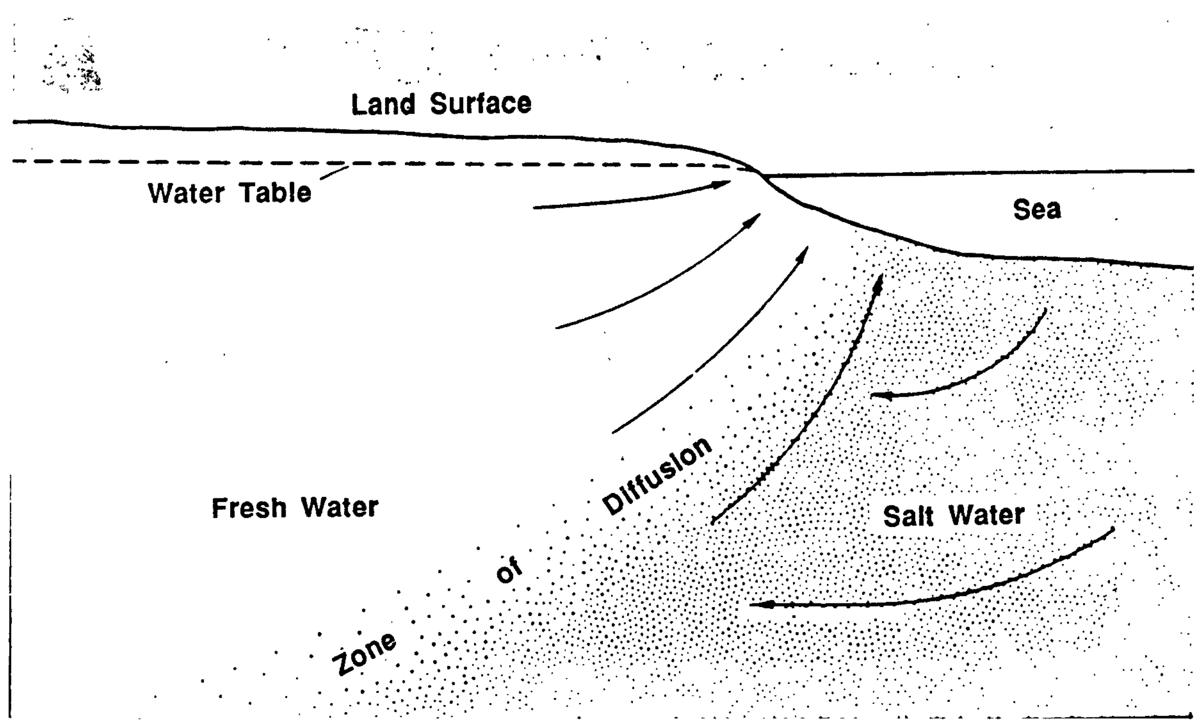


Fig. 8.3. Circulation of Salt Water from the Sea to the Zone of Diffusion and Back to the Sea (after Cooper, 1964).

fluctuations and the Lagrangian time scale tensor have been used to predict anisotropic dispersion coefficients (e.g. Fattah and Hoopes, 1985; Chin, 1986). Based on realistic values of tidal and ground permeability characteristics, it can be shown that the dispersion coefficient decreases from about 100 cm^2/day to 10 cm^2/day at distances of 90 and 275 m from the shoreline, respectively. At a distance of 425 m, the tidally-induced dispersion coefficient would be about the same as molecular diffusion for sodium chloride.

The well known Dupuit approximation is based on the assumption that the hydraulic head is constant along any vertical line throughout the water body and is therefore given by the elevation of the free surface. Strictly speaking, although the Dupuit approximation allows only horizontal flow, it provides meaningful solutions to a number of practical problems.

The entire field of analytical and numerical modeling of groundwater flows is based on Darcy's law which can be expressed as

$$\vec{u} = -\vec{\nabla}\phi \tag{8.3}$$

in which $\vec{u}$ represents the three-dimensional velocity vector, ϕ is the velocity potential

$$\phi = K\left(\frac{p}{\rho g} - y\right) \tag{8.4}$$

$\vec{\nabla}$ is the three-dimensional vector differential operator, and

$$\vec{\nabla}() \equiv \vec{i}\,\frac{\partial}{\partial x}\,() + \vec{j}\,\frac{\partial}{\partial y}\,() + \vec{k}\,\frac{\partial}{\partial z}\,() \tag{8.5}$$

p is the pressure, y is the depth below a datum, K is the hydraulic transmissivity of the medium, and $\vec{i}$, $\vec{j}$, $\vec{k}$ are the unit vectors in the x, y, and z directions, respectively. The continuity equation for an incompressible fluid is

$$\vec{\nabla}\cdot\vec{u} = 0 \tag{8.6}$$

which, when combined with Eq. 8.3, yields the Laplace equation in ϕ

$$\nabla^2\phi = 0 \tag{8.7}$$

In two-dimensional flow, the Laplace equation is also satisfied in terms of the stream function, ψ, so that a field of ϕ and ψ can be constructed to represent the usual flow net with lines of constant ϕ representing lines of equal head and lines of constant ψ being streamlines and everywhere perpendicular to lines of ϕ.

With the governing equations noted above, many solutions to practical problems can be obtained through conformal mapping procedures. For particular geometries, possibly time-varying boundary conditions, and complicated flow boundaries, numerical solutions are applied. The U.S. Geological Survey developed and made available (McDonald and Harbaugh, 1984) a numerical model to simulate groundwater flow in three dimensions.

8.3.2 Discharge through an Unconfined Aquifer

Kozeny (1953) has presented a solution which has been modifed for flow conditions from an unconfined aquifer to a shoreline. The solution is developed in terms of complex variables and a distinct interface between salt and fresh water and yields several results of interest. Denoting x and y as the inland and downward coordinates (see Fig. 8.4), respectively, the interface between fresh and sea water is expressed by

$$y = \sqrt{\frac{2Qx}{\gamma K} + \frac{Q^2}{\gamma^2 K^2}} \tag{8.8}$$

in which Q = freshwater flow per unit length of shoreline, and $\gamma = (\rho_s - \rho_f)/\rho_f$. The horizontal width, x_o, through which the fresh water flows to the sea is

$$x_o = -\frac{Q}{2\gamma K} \tag{8.9}$$

The variation of freshwater head, h, with distance from the shoreline is

$$h = \sqrt{\frac{2\gamma Qx}{K}} \tag{8.10}$$

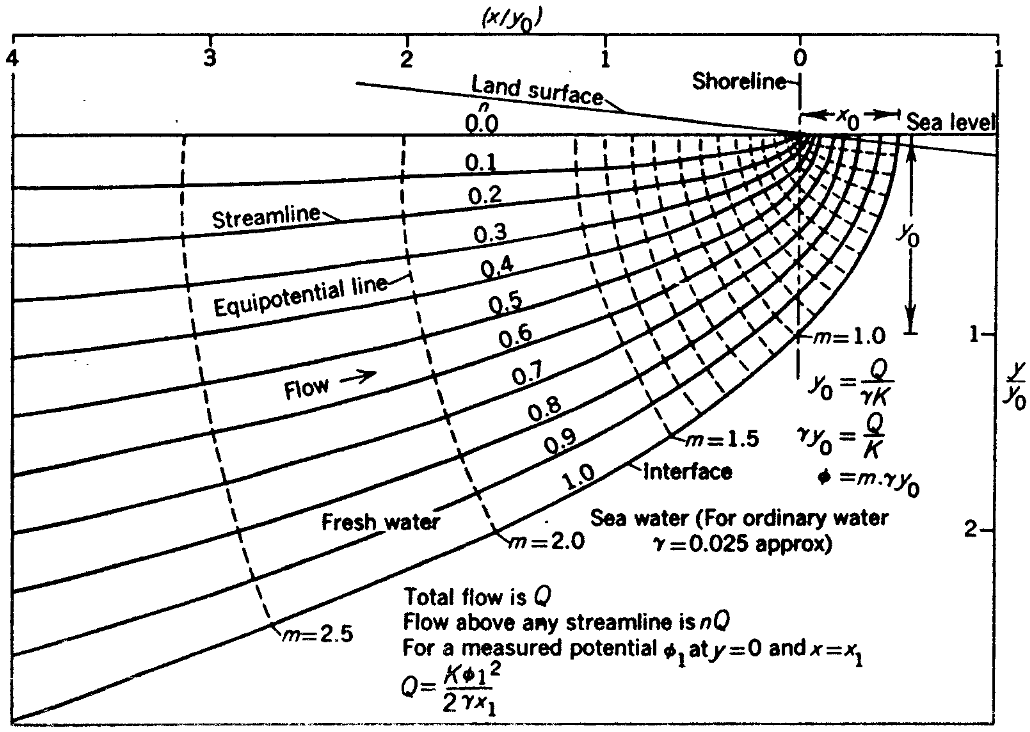


Fig. 8.4. Idealized Characteristics for Unconfined Flow to a Shoreline (after Glover, 1964).

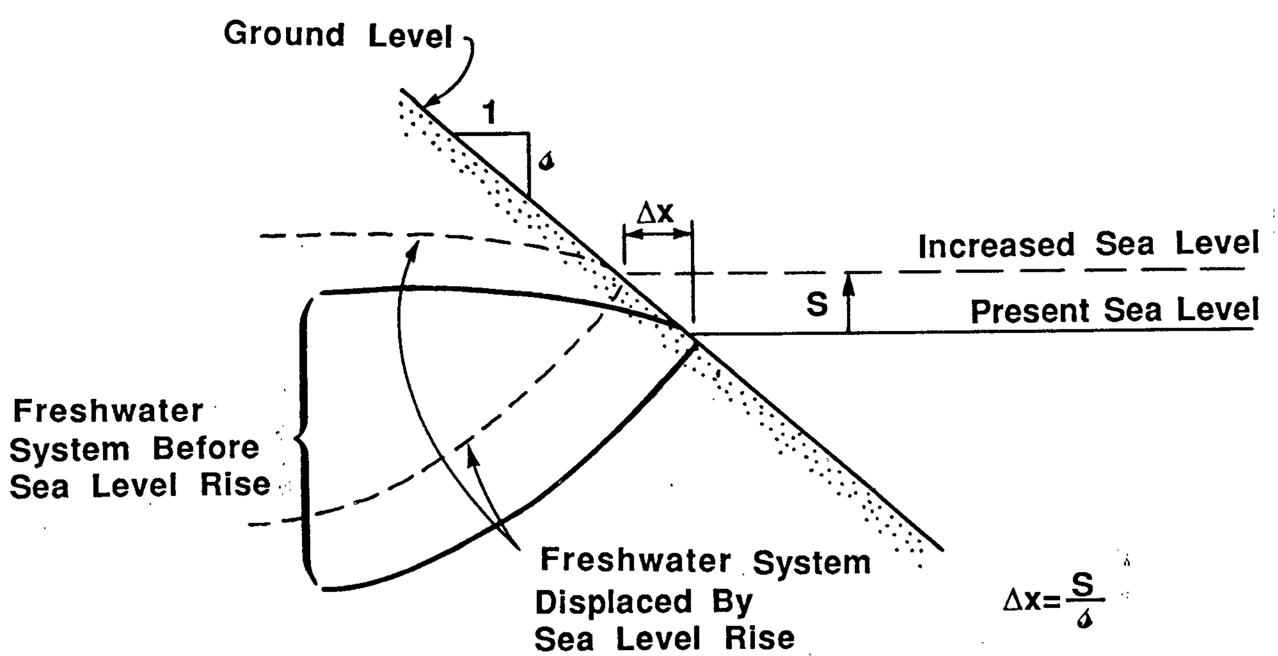


Fig. 8.5. Effect of Sea Level Rise on Equilibrium Groundwater, Highly Exaggerated Vertical Scale.

One result of interest is that if the discharge, Q, diminishes due to a drought, groundwater interception by pumping, or other cause, the outflow region, x_o, is reduced so that the aquifer tends to conserve the freshwater region. Also, from Eq. 8.8, it is possible to calculate the increase in elevation of the salt-freshwater interface if Q is decreased. This is relevant to a pumping field with intakes at a particular depth.

As shown in Fig. 8.5, for an unconfined aquifer, the effect of a sea level rise would be to displace the fresh-saltwater interface landward a distance dependent on the slope, s, of the topography. Thus for a sea level increase, S, the interface would be elevated by a distance S and displaced landward a distance, Δx, where approximately

$$\Delta x = \frac{S}{s} \tag{8.11}$$

For a confined aquifer, which flows to the ocean under pressure, saltwater intrusion into the aquifer is not expected to present the same magnitude of problem as for the unconfined aquifer. As shown in Fig. 8.1b, the only requirement is that the piezometric head at the point of aquifer outcropping must be equal to the depth of water at this point. With an increase in sea level, the rate of outflow can decrease, thereby increasing the piezometric head at the point such that little if any intrusion will occur.

8.3.3 Oceanic Islands

Many oceanic islands are composed of relative permeable limestone with a freshwater layer overlying a saltwater layer. The freshwater layer must be maintained by rainfall or the layer thickness would diminish to zero with increasing time. Referring to Fig. 8.6, the freshwater layer must outcrop at or below the mean sea level.

Considering a circular island of radius, R, with a rainfall recharge rate, W, and employing the Dupuit and Ghyben-Herzberg relations, an approximate freshwater boundary can be determined. The distance, z, to the interface is given by

$$z^2 = \frac{W(R^2 - r^2)}{2K\left[1 + \frac{\Delta\rho}{\rho}\right]\left[\frac{\Delta\rho}{\rho}\right]} \tag{8.12}$$

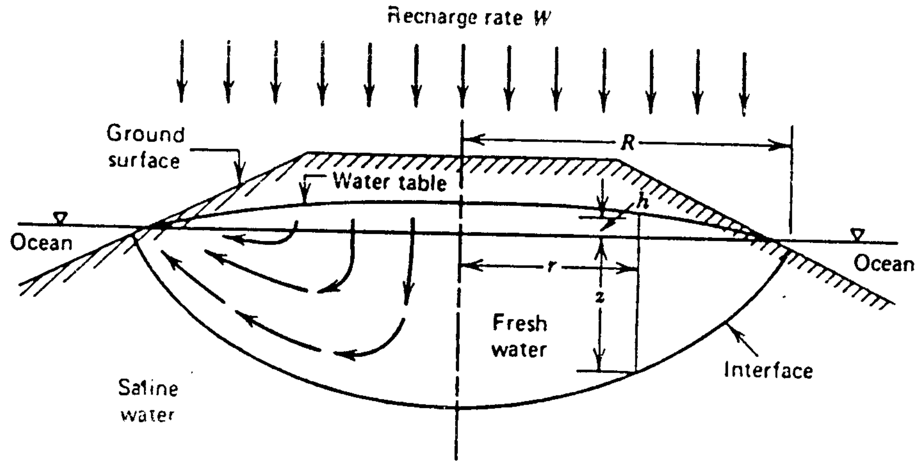


Fig. 8.6. Freshwater Lens Under a Circular Oceanic Island Under Natural Conditions (after Todd, 1980).

in which K is the hydraulic conductivity and the remaining terms are given in Fig. 8.6. Eq. 8.12 presents the idealized situation for a sharp interface; however as noted before, there will always be a fairly gradual transition from fresh to salt water.

The approach yielding the results presented above for a circular island can be applied also to represent a two-dimensional peninsula.

8.3.4 Upconing

One possible effect of groundwater pumping is that "upconing" of a lower saltwater layer may occur in which the decreased pressure caused by pumping draws the salt water into the intake. The maximum discharge, Q_{max}, by a single well to avoid upconing is given by

$$Q_{max} \leq \pi d^2 K(\Delta\rho/\rho) \tag{8.13}$$

in which d is the distance from the bottom of the well to the static interface.

With increasing sea level, assuming that rainfall recharge and other parameters are unchanged, the freshwater region portrayed in Fig. 8.6 will be elevated by the same amount as the sea level rise; and with the same absolute intake depth, the potential for upconing would increase. However, by elevating the well intake point the same amount as sea level rise, the upconing situation would be unchanged to a first approximation from that with the lower sea level. Increased groundwater pumping in amounts greater than the net renewal rate to the aquifer will elevate the interface. An example is at Honolulu where the depths of production wells have decreased gradually from 450 m to 85 m as the freshwater layer thickness has been decreased by excessive pumping.

8.3.5 Single Extraction Well Near a Coast

An analytical solution has been developed by Strack (1976) to represent the flow field in a confined aquifer in the vicinity of a coastline as affected by the withdrawal from a single well, see Fig. 8.7. For the case in which the unperturbed flow per unit length of coastline is q and with the well located a distance x_o from the coast, the critical withdrawal, Q_c, is defined by the following two equations

$$\frac{KD^2}{q_o x_o}\gamma(1+\gamma) = 2\frac{x}{x_o} - \left[1 - \left(\frac{x}{x_o}\right)^2\right] \ell n \left[\frac{1 + x/x_o}{1 - x/x_o}\right] \qquad (8.14)$$

$$\frac{Q}{q_o x_o} = \pi \left[1 - \left(\frac{x}{x_o}\right)^2\right] \qquad (8.15)$$

In the above equations, D is the vertical distance from mean sea level to the aquifer bottom boundary.

8.3.6 Saltwater Barriers

In areas where solution channels exist that convey large quantities of fresh water to sea, the possibility exists of employing underwater dams. This approach has been tested successfully in the limestone caverns of Port-Miou River near Marsielles, France.

8.4 CASE STUDIES

There are two general approaches for reducing saltwater intrusion into coastal aquifers: 1) modifying pumping practice, and 2) construction of flow barriers. The most simple and direct approach is, where appropriate, to reduce groundwater withdrawal during periods of drought when intrusion would

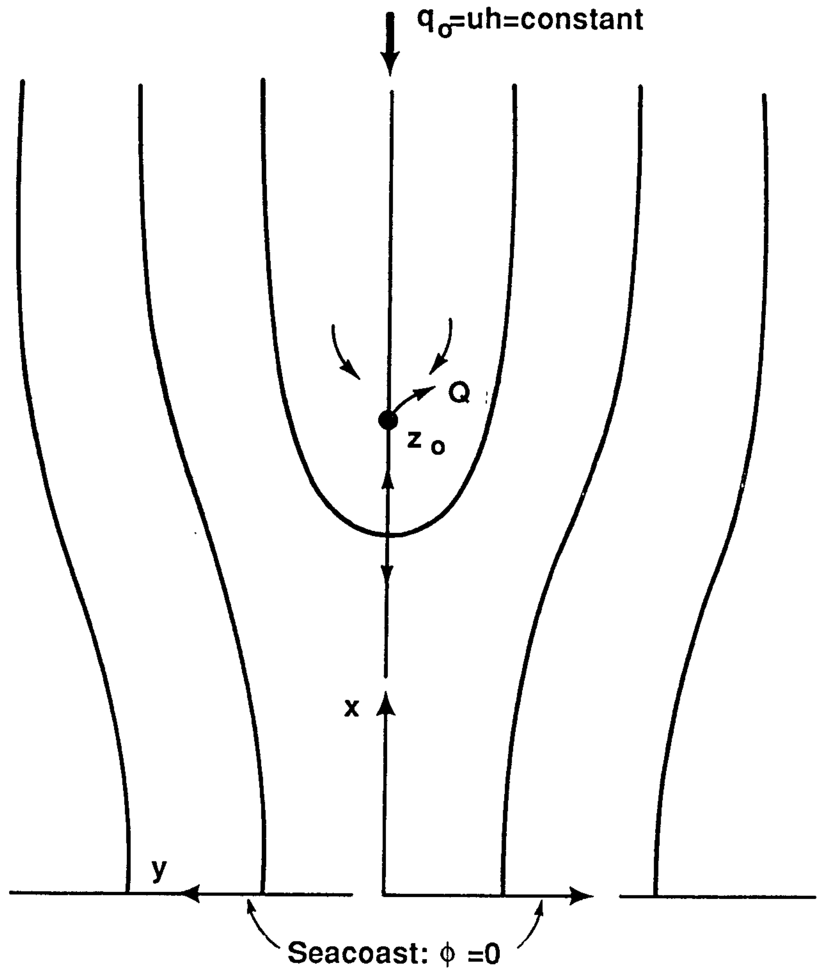


Fig. 8.7. Flow to a Single Well Along a Seacoast.

tend to occur. It is clear that with increased sea level, the frequency and/or duration that intrusion would tend to occur will increase. Unfortunately, drought occurrences are precisely the times that demand on groundwater resources tends to be increased. An alternate approach is to adopt a pumping plan to minimize intrusion through hydraulic or structural barriers.

Five case studies will be presented of approaches to cope with salinity intrusion into aquifers.

8.4.1 Long Island, NY

Todd (1980) has described the effects of excessive pumping near the more populated western end of Long Island including a reduction in the water table to 10 m below sea level. The response has been abandonment of well fields on the western end of the island and an effort to salvage storm water and recharge it into the aquifer through infiltration fields. The present horizontal rate of advance of the saline wedge in southwestern Nassau County varies from 3 m to 60 m per year depending on local pumping rates and recharge by variations in rainfall. Fig. 8.8 presents a cross-section through southwestern Nassau County.

8.4.2 Miami, FL

The initial cause of salinity intrusion in this area was a series of surface drainage canals. These canals both lowered the fresh water table as they were designed to do and also allowed salt water to penetrate up these canals, thereby contaminating the groundwater from the surface. Fig. 8.9 portrays the increase in intrusion from 1904 to 1959. Increased salinity required several well fields to be abandoned in the Miami-Fort Lauderdale area.

Remedial measures have included construction of saltwater barriers in the drainage canals and the establishment of water management areas which serve to pond fresh water for aquifer recharge.

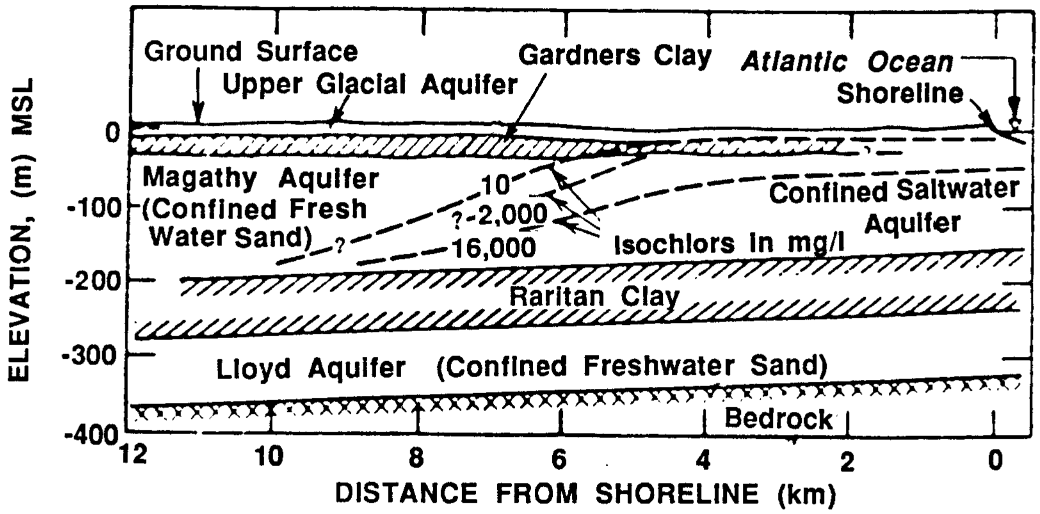


Fig. 8.8 Profile through Aquifer at Far Rockaway, Nassau County, Long Island, Showing Location of Salinity Front as a Result of Pumping (after Todd, 1980).

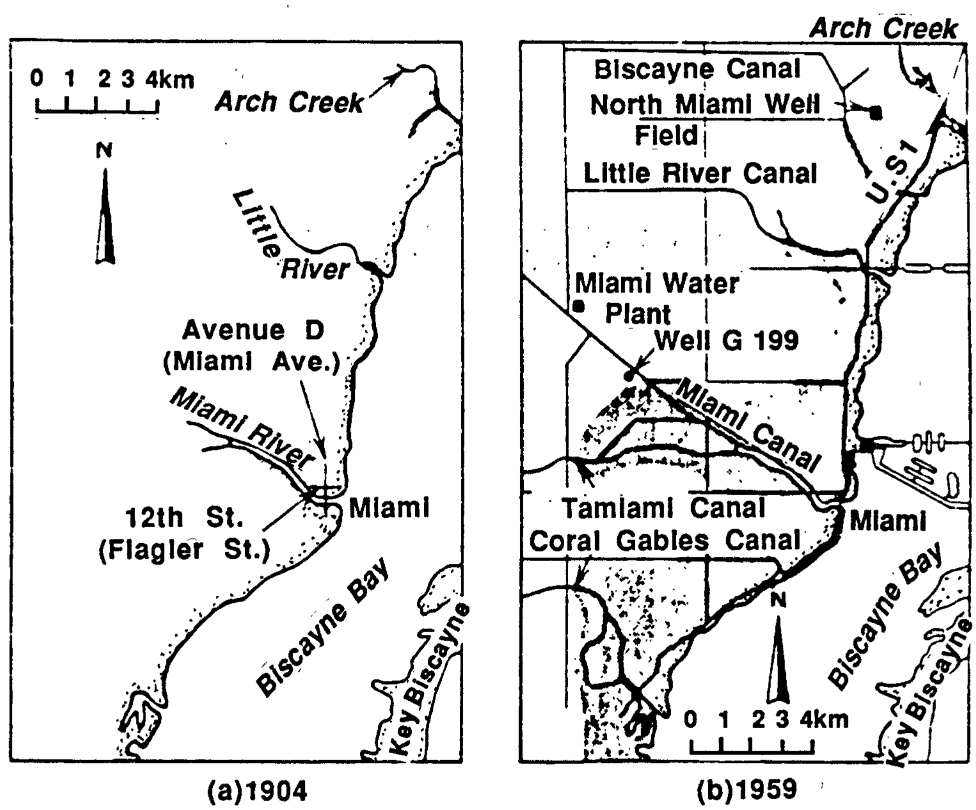


Fig. 8.9. Progressive Saltwater Intrusion in the Vicinity of Miami, FL, 1904 to 1959 (after Todd, 1980).

8.4.3 Los Angeles, CA

Saltwater intrusion was noticed in the early 1930's as a result of water usage for agricultural, domestic and industrial purposes. A water injection barrier project has been implemented along a 11 km portion of the shoreline. A total of 94 injection wells cause a pressure "ridge" in the confined aquifers, thereby establishing a seaward gradient which prevents seawater encroachment. The piezometric head of the barrier is maintained at 1 to 3 m above sea level, requiring approximately 1,500 m^3/day of injected water. Fig. 8.10 presents a cross-section perpendicular to the shoreline which demonstrates the increase in the pressure ridge from 1963 to 1967.

8.4.4 The Potomac-Raritan-Magothy Aquifer System

Lennon *et al*. (1987) have reported on a numerical model study to evaluate the effectiveness of creating a hydraulic barrier to salinity inflow for the Potomac-Raritan-Magothy (PRM) aquifer system. The optimum approach was a combination of extraction and injection wells. The aquifer length (parallel to the Delaware River) considered was 1,520 m and 5 injection wells were

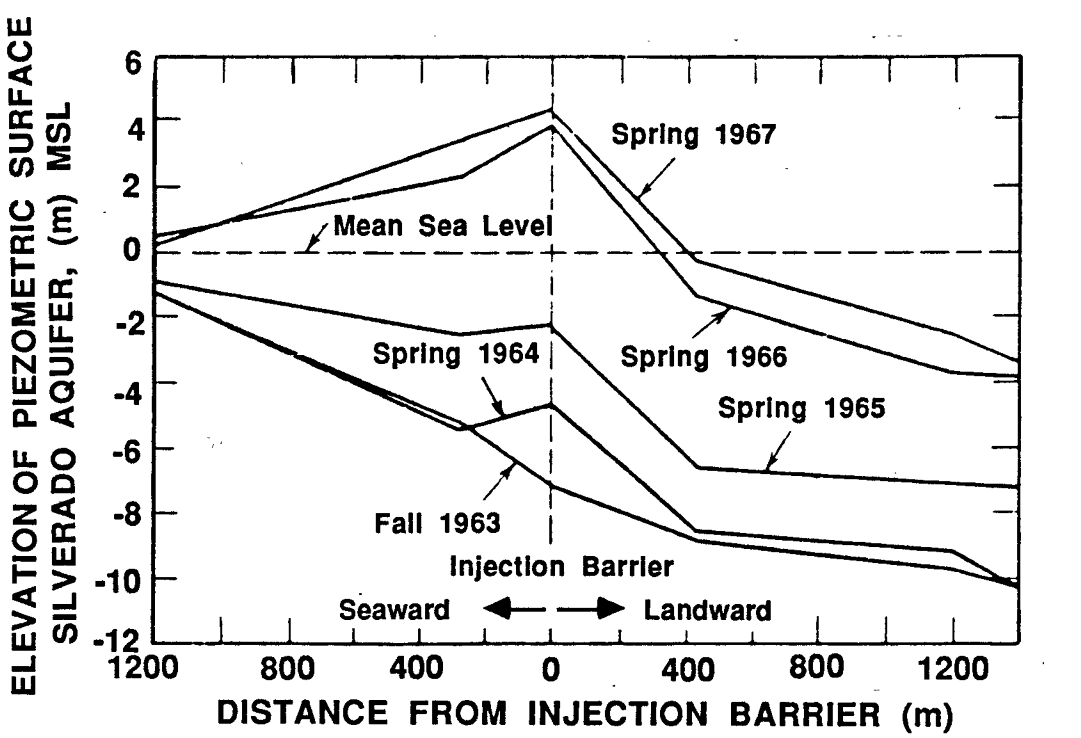


Fig. 8.10. Piezometric Pressure Profiles Perpendicular to the Seawater Intrusion Barrier in Los Angeles County for Various Times after Commencement of Injection in the Fall of 1963 (after Todd, 1980).

located parallel to and 300 m from the Delaware River and the 5 extraction wells were located 600 m from the river. Only 0.014 m^3/s was pumped from each extraction well and reintroduced into the aquifer through the adjacent injection well. This resulted in a freshwater "mound" that performs as a barrier to saltwater inflow. Following the drought, the injection wells are used as extraction wells (0.014 m^3/s) with the water returned to the river, thereby reducing the salinity in the aquifer further.

8.4.5 Okinawa-jima Island

The construction of underground flow barriers through pumping of cement grout near the coastline can create underground impoundments similar to surface reservoirs. In 1978, such an experimental subsurface barrier was constructed in a small buried valley on Miyako-jima Island (Sugio et al., 1987). The barrier was constructed in a very porous limestone aquifer and is 16.5 m high, 5 m wide and 500 m long. Based on field monitoring studies conducted over a four year period, the installation was judged a success; the hydraulic conductivity and porosity were reduced from 0.17 cm/s and 20% to 5×10^{-5} cm/s and 6%, respectively. Plans are underway to construct a much larger barrier at Komesu on Okinawa-jima Island where the annual rainfall is 2,400 mm and occurs during a nine month period. With the present high permeability, much of this fresh water flows to the sea and saltwater intrusion tends to occur during the remaining three months when heavy pumping is conducted for irrigation purposes.

Based on numerical modeling with a particular barrier design of 5 m thickness, it is concluded that with the barrier no salinity intrusion will occur for a 60 day period during the drought and that if a longer period is desired, the barrier thickness must be increased or the pumping rate (6,000 m^3/day) must be decreased.

8.5 RESEARCH NEEDS

Given the ambient flow conditions in a coastal aquifer, the transmissive and porosity properties of the aquifer and various scenarios of extraction "demand" on the aquifer, it appears that the characteristics of the salinity intrusion, including any time dependencies can be predicted reasonably reliably by state-of-the-art numerical modeling. However, at present the effects of sea level rise on saltwater intrusion have been examined only for special cases. Any decisions by those responsible for planning related to the need to relocate well fields, modify usage, implement remedial measures, etc. must be based on realistic estimates of the effects of sea level rise and other causes such as increases in extraction rates.

A research program is recommended which would develop and exercise rather simple models specifically with the objective of illustrating the relative effects of sea level rise. The remaining efforts would be much more comprehensive and concentrate on evaluating the reliability of numerical models for predicting salinity intrusion and on case studies of areas where the potential for salinity intrusion is high. These studies would focus on the applications of models, parameterized and calibrated for the local aquifer characteristics and using various sea level rise scenarios to predict effects on extracted water and on the need for and effectiveness of various types of remedial measures.

9. UPRIVER SALTWATER PENETRATION

9.1 INTRODUCTION

An estuary by definition is a semi-enclosed water body in which sea water and fresh water from river mix under tidal action. Sea water is thus diluted measurably (Cameron and Pritchard, 1963), and, in some cases, penetrates in the form of a saline wedge upriver. In other cases the waters are vertically well mixed, and salt water penetration occurs without the presence of a distinct wedge. Water bodies which do not receive fresh water tend to be highly saline, with salt concentration equal to that in the sea. In some cases, e.g. in parts of Florida Bay, in southern Florida, excessive evaporation during the hot and dry season renders the waters super-saline, with salinity exceeding that in the sea (Atlantic Ocean).

The three main parameters which control the degree of salt penetration are the river runoff velocity, water depth and tidal range in the sea. Increasing the tidal range or the depth, or decreasing the runoff will increase penetration. In urbanized areas, withdrawal of fresh water and dredging of deeper channels for vessel navigation are important issues; hence the effect of reduced runoff as well as the effect of increased depth on salinity intrusion have been investigated by scientists and engineers. The effects of reduced runoff has, for example, been recently investigated in the Myakka and other rivers near Florida's Gulf of Mexico coast.[1] A serious problem of this nature occurred in the Delaware River Basin in the 1960's due to drought. During the worst period, the salt front advanced 53 km up the river and forced some industries near Philadelphia to seek water from a municipal system that imported water from the Susquehanna River Basin (Hull and Titus, 1986).

The influence of increased depth is analogous to what would occur (and has occurred) in the event of a sea level rise. The propagation of tide up the estuary is affected in this case. On the other hand, enhanced fresh water withdrawal only partially simulates the sea level rise effect, since reduced runoff does not influence tides as significantly as does an increased water depth.

[1]Ernest Estevez, Mote Marine Laboratory, Sarasota, Florida, personal communication.

9.2 LITERATURE REVIEW

The significance of salt water penetration has been a matter of common knowledge among scientists and engineers for a long time, but the entire process was placed on a firm physical footing relatively recently. Perhaps the most important work, at least in the United States, was carried out during the post-second world war period by Keulegan at the National Bureau of Standards, by Ippen at M.I.T., and later by Harleman also at M.I.T., in cooperation with the U.S. Army Engineer Waterways Experiment Station at Vicksburg, Mississippi. These works, summarized in Ippen (1966), were primarily laboratory-based, and sought to understand the intrusion process through the development of basic, analytic formulations. It is noteworthy that, in a sense, these investigations were successors to the pioneering work of O'Brien (O'Brien and Cherno, 1934), which dealt with the fundamentally similar problem of predicting the rate at which salt water intrudes into fresh water, as in a lock separating a saline water body from one of lower salinity. Partheniades (1971) has reviewed the fundamentals of the salt water intrusion mechanisms (see also Partheniades et al., 1975). Beginning in 1954, the Committee on Tidal Hydraulics of the Corps of Engineers has issued a series of reports pertaining to various theoretical and practical aspects of the estuarine salinity intrusion problem.

Concurrently with the aforementioned basic laboratory-analytic studies, physical models of real estuaries were developed, with the inclusion of salt water intrusion effects. The Corps of Engineers led this effort, and constructed a rather large model of the San Francisco Bay estuarine system in Sausalito, California (Fischer et al., 1979). This is a distorted model (scales 1000 horizontally and 100 vertically), i.e. one in which the water depths are greater than the linear scale set by the aerial (horizontal) dimensions. Distortion is commonly employed in physical models in order to generate sufficient turbulence necessary to satisfy the desired equivalence between model and prototype flow conditions. Distortion, however, essentially requires additional flow resistance to be provided at the bottom by means of artificial roughness elements. Calibration of such a model, particularly one involving two fluids, can be a tedious process. It took several years to make the San Francisco Bay model fully operational. Other major physical models of this nature are those of the Mississippi River and Chesapeake Bay.

Extensive field investigations were carried out in the fifties for understanding salinity intrusion through direct (prototype) evidence, and to calibrate physical models such as the ones noted above. Reference may be made to the investigation by Sir Claude Inglis (Inglis and Allen, 1957) on the Thames in England, and by Prof. Pritchard (Pritchard, 1952) on the Chesapeake Bay. In both estuaries, as elsewhere, studies which began in the fifties (and earlier) have been of an ongoing nature, and have continued until the present time. The reason for this is both a continued interest in the basic aspects of the mechanism of mixing between salt water and fresh water in the real environment (see e.g. Dronkers and van de Kreeke, 1986), and also because new engineering problems continually arise and, therefore, must be examined afresh. Physical models of large estuarine systems such as the San Francisco Bay, Chesapeake Bay, the Mississippi and New York Harbor have been retained, and are used as needed by the Corps of Engineers and other agencies. As fresh input for modification and calibration of such models, prototype studies are conducted, although many now tend to be highly site-specific, given the costs involved in field work.

Computer technology has made it possible to develop sophisticated numerical models for handling estuarine hydrodynamics including salinity intrusion. The early models, in the sixties and early seventies, were typically one-dimensional, simulating cross-sectional average processes in the longitudinal direction (e.g., Harleman _et al._, 1974; Miles, 1977). These were followed by two-dimensional, depth-averaged models, e.g. such as the one incorporated in the TABS-2 system of estuarine numerical models used by the Waterways Experiment Station (Thomas and McAnally, 1985). More recently, fully three-dimensional models have been developed. These models have been applied to a study of New York Harbor[2].

9.3 PHYSICAL PRINCIPLES

Analytic approaches to solve the problem of salt water intrusion depend upon whether the estuary can be treated as stratified, or as partially mixed or fully mixed. Thus, for example, an estuary is classified as "well mixed"

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[2] Allen Teeter, Waterways Experiment Station, Vicksburg, Mississippi, personal communication.

if the salinity over any flow cross-section does not vary by more than 10%; otherwise it is classified as partially mixed (Ippen, 1966). In a fully stratified estuary the salt is contained almost entirely in the lower layer (the wedge). A further quantitative means of classifying and comparing estuaries, and one which requires salinity and velocity only, is due to Hansen and Rattray (1966). Their commonly used diagram relates, for different types of estuaries, a stratification parameter, $\delta s/s_o$, defined as the ratio of the surface to bottom difference in salinity, δs, divided by the mean cross-sectional salinity, s_o, and a circulation parameter, u_s/u_f, the ratio of net surface current, u_s to the mean cross-sectional velocity, u_f.

The penetration of salt water into fresh is essentially a density-driven phenomenon, since salt water is the heavier of the two fluids. Consider a barrier which separates two fluids, one of density ρ (water) with zero salinity (s=0) and another of density $\rho+\Delta\rho$ (salt water), salinity s_o. As shown in Fig. 9.1a, if the barrier were lifted at time t=0, a salt water gravity front (current) would penetrate fresh water at the bottom, and an equal volume of fresh water would be displaced per unit time into salt water, at the top. If now a fresh water flow of velocity, U_o, were imposed as shown in Fig. 9.1b, the intrusion of gravity current would be arrested at some position where the pressure and drag forces in both directions balance. This is the basic mechanism which operates within estuaries, where sea-driven tides are an additional factor. In cases where the mixing potential of the incoming tidal energy is relatively high, the estuary will be well mixed in terms of vertical salinity structure. If not, the estuary will be stratified. Since both the tidal range and runoff vary; the former principally on a synodic basis while the latter seasonally, many estuaries shift between fully stratified and well-mixed conditions over a year period (Dyer, 1973). In some cases, very high runoff can virtually "clean out" the estuary of salt water, e.g. during and immediately following storm runoff, or, for example, during the monsoon season (Fig. 9.2).

During a tidal cycle, the water level rises and falls, and this variation directly influences the length of penetration of salt water; at high water, salt water penetrates further than at low water. It is instructive to look at the basic governing equations and their behavior in this context.

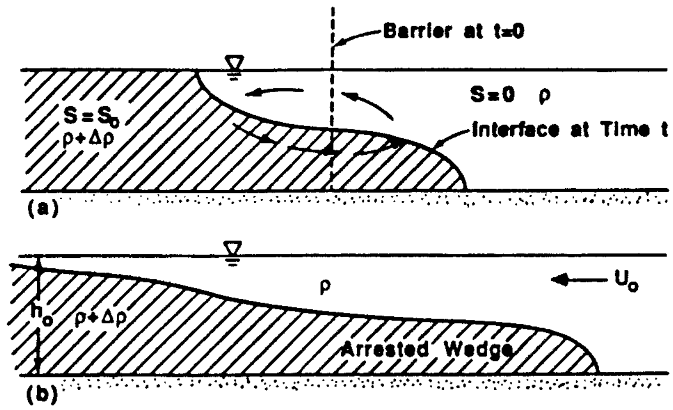

Fig. 9.1. Mechanism of Salt Penetration: a) Development of a Gravity Current, b) Arrested Saline Wedge.

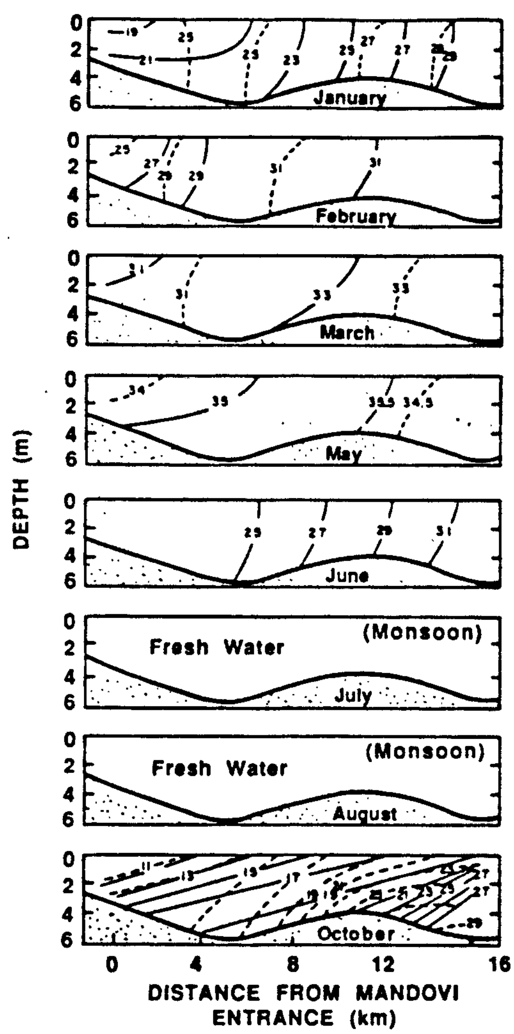

Fig. 9.2. Monthly Salinity (ppt) Distribution in Cumbarjua Canal, Goa, India; —— Ebb; ---- Flood (after Rao et al., 1976).

Ippen and Harleman (Ippen, 1966) considered the basic unsteady, one-dimensional (x,t) mass conservation for salt transport-diffusion, and derived the following expression for the resultant profile of salinity, s(x,t) in a non-stratified estuary:

$$\frac{s(x,t)}{s_o} = \exp\left\{-\frac{U_o}{2D_o'B}\left[N-(N-x)\exp\left(\frac{a_o}{h_o}(1-\cos\sigma t)\right)+B\right]^2\right\} \tag{9.1}$$

where $N = h_o u_o / a_o \sigma$. Here, U_o = fresh water outflow velocity, D_o' = diffusion coefficient, B = an empirical, diffusion-related coefficient, a_o = tidal amplitude, σ = tidal frequency, u_o = maximum tidal velocity at the mouth and h_o = mean water depth.

If the end of the intrusion zone is specified at s/s_o = 0.01, the maximum intrusion length, L_{max}, occurs when $\sigma t = \pi$, and the minimum, L_{min}, when σt = 0. Thus

$$L_{max} = N\left[1-\exp\left(-\frac{2a_o}{h_o}\right)\right] + B\left(3\ \frac{D_o'}{U_oB} - 1\right)\exp\left(-\frac{2a_o}{h_o}\right) \tag{9.2}$$

$$L_{min} = B\left(3\ \frac{D_o'}{U_oB} - 1\right) \tag{9.3}$$

In the above equations, the parameters D_o' and B must be known. In a real estuary, these parameters can be determined by measuring the average salinities at two points in the estuary at low tide.

Examples of average salinity distribution at high and low water slack as shown in Figs. 9.3a,b are based on the experiments in a model tidal channel at the Waterways Experiment Station (Harleman and Abrahams, 1966). This illustration of the difference in salinity intrusion between high water slack and low water slack (tidal range 3 cm) is essentially analogous to what would occur under an equivalent sea level rise. Note that Figs. 9.3a,b clearly show that salinity intrusion is a highly dynamic phenomenon, and that there is a 21.4 m difference in the distance of penetration between the high and low water events. It is thus evident that only a small head is required to cause a significant horizontal movement of the salt water front. Raising the sea level (increasing h_o in Eq. 9.1) would amount to pushing both curves up the estuary. The same would occur if h_o were increased by dredging. Likewise, decreasing, U_o, the outflow velocity, would cause further

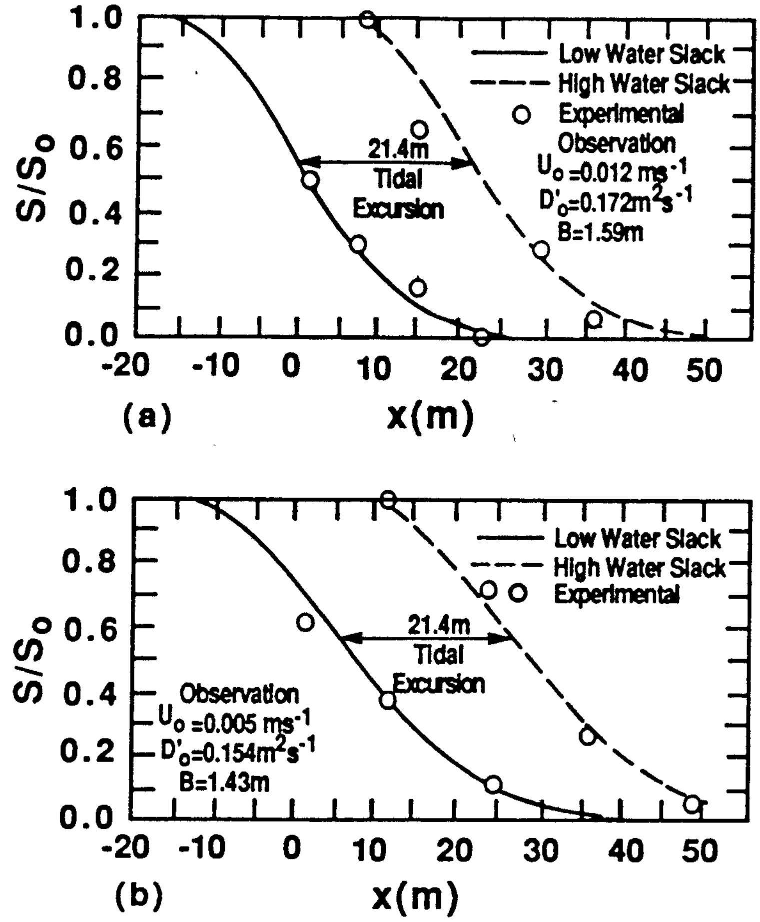

Fig. 9.3. Longitudinal Salinity Distribution in a Model Tidal Channel: a) Test No. 2, b) Test No. 16 (after Harleman and Abrahams, 1966).

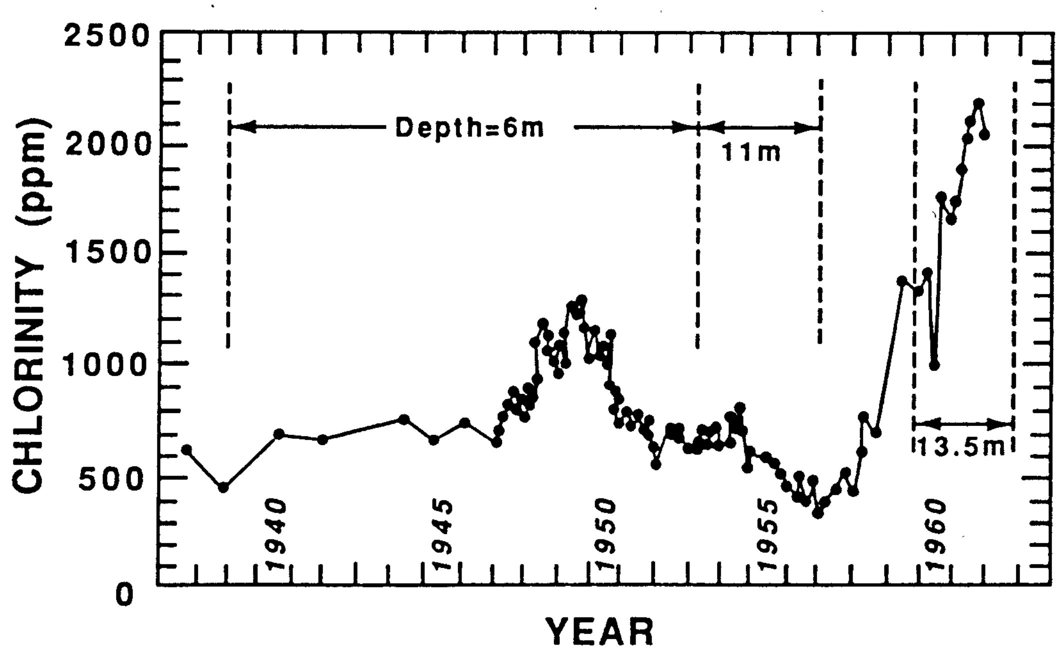

Fig. 9.4. Salinity (Chlorinity) Variation with Years in Lake Maracaibo (after Partheniades, 1966).

penetration. This can be easily shown via Eq. 9.2, for example. A possible secondary effect in the event of a sea level rise could be higher tides at the mouth. In this case a_o would increase, thereby reinforcing the penetration effect produced by increasing h_o.

The aforementioned simple, demonstrative theory of course has major limitations by virtue of the assumptions under which the stated equations were derived. In general, however, many of the restrictions can be relaxed via solution of the fully three-dimensional transport equations numerically. Furthermore, by following a hybrid approach involving a combination of field data, physical model-based data and numerical simulation, it is presently feasible to arrive at realistic descriptions of salinity distributions in the estuary, and such effects as those due to sea level rise can be simulated with a reasonable degree of confidence, particularly in the tide-dominated (as opposed to waves) environment. It should, however, be noted that the costs involved in "full blown" studies remains high, with the result that salinity intrusion effects have been investigated so far, in detail, mainly in highly urbanized estuaries.

9.4 EXAMPLES

It suffices to examine the effects of channel deepening and runoff on salinity intrusion as paradigms (qualitative in the case of runoff) for what would occur in the case of a sea level rise.

A major salinity intrusion problem developed in the Maracaibo estuary, Venezuela, in the sixties (Partheniades, 1966). The problem was traced to the construction, during the previous years, of a deep, 60 km long, navigation channel connecting Lake Maracaibo to the Gulf of Venezuela. A representative salinity (chlorinity) record within the lake during the 1937-63 period is shown in Fig. 9.4. A significant fluctuation of salinity first occurred during the 1947-52 period. This rise in salinity, from typical values of 500-700 ppm during the previous and the subsequent years, to a peak of about 1,400 ppm, is believed to have been due to the occurrence of a dry period with low runoff. However, following the completion of the deeper (from 6 m to 11 m) channel in 1956, the salinity increased relatively rapidly and almost continuously to a peak of 1,500 ppm in 1959. At this time, to accommodate yet larger oil tankers, the channel was further deepened to 13.5 m. This caused a further increase in salinity and, in 1962-63, when the deepening project was completed, the salinity rose to 2,200 ppm. A physical model of the estuary was subsequently constructed to examine the problem in further detail (Brezina, 1975). The ultimate outcome of the channel deepening projects was to turn Lake Maracaibo from a relatively fresh water lake into a brackish water lagoon, with a complete change in life forms[3].

The effect of dredging is commonly evaluated for Environmental Impact Statements. Such a study was carried out as prerequisite to the construction of the Trident submarine base at King's Bay, Georgia. The ocean entrance to this bay is through St. Mary's Entrance, Florida. Fig. 9.5 shows the application of Eq. 9.1 for predicting the high and low water (tidal range 2.2 m) salinity profiles in this prototype case, prior to dredging (Parchure, 1982). Although agreement between theoretical curves and measured data is obviously very rough, the theory, although approximate, does simulate the overall trends suggested by the measurement. Dredging plans called for a deepening of the channel between the entrance (end of jetties) and King's Bay ranging from 0 to 4 m. A 4% rise in the salinity was predicted 20 km up the entrance as a result of dredging (Environmental Science and Engineering, Inc., 1980).

The aforementioned examples are merely illustrative of the basic phenomenon of interest, and are not meant to demonstrate the power of available technology with respect to physical or numerical models and their application. The subject matter has been covered extensively in literature on fluid mechanics and hydraulic engineering. For a fuller treatment, the work of Fischer *et al.* (1979) may be cited. Here, it will suffice to make reference to a recent study conducted to examine the potential effects of deepening the lower Mississippi river on salinity intrusion (Johnson *et al.*, 1987). In Fig. 9.6, the duration of the saline wedge intrusion (days per year) is plotted against distance above the head of a number of passes (distributaries) in the vicinity of New Orleans. Two curves are shown, one for a 12 m deep channel (present depth is 9 m) and another for a 17 m deep channel. These curves were generated numerically. The model was calibrated

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[3] Jindrich Brezina, University of Zulia, Maracaibo, Venezuela, personal communication.

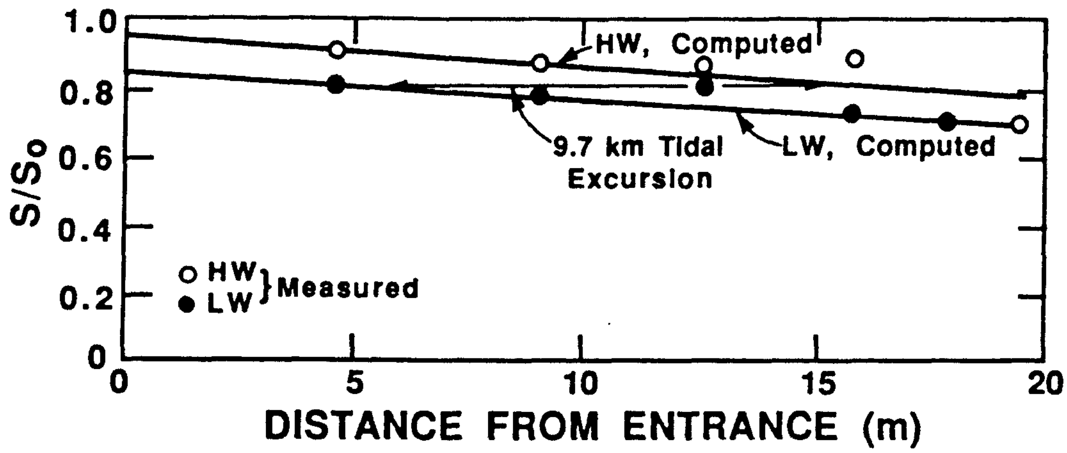

Fig. 9.5. High and Low Water Salinity Profiles through St. Marys Entrance, Florida and Cumberland Sound, Georgia (after Parchure, 1982).

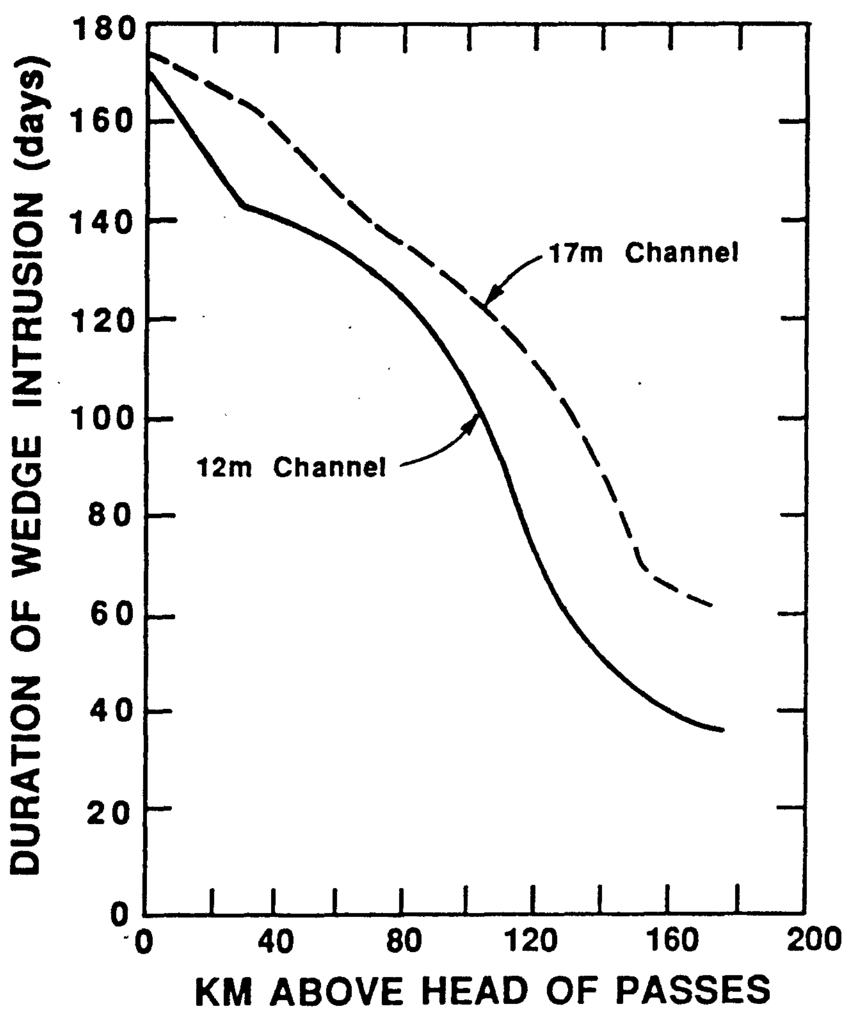

Fig. 9.6. Effect of Channel Deepening on the Duration of Wedge Intrusion in the Lower Mississippi River (after Johnson et al., 1987).

against data obtained during the 1953-54 low-flow period. The results essentially are illustrative of the effect of channel deepening from 12 m to 17 m. Thus, for example, this effect would increase the duration of the wedge at 180 km from about 40 days to about 60 days per year.

9.5 RESEARCH NEEDS

Scientific work in understanding the basic processes of mixing between salt and fresh waters is likely to continue well into the future, in order to steadily improve predictive methodologies, which have at present attained a reasonable degree of sophistication. However, further improvements are highly desirable.

An area which deserves further consideration is mixing under wave action, or combined wave and tide action. At present the effect of waves in this respect is only weakly understood. Much of analysis carried out so far appears to have been directed towards situations dominated by tides alone. It is noteworthy as well that with deeper water associated with sea level rise, waves would be admitted more freely, thereby decreasing fluid stratification by virtue of greater mixing in the vertical direction, particularly in estuary mouths.

Another important research area is related to the development and motion of fronts due to salinity, temperature and sediment density gradients. Understanding the behavior of these fronts is vitally important to a range of water quality and ecological issues.

It must be mentioned that most of the work done to-date appears to have dealt with such effects as those related to channel deepening or changes in upstream river hydrology, rather than sea level rise. As noted, sea level rise can also increase the coastal tide and wave action, particularly if the coastline is rocky, and does not recede, while water depth increases. A similar situation can also arise if the shoreline were erodible, but sea level rise was so rapid as to prevent depths at the mouth from achieving quasi-equilibrium with the hydrodynamic forcing. This would lead to a situation wherein the estuarine mouth would be in deeper water, and where the tide (and waves) would arrive with lesser hinderance due to reduced bottom friction and lesser chance of wave breaking in the offshore waters. In most work carried out so far, the coastal tide (and waves) is typically assumed to practically remain unchanged.

10. SEDIMENTARY PROCESSES IN THE ESTUARINE REGION

10.1 INTRODUCTION

There have been drastic changes in the world's shorelines since the retreat of the last ice age. Over the past ~ 6,000 years, the rate of rise of sea level has been relatively low (0.08 m/100 yr) compared to the period ~ 20,000 - 6,000 BP (0.8 m/100 yr) (Fig 7.1). It is not surprising, therefore, that many of today's estuaries have been "around" approximately in their present configuration only in the past few millennia. Even in the absence of sea level change, estuaries are highly dynamic and in many ways reflect the type of macro-scale processes characteristic of oceans.

Estuarine shorelines change under the action of hydrodynamic forcing and associated sediment transport. Where sediments play a recognizable role, estuaries almost never attain true hydrodynamic/sedimentary equilibrium. Usually there is a quasi-equilibrium characterized by long-term changes in the bottom bathymetry. An important issue posed by potential effects of sea level rise pertains to our ability to predict various facets of estuarine response to sea level rise, including shoreline configuration, bottom sedimentation and marsh development/degradation.

10.2 SHORELINE CONFIGURATION

Much of the knowledge of shoreline changes is based on geological evidence which has been used to develop scenarios for estuarine formation, development and eventual demise. Two edited volumes by Schwartz, Spits and Bars (1972) and Barrier Islands (1973), are collections of important papers in the subject area. Spits and Bars covers seventeen papers, from 1890 to 1971. Barrier Islands covers forty papers, from 1845 to 1972.

More recent set of papers (also of geological nature) edited by Leatherman (1979) shows that there is new emphasis on descriptive modeling of barrier island and inlet morphologic changes, much of it based on holocene shoreline recession evidence. Reference must also be made to a series of papers recently edited by Nummedal et al. (1987) on shoreline response to sea level change.

Our understanding of processes at a particular site, _post facto_, has improved vastly. However, prediction of future shorelines, in any given situation, at best remains a hazardous guess except in a few well defined situations. In many cases therefore, simple modeling based on inundation and the Bruun Rule are used for predicting long-term trends (Kana _et al._, 1984). The problem seems to lie with the fact that our ability to predict sediment motion along the coastline under a changing wave climate on a long-term basis remains rudimentary, and is in fact the subject of major ongoing research effort in coastal engineering. It may take an additional decade or two before confidence in long-term prediction reaches acceptable levels. A part of the difficulty is not so much with relating hydrodynamic forcing to sediment motion, as with synoptic data necessary to obtain reliable (hindcast) hydrodynamic (currents, waves, winds) information. Furthermore, structures interrupt sediment motion. Modeling of sediment-structure interaction is still under research, although some useful modeling work has been done in this context (Kraus, 1983).

In the subject of estuarine mouth or inlet response to changing oceanographic conditions, much more work has been done on sandy inlets than on inlets where the material is fine-grained. In general, simplified description of inlet response can be examined by considering inlet hydraulics as characterized by the repletion coefficient concept (Keulegan, 1967) and inlet sedimentary response via O'Brien's (1969) equilibrium relationship for inlet stability.

The repletion coefficient, K, is defined as

$$K = \frac{A_c}{a_o \sigma A_b} \left(\frac{2ga_o}{F}\right)^{1/2} \tag{10.1}$$

where A_c = inlet flow area, A_b = bay area, σ = tidal frequency, a_o = tidal amplitude in the sea, g = acceleration due to gravity and F = impedance. F accounts for bottom resistance in the channel as well as head losses associated with flow entrance and exit (Keulegan, 1967; Mehta and Özsoy, 1978). Sea level rise will influence several terms, including A_c (increase), A_b (increase, unless bay is bounded by vertical walls), a_o (increase) and F (decrease). Typically, the overall effect will be an increase in K, which means easier flow admittance or better repletion.

The O'Brien relationship between the spring tidal prism, P, and A_c is

$$A_c = aP^m \tag{10.2}$$

where a and m are empirical coefficients which vary somewhat with the prevailing wave climate and other local conditions. The significance of this relationship is that it implies inlet widening and/or deepening with prism. Sea level rise in most cases will increase the prism (greater repletion), to which the inlet flow area will respond likewise. Equation 10.2 is particularly applicable to sandy inlets, as described in section 6. Increasing prism means greater sand flushing ability of the channel. The sand is transported by higher currents out of the channel, both bayward and seaward, to flood and ebb shoals, respectively. With increasing prism, there is likely to be a corresponding increase in the volume of these shoals. Furthermore, as the sea level rises the deltas must grow in elevation to keep up with the rise, implying that any natural bypassing of sand would reduce and that downdrift erosion would increase.

Stabilized inlets will be affected strongly by a large sea level rise. The protective jetties, which retard the ability of the littoral drift to enter the navigational channel and reduce the wave climate in the channel, would become less effective as they are submerged. Also, the stability of the jetties may be reduced due to increased wave heights as a result of sea level rise (National Research Council, 1987b).

The sea level rise scenario imposes a much more gradual change in the inlet/bay system than, for example, channel dredging. There should be enough time for the system to keep pace with water level rise, with the attainment of quasi-equilibrium under the prevailing hydrodynamic forces and sediment movement. Equilibrium is determined jointly by hydraulic conditions characterized by the repletion coefficient (Eq. 10.1) and by sedimentary requirement as per Eq. 10.2 (O'Brien and Dean, 1972).

It is also self-evident that shoreline response as far as inlets are concerned is contingent upon the availability or lack of sediment supply. For example, the barrier islands of the Mississippi-Alabama coast have been migrating and "disappearing" due to lack of sediment supply (Otvos, 1979). See also Fig. 6.1. Over the past ~ 100 years, this factor seems to have been far more important than (absolute) sea level rise.

The East Frisian Islands in Germany illustrate trends of an opposite nature. These islands and the corresponding inlets are believed to have been in existence for the past few millennia (Luck, 1976). Unlike the Mississippi barriers, however the islands have accumulated additional sediment from the West Frisian Islands. Fig. 10.1 shows three of these islands (and four inlets) - Norderney (N), Baltrum (B) and Langeoog (L), plus ends of two - Juist (J) and Spiekeroog (S). It is observed that between 1750 and 1960, the barrier as a whole gained sediment, resulting in greater land area above mean high water (MHW). To some extent, these developments have been influenced by coastal protection works at the eastern ends of Norderney and Baltrum (Kunz, 1987). Luck (1976) has accumulated map-based evidence on the changes at East Frisian Islands and has provided a qualitative explanation for the observed historic behavior. Unfortunately, present day technology precludes the possibility of predicting changes in island chain configuration over even the next 10 years.

10.3 ESTUARINE SEDIMENTATION

It is common for inlets to have sandy beds and upland estuarine waters to consist of fine-grained material. A relatively rapid sea level rise could cause a correspondingly rapid landward migration of the sandy barrier island, thus exposing back barrier fine-grained deposits to open coastal wave action (Everts, 1987). In this way, a marine source of fine-grained sediment is created. Some low coasts lack barriers, for example west Florida north of Tarpon Springs. Barriers have not developed here because of the lack of sandy material, which in turn is due to the limestone formations through which the coastal rivers drain (Shepard and Wanless, 1971). Here again there is an offshore source of fine-grained sediment which generates coastal turbidity mainly during storms.

The significance of a marine source of fine-grained sediment is that this material tends to enter the estuary with the upstream salinity-driven residual current, which typically occurs in the mixing zone of the estuary (i.e. the region where sea water mixes with fresh water). With sea level rise and barrier migration, fine-grained sediment from the sea would become available as a sediment source, stirred up by wave action and transported by the residual current, depositing ultimately at the upland end of the mixing

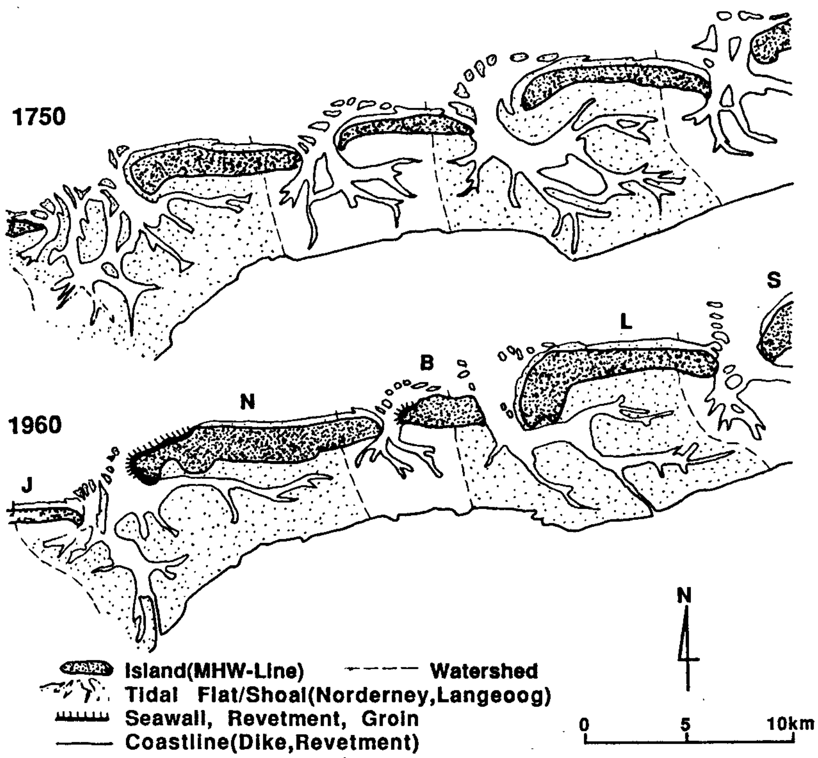

Fig. 10.1. East Frisian Islands in 1750 and in 1960 (after Kunz, 1987).

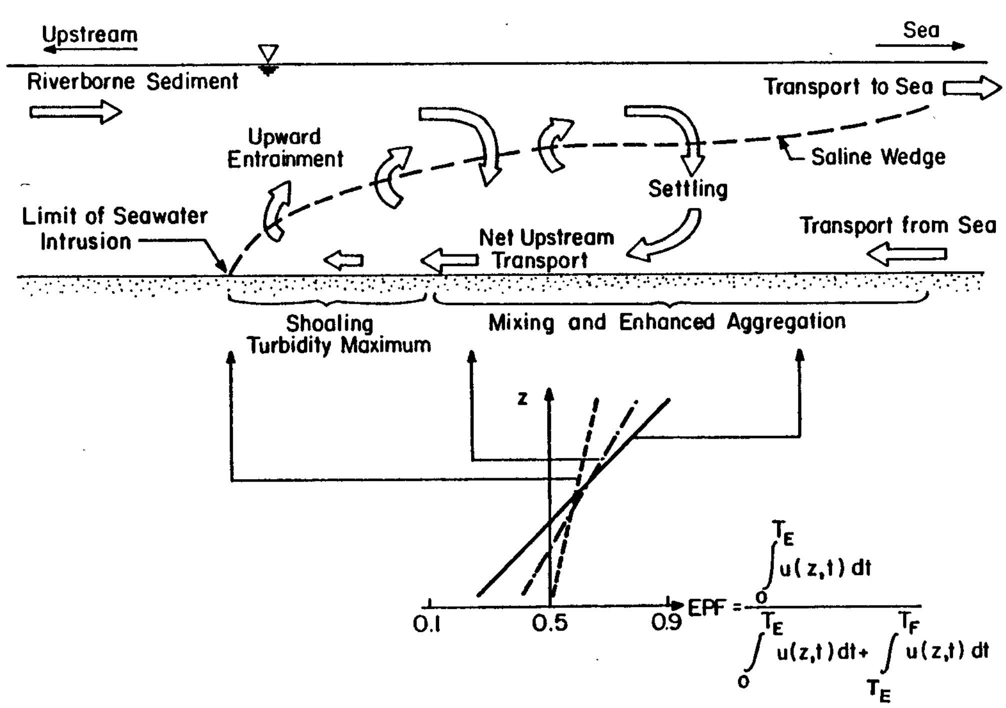

Fig. 10.2. Sediment Transport in the Estuarine Mixing Zone.

zone. In some cases, e.g. in the long navigation channel through the Maracaibo estuary in Venezuela (Partheniades, 1971) and in Zeebrugge harbor, Belgium[4], the present source of fine sediment is marine rather than alluvial.

Fine sediment transport in estuaries is a complex process involving a strong coupling between tides, fresh water flow and the coagulated sediment. This process has been described extensively elsewhere (Postma, 1967; Krone, 1972). In Fig. 10.2, a schematic description is given. The case considered is one in which the estuary is stratified, and a stationary saline wedge is formed. Various phases of suspended fine sediment transport are shown on a tide-averaged basis. In a partially mixed estuary, the description must be modified, but since relatively steep vertical density gradients are usually present even in this case, the sediment transport processes are qualitatively similar.

With reference to this figure, the vertical variation of the horizontal residual flows (tide-averaged flows) can be conveniently described by computing the ebb predominance factor, EPF, defined as

$$EPF = \frac{\int_0^{T_E} u(z,t)dt}{\int_0^{T_E} u(z,t)dt + \int_{T_E}^{T_F} u(z,t)dt} \tag{10.3}$$

where $u(z,t)$ = instantaneous longitudinal current velocity at an elevation z above the bed, T_E = ebb period, T_F = flood period and $T = T_E + T_F$, where T = tidal period (Simmons, 1965). If the strengths of flood and ebb were the same throughout the water column, EPF would be equal to 0.5 over the entire depth of flow. This is almost never the case; EPF is usually less than 0.5 near the bottom, particularly in the saline wedge, and greater than 0.5 in the upper layers. The residual upstream bottom current is due to the characteristic nature of flow circulation induced by the presence of the wedge, which means that the strength of this current will decrease as the limit of seawater intrusion is approached, and is theoretically zero at the limit (node)

[4]Robert Kirby, Ravensrodd Consultants Ltd., Taunton, United Kingdom, personal communication.

itself. Distributions of EPF at three locations - at the mouth, in the wedge and at the node, would qualitatively appear as shown in Fig. 10.2. When interpreted in terms of tidal flows, these distibutions reflect the general observation that in the mixing zone of the estuary flood flows landward at the bottom and ebb flows seaward at the surface.

The trends indicated by the EPF distibutions suggest the dominating influence of hydrodynamics on sediment movement. As noted in Fig. 10.2, riverborne (alluvial) sediments from upstream fresh water sources arrive in suspension in the mixing zone. The comparatively high degree of turbulence, associated shearing rates and the increasingly saline waters will cause sediment aggregates to grow in size as a result of frequent inter-particle collisions and cohesion, and large aggregates will settle. Aggregate settling velocities can be up to four orders of magnitude larger than the settling velocities of the elementary particles. Some of the sediment will deposit onto the bed, and some will be carried upstream near the bottom until times close to slack water when the bed shear stresses decrease sufficiently to permit deposition. The deposited sediment will start to consolidate due to overburden.

The depth to which the new deposit scours when the currents increase after slack will depend on the bed shear stresses imposed by the flow and the shear strength of the deposit. If the currents during both flood and ebb are sufficient to scour all of the new deposit, the net movement will be determined approximately by current predominance. However, if the bed shear stress during ebb is less than sufficient to suspend all of the newly deposited material, a portion of the material will remain on the bed during ebb, and will be resuspended and transported during the predomiant flood flows, resulting in a net upstream transport. Net deposition, i.e. shoaling, will occur when the bed shear during flood, as well as during ebb, is insufficient to resuspend all of the material deposited during preceding slack periods. Some of the fine material that is resuspended will be re-entrained throughout most of the length of the mixing zone to levels above the salt water-fresh water interface and will be transported downstream to form larger aggregates once again, and these will settle as before. At the seaward end some material may be transported out of the system. A portion or all of which could ultimately return with the net upstream current.

Sediment moving upstream along the bottom may also be derived from marine sources, as noted. The strength of this upstream current is often enhanced by the inequality between the flood and the ebb flows induced by the usually observed distortion of the tidal wave. Inasmuch as the low water depth is often significantly less than the depth at high water, the speed of the propagating tidal wave is higher at high water than at low water. This typically results in a higher peak flood velocity than peak ebb velocity and a shorter flood period than ebb period. Such a situation tends to enhance the strength of the upstream bottom current, and the sediment is sometimes transported to regions upstream of the limit of seawater intrusion.

The estuarine sedimentary regime is characterized by several periodic (or quasi-periodic) time-scales. These are: a) the tidal period (diurnal, semi-diurnal, or mixed), b) lunar (spring-neap) cycle, c) yearly cycle, and d) periods greater than a year. Of these, the first is the fundamental period which characterizes the basic mode of the sediment transport phenomenon in an estuary. The second is important from the point of view of determining net shoaling rates in many cases of engineering interest, and by the same token the third and the fourth time-scales are involved in considerations of long-term stability and shoaling in estuaries, as for example due to sea level rise.

Predictive capability for estuarine sedimentation can be illustrated by considering two case studies, Atchafalaya Bay, Louisiana, and Savannah River estuary, Georgia. The Atchafalaya River, a distributary of the Mississippi River, discharges into this bay. In recent years, the delta at the mouth of the river has grown dramatically. A study of the bay and adjacent waters was carried out to predict the rate at which the delta will evolve in the short term (<10 years) and the long-term (50 years), and the manner by which that evolution will affect flood stages, navigation channel shoaling, and the environmental resources of the area (McAnally *et al.*, 1985).

Several factors combined to make the Atchafalaya Bay study unusually complex. They included the long period over which predictions had to be made; the migration of the region of delta growth from lacustrine to estuarine to marine environments; a hydrodynamic regime that is variously dominated by river flows, wind-induced currents, tides, waves and storm surges; and the combined deposition of sediments from the sand, silt and clay classes. The investigation included several separate prediction techniques, including: 1) extrapolation of observed bathymetric changes into the future, 2) a "generic" analysis that predicted future delta growth by constructing an analogy between the Atchafalaya delta and other deltas in similar environments, 3) quasi two-dimensional numerical modeling of hydrodynamics and sedimentation, and 4) use of extensive field measurement of water levels, currents and sediments, and laboratory experiments on sediment samples.

Results have shown a wide range of possible land growth rates for the next 50 years in Atchafalaya Bay. Important results illustrated in Fig. 10.3 indicate sensitivity of delta growth to the subsidence rate (McAnally et al., 1984). These high subsidence rates are caused in part by compaction of thick layers of fine sediments that have been deposited by the Mississippi River and its distributaries over thousands of years.

The extrapolation results shown in Fig. 10.3 were generated by establishing a relationship between past delta growth and forcing phenomena of river flow and sediment supply, then using that relationship in combination with historically recorded flows to project future delta growth. It did not explicitly include subsidence effects. The generic approach assumed that Atchafalaya delta growth could be considered analogous to other deltas growing under similar conditions and matched observed Atchafalaya delta growth with a generic delta growth and decay time history such as shown in Fig. 10.3. The non-dimensionalized generic curve was then taken to represent a possible future growth and decay cycle for the Atchafalaya. Two results are shown--generic least squares fit the generic curve to observed Atchafalaya growth in a least squares approach; generic observed forced the generic curve to pass through the observed 1980 Atchafalaya delta land area. The quasi-two dimensional approach employed a one-dimensional numerical model with two subsidence rates to establish a range of possible results and to define sensitivity to subsidence.

Prediction of bed movement due to sediment erosion-deposition was carried out by the application of the two-dimensional, depth-averaged numerical model in Savannah Harbor, Georgia (Ariathurai et al., 1977). The reach of the estuary under investigation was 6,400 m in length, between stations 1 and 3 (Fig. 10.4). The turning basin, near station 2, was the region of heaviest shoaling. Flow and sedimentary data were collected at the three stations

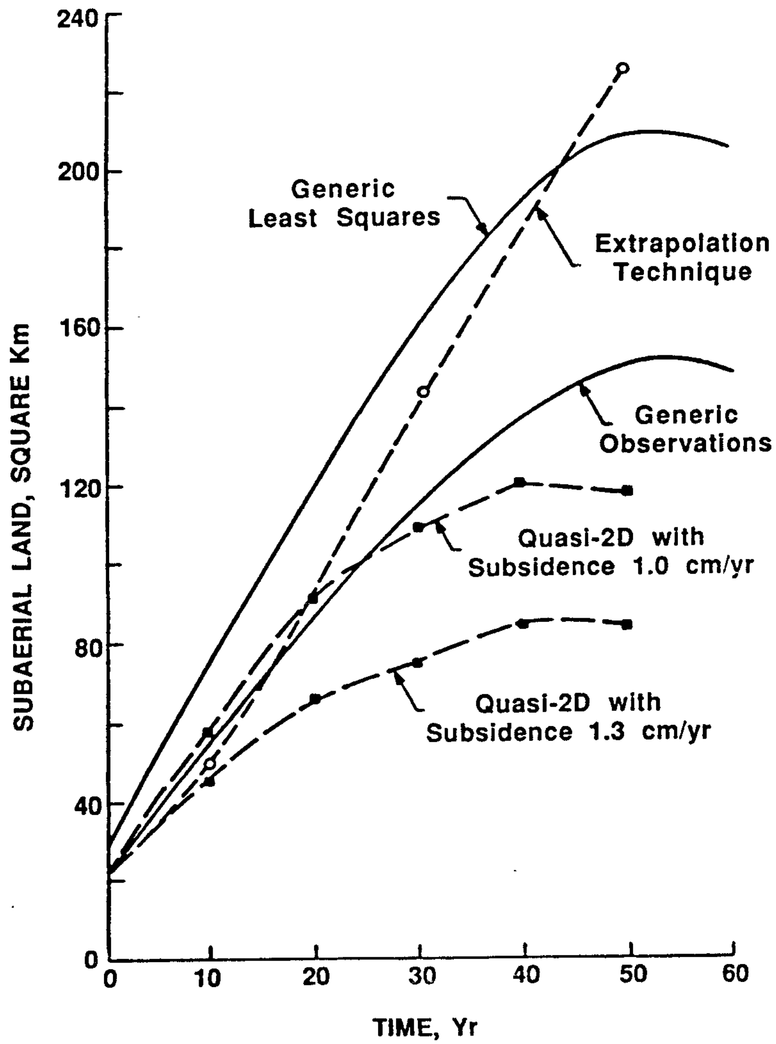

Fig. 10.3. Time-Rate of Subaerial Land Growth in Atchafalaya Bay, Louisiana, Calculated by Different Approaches (after McAnally et al., 1984).

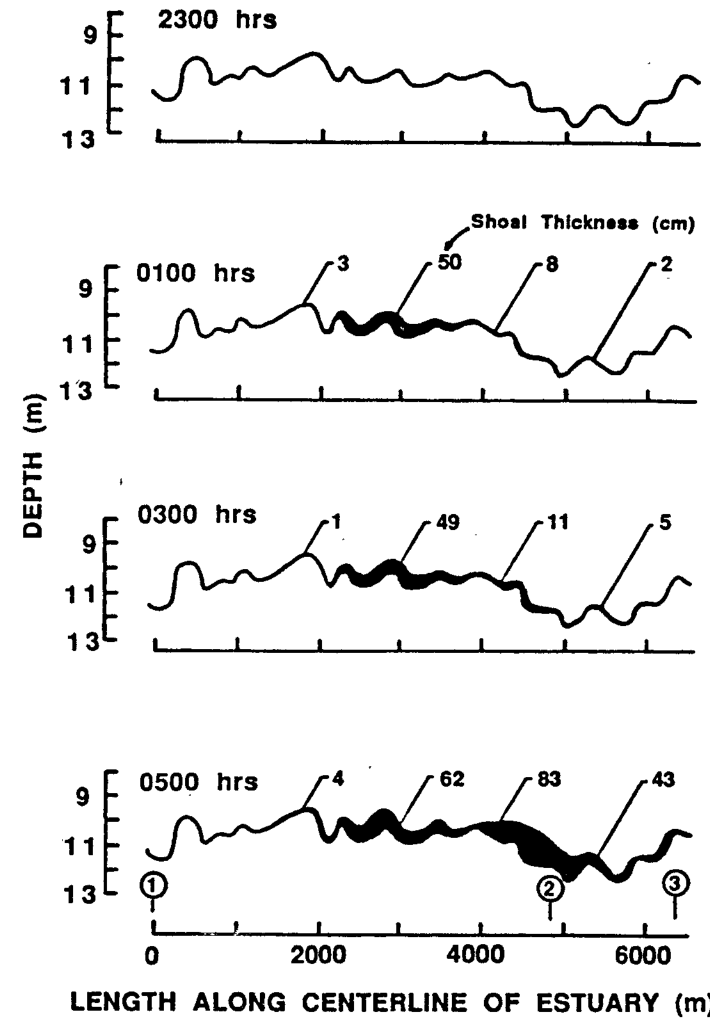

Fig. 10.4. Time-History of Bottom Sediment Movement in Savannah Harbor Estuary, Georgia (after Ariathurai, et al., 1977).

previously. Given the flow field (from measurements), the model was verified against measured suspension concentrations. Fig. 10.4 shows predicted bottom evolution with tide for the period 2300 hr on September 24, 1968, to 0500 hr on the following day. The semi-diurnal spring range of tide during this period was 2.6 m. The initial condition corresponds to the situation at 2300 hr when the ebb current was decreasing but there was no measurable deposition. The bottom topography shown in the figure at that time may therefore be considered to represent the bed which was not scoured by the prevailing current. As observed, heavy deposition occurred by 0500 hr, at slack water following flood, particularly in the region of the turning basin. Measured deposit thickness on the order of 1 m there agreed with model result. Modeling effort such as this one can be used to generate predictive scenarios for sedimentation patterns provided the expected hydrodynamic and sedimentary boundary conditions are well known.

10.4 WETLAND RESPONSE

Shallow bays surrounded by extensive wetlands will expand rapidly in response to a rise both because of the gentle slope and the deterioration of the marshes in response to salinity increases. For example, Barataria Bay, Louisiana, has increased its surface area about 10 to 15 percent over the last century in response to about 1 m of local relative sea level rise (National Research Council, 1987b). In general, however, although wetlands are critically important as a buffer against shoreline erosion, their response to sea level change is complex and not yet fully understood in the quantitative sense.

Wetlands account for most of the land less than 1 m above sea level. These extensive marshes, swamps, and mangrove forests fringe most of the U.S. coastline, particularly along the Atlantic and Gulf coasts. Their estimated original extent in the United States was about 2.0×10^4 km^2 (Hoese, 1967). This area has been significantly reduced through a variety of actions, including an early widespread practice of filling marshlands in urban areas. Wetlands loss has also been caused by other human actions, such as the construction of canals and waterways and the diversion of fluvial sediment to the offshore (National Research Council, 1987b).

Ecological conditions in coastal marshes range from marine to nearly terrestrial. A change in controlling factors, such as water salinity or tidal and wave energy, will cause a displacement in marsh zonation. Generally, coastal marshes are divided into low and high marsh based on their elevation relative to sea level (Redfield, 1972). Since marsh plants are attuned to particular mean water levels, a rise in sea level will shift the distribution of plant species proportionally landward. Beyond this response to variation in relative sea level, however, a more complex set of responses may occur, tied to the type of marsh considered. Thus, anticipated changes in coastal marshes must be assessed within the context of the basic marsh types that characterize the coasts. With respect to the future effects of a rise in sea level, coastal marshes may be broadly divided into backbarrier marshes, estuarine (brackish) marshes, and tidal freshwater marshes (National Research Council, 1987b).

Backbarrier marshes occur along the bayward sides of barrier systems of the Atlantic and Gulf coasts. Studies (e.g. Zaremba and Leatherman, 1986) show that these marshes are formed and destroyed rapidly in such dynamic environments. Maintenance of these marshes therefore appears to be more a function of barrier stability than of the pace of upward growth of the marsh surface, since sediment supplies are ample (Letzch and Frey, 1980). For barriers rapidly migrating landward, there may be a net decline in backbarrier marshes. This has been found to be the case at Assateague Island, Maryland, where sediment blockage by jetties has greatly increased the rate of landward barrier migration (Leatherman, 1983), and the same qualitative result could be anticipated as a result of accelerated sea level rise (National Research Council, 1987b).

Estuarine marshes embrace a wide variety of vegetative species in diverse geologic settings where salinities are less than approximately 30 ppt. These marshes, comprising integral components of major estuarine systems such as the Chesapeake Bay, occur in areas of quiescent waters and ample sediment supply. Accretionary budgets differ widely, as seen from Table 10.1 (Stevenson et al., 1986).

The data of Table 10.1 are plotted in Fig. 10.5 in terms of mean marsh accretion rate against relative sea level rise rate, both in mm/yr. It is noteworthy that with the exception of the three data points from Louisiana and

Table 10.1. Rates of Marsh Accretion and Relative Sea Level Rise (adapted from Stevenson et al., 1986)

Location	Mean tidal range (m)	Relative sea level rise (mm/yr)	Salinity (ppt)	Accretion rate	
				Range (mm/yr)	Mean (mm/yr)
Barnstable, MA	2.9	0.9	20-30	3-8	5.5
Prudence Is., RI	1.1	1.9	28-32	2.8-5.8	4.3
Farm River, CN	1.8	1.9	-[a]	-[a]	5.0
Fresh P., NY	2.0	2.2	26	-[a]	4.3
Flax P., NY	2.0	2.2	26	4.7-6.3	5.5
Lewes, DE	1.3	2.0	25-30	-[a]	4.7(>10)[b]
Nanticoke, MD	0.7	3.2	2-6	4.9-7.2	6.1
Blackwater, MD	0.3	3.9	1-5	1.7-3.6	2.6
North R., NC	0.9	1.9	-[a]	2-4	3.0
North Inlet, SC	1.6	2.2	30-35	1.4-4.5	2.5
Savannah P., GA	3.0	2.5	-[a]	-[a]	11.0
Sapelo Is., GA	2.1	2.5	30-35	3.5	4.0
Barataria, LA[c]	0.5	9.5	<1->15	5.9-14.0	7.2
Fourleague, LA	0.3	8.5	10-20	-[a]	6.6
L.Calcacieu, LA	0.6	9.5	15	6.7-10.2	7.8

[a]Not reported.
[b]Lower value obtained by dating with Lead-210, higher with Cesium-137.
[c]Values based on fresh, brackish, intertidal and water marshes.

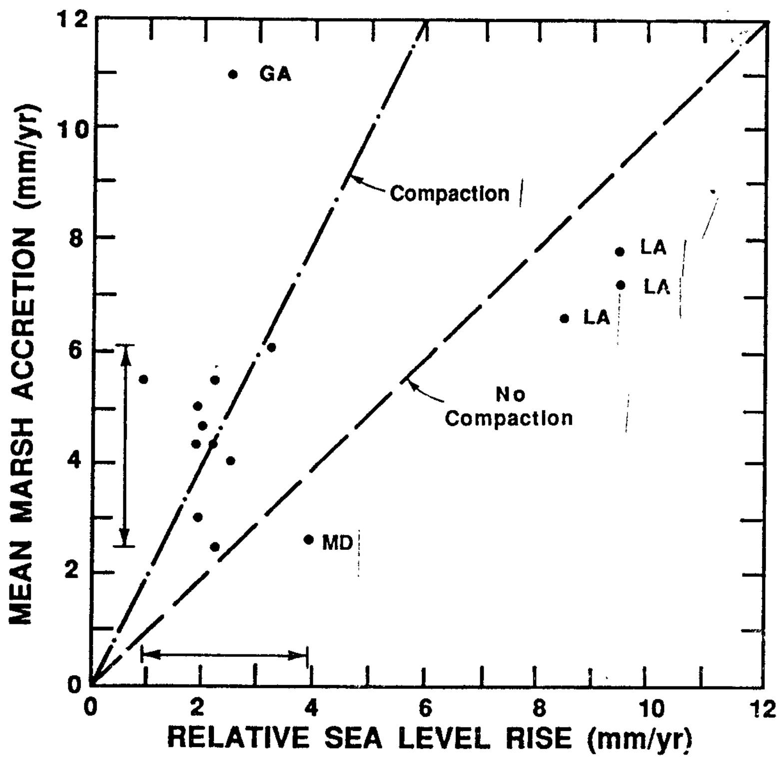

Fig. 10.5. Relationship between Sea Level Rise and Marsh Level Rise Rates (based on data compiled by Stevenson et al., 1986).

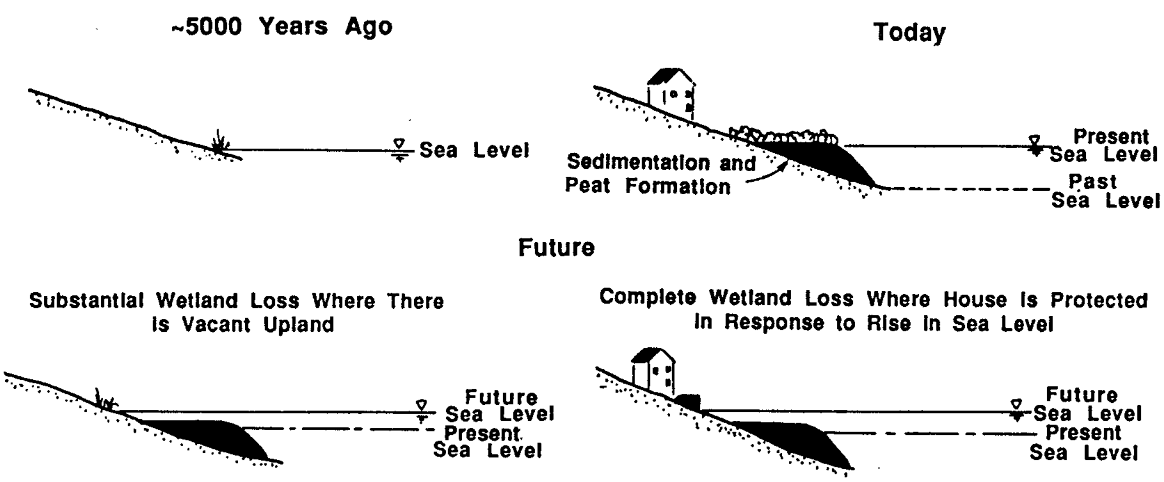

Fig. 10.6. Marsh Evolution with Sea level Rise (after Titus, 1986).

one in Georgia (Savannah R.) the remaining data points are confined within a relatively narrow domain bounded by a range of 0.9 to 3.9 mm/yr for sea level rise and 2.5 to 6.1 mm/yr for marsh accretion. The corresponding mean values are 2.3 mm/yr and 4.3 mm/yr, respectively. Thus overall marshes seem to have accreted at almost twice the rate of sea level rise.

The apparent discrepancy between sea level rise rate and accretion rate is most likely to be due to the effect of compaction. The rate of accretion can be thought of as the initial rate of sediment deposition, which depends on the ambient suspended sediment concentration and the sediment settling velocity. In the absence of compaction, and noting that usually in the long run marsh level can at most keep pace with sea level but not rise faster, the maximum marsh accretion rate must equal sea level rise rate. This is indicated by the 45° dashed line (no compaction) in Fig. 10.5. In general, however, compaction cannot be ignored. Laboratory tests on the deposition of relatively thin fine-grained sediment beds (Dixit, 1982) show that the density of the initial deposit (dry sediment mass per unit volume) tends to increase from about 0.05 - 0.1 g/cm^3 to between 0.2 - 0.3 g/cm^3 after a few days of relatively rapid consolidation. Beyond this period, further increase in density is very slow. Although there would be a significant difference between these test results and marsh compaction in the field, it is worthwhile to examine the implications of compaction based on the laboratory evidence, qualitatively. Thus, if we assume a two-fold increase in density, the line shown in Fig. 10.5 (compaction) will result. Several data points tend to corroborate the observed linear trend. It can thus be surmised that compaction effects are significant in marshes.

Tidal freshwater marshes are located in the upper reaches of estuaries and other areas where ambient salinities are less than about 5 ppt. The effects of rising sea levels will be saltwater intrusion and the eventual dominance of higher salt-tolerant plants. However, the effects of canalization on tidal freshwater marshes in the Mississippi delta demonstrate that dramatic increases in salinity over a comparatively short period exceed the capability of these marshes to adjust so that rapid losses ensue (National Research Council, 1987b).

The complexity of marsh response to water level and associated factors has meant that predictive modeling effort has been limited. Most modeling to-date has been of descriptive nature, as illustrated in Fig. 10.6 (Titus, 1986). Coastal marshes have kept pace with the slow rate of sea level rise that has characterized the last several thousand years. Thus, the area of marsh expanded over time as new lands were inundated. If in the future, sea level rises faster than the ability of the marsh to keep pace, the marsh area will contract. Construction of bulkheads along lagoonal banks may prevent new marsh from forming and result in a total loss of marsh in some areas.

Krone (1985) used a simple deposition model to predict the response of marshes to sea level rise in the San Francisco Bay area. As water leaves a channel and flows onto the marsh surface during a rising tide, the concentration, C, of suspended sediment diminishes as sediment aggregates settle to the marsh surface. As new sediment-laden water from the channel mixes with the previously flooding water, however, the added sediment increases the concentration. This process continues until the tide reaches its maximum elevation after which only deposition occurs while the water drains from the marsh. A mass balance of this process leads to

$$(y_w - y_m)\frac{dC}{dt} + W_s C - C_o\frac{dy_w}{dt} = 0 \qquad (10.4)$$

where y_w and y_m are the elevations of the water surface and marsh surface relative to a selected datum, respectively, W_s is the settling velocity of the suspended aggregates, C_o is the concentration of suspended solids in the flooding waters, and t is time. $C_o = C_o$ during a rising tide, and $C_o = 0$ during a falling tide.

Laboratory tests on San Francisco Bay muds showed that the median settling velocity (by weight) in cm/s of the suspended aggregates is described by

$$W_s = kC^{4/3} \qquad (10.5)$$

where k was found to be 110 when the concentration is in g/cm^3. Equations 10.4 and 10.5 are combined to solve for concentration C through a finite difference scheme, and from that the rate of growth of marsh elevation, y_m, given the density of the marsh soil. The value of the ambient concentration in the channel, C_o, was obtained by calibrating the calculated y_m change against measurement of the same at specific sites within the bay.

As would be expected, the rate of rise of marsh elevation is strongly dependent on C_o, as demonstrated by computations for two values of C_o in Fig. 10.7. The computations begin with an initial marsh level (0.15 m) in 1930 when a levee was removed to allow tidal waters to flood the marsh area. The figure shows a more rapid rate of rise during the early period, when the marsh was most frequently flooded, and a slowing rate as a steady rate of rise was approached.

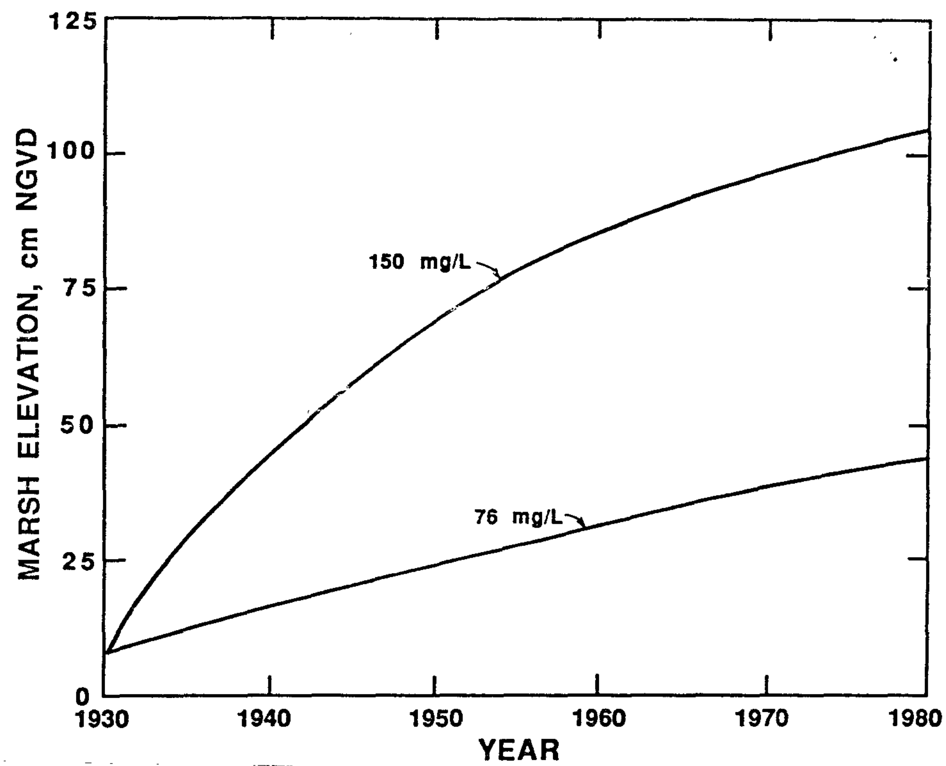

Fig. 10.7. Effect of Suspension Concentration on Marsh Elevation Rise and Sea Level (after Krone, 1985).

10.5 RESEARCH NEEDS

Present day capability in predicting the evolution of the morphology of the estuarine mouth and adjacent shorelines due to sea level rise is limited. So is our ability to quantitatively evaluate wetland response to sea level rise. Predictive capabilities for sedimentation within the estuary are better, although in areas where high suspension concentrations occur, the physics is poorly understood. Furthermore, the precise nature of chemical and biological variability is not clearly known; hence we are not in a position to establish the quantitative significance of physical versus chemical versus biological control in estuarine sedimentation processes.

As with most other areas, the basic research issues are not particularly related to sea level change, but with the fundamentals of physics, chemistry and biology, as they interact and influence estuarine processes. Advances in knowledge have been diluted by too much site-specific and empirical work. A part of the problem appears to lie with the fact that a great deal of effort has been directed to address specific problems unrelated to basic questions posed by scientists.

In recent years several scientific reports have identified relevant research areas which must be tackled for ultimately improving our predictive capabilities (e.g. North Carolina Sea Grant, "Geophysics in the Environment"; MSRC's "Transport of Fine-grained Sediment"; C. Officer's book; Chesapeake Bay's program; NOAA's estuary program plan)[5]. Greater attention needs to be paid to research requirements outlined in these reports. Thus, for example, in a recent report issued by the National Research Council (1987a), research areas in fine sediment transport have been clearly identifed.

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[5]Henry Bokuniewicz, Marine Sciences Research Center, SUNY, Stony Brook, New York, personal communication.

11. COASTAL ECOSYSTEMS

11.1 INTRODUCTION

As sea level continues to rise, perhaps at an anthropogenically accelerated rate, existing coastal ecosystems will become submerged and saline water will move inland. Furthermore, tidal amplitude and wave energy may increase due to submergence of protective nearshore reefs and sandbars and other effects (see sections 4 and 5). Water level, water motion, and salinity are principal determinants of the type, nature, and function of coastal ecological systems. Therefore, changes caused by rising sea level could cause dramatic changes in coastal ecosystems.

Coastal ecosystems include open water systems, submerged benthic (bottom) systems, and intertidal systems. Open water systems consist of plankton (suspended organisms transported by the current), neuston (organisms dwelling in or near the surface film), and nekton (actively swimming organisms). Benthic systems include seagrass ecosystems and ecosystems of unvegetated sediments. Mud and sandflats that are periodically exposed, and marshes and mangroves are common intertidal systems. Each of these coastal systems is valued for its contribution of food and cover to the production of a diversity of living coastal resources (Haines, 1979; Peterson, 1981; Zieman, 1982; Odum *et al.*, 1982; Boesch and Turner, 1984; Seaman, 1985). It is likely that a mix of coastal habitats is more important to these resources than any one system alone.

Ecological production (i.e. primary production) in coastal zones is equal to or greater than that obtained with the best mechanized agriculture, yet without the subsidies of plows and chemicals (Odum, 1971). Coastal zones have been said to have natural energy subsidies that together with sunlight account for this very high level of production, namely, the water movement caused by tides, winds, and freshwater discharge (Schelske and Odum, 1961; Odum, 1980).

11.2 ECOSYSTEM RESPONSE

As sea level rises, open water and submerged benthic communities may not be as productive as they are today in relatively clearwater coastal zones. Increases in tidal range and wave energy may cause an increase in turbidity from suspended sediments. Increased turbidity will reduce the growth rate of light-limited seagrasses (Orth and Moore, 1984; Keesecker, 1986). Intertidal marsh plants, however, may grow better as a result of greater energy subsidies from greater water movement.

The organisms that occupy the three general types of coastal habitat (open water, submerged bottom, and intertidal zone) differ depending on average conditions of salinity, water level, light, temperature, dissolved oxygen, waves, and current (Remane and Schlieper, 1971). Gradients of these factors intersect in coastal zones to give a wide variety of microhabitats. Rising sea level should reposition these gradients (Browder and Moore, 1981).

Therefore, at any one location, a succession of ecosystems could co-occur with the changes caused by rising sea level. These changes would depend on the rate of sea level rise and the rate of change of marsh level due to sedimentation or erosion. In intertidal freshwater zones, encroaching salinity will kill bottomland hardwood forests (cypress and gums) and freshwater marsh plants (Odum et al., 1984). These will be replaced by brackish-water grasses and shrubs (Stout, 1984), such as sawgrass (Cladium jamaicense), giant cutgrass (Zizaniopsis miliaceae), giant cordgrass (Spartina cynosuroides), and saltmarsh bullrush (Scirpus robustus). These brackish plants will then be replaced by more saline species of salt marsh plants (Teal, 1986), such as saltmarsh cordgrass (Spartina alterniflora) and black needle rush (Juncus roemerianus). In south Florida, mangroves (Rhizophora mangle, Avicennia germinans, and Laguncularia racemosa) should replace brackish grasses (Odum et al., 1982). Salt marshes and mangroves will be replaced by intertidal mud and sand flats and then by open water over submerged benthic ecosystems (Peterson and Peterson, 1979). If light can penetrate to the bottom and currents are not too erosive, seagrasses will probably develop on submerged sediments.

To survive, organisms must be able to tolerate or avoid unfavorable conditions. Frequency, amplitude, and regularity (predictability) of fluctuations influence the types of sedentary organisms (e.g. vascular plants, oysters and clams), as well as the ability of vagile organisms (e.g., crabs, fish, and shrimp) to use these habitats. Temporal variation in the coastal environment is most often caused by tides, storms, and seasons. Few species are able to withstand the simultaneous fluctuations in salinity, oxygen, and temperature that occur in many estuaries (Deaton and Greenberg, 1986), though those few that are well-adapted may be able to be very productive because of energy subsidies. Rapid changes in environmental conditions, however, can prevent the full development of an ecosystem. If conditions continually change before any one group of organisms becomes established, then ecological production may be restrained (Montague et al., unpubl. manuscr.).

Although rising sea level will cause a succession of ecosystems, the rate of rise is probably not rapid enough to prevent full ecosystem development in coastal zones. The seeds and larvae of plants and animals that occur at different points along environmental gradients are widely distributed in estuarine waters. Marshes can become established within one to five years following a sudden appearance of a favorable environment (Montague, unpubl. data). Mangroves may take at most 15 to 25 years to become fully developed (Odum et al., 1982). The rate of sea level rise may or may not be gradual enough to create a significant long-term lag time in the development of successive ecosystems.

Despite the possibility of a timely replacement of coastal ecosystems, whenever a large mass of existing organisms dies, short-term perturbations may accompany the transition to a new ecosystem. Dead plant matter, for example, may temporarily increase in coastal waters and sediment, formerly held in place by roots and rhizomes, may become unstable (Montague, 1986). The added dead matter will decompose, and so may reduce dissolved oxygen levels sufficiently to cause local fish kills. Destabilized sediment will increase turbidity in coastal waters, which will most likely reduce seagrass and phytoplankton production and may foul the filtering mechanisms of filter feeders such as clams and oysters.

11.3 RESEARCH NEEDS

The responses of some of the more prevalent coastal plants and animals to environmental changes are known. As predictions of the effects on sea level rise on ecologically important variables improve, better predictions of the nature and timing of ecological changes can be made using existing information. For the best predictions, however, new ecological and physiological information may also be required. The most important information needs can be identified with the aid of a literature synthesis and subsequent simulation model. The model will simulate ecological responses to predicted environmental changes (e.g., rising water level, encroaching salinity, increased tidal range and wave energy). Estimates of functions and parameter values will be based on the best available information, but some guesswork is anticipated. Uncertainty in ecological generalizations can be explored by a sensitivity analysis of the model. Needed information can be ranked according to a combination of the uncertainty involved in an estimate and the sensitivity of the model to changes in the estimate. Thus, not only can the most important information needs be identified, but also the consequences of a lack of this knowledge can be demonstrated with the model (Montague et al., 1982).

Although literature synthesis and exploratory models will facilitate the identification of specific research needs, the most relevant research will undoubtedly include several general areas. The time required for full development of subtidal, intertidal, and very nearshore supratidal ecosystems should be established with greater certainty. Under the most rapid sea level rise scenarios, conditions may not remain constant long enough for full development of an ecosystem. If so, the production of fish and shellfish and the stability of shorelines may decline.

Knowledge of the major regulators of the production of principal animals and plants is essential for coupling predictions of ecological changes to predictions of physical changes. Physiological factors that determine the type and productivities of organisms in the coastal zone include: light (turbidity), temperature, nutrients (including CO_2), salinity, water level, and biochemical oxygen demand (BOD). All of these will be influenced by sea level rise, global warming, and increased levels of atmospheric CO_2.

Physical uprooting and erosion of present ecosystems should be a major agent of ecological change. Predictions are needed both for shores and for tidal creeks. Knowledge of the resistance to erosion of these systems is also required.

The aerial extent of the intertidal zone and of sufficiently well-lit subtidal zone are of direct importance to the production of plants that provide food and cover to aquatic and nearshore animals. If coastal topography is steeper just inland from the present shoreline, then the aerial extent of the intertidal zone should be greatly reduced as sea level rises into the steeper areas. If turbidity increases, the aerial extent of well-lit subtidal zone will likewise be reduced. Both types of area may subsequently expand however, as sediment becomes trapped by intertidal vegetation. Knowledge of nearshore topography and predictions of tidal range are essential to predictions of aerial extent, but so too is an understanding of the level of suspended sediments to be expected and the trapping rate of sediment by subtidal and intertidal plants and microbes (Montague, 1986).

Intertidal marshes and mangroves have been highly touted as good habitat for the growth of juvenile fish and shellfish of commercial and recreational importance. In addition, exchange of materials between the marsh and the estuary is believed to control supplies of nutrients in adjacent estuarine waters. Not all marshes are equivalent in their habitat value, however, and not all exchange significant quantities of materials with surrounding waters (Montague et al. in press). Perhaps the most important factor in the accessibility of marshes to organisms, and in the exchange of materials, is the density of tidal creeks (Zale et al., 1987). The density of tidal creeks can be defined as the ratio of length of edge of tidal creeks to surface area of marsh. Knowledge of the influence of creek density on habitat utilization and material exchange may be essential to understanding the relative value of marshes that develop in response to sea level rise. Comparative studies of the effects of creek density have never been reported, however.

Empirical understanding of the kinds and extent of ecosystems that may result from sea level rise can be enhanced considerably by three areas of study: first, by paleoecological analysis of cores from various coastal ecosystems (to assess response to past sea level rise); second, by analysis of ecological zonation along gradients of salinity and elevation, which should reflect kinds of ecosystems to be expected as salinity encroaches and water becomes deeper (Kurz and Wagner, 1957); and third, by greater knowledge of the variation of environmental conditions under which each major type of system can now exist. Detailed physiometric studies of coastal ecosystems are limited to a few areas, usually near marine research laboratories. Results are often extrapolated to other sites. A given set of predicted environmental conditions, however, may not match those of these few study sites. Each type of coastal system may exist in a much broader range of environments than is now documented, and gradual changes probably occur between system types. Greater regional knowledge of the variety of ecosystem types, and of the variety of environments that support the same ecosystem, will enhance the resolution of empirical ecological predictions.

12. SUMMARY OF RESEARCH NEEDS

In this section research needs are summarized in a tabular form. There is a table for each of the ten research topics described in sections 2 through 11. It should be pointed out that the research needs as discussed in these sections are broad-based comments in some cases, while the tables in this section list specific research issues which must be addressed for improving upon the state-of-the-art understanding.

A special issue ranking procedure is used to determine the importance of each research issue or task to CO_2 - climate interactions. The results of applying this procedure are presented in Tables 12.1-12.10. A brief explanation of each item follows.

Research Issue or Task

While "issue" refers to a broad research category, "task" is more specific research. At times it is difficult to distinguish a task from an issue. In general, however, a task may be addressed in a shorter period (e.g. 0-2 years) than as issue (e.g. 6-10 years).

Uncertainty: statistical

a) Coefficient of variation of available data (standard deviation divided by mean in percent)
   - (1) low $\leq$ 100%
   - (2) medium 100% - 200%
   - (3) high > 200%
   - (4) unknown

b) Sample size (does available data contain all possible values?)
   - (1) all
   - (2) < all
   - (3) > none
   - (4) unknown

c) Range (difference between largest value of data and smallest, divided by the mean in percent)

(1) < 50%

(2) 50 - 200%

(3) > 200%

(4) unknown

Note that items a, b and c are listed in the tables in that order.

Uncertainty: subjective

known - both direction and magnitude of an impact are known based on data

uncertain - direction is known but magnitude is unknown

unknown - data exist but neither direction nor magnitude can be determined

no data - no data exist on the topic

Time - the estimated time required to yield significant progress on a research issue

(5) 0-2 years

(4) 3-5

(3) 6-10

(2) 11-20

(1) > 20 or unknown

Cost - estimated cost per year to perform the research

(7) < $125 K

(6) $125-250 K

(5) $250-500 K

(4) $500-1000 K

(3) $1-10 M

(2) > 10 M

(1) unknown

Rate - the rate at which research results are expected to accrue

(3) flat

(2) linear

(1) step increasing

Impact - the degree to which resolving the research issue will independently lower the uncertainty of the resource response. Based on time of study and number of other related questions which must be answered.

Number of related topics required	time (years) 0-5	6-10	11-20	> 20
none	high	high	med.	unknown
1-3	high	med.	low	unknown
> 3	med.	low	low	unknown

Sensitivity - variability of resource issue in response to anomalies in climate and CO_2

(1) low

(2) medium

(3) high

(4) unknown

Data - status of available or required data for the research issue. Ranking (time, cost, effort) is from 1 (highest) to 6 (lowest) for large, medium and small amounts of data. Numbers given here are examples.

Status	Volume of Data Large (L)	Medium (M)	Small (S)
existing (E)	3	6	6
need to be developed (D)	1	2	3

Consensus - percentage of authors writing on the resource topic that identify this particular research issue or task as necessary to resolve the impact on the resource. The percentage indicated is in general not based on the number of papers cited in this report for any particular task area. In arriving at a percentage, we have exercised a subjective judgement based on: 1) our knowledge of research groups active in the area, 2) diversity of known opinions and 3) journal articles, conference proceedings and other publications in coastal engineering and related fields. The latter are well in excess of literature cited. Literature of the past five years has been given extra weighting in relation to older material.

Table 12.1. Estimates of Eustatic Sea Level Rise

Research Issue or Task	Uncertainty		Time	Cost	Rate	Impact	Sensitivity	Data	Consensus
	Statistical	Subjective							
Apply post glacial rebound estimates	2: 100% 2: < all 2: 150%	Known	5	6	2	high	3	E: M: 5 D: S: 3	70%
Develop improved understanding of noise in tide gage records	2: 150% 3: > none 2: 150%	Uncertain	4	6	2	high	3	E: S: 6 D: M: 2	40%
Improved methods of analysis using world-wide tide gage data	2: 150% 2: < all 2: 100%	Uncertain	5	7	2	high	3	E: L: 4 D: L: 3	60%
Compaction measurements	2: 150% 3: > none 2: 100%	Unknown	5	5	2	high	3	E: S: 5 D: L: 1	70%
New tide gage installations	2: 100% 3: > none 2: 100%	Unknown	3	4	2	med.	3	E: L: 4 D: L: 2	80%
Satellite altimetry (water level measurement)	2: 200% 4: unknown 4: unknown	No data	2	3	1	med.	3	E: S: 5 D: L: 2	70%

Table 12.2. Compaction Effects

Research Issue or Task	Uncertainty		Time	Cost	Rate	Impact	Sensitivity	Data	Consensus
	Statistical	Subjective							
Design and install compaction devices and tide gages	1: 50% 1: all 1: 50%	Known	5	5	3	high	1	D: S: 3	70%
Tie tide gages to satellites (water level measurement)	1: 50% 1: all 1: 50%	Known	4	6	1-2	high	1	E: S: 6 D: S: 3	80%
Compaction studies in coastal areas	1: 50% 2: < all 2: 100%	Known	4	5	2	med.	2	E: S: 6 D: M: 2	60%

Table 12.3. Tidal Range Effects

Research Issue or Task	Uncertainty		Time	Cost	Rate	Impact	Sensitivity	Data	Consensus
	Statistical	Subjective							
Basic analysis of water level data	2: 100% 1: all 2: 200%	Known	3	7	3	low	1	E: L: 3 D: L: 1	50%
Development of synoptic hydrographic (runoff) and mete-orologic (wind) data	2: 150% 2: < all 2: 150%	Uncertain	3	7	2	med.	1	E: L: 3 D: L: 1	50%
Circulation and mixing processes in ocean and estuaries	2: 200% 2: < all 2: 200%	Uncertain	2	4	2	med.	3	E: M: 3 D: M: 2	80%
Boundary layer processes (bottom dissipation, wave-current interaction)	2: 150% 2: < all 2: 150%	Uncertain	3	5	2	high	1	E: S: 3 D: S: 2	80%
Non-astronomical forcing (local, remote)	2: 200% 2: < all 2: 200%	Uncertain	3	6	2	med.	3	E: S: 3 D: M: 2	75%
Air-sea interaction (surface stress)	2: 150% 1: all 2: 150%	Uncertain	3	5	1	med.	3	E: M: 3 D: L: 2	30%
Numerical hydrodynamic modeling	2: 100% 1: all 2: 150%	Known	4	6	2	high	1	E: M: 3 D: M: 3	90%

Table 12.4. Storm Surge and Wind-Wave Response

Research Issue or Task	Uncertainty		Time	Cost	Rate	Impact	Sensitivity	Data	Consensus
	Statistical	Subjective							
Storm surge measurements	2: 200% 3: > none 2: 100%	Uncertain	4	5	1-2	high	1	E: S: 6 D: M: 2	100%
Wind-wave generation	2: 200% 2: < all 2: 200%	Uncertain	3	5	2	high	4	E: M: 6 D: M: 2	90%
Wind-wave damping	2: 150% 2: < all 2: 100%	Uncertain	4	6	2	med.	1	E: S: 6 D: M: 2	60%
Directionality of wind-wave spectra	2: 200% 2: < all 1: 50%	Uncertain	4	6	2	med.	4	E: L: 3 D: S: 3	70%

Table 12.5. Interaction with Natural Features and Constructed Works

Research Issue or Task	Uncertainty		Time	Cost	Rate	Impact	Sensitivity	Data	Consensus
	Statistical	Subjective							
Wave refraction/ diffraction modeling	1: 50% 2: < all 1: < 50%	Known	5	7	2	high	1	E: S: 6 D: M: 2	80%
Breaking wave studies	2: 120% 1: all 1: < 50%	Uncertain	4	6	2	high	1	E: L: 3 D: M: 2	90%
Undertow and longshore current modeling	1: 80% 2: < all 2: 100%	Known	4	6	2	high	1	E: S: 6 D: M: 2	95%
Sediment entrainment studies	2: 200% 3: > none 2: 200%	Uncertain	4	5	1	high	1	E: S: 6 D: M: 2	100%
Shoreline planform and profile response modeling	2: 150% 2: < all 2: 150%	Uncertain	4	6	3	med.	1	E: M: 6 D: M: 2	90%
Performance and costs analysis of response alternatives	2: 200% 3: > none 3: > 200%	Uncertain	4	6	3	med.	1	D: M: 2	90%

Table 12.6. Shoreline Response Modeling

Research Issue or Task	Uncertainty		Time	Cost	Rate	Impact	Sensitivity	Data	Consensus
	Statistical	Subjective							
Isolation and extraction of anthropogenic effects	2: 150% 3: > none 3: > 200%	Uncertain	5	7	2	high	3	E: S: 5 D: M: 2	70%
Regional correlation of shoreline and sea level change rates	2: 200% 3: > none 3: > 200%	Uncertain	5	7	2	high	3	E: L: 6 D: L: 3	40%
Quantification of cross-shore sediment budget components	4: unknown 4: unknown 4: unknown	Unknown	4	4	2	high	3	E: S: 5 D: L: 1	60%
Shoreline monument system for coastal states	2: 100% 3: > none 2: 150%	Unknown	4	4	3	high	2	E: S: 6 D: L: 2	50%
Obtain data from continental shelf	4: unknown 3: > none 4: unknown	Unknown	4	4	2	med.	3	E: S: 5 D: L: 2	65%
Conduct regional shoreline profiling using laser technology	2: 200% 4: unknown 4: unknown	Uncertain	3	4	1	high	3	E: S: 3 D: L: 3	70%

Table 12.7. Saltwater Intrusion

Research Issue or Task	Uncertainty		Time	Cost	Rate	Impact	Sensitivity	Data	Consensus
	Statistical	Subjective							
Develop and apply simple analytical models of sea level rise effects on groundwater	2: 100% 3: > none 2: 200%	Uncertain	5	7	2	high	3	E: M: 6 D: S: 3	70%
Evaluate reliability of numerical models	2: 200% 3: > none 2: 150%	Uncertain	4	4	2	med.	3	E: S: 5 D: L: 2	75%
Apply numerical models to intrusion prone areas, emphasing probable response options	3: > 200% 3: > none 2: 200%	Uncertain	3	4	2	med.	3	E: S: 6 D: L: 2	75%

Table 12.8. Upriver Saltwater Penetration

Research Issue or Task	Uncertainty		Time	Cost	Rate	Impact	Sensitivity	Data	Consensus
	Statistical	Subjective							
Mixing processes in estuaries (stratification, fronts)	2: 100% 2: < all 2: 150%	Known	3	4	2	high	2	E: S: 3 D: S: 3	90%
Hydrodynamics (wave-current interaction)	2: 150% 2: < all 2: 150%	Known	3	7	2	high	2	E: S: 3 D: S: 3	50%
Bottom boundary layer (bottom friction)	2: 150% 1: all 2: 150%	Known	4	6	2	high	1	E: S: 3 D: S: 3	80%
Numerical transport modeling	2: 100% 1: all 2: 100%	Known	4	7	2	med.	1	E: S: 3 D: S: 3	80%

Table 12.9. Sedimentary Processes in the Estuarine Region

Research Issue or Task	Uncertainty		Time	Cost	Rate	Impact	Sensitivity	Data	Consensus
	Statistical	Subjective							
Nearshore wave-current-sediment interaction	2: 200% 2: < all 3: > 200%	Uncertain	4	6	2	high	1	E: S: 4 D: M: 2	80%
Shoreline response modeling including sediment-structure interaction	2: 200% 2: < all 3: > 200%	Uncertain	3	5	2	high	1	E: M: 4	80%
Hydrodynamics and stability of tidal entrances	2: 150% 2: < all 3: > 200	Uncertain	4	6	3	med.	3	E: L: 2	50%
Fundamentals of cohesionless and cohesive sediment transport in estuaries	2: 150% 1: all 2: 150%	Known	3	5	2	med.	1	E: M: 4 D: M: 3	85%
Estuarine sediment transport numerical modeling	2: 100% 2: < all 2: 150%	Known	4	7	2	med.	1	E: S: 3 D: S: 3	80%
Response of wetlands to flow and sediment dynamics	3: 250% 2: < all 2: 200%	Uncertain	3	5	2	high	3	E: S: 4 D: M: 3	80%
Quantitative modeling of wetland formation, growth and destruction	3: 300% 3: > none 3: > 200%	Uncertain	3	6	2	high	1	E: S: 4 D: M: 3	60%

Table 12.10. Coastal Ecosystems

Research Issue or Task	Uncertainty		Time	Cost	Rate	Impact	Sensitivity	Data	Consensus
	Statistical	Subjective							
Synthesis of present knowledge of coastal ecological responses to physical changes	3: 250% 2: < all 3: > 200%	Uncertain	5	7	1	high	1	E: M: 4	90%
Exploratory models of ecological responses to multiple, simultaneous physical changes	2: 200% 2: < all 2: 50-200%	Uncertain	5	7	1	high	1	E: M: 4 D: S: 4	50%
Time scale of response of coastal ecosystems to change	2: 150% 2: < all 2: 50-200%	Uncertain	2	6	2	med.	3	E: M: 4 D: M: 2	90%
Growth controlling phenomena involving principal species of coastal animals and plants	2: 150% 2: < all 2: 50-200%	Uncertain	3	6	2	high	3	E: L: 3 D: M: 2	90%
Paleoecology of cores from major costal ecosystems	1: < 100% 2: < all 2: 50-200%	Uncertain	3	6	2	high	1	E: S: 4 D: L: 1	90%
Aerial extent of existing major types of coastal systems	1: < 100% 2: < all 2: 50-200%	Uncertain	4	5	1	high	3	E: M: 4 D: M: 2	90%
Comparison of existing marshes of various densities of tidal creeks	3: 250% 3: > none 3: > 200%	Uncertain	4	5	2	high	3	E: S: 3 D: L: 1	50%
Detailed quantitative physiometric descriptions of major coastal ecosystems where they now exist	1: < 100% 2: < all 2: 50-200%	Uncertain	4	7	2	high	3	E: M: 3 D: L: 2	80%

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