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MODES OF HOLOCENE COASTAL PROGRADATION

GULF OF CARPENTARIA

EUGENE GARTH RHODES

Thesis submitted for the degree of
Doctor of Philosophy
at the
Australian National University

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Except where otherwise acknowledged, this thesis represents my own work.

Eugene Garth Rhodes

Eugene Garth Rhodes
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Chenier and beach-ridge plains are both examples of prograded coasts, and various models have been proposed for their development. In the case of the chenier plain, the Louisiana coast has provided the prime example where deposition is linked to variation in sediment supply from the Mississippi delta. The study of beach-ridge plains has been more widespread, although Galveston Island and the Nayarit (Mexico) examples have been important in the understanding of such coasts. In most examples of prograded coasts, relative sea level and climatic change have been suggested as important factors controlling coastal deposition.

Holocene chenier and beach-ridge plains in the Gulf of Carpentaria, a shallow epi-continental sea in northern Australia, exhibit a wide range of depositional environments distributed over broad coastal plains. The depositional environments produce distinctive facies which are preserved in the stratigraphy of the prograded coast. Some of these facies have a well-defined relationship to sea level, and also contain datable shells. Therefore, a record of upper Holocene sea level changes remain preserved in the morphostratigraphic record of these prograded plains.

Relative sea level change during the last 6000 radiocarbon years has been different for the chenier and beach-ridge plains. The chenier plain, which occurs on the southern margin of the Gulf, shows a relative fall of 2.4m since the mid-Holocene. In contrast, the beach-ridge plain, on the eastern margin of the Gulf, has experienced only 1.5m of fall during the same period. These differences appear attributable to hydro-isostatic response to water loading of the continental shelf during and after the Postglacial Marine Transgression.
Coastal deposition since the mid-Holocene has occurred in an episodic manner on both the chenier and beach-ridge plains. The modes of deposition are different with mudflats constructed on the chenier plain and beach ridges developed on the beach-ridge plain during periods of high terrigenous input. During periods of lower terrigenous input, cheniers are constructed on the chenier plain whilst a period of non-deposition occurs on the beach-ridge plain. In the Gulf of Carpentaria, such a variation in sediment supply has occurred at least four times since 6000 years B.P. In areas outside the Gulf of Carpentaria, similar episodic phenomenon can be shown in areas such as Broad Sound, Queensland, where sufficient radiocarbon dates have been obtained on coastal sediments.

Climatic change in the form of variation in rainfall intensity (or storminess) is suggested as the forcing function in this model. At present, insufficient data on upper Holocene climatic change exist within Australia for regional correlation of such events. However, there appears to be a correlation between progradation of the beach-ridge plain during the period 6000-4800 years B.P. and increased cyclogenesis accompanied by a pluvial in eastern and southern Australia. On a global scale, glacial advances in the northern hemisphere also appear to be in phase with this increase of sediment supply to the Gulf of Carpentaria. Furthermore, there is a general agreement between a wetter period in southern Australia after 2000 years B.P. and a long stage of beach-ridge progradation in the Gulf of Carpentaria during the time 2300-600 years B.P.
INTRODUCTION

The accretion of shorelines adjacent to the seas and oceans of the world has concerned man since prehistoric time. Prograding shorelines either presented advantage or inconvenience, but rarely has man considered the broadening of a coastal plain with ambivalence. Coastal accretion and nearshore deposition closed navigable channels and altered established life patterns as often as it created useable or desirable real estate.

If one is to seek an answer to the question "under what conditions are coasts prograded and strandlines preserved?" then some broad definitions concerning prograded coasts must be set forth. Progradation implies deposition and Curray et al. (1969) divided depositional coasts into two major categories: (a) chenier plains, and (b) beaches and barriers. Beaches and barriers were further subdivided into transgressive, stable and regressive coastal types. If discussion is to be limited to progradation, and by definition most transgressive and stable coasts do not prograde, then prograding coasts may be subdivided in the same manner as depositional coasts: (a) chenier plains, and (b) beaches and barriers. Furthermore, beaches and barriers may prograde as a series of beach ridges and the resulting landform is called a beach-ridge (or strand) plain.

As chenier and beach-ridge plains represent the major sub-divisions of prograding or regressive coasts, then obviously both types of coasts offer a source of information concerning modes of progradation and preservation of strandlines. The eastern Gulf of Carpentaria presents these two types of coast juxtaposed on the same epi-continental sea,
offering potential for geomorphic comparisons which are accessible in few other locations.

The recognition of cheniers by Russell and Howe (1935) and the identification of chenier plains throughout the world by Price (1955) helped stimulate geological interest in the modern environments of deposition as well as their ancient analogues (eg. shoestring sands). Much less has been written on beach-ridge plains, although interest in their genesis extends back to Gilbert (1890). A full review of previous work is best accomplished by considering the two morphologies separately.

CHENIER PLAINS

Definition, morphology and composition

Several long, narrow, sandy ridges run roughly parallel to the coast of southwestern Louisiana. Rising slightly above surrounding marshes, lakes, and watercourses, all essentially at sea level, these low ridges form the most conspicuous topographic features of the region. Sharply localized, well drained, and fertile, they support naturally a luxuriant vegetational cover in which large evergreen oaks form so striking a part that, quite deservedly, the ridges have been called cheniers by their Creole inhabitants. Many of the cheniers have been cleared. Cotton fields cover large parts of the highest and most accessible; those lower and more distant form bases for grazing, trapping, hunting, and fishing. Where they are served by roads or navigable waters, the population is relatively dense and prosperous. (Russell and Howe, 1935, p.449)

The description above was extended by Curray (1969) from a Louisiana origin to cover all similar elongate bodies of sand and/or shell stranded on a coastal mudflat or marsh. Cheniers are oriented parallel or sub-parallel to the modern coast and are widely separated from adjacent ridges. Where these ridges are distributed across a broad plain, the feature is known as a chenier plain (Price, 1955; Byrne et al., 1959).
Individual chenier ridges vary from 50m to 500m in width and can extend for several tens of kilometers (Price, 1955; Hoyt, 1969). Deposits of this type are common in hot-wet regions where large quantities of fine sediments are carried by rivers (Brouwer, 1953; Davies, 1972). Crests of cheniers may extend above the highest normal level of swash runup, reaching levels where storm surges or similar meteorological variations in sea level temporarily create higher water levels. These stranded ridges are frequently vegetated with upland or non-halophytic vegetation and in some areas, have become locations for human settlement.

Ridges on the southwestern Louisiana coastal plain are composed of sand and shells with minor quantities of silt and clay. Shells composed up to 22% of the Louisiana ridge deposits (Byrne et al., 1959), although Brouwer (1953) noted ridges on the Surinam coastal plain which consisted entirely of sand. Cook and Polach (1973) found ridges in Broad Sound, Queensland, constructed of predominantly sand and gravel. The sand fraction contained quartz grains and small carbonate fragments, and the gravel fraction was dominated by whole or broken bivalve shells with a mixture of terrigenous pebbles, cobbles and mud balls.

The intervening mudflat or marsh can be a broad featureless, unvegetated flat similar to those of Northern Australia (Thom, et al., 1975), or may be a vegetated marsh surface such as the Louisiana plain. Cheniers may be separated by distances of only several hundred meters or up to 4-5 kilometers on either sort of plain. Total width of a chenier plain may be as wide as 20-50 km as in the Louisiana or Surinam examples whilst Otvos and Price (1979) identify a smaller "bayhead chenier plain" with a typical width of 1-10 km. Cheniers may be located within mangrove swamps, although their occurrence within this environment remains to be catalogued fully.
Many workers have noted that bases of chenier ridges rest conformably on nearshore or littoral facies (Gould and MacFarlan, 1959; Byrne et al., 1959; and Hoyt, 1959) whereas earlier workers (Russell and Howe, 1935; Brouwer, 1953) felt that such ridges were deposited over marsh or swamp deposits. Coleman (1966) showed that Louisiana cheniers may rest on either marsh or nearshore sediments as in the case of Pecan Island which rests on marsh sediments at its western end and nearshore sediments near its eastern end. Figure 1-1, reproduced from Coleman demonstrates this change along the length of Pecan Island. In another temperate location, at mouth of the Essex, England, Greensmith and Tucker (1969) observed shell ridges resting on a silty mud base containing plant debris, from which they inferred that deposition occurred over marsh sediments. In a very different environment, the arid Gulf of California, Thompson (1968) described gravel and shell ridges which form on fine-grained intertidal mudds. In a tropical estuary such as Broad Sound, Queensland, Cook and Polach (1973) and Cook and Mayo (1978) showed cheniers to be resting on mangrove mud.

A summary of alternative stratigraphic models is presented in Table 1-1. It appears that no generalized stratigraphic model may be derived from these limited examples, particularly in the absence of published drill hole data from the coast of Surinam, one of the major chenier plain examples.
Figure 1-1 Cross-sections of Pecan Island, central Louisiana, U.S.A., demonstrate that Louisiana cheniers may have distinctly different stratigraphy at various sites along the same ridge.
Table 1-1

<table>
<thead>
<tr>
<th>Ridge Composition</th>
<th>Basal Sediments</th>
<th>Example</th>
<th>Comments</th>
</tr>
</thead>
<tbody>
<tr>
<td>Sand and shell</td>
<td>Marsh</td>
<td>Louisiana</td>
<td>Bases above mean sea level</td>
</tr>
<tr>
<td>Sand and shell</td>
<td>Subtidal silts and clays</td>
<td>Louisiana</td>
<td>Bases 3-4m below sea level</td>
</tr>
<tr>
<td>Gravel and shell</td>
<td>intertidal muds</td>
<td>Gulf of California</td>
<td>Bases within intertidal range</td>
</tr>
<tr>
<td>Sand and gravel</td>
<td>Mangrove muds</td>
<td>Broad Sound Australia</td>
<td>Bases within intertidal range</td>
</tr>
</tbody>
</table>

Determination of chenier plain histories

Rates of chenier plain progradation can be determined from chronologies based on datable materials incorporated in either chenier ridges or the underlying mudflat facies. Datable materials from mudflat environments (with the exception of in-situ mangroves and in-situ shellfish) cannot be assumed to date the time of deposition of the mudflat facies (see Chapter 4). Similar uncertainties apply to datable material in strandlines. However, dates from a sequence of strandlines can show progradation trends and several field replicates may be taken from different locations on the strandline and used to establish the trace of time lines along a prograded plain.

Some workers, Gould and MacFarlan (1959), Cook and Polach (1973) and Schofield (1960) have attached a radiometric chronology to chenier plain development. In general, these studies provide only gross average rates of progradation. Gould and MacFarlan and Schofield inferred ages of chenier ridges by dating freshwater and brackish water organic deposits beneath these ridges. This approach provides only a maximum age for the ridges and therefore may not provide the best age structure of a
prograded plain, a point reinforced by the discussions of Thom et al. (1978) in their study of prograded barriers on the coast of New South Wales, Australia. Approximate rates of progradation can be calculated from chenier plain cross sections associated with these studies. A linear rate of shoreline progradation is usually assumed between age samples or sample clusters. Table 1-2 presents rates of coastal progradation from the above studies.

Table 1-2

<table>
<thead>
<tr>
<th>Locality</th>
<th>Distance between dated points</th>
<th>Age range</th>
<th>Progradation Rate</th>
<th>Comments</th>
</tr>
</thead>
<tbody>
<tr>
<td>Louisiana, U.S.A.</td>
<td>12500m</td>
<td>2800-0</td>
<td>4.45m/yr</td>
<td>From Gould and McFarlan, 1959</td>
</tr>
<tr>
<td>Broad Sound, Qld Australia</td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>Charron Point</td>
<td>400m</td>
<td>4500-1400</td>
<td>0.12m/yr</td>
<td>From Cook and Polach, 1973</td>
</tr>
<tr>
<td>Hoogly-Waverley</td>
<td>900m</td>
<td>5500-1000</td>
<td>0.20m/yr</td>
<td>using dates in Cook and Mayo, 1978</td>
</tr>
<tr>
<td>Firth of Thames, New Zealand</td>
<td>1350m</td>
<td>3900-0</td>
<td>0.29m/yr</td>
<td>From Schofield, 1960</td>
</tr>
</tbody>
</table>

Formation of chenier ridges is generally presumed to be episodic as the ridges themselves are discrete features. Progradation of a chenier plain as a whole is probably continuous, although varying in rate with certain factors such as wave climate and sedimentary fluctuations. Various explanations for the episodic nature of chenier plains have been developed from several examples.

Some early workers such as Geijskes (1952), Brouwer (1953) and Price (1955) attributed the episodic appearance of the chenier plain to the periodic variations of sediment availability in the nearshore zone. It was with the aid of chronologies established by C-14 dating that researchers began to look beyond the chenier plain itself for a more specific explanation of episodic development. Episodic occurrence of
chenier deposits in the marginal deltaic plain of the Mississippi were interpreted by Gould and MacFarlan (1959) as response to shifts in the location of the mouth of the Mississippi River. In this case, C-14 dating of deltaic channels of the Mississippi River and the Louisiana chenier plain showed that changing positions in the location of the mouth of the Mississippi River provided variation in sediment supply over lengthy period (hundreds to thousands of years). A similar explanation was invoked by Cook and Polach (1973) who linked chenier ridge development to a variable supply of sediment from nearby estuaries. Schofield (1960) related periodic fluctuations in the debatable sea level curve of Fairbridge to the radiometric date determinations of cheniers. Mass mortality of shellfish was correlated with the occurrence of ridges on the Essex plain by Greensmith and Tucker (1969). On the coast of Surinam, recent work by Augustinus (1978) has confirmed earlier opinions concerning that coast, that episodic occurrence is due to longshore movement of mud deposits. Most of these mechanisms, to include fluctuations in eustatic sea level exist as yet unproven alternatives to variations in deltaic discharge. It is only the Surinam example which appears to have sufficient evidence to support a hypothesis distinctly different from that related to deltaic processes. Augustinus (1978) presents field evidence to show that regular migrations of shoals away from the Amazon River permits the alternation of chenier ridge and mudflat formation. Otvos and Price (1979) have reviewed earlier work on the Surinam plain which also suggested such a model, however the external driving mechanism for such mudflat migrations has not been identified.

The problem of genesis

As shown above, there are several different models of chenier plain
formation, each purporting to explain episodic formation of chenier ridges in a context of more or less continuous progradations. Within this context of progradation it has been suggested that the formation of a chenier ridge is the result of a special set of physical and geologic conditions.

The southwestern Louisiana coastal plain has contributed most of the information supporting present models of chenier plain development. Russell (1953) advanced the now familiar view that this chenier plain developed in response to shifts in sediment distribution by the Mississippi delta. Gould and MacFarlan (1959) were able to correlate major shorelines with sub-delta development through their detailed sedimentary and chronologic framework for this coast. In a related work, Byrne, et.al. (1959) described the Louisiana cheniers as features formed during periodic halts of coastal progradations due to a shifting source of fine-grained sediment in the Mississippi delta. Gould and Morgan (1962) further consolidated the model by reiterating that the Louisiana chenier plain was a result of pulsations of fine sediment caused by shifts in river mouth location. This model was developed for a coast exhibiting either submergence of sea level stability during most of the period of progradation (Coleman, 1966). However, relative sea level change in relation to chenier plain genesis has not appeared to contribute significantly to either the model developed from Louisiana examples or to most departures from that model based on other examples.

Formation of chenier plains in other parts of the world have been attributed to variations on the above model and, in a few cases, distinctly different alternatives. Sea-level oscillations of the Fairbridge sea level curve were correlated with chenier ridges on the chenier plain at Firth of Thames, New Zealand by Schofield (1960). In a model similar to deltaic diversion, Thompson (1968) suggested that
chenier plain development in the Gulf of California may have resulted from Colorado River diversions during the upper Holocene. Greensmith and Tucker (1969) postulate mass mortality of shellfish as being partly responsible for the construction of cheniers at Essex, England. Cheniers formed of coarse-grained bioclastic carbonates in northwestern Qatar of the Persian Gulf are described by Shinn (1973) as developing from spits which were constructed in response to longshore transport. However, Shinn neglected to indicate the controlling factor for the development of such spits.

In another variation from the Louisiana model, Cook and Polach indicate that a periodic decrease of sediment supply from adjacent estuaries permitted the winnowing of mangrove muds at Broad Sound, Queensland, thereby providing coarse material for chenier ridge construction. However, the reason for variations in sediment supply from such estuaries was not established. Furthermore, the use of the word "winnow" causes some confusion. To authors such as Gould and MacFarlan and others who quote their work, "winnowing" is the process of removing the coarse fraction from a muddy facies in which either sand or shell is present. Cook and Polach (1973) describe the process as follows: "Cheniers reflect periodic interruptions to this sequence, forming when the mangrove sediment is exposed to erosion and the shell material is winnowed out" (p.262). In such cases a distinction should be made between winnowing and sorting.

The differentiation between winnowing and sorting was effectively clarified by Todd (1968) who showed in a semi-quantitative model that separation of coarse sediments from unsorted muds is a two-step process. Firstly, clay and silt is removed by winnowing and transported elsewhere (offshore in Todd's model). Secondly, the coarse lag deposits are further sorted and concentrated by wave action into chenier and beach deposits.
This two-step process cannot occur if the nearshore waters are heavily laden with suspended sediment, an observation which fits with the traditional interpretation for the origin of the Mississippi chenier ridges. Under heavily laden conditions, nearshore waters tend to deposit a poorly sorted detritus, inhibiting both winnowing and sorting processes. Todd summarized his findings: "Because the increased density and viscosity of the overloaded currents increases the effectiveness of traction and slightly reduces or leaves unchanged the efficiency of suspension, the differential transport capacity of longshore currents is reduced. The result is a moderately rapid accretion of unsorted clayey interchenier debris" (p.744). Conversely, according to Todd, nearshore waters which are in motion due to either longshore transport or wave activity, or both, and which are sediment starved, are extremely effective in removing the fine fraction and sorting the coarse fraction into cheniers and beaches.

In a more distinctive alternative to the model from Louisiana example, recent workers on the Surinam coast have identified a periodic movement of mud shoals or "sling mud" (Augustinus, 1978) along the coast as the primary factor responsible for chenier plain development in that area. Otvos and Price (1979) have shown that as early as 1832 Lyell identified the Amazon River as the source of mud for coastal progradation along that coast. According to Otvos and Price, more recent works have confirmed this movement of mud to be cyclic with short-lived climatic changes perhaps controlling erosion and accretion.

In summary, a number of attempts have been made to explain both the mode of deposition and the means of initiation of chenier ridge construction since Russell and Howe (1935) first described the ridges of southwestern Louisiana. Curray (1969) summarized the general thrust of most models by stating that "a fluctuation in the rate of supply of
sediment to a coastline is a requirement." It appears that in the absence of a large supplier such as the Mississippi River, which may be turned on or off by altering the location of discharge, that a number of alternative mechanisms can be advanced. These include fluctuations in sea level, variations in storminess, climatic change and mass mortality of shellfish. A universal model is not yet available to explain the origin of cheniers. To assist in the creation of such a model, a chenier plain complex was selected for detailed morphostratigraphic study in the Gulf of Carpentaria.

**BEACH RIDGE PLAINS**

Beach or dune ridges are usually found in greater or less degree of perfection on any prograded shoreline, and they are of so much importance, not only in showing the successive stages of development of the forms which possess them, but also, as will presently appear, in showing whether a coast has remained stable or experienced changes of level during their formation, that it is pertinent to enquire somewhat fully into their origin and significance. (Johnson, 1919, p.404)

Definition, morphology and composition

Johnson mentions both beach and dune ridges; this apparent combination of two different landforms in the same discussion requires some clarification. On many coasts, after initial deposition of a beach ridge by marine processes, aeolian deposition builds a low dune whose morphology may be concordant with the marine constructed beach. This aeolian cap or mantle fosters the use of the term dune ridge for a landform whose primary formation is not due to aeolian activity. Discussions in this section will be focussed on the marine, or beach component of such ridges.
Beach ridges are successive strandlines which have been deposited seaward of their predecessors on a prograding shoreline. They are usually separated by narrow swales of varying width and depth depending on how closely the successive ridge crests have been formed. When a broad sequence of parallel or sub-parallel beach ridges has been deposited, their surfaces constitute a beach-ridge plain. Beach-ridge plains may form a barrier island backed by a lagoon or estuary or merely a strand plain between two headlands.

Individual ridges may have amplitudes less than 1.5m (Curray et al., 1969) or as much several meters (Nossin, 1965). Curray et al. described a strand plain at Nayarit, Mexico where ridge spacing varied from 15m to 200m crest to crest and individual ridges averaged 50m in width. The Nayarit plain is over 15km wide with as many as 280 ridge crests. Although some of the Nayarit ridges could be traced for 50km, individual ridges on other coastal plains are continuous only over several kilometers. Beach ridges described by Thom et al. (1978) on the New South Wales coast have crest to crest distances of 20m to 100m with crests continuous across bedrock embayments whilst the entire plain may be only a few hundred meters to 3km wide. Elsewhere, as on the Dutch coast (van Straaten, 1965), beach-ridge plains may be 8-10km wide.

Beach-ridge plains present an undulating form in section, and have crests which may be capped with aeolian material. Individual ridges generally show a gently dipping seaward slope with a more steeply dipping landward slope. Similar to chenier crests, beach-ridge crests can be be deposited above the level of highest normal wave runup during periods of meteorologically raised sea levels (Psuty, 1966). When capped with a foredune before further progradation moves the active beach seaward, ridge crests can attain heights in excess of 10m. In areas of massive aeolian activity such as the Newcastle Bight, New South Wales,
Thom et al., 1978) evidence of beach-ridge morphology may be partially obliterated by dune fields. Thom et al. discuss at length morphologic and stratigraphic models for such barriers capped with large transgressive dune fields in eastern Australia.

Beach-ridge plains may be associated with local progradation such as cuspatc headlands like Dungeness (Lewis, 1932; and Hey, 1967) and marine plains of limited extent in small embayments (Davies, 1961; Nossin, 1965; and Thom et al., 1978), or they may indicate regional progradation such as Galveston Island (Bernard et al., 1962) or strand plains such as described by Curray et al. (1969) in Mexico and Smart (1976) in North Queensland. Where "sets" of ridges can be distinguished by sub-parallel or oblique relationships to neighboring sets (Davies, 1957; Psuty, 1966; Curray et al., 1969; Smart, 1976; and Thom et al., 1978) these disconformities may represent periods of cut or non-deposition followed by fill during later progradation of the plain.

The bases of the regressive beach ridges studied by Bernard et al. (1962) rest on shelf muds 9-12 m below low water line. Curray et al. 1969 shows the Nayarit, Mexico strand plain to be resting on littoral sands with a facies change 5-8m below sea level. Similarly, the regressive barriers described for eastern Australia are composed of a beach-ridge facies which conformably overlies 5-20 meters of littoral deposits called nearshore shelly sands by Thom et al. (1978). These examples are similar to van Straaten's description of the Dutch coastal deposits. The Dutch sequences exhibit a beach or nearshore facies which extends below sea level, with a marked fining downward until a clayey deposit is encountered at approximately 7 meters below mean sea level. This fine-grained facies is interpreted by van Straaten as bottom sediments deposited in an open sea environment.

Beach ridges are formed from materials as coarse as the shingles of
Dungeness (Steers, 1969; Hey, 1967) or as fine as the well sorted quartz sand with shell fragments in eastern Malaya and New South Wales (Nossin, 1965a; Thom et al., 1978). Although most sandy beach ridges are composed of quartz sand and/or rock fragments, (Curray et al., 1969; Psuty, 1966), calcium carbonate as bioclastic constituents can vary from 5% (Thom et al. 1978) to greater than 50% (Smart, 1976).

The internal structure of beach ridges has been described by a number of workers. Van Straaten (1965) showed that strandline deposits on the Dutch coast contained cross-bedded dune sand grading downward to horizontal laminations of slightly coarser beach material at or near sea level. Laminations in this lower zone sometimes had a slight inclination seaward, similar to the slopes of modern beach faces (2.5 deg.). Grain size then decreased downward until clay beds were encountered several meters below sea level. Beach ridges in the Tabasco plain showed both seaward and landward dipping beds with landward dipping strata predominating in the upper sections of the ridges (Psuty, 1966). Lower and frontal (seaward) portions of ridges contained dominant seaward dipping beds. Curray et al. (1969) found that ridge sands were not well laminated on the Nayarit coast, Mexico, although grain size varied with environment of deposition. Finest sands were found in the shoreface deposits with a coarsening upward into beach and dune sands. Jennings and Coventry (1973) studied the internal structure of small barrier islands in the Fitzroy estuary, Western Australia and found foreslope (seaward dipping beds) were rare in contrast to topslope (nearly horizontal) or backslope (landward dipping) beds. Stratigraphic investigations of New South Wales barriers, Australia, (Thom et al., 1978) show that beach sands exhibit low angle seaward dipping beds, locally enriched with heavy minerals and shell fragments, and overlying nearshore sands which fine downwards. Beach-ridge plains have also been
shown to grade laterally into chenier plains such as in the Louisiana examples. (Price, 1955). Reasons for such change have been given as diminished supply of silt and clay away from the Mississippi (Hoyt, 1969) or changes in wave energy adjacent to steeper offshore profiles, with steeper profiles promoting beach-ridge formation (Todd, 1968). Otvos and Price (1979) in reviewing examples of such lateral changes in morphology supported variation in wave energy as the dominant controlling factor.

Determination of beach-ridge plain histories

It appears that large series of extensive ridges must represent a long time interval, and that 25 to 50 years is not an improbable figure for the time required to build such prominent ridges as are characteristic of the Dungeness, Dars and Swineforte areas. (Johnson, 1919, p.439)

Without the aid of radiometric dating, early workers such as Johnson were able to deduce from historical accounts the approximate progradation rates for some classic beach-ridge plains. Johnson established a growth rate of 5.5-6 yards per year (5m/yr) for Dungeness. From other historical descriptions he reports 1500 meters of progradation in two hundred years (7.5m/yr) for the Swineforte complex.

Since these early efforts to quantify rates of progradation, increased use of radiometric and historical chronologies have established rates of progradation for barrier and beach-ridge coasts. Thom et al. (1978) summarize the principal studies using radiocarbon chronology. A lack of stratigraphic data precludes volumetric evaluation of progradation in Allen (1965), Nossin (1962, 1965a and b) and Moore (1960). Table 1-3 summarizes data from Thom et al. and other sources with respect to rates of barrier and beach-ridge accretion. It appears that the historical rates quoted above are high in comparison with middle to upper Holocene rates for Australia which range from 0.5m to 0.7m/yr. The
Galveston Island rate of 1.1 m/yr is distributed over a long time span and is probably a minimum. Such is also the situation with the Nayarit progradation which could have occurred in discrete episodes rather than over a continuous time span as suggested in Table 1-3. Only the Scheveningen rate from the Netherlands approaches those of the Dungeness and Swinepforte complexes.

<table>
<thead>
<tr>
<th>Locality</th>
<th>Distance between dated points</th>
<th>Age range</th>
<th>Progradation rate</th>
<th>Comments</th>
</tr>
</thead>
<tbody>
<tr>
<td>S.E. Australia</td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>Moruya</td>
<td>675</td>
<td>5700-2300</td>
<td>0.5 m/yr</td>
<td>Thom et al.,</td>
</tr>
<tr>
<td>Fens</td>
<td>300</td>
<td>5000-3750</td>
<td>0.6</td>
<td>1978</td>
</tr>
<tr>
<td>Wonboyn</td>
<td>1100</td>
<td>5480-1400</td>
<td>0.4</td>
<td></td>
</tr>
<tr>
<td>Woy Woy</td>
<td>3300</td>
<td>6240-1380</td>
<td>0.7</td>
<td></td>
</tr>
<tr>
<td>Galveston Is. U.S.A.</td>
<td>4000</td>
<td>3500-0</td>
<td>1.1</td>
<td>Bernard et al., 1962</td>
</tr>
<tr>
<td>Nayarit, Mexico</td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>Section A</td>
<td>10000</td>
<td>3600-0</td>
<td>2.8</td>
<td>Curray et al., 1969</td>
</tr>
<tr>
<td>Section B</td>
<td>9500</td>
<td>3600-0</td>
<td>2.6</td>
<td></td>
</tr>
<tr>
<td>Section C</td>
<td>5000</td>
<td>3600-0</td>
<td>1.4</td>
<td></td>
</tr>
<tr>
<td>Section D</td>
<td>12000</td>
<td>4500-0</td>
<td>2.7</td>
<td></td>
</tr>
<tr>
<td>Scheveningen, Netherlands</td>
<td>8000</td>
<td>4800-2850</td>
<td>4.1</td>
<td>van Straaten, 1965</td>
</tr>
<tr>
<td>Bay of Plenty New Zealand</td>
<td>9000</td>
<td>5300-0</td>
<td>1.8</td>
<td>Chappell and Pullar in press</td>
</tr>
<tr>
<td>Dungeness</td>
<td>?</td>
<td></td>
<td>5.0</td>
<td>Johnson, 1919</td>
</tr>
<tr>
<td>Swinepforte</td>
<td>200</td>
<td></td>
<td>7.5</td>
<td>Johnson, 1919</td>
</tr>
</tbody>
</table>

Workers such as Gilbert (1890) felt that beach ridges are inherently episodic in their mode of development. For example, Gilbert (1890, p.57) said: "each of these [ridges] is referable to some exceptional storm, the waves of which threw the shore drift to an unusual height". Similar views have been reiterated by later workers such as Psuty (1965) who described the process of ridge building on the Tabasco, Mexico coast where individual storms forced berms to migrate landward until they...
reached a stranded position and were no longer modified by wave runup. Thom (1964) referred to the Tabasco example whilst comparing that coast with the east coast of Australia. Thom did not advocate that each ridge was a single event, but suggested that at some locations a combination of wave and later wind activity constructed ridges, whilst at other sites only storm waves were responsible for ridge formation.

Episodic development of beach ridges may also refer to the longer term stages in which some prograded plains appear to develop. Such episodes may be indicated by groups or sets of ridges apparent in the morphology of the plain. Most authors consider discordant intersections of groups of ridge crests to be prima facie evidence of periodic development, (Johnson, 1919; Davies, 1957; Psuty, 1966; Curray et al., 1969; and Thom et al., 1978). Additional indicators of such episodic development may include differences in soil development, degree of carbonate leaching, historical and anthropologic data and radiometric dating.

The problem of genesis

As in the case of the chenier plain, explanations for episodic progradation (regression) of the beach-ridge plain are closely tied to the origin of the plain. Change in location or quantity of fluvial discharge is the most frequently used explanation for discordant ridge development (Psuty, 1966 and Allen, 1965). Psuty attributed episodic development to a seasonal variation in the short term and changes in the distribution of fluvially supplied sediment in the long term. Allen noted apparent periods of erosion and reorientation which could be identified by discordance between ridges and only explained in terms "of
variation in sediment supply, direction of tidal feeder channels and climatic factors". Variation in storminess has been suggested by Davies (1957), Moore (1960) and Thom (1978). Davies advocated a relationship between storminess and the absence of beach-ridge accretion. Moore observed a variation in levels of arctic beach-ridge deposition which, when dated by archaeologic evidence, indicated that minor eustatic variations in sea level probably enhanced the preservation of ridge sets. Thom has advocated that discrete periods of increased cyclogenesis are responsible for the episodic behavior of barrier building (or lack of it) in eastern Australia during the upper Holocene. Curray et al. (1969) found evidence suggesting that climatic change during the late Holocene was responsible for changing the direction of longshore transport. These changes were reflected in the beach-ridge morphology by distinct unconformities and realignments of the coast. A similar climatic explanation was invoked by Smart (1976) in his explanation of periodic progradation at Cape Keer Weer, North Queensland. Smart felt that increased aridity during the mid-Holocene caused higher rates of upland erosion and therefore higher fluvial sediment input.

Workers such as Johnson (1919) was concerned less with the episodic nature of ridge formation and more with the conditions which fostered ridge development. According to Johnson beach-plains could be grouped into three genetic types.

a. exceptional storm deposits from longshore current supply
b. offshore bar migration supplied by sea bottom material
c. spit extension supported by longshore drift.

Each of these cases entails accretion to the beach face, either of a conformable veneer or, more characteristically, of a new beach ridge constructed of sediment from within or beyond the surf zone. This has been purported by some to be the result of a single high energy event.
(Gilbert, 1890). However, others such as Davies (1957) agreed that the pebble ridge complexes such as Dungeness in southern England may be explained as storm wave deposits, but stated that multiple ridges on sandy coasts are not entirely analogous to gravel ridges. Davies (1957) indicated a different response to storm waves for such coasts. In his model, decreased storminess encouraged accretion on the berm which was developed into a foredune by the stabilizing effect of vegetation. As an advocate of vegetation acting as a primary stabilizer and sand trap, McKenzie (1958) differed from the views of Davies, and suggested that vegetation encroached seaward from a previously developed foredune. The Davies' view of initial foredune colonization was supported by Bird (1960) who noted that ridges on an actively prograding portion of Ninety Mile Beach, Victoria, showed vegetation on foredune crests, but not in the intervening swales. Thom (1964), reviewing this controversy, stated that ridges in the Gulf of Mexico and similar environments developed during periods of meteorologically raised sea levels and were later preserved by vegetative action (see Psuty, 1966). In his review, Thom found evidence from both the Tabasco and eastern Australia examples to partially support the model proposed by Davies (1957), but presented further evidence from the Newcastle area to support a wave-wind model in other cases.

Leont'ev and Nikiforov (1965), Allen (1965), Nossin (1962, 1965a and b) and Wright (1970) limited their discussion to the source of the sediment and left the mode of ridge development unexplained. The erosion of rocky headlands and fluvial input provided material for progradation of Nossin's beach-ridge plain in eastern Malaya. Leont'ev and Nikiforov (1965) indicted that shoreward transport of transgressed strand deposits yielded reworked terrigenous material for barrier progradation as sea level receded. Beach-ridge morphology and granulometric parameters were
a function of longshore distance from sediment source in Wright's (1970) study of the mouth of the Shoalhaven River, New South Wales. Allen found that redistribution of coarse fluvial sands into extensions of spits by tidal, wave, and longshore current processes explained the development of parallel and subparallel ridges forming barrier islands of the Niger delta.

Some workers integrate their discussion of initiation with that of mode of deposition, for example; Davies (1957) stated that over flattened off-shore profiles contributed to the initiation of beach-ridge building. However, he supported the views of Johnson (1919) that periodicity in ridge building cannot be correlated with any succession of cyclic events. Thom et al. (1978) described a series of radiometrically dated ridges exhibiting episodic development, noting that progradation was initiated at the time of postglacial stillstand in eastern Australia 6000-6500 years ago and continued locally until approximately 1200 years ago. According to Thom et al. the cessation of progradation on the New South Wales coast was due to a decreased availability of nearshore sand deposits as coastal plains prograded and removed sand from nearshore systems.

Discussion of the genesis of individual ridges has a connection with the longstanding controversy concerning sand barrier genesis. Regressive ridges are a ubiquitous component of such barriers and their development is essential to the existence of a prograded sandy shoreline. Hoyt (1969) amply reviewed the theories of barrier formation to that time, of which many are still accepted, especially as more examples of recent barriers enter the literature. There is no reason to suggest that Australian beaches are formed by a mode of deposition different from other examples. Therefore, all the traditional factors such as shoreward movement of offshore bars, spit elongation, storm deposition, emergent
Figure 1-2  Two widely separated locations in the Gulf of Carpentaria are used to study chenier and beach-ridge plains.
coastal conditions, vegetation as a constructional agent and wave-wind processes in combination may play some contributory role in the formation of prograded beach ridges. At one locality, some factors may dominate, however, the rank ordering of these factors may be more precisely characterized by longterm process-response observations supplemented with nearshore dynamics studies. Furthermore, dominant factors may vary through time, particularly when feedback mechanisms of climatic or sea level change are functional. Although the relative importance within a small group of factors may vary, it is likely that such a group of parameters which favor progradation on the Tabasco coast, the Gulf of California, and the Dutch coast are functional to some extent on the coasts of Australia provided the landforms can be shown to be the same. The occurrence of chenier plains adjacent to beach-ridge plains is infrequent, but the southern and eastern coasts of the Gulf of Carpentaria present these two morphologies side by side on a regionally prograding, tectonically stable coastline. The Gulf is a large epi-continental sea (thereby protecting the coast from waves generated in distant areas) making it an ideal area for the investigation of the morphology, stratigraphy, and chronology associated with these two modes of coastal development.

RESEARCH OBJECTIVES

Rationale - Why the Gulf of Carpentaria?

The shoreline of the Gulf of Carpentaria (Figure 1-2) exhibits over 1500 kilometers of prograded coast. Rivers carrying sediment into the Gulf of Carpentaria drain a basin of over 0.5 million square kilometers. This large watershed responds in a near catastrophic manner during each
annual wet season as 1000mm-2000mm of precipitation are dumped on the area in a period of 4 months (see Chapter 2).

The area is tectonically stable; Doutch (1976) was unable to detect major tectonic disturbances since the Pliocene. Furthermore, the close topographic accordance between Holocene and Pleistocene strandlines throughout the basin suggests continued stability during the last 120,000 years (see Chapter 2).

The onshore and offshore topographic gradients are low. Gradients of the coastal plain are less than 1:2500; offshore gradients are similar with nearshore slopes rarely exceeding 1:1500. This broad uniform topography provides a base on which nearshore and coastal deposition may proceed. The large annual sediment yield from rivers promotes rapid seaward advance of the coastal plain thereby providing a thin but laterally extensive Holocene sedimentary record.

Inspection of the coast at selected sites between Weipa and Burketown at the beginning of the fieldwork in 1975 confirmed morphologic disconformities in both the chenier plain and beach-ridge type of coast with a sharp and horizontal facies change at the base of the beach and chenier ridges. Further inspection of the chenier plain during the reconnaissance phase revealed that base levels of chenier ridges varied with distance from the sea. The assumption was made that past relative sea levels could be established by quantifying the variation in elevation of this abrupt facies change across the prograded plain. An analogous base level was later confirmed in the multiple beach ridges, but was only accessible from drill hole data. Finally, both types of coastline provided ample quantities of easily obtained and well preserved marine shells for C-14 dating purposes.
Experimental design

Most previous studies dealing with prograding coasts have focused specifically on one type of coast. Gould and MacFarlan (1959) and Cook and Polach (1973) discussed chenier plain development, whereas Psuty (1966), Nossin (1962, 1965a and b), and Curray et al. (1969) discussed beach-ridge plains. Van Straaten (1965), Bernard et al. (1962) and Thom (1978) studied prograded barriers of limited regional extent. Furthermore, some of these studies concluded that the major factors responsible for progradation are ones which no longer operate. For example, major engineering works have disrupted the natural processes of the Mississippi delta system, whilst in Australia, Thom et al. have inferred from age structures of prograded barriers that coastal progradation in some areas ceased approximately 1200 years B.P.. The Gulf of Carpentaria offers a single depositional basin with varying coastal morphology and which appears to be presently undergoing regional deposition.

The primary objectives of this study were:

a. to document the morphology and stratigraphy of the two depositional types of coast. This was accomplished by sampling and describing modern environments in conjunction with stratigraphic drilling along transects normal to the coast.

b. to establish geologic histories through careful dating and infer rates of Holocene progradation in both types of coastal morphology.

c. to evaluate factors of environmental change, especially Holocene land/sea level changes, on the formation and periodicity of beach ridges and cheniers. This required precise levelling of modern and relict landforms in relation to the present tidal datum.

d. to explain the origin of beach ridges and chenier plains adjacent to
the Gulf of Carpentaria. Sedimentologic comparisons between modern environments provided the basis for interpreting the stratigraphic record. Additionally, it was necessary to observe the variation in response of the coast during summer, winter and catastrophic events. Appendix A deals with laboratory and associated data reduction techniques.

e. to compare the depositional sequence and modes of development of Carpentaria strandline features with similar depositional coasts elsewhere, further discussion of which is contained in Chapter 7.
CHAPTER 2
REGIONAL SETTING

GENERAL

The Dutch yacht "Duifken" commanded by Wm. Jansz, entered the Gulf in 1605 in search of raw materials for the Dutch East India Company. He proceeded only as far as Cape Keer-Weer before turning back (Heeres, 1899). This expedition was followed by a later one in 1623. Two Dutch vessels the "Pera" and the "Aernem", commanded by Jan Carstensjoon visited the eastern Gulf in April and May of that year. They were unimpressed:

... in our judgement this is the most arid and barren region that could be found anywhere on earth;...

(Carstensjoon's Journal, in Heeres, 1899, p. 39).

Matthew Flinders sailed from Port Jackson in 1802 to explore Torres Strait and the coasts of the Gulf of Carpentaria. For nearly three months, he charted the Gulf coasts, recording data which appears on modern nautical charts.

Renewed interest in the interior of Australia prompted the Stokes expedition of 1841, which revisited some of the localities reported by Flinders and explored further inland with shore parties. The expeditions' winter visit to Carpentaria when the country is most accessible provided a misleading impression which was to lead later explorers and settlers to misfortune (Bauer, 1959). Stokes' party enjoyed dry overland travel with easily crossed rivers, experiences which are not available during the wet season in this area.

The Gulf country was ignored by early settlers until a desire for new and larger grazing properties renewed interest during the late 19th century. Since that time the Carpentaria lowlands and western Cape York
Peninsula have attracted optimistic outsiders for over half a century. It was Leichhardt in 1844 who completed the first exploration into the lower Gulf country. Although his botanist, Gilbert, was killed by hostile Aboriginals on the lower Mitchell River, he continued the journey with comparative ease. Later, Mitchell and Kennedy (1845-8) pursued unsuccessfully the "Plains of Promise" described by the earlier explorer, Stokes. The first to report unfavourably (and perhaps accurately) on the Gulf country was Gregory after his 1848 expedition (Bauer, 1959). However, his cautions went unheeded and the debacle of the Burke and Wills expedition in the early 1860's precipitated more exploration in the form of further expeditions in the search of those unfortunates. By the middle of the 1860's the Gulf country was open for settlement by individuals who refused to "knuckle under" to the inhospitable alternation of dryness and deluge.

The area is little more accessible today than when Leichhardt first visited the Gulf. Surface access is limited to the dry season (March through November) and the only "all weather" road passes through Cloncurry and thence north to Normanton, the wet season frontier of the north. A railway which connects to nowhere links Normanton with Georgetown as a result of goldfield illusions. The remainder of the towns and stations are linked by dirt roads, some formed and graded, others little more than wheel marks on the grasslands.

Access by boat is limited to shallow draft vessels (4m or less) into the prawn fishing port of Karumba, although Aurukun, Edward River, Kowanyama and Burketown can be visited by small boat with patience and luck. Weipa, the bauxite loading port is the only deep water anchorage in the eastern Gulf.

Air travel remains the surest mode of transport although many air strips are rendered useless by heavy rains in the summer. Some stations
in Cape York Peninsula remain isolated from all means of access for 3 to 4 months of each year.

CLIMATE

The area has a hot moist climate with marked dry season (Koppen classification AW, Australian Atlas of Resources). Other workers such as Pedley (1971) refer to the climate as monsoonal, although the seasonal variation of weather fails to meet the strict definition of the term.

Rainfall varies from 800 mm in the southern Gulf to an excess of 2000 mm in the northern areas of Cape York Peninsula (Australian Pilot Vol V, and Australian Atlas of Resources). Figure 2-1 presents an isohyet map for the area. Most precipitation occurs during the summer months of December through March with abrupt cessation during the end of March. Tank evaporation is approximately 750-1000 mm per annum and summer temperatures maximums range from 29 deg-35 deg C with winter minimums in the 15 deg-21 deg C range (Pedley, 1971).

The dry winter season is characterized by strong southeast winds blowing off the interior with occasional local sea and land breezes. The wet summer is dominated by northwest winds which bring moist air in from the Gulf (Figure 2-2). Summer rainfall tends to be highest in coastal areas, gradually diminishing landward. The area is subject to cyclonic disturbances during the summer season and these have an important effect on the coastal plain (see section on physical oceanography).
Figure 2-1 Isohyet map of northeastern Australia (modified from Australian Atlas of Resources). Rainfall is shown in mm/year.
Figure 2-2 Wind roses showing dominant wind directions during the months of January, April, July and October. From left to right stations are Darwin, Groote Eylant, and Weipa. The number of observations is shown in the upper figure, the percentage frequency of variable winds by the middle figure and calms by the lower figure on each rose (from Australia Pilot, Vol. V)
HYDROLOGY

The marked wet/dry seasonality of the area causes most Gulf rivers to be ephemeral. Only the major tributaries of the Archer, Kendall, Mitchell, Gilbert, and Leichhardt rivers maintain surficial flows during the dry season. Smaller river systems become little more than a series of cut-off lagoons dotted along the river's course. However, during the summer, overbank flow is frequent in the tidal sections of the river channels, and is especially common when heavy rains coincide with spring tides. In the upland areas, summer rainfall produces runoff at least equivalent to river channel capacity. According to Simpson and Doutch (1977), present day flooding has very little erosional or depositional effect on the interfluve areas in the southern Carpentaria Plains. These authors suggest that the upland drainage system has been adjusted to higher rates of discharge sometime in the past, perhaps prior to 60,000 years BP.

Regardless of the extent to which present fluvial channels are adjusted to runoff, there is ample field evidence to show that the Gulf rivers move massive amounts of material in both bedload and suspended toward the Gulf. Few of the rivers have been both gauged and rated, thereby precluding any accurate estimate concerning the annual discharge or sediment load. Nimmo (1947) estimated that the Queensland watershed of the Gulf provided $351 \times 10^9$ cubic meters per year with less than 4% of that amount reaching the Gulf. Nimmo’s estimate of loss is probably incorrect, because field observations indicate that a much higher percentage of runoff reaches the estuarine systems after an initial saturation by 200-300mm of rain.
PHYSICAL OCEANOGRAPHY

Eastern gulf ports show an increase in tidal range from north to south. Weipa has a tidal range of 2.4m whilst Karumba at the southern end of the Gulf has a 3.2m range (Figure 2-3). Tides vary from semi-diurnal to fully diurnal and exhibit a marked response to changing meteorological conditions. Radok (1976) has shown that variation in mean sea level is most pronounced in shallow, semi-enclosed basins such as the Gulf of Carpentaria. Although variation due to barometric changes certainly contributes to the departure of observed tidal levels from predicted levels, it is wind stress which is most significant in the Gulf of Carpentaria. According to Radok, it is the nearly uni-directional northwest winds during the monsoon season (December through March) which is responsible for large variations in mean sea level in the Gulf. This annual change in sea level is approximately 0.9m at Edward River (Figure 2-4) and more than 1.0m at Karumba (Munro, 1972). Even during the mild uniform weather of the dry season, meteorological effects may be significant. Figure 2-5 shows observed variations from predicted levels for Edward River on two occasions during 1976.

Few permanent reliable benchmarks of vertical control exist in the area of the Gulf. For purposes of field measurement it was necessary to identify a datum other than mean sea level which has been shown to vary. A level called prediction datum was selected because it is readily recoverable from tidal observations. This datum approximates Indian Spring Low Water and all discussions of absolute elevations on the coastal plains are with reference to this level. Occasionally, reference to mean monthly or annual sea level will be necessary. Mean sea level for any period is taken as the arithmetic average of all hourly water levels for that period. However, the less ambiguous prediction datum is
Figure 2-3  Summary of tidal parameters for Weipa, Edward River, and Karumba stations. Harmonic constants are from Easton (1970) and M. Greig (pers. comm. 1980)
the preferred reference plane.

Little is known about the oceanography of the Gulf. Rochford (1966) reported physical and chemical properties from Gulf waters during brief cruises in 1964. He identified three discrete water masses which affected the Gulf from the Arafura and Coral seas and noted the marked effect of fresh water runoff on salinity in the southeastern Gulf during the summer season. Tidal and non-tidal currents were studied by Cresswell (1971) who showed that dominant tidal circulations probably moved clockwise in the basin with velocities up to 1.5 knots. Non-tidal currents are almost entirely wind driven and these move in accordance with stress created by northwest winds in the summer and southeast winds in the winter. Cresswell's observations showed wind driven currents along the eastern side of the Gulf could reach 4 miles/day toward the north. On this limited base of "ground truths" Teleki et al. (1973) utilized ERTS imagery to extend Cresswell's model and map sediment dispersal. The ERTS data showed sediment transport to be toward the south during January, changing to negligible transport during August with northerly movement observed in November.

Northern Australian coasts are prone to storm surges associated with major cyclonic disturbances (Hopley 1973a and b). The southern and eastern coasts of the Gulf of Carpentaria are no exception. In this area storm surges appear to be a composite effect of sea water pushed against the coast by strong onshore winds in conjunction with high fluvial discharges. If a higher than normal sea level is maintained for 2-3 days, the coast begins to respond to "tidal damming" of the fresh water runoff. The result is shown in Figure 2-6 which depicts a storm surge associated with cyclone Ted (November 1976). The "tidal damming" effect and subsequent overbank flooding of the Karumba coastal plain is described by Simpson and Doutch (1977) from observations in 1974.
Figure 2-4  A plot of mean annual sea level and mean monthly sea level is based on hourly observations at Edward River during 1976.

Figure 2-5  Comparisons of predicted and observed tides at Edward River on two occasions in 1976 show considerable difference between the curves. Predicted curves contain corrections for monthly variations in sea level. Most of the difference between predicted and observed curves is attributed to local meteorological conditions.

Figure 2-6  Storm surge associated with cyclone Ted during November 1976 at Karumba caused raised water levels of 2.0m at the coast.
The shallow bathymetry of the Gulf (maximum depth less than 80m) precludes development and propagation of deep water waves. Most of the nearshore zone is less than 20m in depth and the resulting wave regime is marked by short steep waves generated locally. Shipboard observations during a 10 day period (March, 1977) showed that wave periods averaged 4-6 seconds with 1-3m amplitudes. Limited shore observations during the summer season confirm the prevalence of short steep waves with amplitudes of less than 2 meters. Winter wave activity on the southern and eastern shores of the Gulf is limited due to the dominant southeast winds which blow offshore. Land and sea breezes during the winter create small wind chop, affecting tide levels more than wave regime.

Traditional views of ancient shallow epeiric seas have held that these basins were tideless and subject to only moderate wave activity (Shaw, 1964; Friedman and Sanders, 1978). Klein and Ryer (1978) attempted to refute the view of Shaw (1964) by demonstrating that Holocene analogs of ancient seas actually have effective tidal movements. However, they erroneously identified the Gulf of Carpentaria as among the examples which do fit the traditional model. It does not, and as demonstrated previously, has a tidal range which increases toward the southern end. In addition, wave activity may be shown to be important in modifying the shoreline of the Gulf despite its gentle nearshore gradient.

On an oceanic, wave-dominated coast with a steep nearshore profile such as southeastern Australia, Wright (1976) has demonstrated that over 95% of deep water wave power is transmitted to the inshore zone. Wright compared this situation with other coastal areas, especially ones with flatter nearshore profiles, and showed that on coasts where nearshore gradients range from 1:1000 to 1:1200, only about 16% of the deep water wave power reached the inshore zone. Using computer modelling techniques presented by Goldsmith et al. (1974) similar computations may be made for
the Gulf of Carpentaria. These computations may not permit full comparisons with the results of Wright because (1) the nearshore bathymetry of the Gulf is poorly charted, (2) insufficient wave data exists to properly characterize one year's wave spectrum and (3) the shallow nature of the Gulf precludes the development of deep water waves. However, assuming that northeasterly waves of approximately a 6 second period and 1.5m height are representative of normal wet season conditions, then some comparisons may be made between wave power arriving at the inshore zone in the Gulf with those examples shown by Wright (1976). In addition, the same computations may be used to predict the relative increase in wave power at the shore during storm surges (Figure 2-7).

Wave power attenuation due to bottom interference is high in the Gulf of Carpentaria. Assuming that input levels are as high as 31 watts/cm of wave crest for the 6 second wave described above, then approximately 14% of this wave power reaches the inshore zone. However, with the addition of a 2.0m storm surge, up to 24% of this wave power reaches the inshore zone, an increase of by a factor of 1.73 over normal conditions. Temporarily raised water levels in shallow seas such as the Gulf may grossly alter the amount of wave power available for beach modification. If ancient epeiric seas may be judged by their Holocene analogues then traditional views of the processes operative in such environments might require revision.

REGIONAL GEOLOGY AND GEOMORPHOLOGY

Recognition of the Karumba Basin as a separate entity from the Carpentaria Basin was proposed by Doutch (1976). The Bulimba Formation of late Cretaceous to early Pliocene age rests unconformably on the lower
Coastal Region | $P_o$ | $\Delta P$ | $\Delta P'$
--- | --- | --- | ---
New South Wales, E. Australia | 178 | 6 | 3.4
Brazil, Atlantic Coast | 93 | 27 | 29
Cape Henry-Cape Hatteras, Atlantic Coast, U.S.A. | 35 | 17 | 48
Florida, Gulf of Mexico, U.S.A. | 2.2 | 1.3 | 58
Georgia "Sea Islands", Atlantic Coast, U.S.A. | 10.6 | 8.9 | 84
NW Queensland, Gulf of Carpentaria normal conditions | 30.8 | 26.5 | 86
storm surge 2.0m | 31.1 | 24.1 | 77

where:
- $P_o$ = deep water wave power in watts per cm of wave crest.
- $\Delta P$ = average absolute wave power loss in watts per cm of wave crest.
- $\Delta P'$ = average wave power loss in percent.

Figure 2-7 Wave refraction diagram for a portion of the southern Gulf using a 6 second, 1.5m high, deep water wave approaching from the northwest (315 deg.). Wave power loss due to friction is compared with the results of Wright (1976) and shows that although considerable power is lost due to the gentle nearshore gradient, a small increase in depth greatly increases wave power available at the inshore zone. Deep water wave power for the Gulf of Carpentaria appears higher than that for the Georgia or Florida examples because the wave selected for modelling in the Gulf does not represent average yearly significant wave power.
Cretaceous Rolling Downs Group of the Carpentaria Basin (Figure 2-8). The tectonism initiating deposition in the Karumba Basin is believed to be associated with the separation of Australia from Antarctica (see Doutch, 1976; Falvey, 1974; and Boeuf and Doust, 1975).

After deposition of the Bulimba formation unconformably on Carpentaria Basin sediments, planation and lateritization in the early Tertiary, probably Paleocene, (Doutch, 1976), produced the Aurukun surface. Later weathering of this same surface in the Pliocene finally resulted in the bauxite ore deposits mined at Weipa today (Grimes, 1977; Smart, 1977; and Doutch, 1976).

Faulting and warping in the late Tertiary produced the Gilbert-Mitchell trough (Figure 2-9), as well as uplift in the central Cape York Peninsula which provided sediment for the deposition of the Wyaaba Beds (Smart et al., 1972). Both continental and marine deposition continued during the late Tertiary development of the basin, which Doutch (1976) claims was coincident with and related to the Oligocene orogeny in New Guinea resulting from the collision of the Australian and Pacific plates.

The presence of a marine facies in the Wyaaba deposits supports the idea that the Gulf of Carpentaria assumed something like to its present form by the end of the Tertiary period, thereby providing the base level controlling erosion since the Pliocene (Doutch, 1976). The Kendall surface, distinguished by silicification and deep weathering, marked the end of this second period of planation (Grimes and Doutch, 1978). In the northwest sector of the peninsula the Kendall surface cut across the previously weathered Aurukun surface, thereby enhancing the development of the bauxite horizon. Later in the Pliocene, in the south and east of the Karumba Basin, the Kendall surface and its southern equivalents were removed to produce units such as the Yam Creek beds, Falloch beds and
Figure 2-8 A diagram modified from Doutch (1976) shows the generalized sequence of tectonic, depositional and weathering events during the Cenozoic history in the Karumba Basin.
Figure 2-9  Simplified geology of the Karumba Basin (from Doutch, 1976).
others (Powell et al., 1976). Doutch (1976) tentatively named the resulting ferruginized surface of this event the Campaspe surface.

Deposition in the Basin from the late Pliocene through to the present has been characterized by a series of alluvial fan units. These units can be divided into five episodes of fan growth influenced by climatic and sea level changes (Grimes and Doutch, 1978). The first two fan stages (Figure 2-8) are responses to tectonic, eustatic, and/or climatic events change (Grimes and Doutch, 1978). A sea level maximum during the Pleistocene (110-120,000 years BP) is associated with the third stage. This sea level stood a few meters above present sea level (Doutch, 1976, Smart 1976, and Chapter 5 this thesis). A eustatic drop in sea level during the last glacial probably reduced the Gulf of Carpentaria to dryness or at least to an inland lake. From approximately 37,000 to 11,000 BP (Smart 1977) subaerial exposure of older marine sediments promoted calcretion and lateritization. The Holroyd Surface (Doutch, 1976) developed during this period and is a climatic marker event which separates the first three fan units from the last two episodes of alluvial deposition (Figure 2-8). The Holroyd surface is characterized by aeolian deflation and deposition producing silt and sand dunes and choking drainage systems.

Rainfall increased following the development of the Holroyd surface and coincident with an initiation of a rise in sea level 11-12,000 years BP (Doutch, 1976). Renewed alluvial and coastal deposition resulted. This sea level rise proceeded until 6-7,000 years BP at which time sea level stood at or near its present level. Deposition of major fan units (Smart et al., 1978) continued in the face of rising base level. The final episode of deposition probably accompanied a fall in sea level since 6000 BP (Smart, 1976). Smart suggests that the marked increase in upland erosion and therefore coastal deposition since 4000 years BP can
be linked to increased aridity in northern Australia.

Physiographic nomenclature adopted by Ingram (1973), Needham and Doutch (1973 a and b), Simpson (1973), Grimes (1977), Smart (1977) was originally proposed by Doutch et al. (1973). Three of the physiographic units, the Holroyd, Mottle and Karumba plains, presented by Doutch et al. are subdivisions of the Carpentaria Plains described by Twidale (1966). Twidale used the term Karumba Plain to label the marine plain which lies seaward of the upland units from the southern Gulf to the Jardine River. Figure 2-10 from Smart et al., (in press) summarizes the physiographic units used in this discussion.

COASTAL GEOMORPHOLOGY AND MARINE GEOLOGY

The coastal lowlands and nearshore zone of the Gulf of Carpentaria remained largely unstudied until the Bureau of Mineral Resources and Geological Survey of Queensland mapped the area in 1969-1973 whilst preparing the 1:250,000 geological map series. Earlier workers who pursued geomorphologic investigations did little to integrate the coastal landforms with the time-equivalent fluvial landscapes.

The effect of summer flooding on the coastal plains was discussed by Whitehouse (1944). Whitehouse studied the development of meander channels in the lower reaches of the Gulf rivers and noted the centrifugal drainage typical of low relief coastal plains. Twidale (1956) identified distinct coastal landforms (i.e. dunes, mudflats, beach ridges and lagoons), but his analysis of topographic relationships along the Karumba transect was incomplete due to the lack of precise tide and levelling data. Twidale suggested recent emergence of approximately 20 feet with an associated dissection of terraces to explain the morphology.
Figure 2-10  Summary of physiographic units of Karumba Basin (from Smart et. al., in press).
of the Karumba Plain. A reconnaissance of the entire west coast of Cape York Peninsula by Valentin (1961) led to a division of the coast into distinctly different types. In addition, Valentin (1961) recognized two sets of beach ridge deposits, distinctly different in age. Auger holes through the beach ridge sequence at Edward River by Whitehouse (1963) showed that the youngest beach ridges rested on mud.

The first radiocarbon dating of strandline deposits from the Gulf was reported by Twidale (1966). This author re-iterated his earlier findings concerning physiographic relationships on the Karumba Plain and added stratigraphic data from Ingram (1973) who described the drilling of AAO No. 8 (Karumba). Twidale reported an age of 3320±125 years B.P. (Geochron GX0305) for shells recovered from the location of drillhole AAO No. 8.

Galloway et al. (1970) prepared a land systems survey report (CSIRO) of central Cape York Peninsula and summarized much of the work to that time. Research associated with the CSIRO Fisheries and Oceanography Division investigations of the tropical prawn provided a general characterization of nearshore and offshore sediments and morphology (Munro, 1972). In a follow up to Munro's work, Rhodes (1978) mapped the bottom sediments of the eastern Gulf noting a strong lithologic contrast between inshore and offshore sediments.

The deeper portions of the Gulf have been sampled by piston coring by Phipps (1966). He reported that the maximum sediment cover was 4.5 meters overlying a bedrock or laterite surface. In a later work, Phipps (1970) described this cover as consisting of both marine and non-marine facies. C-14 dating of these sediments established the oldest datable materials as 16,650-19,500 years BP. Phipps postulated that the alternation of marine and non-marine facies was associated with a rising sea floor concurrent with sea level rise, followed by a depression of the
sea bed to its present depth during the Holocene. Smart (1977), who reviewed data on offshore sediments in the Gulf and presented results of Bureau of Mineral Resources drilling at Cape Keer Weer Weer, offered an alternative interpretation of the data reported by Phipps (1979). Smart described the Gulf as a lacustrine environment separated from the Arafura Sea by a sill until flooded by the rising seas in the upper Pleistocene, thereby suggesting that tectonism was unnecessary to explain the Gulf bottom stratigraphy observed by Phipps.

Most recently, a large barrier island and chenier plain complex at Cape Keer-Weer (Figure 1-2) was investigated by Smart (1976). Precise topographic and stratigraphic control in conjunction with radiometric dating provided evidence for episodic progradation at Keer-Weer. Figure 2-11 summarizes Smart's stratigraphic findings which will be compared with those of this study in Chapter 4.
Figure 2-11  Stratigraphic sections across the coastal plain at Cape Keer Weer, 100km north of Edward River, see Figure 1-2 for location, (from Smart, 1976).
CHAPTER 3
MODERN SEDIMENTARY ENVIRONMENTS

GENERAL

The study of modern environments is frequently pursued to provide examples for use in the interpretive study of the stratigraphic record. A most desirable study situation is where there is continuity from modern environments into the deposits of their ancient counterparts. In such situations, the geometric and time-transgressive natures of facies boundaries are most clearly recorded. the Gulf of Carpentaria chenier and beach-ridge plains have this quality.

A sedimentary environment is "a part of the earth's surface which is physically, chemically and biologically distinct from adjacent areas" (Selley, 1970, p.1). An environment may be classified as one of erosion, equilibrium or deposition. Further classification can identify the environment as continental, transitional and marine (Selley, 1976). Since submarine depth zones are often difficult to establish from observations of the geologic record, Selley suggests that classification of marine environments ignore depth, and consider lithologic, paleontologic and sedimentary features.

The term "facies" as used by Gressly (1838), meant that similar lithological characteristics were linked to the same paleontologic assemblage. Teichert (1958) provided a review of the facies concept whilst Selley (1970) refined the definition of facies to include five parameters: geometry, lithology, paleontology, sedimentary structures and paleocurrent pattern. All of these parameters with the exception of paleocurrent patterns will be used to characterize the facies discussed in this chapter. Most sedimentary deposits of the Gulf of Carpentaria exhibit few current indicators thereby limiting the use of this parameter.

The low-lying land surrounding the Gulf of Carpentaria permits tidal
incursion beyond the coastal plain in most of the large rivers. There is a necessity to differentiate the portions of the Gulf rivers subjected to fluvial processes from portions subjected to tidal processes. In addition, many of the Carpentaria rivers are ephemeral and exhibit gross hydrologic contrasts between wet and dry seasons. Therefore fluvial environment makes reference only to those portions of rivers, streams, and creek channels which are beyond tidal influence throughout the year. In all cases, this places the fluvial environment on the upland surface where overlapping and anastomosing alluvial units provide a great variability in lithology. Only active channel deposits will be discussed because it is these sediments which are in direct transport to the tidal system and later, the open coast.

The tidal reaches of the rivers may be called estuaries only in the broadest sense of the term. "An estuary is a semi-enclosed coastal body of water which has free connection with the open sea and within which sea water is measurably diluted with freshwater from land drainage" (Pritchard, 1967). Difficulties arise with the application of this definition to the Gulf coastal areas when consideration is given to seasonal variation. During periods of low discharge in the dry season, there is minimum dilution of sea water by fresh. The estuaries of the Gulf are little more than tidal creeks and inlets. Conversely, high discharge during the wet season extends freshwater to the mouth of the larger rivers and tidal reversal is not observed (Staples, 1979). The river merely rises and falls in association with semi-diurnal tides.

Given these constraints, an arbitrary assignment of the term estuary is made to the portions of the Carpentaria rivers within the reach of dry season tidal influence. No differentiation is made in the case of smaller tidal creeks which empty directly into the sea. These bodies of water experience the same hydrologic and sedimentologic processes as the
larger rivers, and can therefore be considered small estuaries. In addition, many of these creeks have an ephemeral extension onto the upland surface, and therefore connected to the fluvial environment during the wet. However, some distinction must be made with regard to the subsidiary tidal creeks which branch off estuaries in order to drain the interfluve areas of the mudflat. Although these creeks represent an hydrologic extension of the estuary, they are of geologic significance to the mudflat. Such creeks drain water from the mudflat during the wet and play an important part in regulating mudflat processes during the dry. Therefore these creeks will be considered when discussing the high-tide mudflats.

Discussion of high-tide muds of the mudflat environment involves use of a term which is not totally appropriate in the Carpentaria area. The annual change in sea level which occurs between wet and dry seasons places the unvegetated mudflat in a supratidal position during the dry season. However, the inundation of these flats for 3-5 months of the year under wet season conditions is sufficient to turn the facies into a high-tide mud deposit.

Aeolian processes active on the high-tide muds during the dry season vary along the coast of the Gulf. On the southern chenier plain, this aeolian activity creates a dune environment in which a silt-clay facies is found. Although the material moves in sand and silt-sized clasts and pellets, granulometric analysis indicates a silty clay texture after deposition. The term "aeolian silts and clays" best describes this facies. On the more northern beach-ridge plain, aeolian mobilization of high-tide flats is uncommon.

Most facies used to outline modern deposition in the Gulf may be labelled by descriptive terms whose connotations with respect to environments of deposition are obvious. The study of these environments
provides an effective key to understanding the Holocene coastal stratigraphy, a record which is preserved laterally and vertically. The effective characterization of modern environments in a basin over 0.75 million square kilometers in area can be frustrated not only by the size, but the possibility of overgeneralization. Some facies may exhibit a within-facies variation of lithology and sedimentary structure which exceeds the between-facies variation. Occasionally, this variation has a seasonal component and in such cases an attempt has been made to catalogue these seasonal factors.

Characteristics of some facies are similar but not identical on both the chenier plain and beach-ridge plain. Other facies are present on one type of coast, but absent on the other. Therefore, the two different types of coast will be considered separately. A concluding section of this chapter bridges both coastal types and deals with the fingerprinting of facies by analytical means: shell assemblages, clay mineralogy and sand fraction size analysis.

The size of the study area and the diversity of modern environments necessitated some economies. Following the example of Thompson (1968) small but representative numbers (5-10) of samples were subjected to various analyses keyed to the environment. For example, size analysis of the sand fraction of a point bar mud may not be as essential as the analysis of sand in a tidal delta sand. Analysis of modern sediments were limited to those parameters which might prove distinctive when studying the Holocene morphostratigraphic record.

Granulometric statistics were determined by the method of moments applied to size analysis done by a settling tube. Roundness is described according to Powers (1953), sorting is classified by verbal classification of Folk (1965) and mineralogy of the coarse terrigenous fraction was determined by binocular and thin section examination. Other
analytical techniques which have been used (X-ray diffraction and radiographs, carbonate, soluble salt, and organic determinations) are discussed in Appendix A.

CHENIER PLAIN ENVIRONMENTS AND FACIES

The chenier plain is divided into 6 major depositional environments; Fluvial, Estuarine, Mudflat, Strandline, Nearshore and Dune. Table 3-1 presents the environment-facies relationships used in discussing the chenier plain.

<table>
<thead>
<tr>
<th>ENVIRONMENT</th>
<th>FACIES</th>
</tr>
</thead>
<tbody>
<tr>
<td>FLUVIAL</td>
<td>upland channel sands</td>
</tr>
<tr>
<td>ESTUARINE</td>
<td>tidal channel sands</td>
</tr>
<tr>
<td></td>
<td>tidal point-bar muds</td>
</tr>
<tr>
<td></td>
<td>tidal overbank muds</td>
</tr>
<tr>
<td>MUDFLAT</td>
<td>high-tide muds</td>
</tr>
<tr>
<td></td>
<td>intertidal organic muds</td>
</tr>
<tr>
<td></td>
<td>low-tide muds</td>
</tr>
<tr>
<td>STRANDLINE</td>
<td>shelly chenier sands</td>
</tr>
<tr>
<td>NEARSHORE</td>
<td>subtidal muds</td>
</tr>
<tr>
<td></td>
<td>ebb-tidal delta sands</td>
</tr>
<tr>
<td>DUNE</td>
<td>aeolian silts and clays</td>
</tr>
</tbody>
</table>

Figure 3-1 shows the relationships of modern environments outlined in Table 3-1 as they occur on a portion of the chenier plain. Figure 3-2 provides the locations of sample and photo sites used to characterize the facies resulting from these modern environments.

FLUVIAL ENVIRONMENTS, CHENIER PLAIN

The chenier plain between Burketown and Karumba is traversed by six major rivers which drain a watershed of approximately 75,000 square kilometers (Figure 2-10). Mesozoic mudstones in the watershed (Grimes
Figure 3-1. Modern environments on the chenier plain. Facies occurrence is shown in this generalized view of the plain.
Figure 3-2. Location of samples and photographs are shown here for the chenier plain.
and Doutch, 1978) provide fine-grained sediment for the suspended load of the rivers. Closer to the coast, the reworking of the Armraynald and Wondoola Plains sediments (Chapter 2) provide additional clay-sized material. Only the Norman River has access to sandy sediments in the southeastern portion of its drainage basin where the Norman and its tributary the Clara River gather sandy material from the Millungera and Claraville Fans (Figure 2-9).

The dry channels of the Bynoe and Flinders rivers during the dry season show a channel floor cut in the sandstones of the Normanton Formation. Channel lag deposits of cobbles and pebbles with occasional sand waves of medium to coarse sand litter the otherwise clean rock bottoms. The predominance of silt and clay in all facies of the chenier plain further confirms that fluvial sands are minor components of the chenier plain. Sampling of sand in transit to the sea was therefore limited to that which was already in the tidal or estuarine portions of the rivers and will be discussed in the next section dealing with the estuarine system.

ESTUARINE ENVIRONMENT, CHENIER PLAIN

Tidal channel sands

Fathometer profiles and bottom sampling in the Norman River between Normanton and Karumba (a river distance of 85km) permits description of the tidal channel sand facies in the estuarine portion of the Norman River. Cross channel profiles and locations of channel bottom samples are shown in Figure 3-3. In general, the data show an upstream clean bedrock channel becoming filled with sediment of decreasing grain size in a downstream direction.

The Norman River remains intertidal beyond Normanton, however, water depth at the Normanton bridge is less than two meters at low tide in the
dry season (Station 8, Figure 3-3). Grab sampling was unsuccessful at this location and yielded only broken fragments of ferruginized sandstone 8-10 centimeters in diameter. The irregular channel topography is probably inherited from scouring of the weathered bedrock.

Further downstream, at stations 6 and 7, there exists a smooth bottom of mixed silt and sand. Station 6 is unique, because it is one of the few locations between Karumba and Normanton where bedforms are exposed in the intertidal zone. Lunate megaripples oriented in an ebb direction with ripple lengths of 1.0 -1.5m and 0.2m amplitude cover the sandy point-bar at this site. Megaripples appear to extend down the point-bar ramp toward the center of the channel, although water turbidity limits complete observation.

Fathometer profiles at station 5 suggest large ebb-oriented sand waves of 0.5-0.7m amplitude on the floor of the channel. Channel sands here are muddy, with a 20% silt and clay. Although the thalweg topography at station 5 is believed to be sand waves, the topography at station 4 appears to be irregular bedrock, with sands infilling scoured holes. Grab sampling at station 4 was unsuccessful with the sampler striking rock several times.

Stations 3 to 1 are typical of lower estuarine channels with the channel broadening rapidly to over 1100 meters at station 1. The channel bottom is covered by muddy sands with no evidence of bedrock outcrop. Bottom topography is probably due to large bedforms created by ebb scouring. There is fining toward the mouth, where at Station 1 a deep basin up to 8m deep at low tide (Figure 3-3) collects fine material inside the ebb tidal delta. These basins are typical of estuarine mouths along the coast. They tend to fill with mud during the lower tidal current velocities of the dry season.
Figure 3-3. The Norman River between Normanton and Karumba is dominated by tidal processes. Cross section and thalweg profiles are shown here at 8 locations over the 85km distance.
Although there is an increase in muddiness of channel sands toward the sea, it must be noted that sampling was conducted only in the dry season. During these months, the estuarine waters remain turbid as silts and clays are redistributed in the estuarine system. Channel sands and coarser lags deposited during the wet may become mantled with fines during the dry. Thus the mud content of channel deposits could in part represent a sampling artifact.

In summary, the thin sands and gravels of the tidal channel bottom are underlain by an irregular bedrock topography which is scoured into prominent holes and troughs. The sands are medium-grained, angular to sub-angular with pitted surfaces typical of material derived from upland weathering. They are poorly sorted and composed of 90-95% quartz grains, 4-6% weathered rock fragments presumably of the Normanton Formation, and 1-4% pisolithic ironstone grains. The gravels are pebble to granule-sized and equally composed of large quartz and laterite particles. Carbonate is absent near the upland surface and increases to 6-10% near the mouth of the estuary. Carbonate material consists of broken molluscan fragments of indeterminate species mixed with whole shells of *Cerithidea largillierta*, *Nerita lineata*, *Caecella* sp., *Melarapte scabra* and *Melampus* sp. Shell content can be high (up to 20%) in sediments immediately downstream from chenier ridges which are being reworked.
Tidal point-bar muds

With the exception of the sandy point-bar deposit with megaripples near station 6 on the Norman River (Figure 3-3), point bars in the estuarine system of the chenier plain are composed of dark brown homogenous muds. This characteristic makes them atypical when compared with traditional facies descriptions of point bars. Discussions in the literature of point-bar deposits are usually typified by examples from fluvial and not tidal environments. Visher (1965) described the point-bar facies as a laminated deposit of sand and silt which graded upward to a ripple cross-bedded zone of finer sand and silt. Selley (1970) described point-bar deposits as characterized by micro-laminated and flat-bedded fine sands which grade upward into the floodplain facies. Reineck and Singh (1973) stated that the lithology of point-bar deposits depends largely on the material available. They suggested that a wide range of material permits upward decrease of grain size in the point-bar sequence.

Point bars on the Carpentaria chenier plain are composed of fine muds, several meters thick, probably as thick as the channel is deep. They have a smooth, featureless surface except for tracks and burrows. This surface extends down the point bar ramp toward the thalweg. The transition to channel sands is difficult to locate due to high water turbidity and inadequate sampling. The facies exhibits no sedimentary structures other than burrows and tracks by gastropods, crabs and worms. Point bar deposits approach viscous mixtures of mud and water, unable to support the weight of a man. Crab burrows appear to be infilled during each tidal inundation and side-slumping into steep sided channels is common.

Texturally, the facies contains less than 10% sand, up to 50% clay, most of this latter fraction in the sub-micron size range. Figure 3-4
Figure 3-4. Representative sampling of point-bar muds shows them to be mostly silt and clay with minor traces of calcium carbonate and organic material.
summarizes the lithologic qualities of these sediments. Sand-sized particles are all quartz grains or finely broken shell material. Clays occur in non-specific assemblages with variable amounts of amorphous material mixed with illite, montmorillonite and kaolinite.

Tidal overbank muds

Reineck and Singh (1973) describe levee deposits as ribbon-like sinuous bodies, triangular in cross-section and composed of finer grained material than point bars. However, on the chenier plain, levees are mound-like and lobate. They constitute a minor environment of deposition occurring on the outside or outbank of meander curves. Levees are composed of a clay-silt facies called the overbank muds.

Visher (1965) stated that levee deposits cannot always be differentiated from the upper point bar or flood plain deposits on which they rest. Selley (1970) combined levee, flood basin and swamp deposits in one category, describing them as fine-grained sands, clays and silts which are inter-laminated, cross-laminated and characteristically dessication cracked.

Tidal overbank deposits on the Carpentaria chenier plain exhibit characteristics of both the levee and crevasse splay deposits described by Reineck and Singh (1973). Tidal overbank muds are laminated light brown muds with few other sedimentary structures. Occasionally, they contain layers of fine sand 3.0 - 5.0cm in thickness which terminate on the mudflat in a small slipface. Although Kukal (1970) noted graded bedding as common to overbank deposits, none was observed on the chenier plain. The entire facies averages 1.0m in thickness. It occurs only as thin deposits on the outside of meander curves as shown in Figure 3-5. The facies grades into a thin layer of silts which merge with the mudflat surface 100-200m from the channel bank.
Tidal overbank muds are coarser than point bar muds, containing 40-60% fine sand with little clay, probably as a result of the high stages of flow associated with overbank deposition. Calcium carbonate is absent in levee deposits near the upland surfaces, but rises to 2-4% near the sea. Carbonate is in the form of broken pelycypod shells, too small for species determination. Organic carbon is present from root material of grasses growing on the levees and ranges from 1.0 to 2.0 percent. Soil salinities are low, 3-6 parts per thousand, encouraging nonhalophytic vegetation.

MUDFLAT ENVIRONMENTS, CHENIER PLAIN.

The literature contains many examples of mudflat environments zoned according to their topographic relationship to mean sea level. This topographic classification usually corresponds closely to a zonation of surface sediments by textural variation. Thompson (1968) presented an example of sediment zonation according to topography and morphology for an environment similar in many ways to the Gulf of Carpentaria.

The most extensively studied tidal flats (waddens) occur on the coasts of Northern Germany, the Netherlands, and Denmark (Kukal 1970). Reineck and Singh (1973) reviewed the impressive geographic variation of tidal flats pointing out that warm and temperate climates permit colonization by halophytic plants to develop salt marshes. Further reviews have been compiled by Ginsburg (1975) and Klein (1976) whilst Evans and Collins (1975) have taken a much needed step toward the quantification of sediment flux on intertidal deposits. However, these dominantly North American and European examples offer little preparation to deal with the mudflats of the Carpentaria chenier plain. Previous investigations on chenier plains are also of little assistance; Byrne et al. (1959), Greensmith and Tucker (1969) and Schofield (1960) reported a
marsh facies established in the supratidal zone with unvegetated mudflats limited to the intertidal zone.

The unvegetated flats discussed in most previous studies are all associated with arid coastal areas, but few have been discussed in the context of chenier plains. In the study of chenier plains, only Cook and Polach (1973) described extensive unvegetated flats at Broad Sound, Queensland. These might be similar to the Carpentaria mudflats. Broad Sound mudflats are above the reach of normal tides, and Cook and Mayo (1977) grouped them together with supratidal grasslands in a unit called "supratidal mudflats and coastal grasslands".

Coleman and Wright (1978) in their discussion of the macrotidal Ord River estuary identified high-tidal flats similar to the high-tide mudflats of Carpentaria. In the lower Ord system these flats are "the most extensive depositional feature within the delta complex" (Coleman and Wright, 1978, p. 637). The unvegetated, partially evaporitic surface of the Ord mudflats suggest they are morphologically and chemically analogous to the Carpentaria flats.

Kukal (1970, p. 264) has stated that two factors are distinctive of mudflats; 1. "Periodic desiccation of large areas of sediments during low tide", and, 2. "Marine origin of the sediments". An immediate marine origin for the Carpentaria mudflats is assumed. Although the flats are composed of terrigenous material introduced by local rivers, the Gulf has provided the shallow settling basin with wave and tidal action to accrete the flats. However, the issue of periodic desiccation is complicated by the annual change in mean sea level. Loosely speaking, the Gulf has one high tide lasting for 3-4 months with a low tide of 4-5 months duration for each year. The remaining intervening periods could be considered ebb or flood stages. Through use of the argument outlined above, the terms high-tide flat can justifiably be assigned to the extensive unvegetated
flats common to the chenier plain with the use of high-tide muds for the facies deposited on these flats.

The remaining mudflat environments can be separated into two additional environments and facies. Figure 3-1 delineates the facies distribution of the intertidal organic muds and the low-tide muds of the chenier plain. Each of these mudflat facies will be discussed in turn after description of the high-tide muds.

High-tide muds

High-tide flats are the dominant sub-aerial environment of the chenier plain, comprising 50-70% of the surface. Relief is slight, sometimes only 3.0cm in 1km with no regional slope. This environment produces a facies termed the high-tide muds. The high-tide flats could more appropriately be called a non-depositional environment. Lack of evidence for net erosion or deposition is unequivocally linked to the seasonal characteristics of this environment. Discussion of this sediment balance can only proceed after clarification of the seasonal variation.

As demonstrated in Chapter 2, the southern and eastern shores of the Gulf of Carpentaria experience an annual change in sea level of the 1.0 - 1.5 m range between summer and winter. Inspection of tidal observations for Karumba showed that for 180 day period 1 October – 30 March, 1965, the summer months of higher mean sea level, there were approximately 80 astronomically produced indundations of the high-tide flats. When these events are supplemented by meteorological events, the result is a saturated, poorly drained mudflat for almost 5 months. Water salinities in the wet season are as low as 2-4 ppt in the estuaries (Staples, 1979). The mudflat surface shows locally extensive scouring like that in Figures 3-6 and 3-7 due to ebb-tidal flow and overbank drainage of major rivers.
Figure 3-5. Tidal overbank deposits occur on the outside or cutbank of meander curves. Grasses on the top are similar to those on the stabilized clay dunes, mostly *Enneapogon arenicola*. The heavily cracked and fissured character of the high-tide muds may be seen below the scale. Scale is 0.5m.

Figure 3-6. Ebb-tidal flows after high spring tides of the wet season, scours the high-tide flats and produces the ebb-oriented features on the surface of the flat.
As the flats drain, especially at the end of the wet season, internal drainages develop in the scoured areas, producing saline "pans" such as that shown in Figure 3-8.

Blue-green filamentous algae cover large areas of the saturated mudflats during the wet. The mats are mono-specific *Lyngbya aestuarii*, an aquatic and terrestrial algae common to salt marshes and salt flats (I.J. Powling, written comm. 1976). *L. aestuarii* is an unbranched, sheathed filament, approximately 13 microns wide and characteristically grows in mats. Some of the mats are washed off the flats during unusually strong ebb tidal flows or floods, and a berm of rolled algal mats is deposited on the active beaches.

Algal mats are frequently associated with carbonate systems (Kukal, 1970; and Logan, 1961). Illing et al. (1965) reported *Lyngya aestuarii* on the sabkha of the Qatar Peninsula. The occurrence of extensive algal matting on the Carpentaria noncalcareous mudflats may suggest that algal laminae play a more important part in terrigenous systems than generally realized. Conybearse and Crook (1968), whilst relating modern environments to the rock-record, were careful to point out the association of noncalcareous algae to tidal flats bordering tidal marshes. According to Conybeare and Crook, this type of deposition is common in modern environments such as the Dutch coast of the North Sea.

The cessation of summer rains and the lowering of mean sea level at the beginning of winter isolates high-tide muds from further inundation. Evaporative processes convert the high-tide flat to an arid, hypersaline wasteland by mid-winter. The flat becomes deeply cracked as in Figure 3-9a and efflorescence of salts contributes to the mechanical disturbance of the sediment. Where algal mats are well developed and not removed by ebb tide or flood scour, a fluffy layer of clay, salt crystals and dried algal mat protects the surface (Figure 3-9b). Soluble salt content of the
Figure 3-7. Wet season runoff, especially after cyclones causes erosion through older chenier ridges. The water-filled depressions shown here are 1-2m deep below the bases of chenier ridges.

Figure 3-8. A closed depression on the high-tide flat shows the relict stumps of an unfilled tidal creek and the 1-3cm thick salt crust which develops in such depressions during the dry season.
layers shown in Figure 3-9b can be as high as 9\%, equally divided between NaCl and soluble sulphates. Insoluble or partly soluble sulphates (gypsum) vary from 4 to 7\% with organic material in the range of 2-3 \%. The surface is quite unlike the "chaotic" muds described by Thompson (1968) adjacent to the Gulf of California. Thompson's "chaotic" muds developed in an arid terrigenous system, and contained up to 80\% gypsum.

No systematic investigation of groundwater table oscillations was undertaken on either the seasonal or tidal cycle basis. However, it is apparent from the study of similar environments on playa lakes in central Australia that seasonal variation of groundwater table plays an important role in the salt efflorescence at the surface and the formation of gypsum in the lower layers (J.M. Bowler, pers. comm. 1978).

Groundwater (fresh) was noted at 1-2 meters below the surface in holes drilled in the dry season on the high-tide flat. Drying of the high-tide muds extends 30-40cm below the surface and the flats are easily trafficable by vehicles. Occasional unbroken algal/salt skins retain moisture in depressions on the flats or on the flanks of chenier ridges. In the case of the chenier ridges, ground water is seeping out of the perched storage layer above the level of the flat. The muds remain moist and plastic for the duration of the dry season when protected by such algal skins.

The X-radiograph in Figure 3-9b illustrates the effect of seasonal wetting and drying on the upper portions of the high-tide muds. The top 2-5mm show fine lamination due to the previous wet season's growth of algae and settling of silts and clays. Clay particles are aligned parallel to the surface of deposition suggesting a long undisturbed settling period during tidal flat inundation. Below this upper crust is a 2-3cm thick zone of slightly more disturbed clay and silt. Lamination is not so evident because this zone has been subjected to repetitive
Figure 3-9. Dry season desiccation of the high-tide muds produces wrinkled algal mats which cover the heavily cracked surface shown in (a). In (b), an X-radiograph across the section shown in (a) indicates the results of repeated wetting and drying. Laminations are present only in the upper layers where salt wedging and mechanical disturbance due to drying has not destroyed sedimentary structures remaining from the wet season.
seasonal wetting and drying. Algal fibers are less well preserved and the effect of salt wedging is marked. The next 4-10 centimeters contain large cracks and void spaces. The material is highly oxidized and disturbed by several years of mechanical activity associated with dessication. At the edge of polygonal blocks, deep V-shaped cracks extend 20-30cm into the sediments. Within these blocks, which are 10-20cm across, are horizontal partings which extend across the block and are frequently concave upward as shown in Figure 3-9b.

The high-tide muds contain considerable silt, sometimes up to 50% as shown in Figure 3-10. However, Figure 3-10 also shows that of the five samples selected for sub-micron size analysis (10 phi or smaller), several samples contained over 40% sub-micron clay. Organic content is variable, as a function of algal development. Recently deposited muds contain 1.5% organic matter by weight. Carbonate is absent or less than 1%, usually in the form of very small broken pelycytopod shells. The presence of fine ironstone pisolites washed onto the flat from the upland surface indicates a possible source of iron which forms characteristic mottling appearance in the upper 30cm.

Aeolian deflation of high-tide muds in late winter removes much of recently deposited algal and clay layers. These fine laminations contribute to the presence of organic material and oriented clay particles in the clay/silt dunes. This annual removal of recently deposited sediment is the major erosional process maintaining the elevation and form of the mudflat surface. Scour by ebb tide and flood flows such as shown in Figure 3-6 is important at the seaward margin of the flats and adjacent to the major rivers. However, aeolian deflation remains the key factor in uniform erosion of the surface everywhere on the chenier plain.

The critical difference between aeolian deflation and hydraulic
Figure 3-10. A lithologic summary of high-tide muds shows these to be somewhat finer than point-bar muds, with a tendency for clay to dominate. They represent a mixture of upland clays resulting from tropical weathering.
scouring is the respective destinations of materials removed. Aeolian deflation "locks up" windblown silt and clay into dunes (discussed later in this section) thereby precluding further tidal and fluvial transport of this sediment. Mud scoured by tidal or overbank flow is merely moved elsewhere on the chenier plain or into the subtidal zone where it can be redistributed by tide and wave processes.

Tidal creeks are sub-environments of the high-tide flats which do not produce a significant depositional facies. They are essential for the drainage of tide and flood waters from the flats and they frequently form the nucleus for aeolian accretion. The creeks are ephemeral, in that they are easily infilled by aeolian material and mangrove vegetation on their banks dies quickly when cut off from regular tidal inundation. The mangrove stumps in Figure 3-8 are remnants of an infilled creek, the only surficial evidence of its existence after being filled by windblown silt and clay.

A transitional mudflat separates the high-tide flat from the intertidal organic mudflat. Between Wildhorse Creek and Disaster Inlet (Figure 3-2) this flat is a debris strewn surface 5-10km wide with a seaward slope of 1:4000 (Figure 3-11). It is introduced here only because it is the next morphologic unit in a description progressing to the seaward. However, the facies deposited here represents a landward extension of the low-tide muds and will be discussed further in the section dealing with that unit.

Intertidal organic muds

Intertidal organic muds are a vegetation defined facies on which mangroves grow located between the high-tide flats and the low-tide flats (Figure 3-12). The organic content of these muds is much less than their name implies. The material on which mangroves grow contains only 2-3%
Figure 3-11. The debris strewn surface of the transitional mudflat is considered a landward extension of the low-tide flat. Vegetation in the background of this photo is on a chenier ridge near Pandanus Yard.

Figure 3-12. Several facies are shown in this aerial of the seaward edge of the chenier plain. Taken at mid-tide in the dry season, low-tide muds are inundated. However, intertidal organic muds are exposed, with a transitional mudflat joining them with the high-tide mudflat. Note the discontinuous modern chenier ridges formed on the transitional mudflat. Meander patterns of Wildhorse Creek are visible in the background.
organic material with greater amounts when large amounts of degraded wood fragments are present. The organic muds are rarely more than 0.3m thick and grade downward to alternating layers of sandy silt and clay, typical of the low-tide muds. Pneumatophores dominate the surface surrounding mangroves, but peat development has not been observed in an exposed coastal mangrove fringe. Stratigraphy presented in Chapter 4 will substantiate the absence of this facies from the Holocene stratigraphic record.

The environment is of ecological importance because it provides a habitat for a distinctive assortment of molluscan. Table 3-2 presents molluscan assemblage which constitutes a portion of the shell assemblage found in cheniers forming landward of the mangrove fringe.

Table 3-2

<table>
<thead>
<tr>
<th>Littorina scabra</th>
<th>Terebralia sulcata</th>
<th>Cerithidea largillieri</th>
<th>Nerita crepidularia</th>
<th>Melarapte scabra</th>
<th>Cerithium sp.</th>
<th>Cassidula sp.</th>
<th>Salinator Fragilis</th>
<th>Melaraphe scabra</th>
<th>Iravadia sp.</th>
<th>Stenothyra sp.</th>
<th>Telescopium telescopium</th>
</tr>
</thead>
</table>

Few of the above molluscan are incorporated in mangrove substrate after death. Instead, they are removed from the organic intertidal muds and transported across the transitional mudflat to strandline deposits immediately landward of the mangrove fringe. Therefore, relict strandlines give evidence of either the absence or presence of seaward mangrove fringes by their fossil assemblages.

Coastal mangrove fringes fronting the chenier plain lack species
zonation. *Avicennia* sp. presents an almost monospecific colony with some intermingling of *Ceriops*. *Rhizophora* grows in a more protected location at the mouths of estuaries, and it is only in such estuarine conditions that zonation is present.

Intertidal organic muds are frequently transgressed by silty and shelly sands from the low-tide flats. Figure 3-13 shows local mangrove die-back associated with this sort of onshore movement. This locality at the mouth of Wildhorse Creek has been an area of considerable change during the last 25 years. Visits to similar sites near the entrances to the Flinders and Bynoe rivers suggest that once the sand layers have transgressed the intertidal organic muds, killing the mangroves, the muds and mangrove stumps are quickly torn up by wave activity and all evidence of mangrove colonization is erased.

Data gathered from air photos of the mouth of Wildhorse Creek taken in 1951, 1966, and 1976 permit the summary of historical change shown in Figure 3-14. This summary emphasizes the relationship between changes in strandline development and the extent of mangrove fringe changes. Variations in the channel bank mangroves inside the estuary were minimal although the mid-channel island has become a fully established swamp. Landward migration of the chenier ridge west of the channel entrance is apparently responsible for changes in the extent and form the mangrove swamp.

The importance of onshore movement of coarse material (sand and shell) is utmost in causing mangrove die-back. This observation is supported repeatedly by inspection of other die-back sites external to the chenier plain (mouths of Staaten, Gilbert, and Mitchell rivers). Aerial inspection of the entire coast from Wildhorse Creek, east to Karumba and north to Weipa after cyclone Ted (December 1976) showed mechanical damage to mangroves was either limited or absent, but the
Figure 3-13 A mangrove forest at the mouth of Wildhorse Creek has been killed by the migration of a sand sheet across the low-tide flat.
Figure 3-14. A summary of 25 years of change, documented by three sets of aerial photos indicates the rate of change experienced by some mangrove swamps on the open coast.
onshore migration of sandy material was the single most obvious characteristic of coastal modification.

Controversy has surrounded the discussion of mangroves as a "land builder". The works of Vaughan (1910, Davis (1940) and Van Steenis (1941) are examples of opinions which tend to support the concept that mangroves have an active role in the progradation of coastal swamp or mudflat. In contrast, Watson (1928) held the view that mangrove colonization was only possible after the deposition of a suitable substrate. Quoting Watson, and referring to his own findings on the Tabasco, Mexico coast, Thom (1967) took the view that a two-way relationship exists between plants and landforms. However, Thom examined mangroves in a deltaic setting, a context which may not be entirely applicable to the Gulf of Carpentaria. Thom et al. (1975) and Thom (1975) contain further discussion of mangroves with respect to strandline communities. Thom et al. (1975) found that wave energy was readily transmitted through Avicennia permitting beach deposits to occur landward of mangroves, a process also noted by Jennings and Coventry (1973) whilst studying the Fitzroy estuary, Western Australia.

Mangrove development in three contrasting environments, the macro-tidal, seasonally dry Cambridge Gulf, the micro-tidal wet deltaic coast of Tabasco, Mexico and the low wooded islands of the northern Great Barrier Reef were compared by Thom (1975). His findings represent the best information to date concerning the relationships between mangrove ecology and geomorphology:

Given a climatic-tidal environment and a pool of mangrove species, each of which possess a certain physiological response to habitat conditions, it is considered that the history of land surface and contemporary geomorphic processes together determine the nature of the soil surface in which mangroves grow. Attributes of the substrate (...) are to a large extent a function of past and present geomorphic processes (Thom, 1975, p478).
The absence of intertidal organic muds as a relict or stratigraphic facies handicaps geomorphic study of coastal development in the Carpentaria chenier plain. It is a handicap with which few other workers in mangrove regions have had to cope. Mangroves on deltaic coasts produce a distinctive morphostratigraphic and stratigraphic facies. Investigations such as the one by Cohen and Spackman (1977) in Florida show the use of mangrove facies for paleoenvironmental reconstruction. They have developed fingerprinting guidelines for modern mangrove peats in southern Florida which are sufficiently distinctive to identify dominant species in ancient deposits.

Mangrove facies along open or semi-exposed coasts have been less extensively studied (Jennings and Coventry, 1973; Cook and Polach, 1973; and Cook and Mayo, 1977). These studies located in the Fitzroy estuary, Western Australia and Broad Sound, Queensland, revealed a distinct facies associated with mangrove growth (see also Thom et al., 1975). This distinction was based on a large amount of woody material incorporated in a dominantly silt and clay facies. Both these areas are macro-tidal in contrast to the Gulf of Carpentaria which is meso-tidal and wave-dominated. Wave dominated environments may seriously disadvantage the production and preservation of a highly organic facies. Other factors such as rapid progradation and sedimentologic instability could contribute to the absence of a coastal mangrove facies on the chenier plain.

The role of mangroves, not only on the Carpentaria coast, but on other coasts, is obviously not fully appreciated. Further attention may need to be devoted to zonation according to tidal inundation, passage and/or dissipation of wave energy during the wet season and the effectiveness of the mangrove as a sediment filter during ebb tide drainage of the high-tide flats. Figure 3-15 is a LANDSAT image showing
some of the continuous nature of mangrove fringe between the Bynoe River and Gore Point. With the exception of inlets associated with the Flinders River, Wildhorse Creek, Morning Inlet, and Disaster Inlet, there is no gap in the fringe for 100 km. The high-tide flat landward of this fringe is up to 50 km in width and therefore collects and stores considerable upland runoff and direct precipitation during the wet season. Scour processes have already been shown to be important in reworking sediment on these flats. This suspended mud is probably moved seaward during the draining of the flats. Quantitative assessment of mud accretion in this mangrove fringe is complicated by its inaccessibility during most of the year.

The foregoing does not constitute an endorsement of mangroves as the land builder on the chenier plain. It does call attention to a possible role of mangroves as a sediment trap, a concept which has here-to-fore been hinted, but rarely described or quantified with respect to mangrove swamps on an open coast. It is possible that mud accretes at the mangrove fringe as it moves seaward off the mudflats. However, such accretion might proceed with or without the presence of mangroves provided topography of the mudflats remained the same.

Low-tide muds

Low-tide muds are the facies produced on the low-tide flats (Figure 3-1). These flats occur seaward of cheniers and mangrove fringes and are subject to tidal inundation throughout the year. They slope seaward approximately 1 m per km and may be gently undulating and ripple marked when sandy but flat and featureless when muddy.

Ripple marks are common during the winter months on sandy portions of the low-tide flat. Ripple length varies from 0.1 m to 0.2 m with ripple height in the range 1 - 3 cm. Crests are curved and strongly bifurcated
as shown in Figure 3-16. Ripples are symmetrical to slightly asymmetrical and often have rounded crests due to modification during ebbing tides.

Water is ponded in shallow troughs on the flats during low tide. Sediments in these troughs tend to be slightly muddier than on the better-drained crests such as in the foreground of Figure 3-16. Water birds forage in and out of these troughs during low tide leaving claw marks superimposed on the ripple marks.

Low-tide flats in the eastern portion of the chenier plain near the mouths of the Norman, Bynoe and Flinders rivers show little burrowing and no evidence of live shellfish. However, between Disaster Inlet and Morning Inlet (Figure 3-2), limited colonies of *Mactra* sp. are present on the low tide flat, especially in sandy zones. In addition, gastropods, *Thais costata* and *Tympanotonus lavardi* are common; they are probably distinctive of this environment. *Tympanotonos*, which is smaller of the two, is an aggressive "traveller" at low tide and ploughs small circuitous tracks in sands and silts several millimeters wide and deep. These tracks are similar to those produced by worms in marine sediments, supporting the view of Conybear and Crook (1968) that many tracks in the rock record attributed to worms, may actually be from gastropods. Whole shells are rarely deposited in this environment, instead they are transported across the low-tide flat to either the mangrove fringes or beach. Near Pandanus Yard, east of Disaster Inlet (Figure 3-2), the transitional mudflat between the mangrove fringe and the chenier ridge, a distance of 500m, is littered with bioclastic material (Figure 3-17). The substrate below this surface contains no similar quantity of shell material.

Low-tide muds show a considerable variability in texture and color. They range from light brown, fine sandy silts to dark brown, silty clays.
Figure 3-16. Low-tide flats near the Karumba transect show the slight undulations in the surface and the low seaward gradient. The low tide swash line is over 500m beyond the man shown walking on the flats.

Figure 3-17. A transitional mudflat near Pandanus Yard immediately landward of the mangrove fringe shows a shell littered surface on which seedling and juvenile *Avicennia* sp. are growing.
with 40% of the clay in the submicron range. This variation is often a seasonal one, with different textures occurring at the same place in different months. For example, Figure 3-18 shows a shallow section typical of the alternate layering of coarse and fine sediments associated with low tide muds. Sandy layers are thinner (1-4cm) than muddy layers which are 2-6cm in thickness. Sections in the mudflat show combinations of flaser and wavy bedding. The preservation of ripple forms on the upper surface of the sandy layer probably occurs during rapid accumulation from muddy waters at the beginning of the wet season. Silt and clay deposition proceeds during this period of high terrigenous input in the wet, followed by deposition of sandy material in the dry. This sequence is similar to the model suggested by Reineck and Singh (1973) for deposition of flaser bedding in other tidal environments. However, the period of alternation is an annual one in this case, rather than being diurnal. The lengthy period of alternation probably makes the layers more continuous and similar to wavy bedding. In addition, there is superimposed on the annual cycle of events, an alternation of coarse and fine sedimentation with a period expressed in years. Local mangrove die-back in sites such as those near Wildhorse creek and the Bynoe river may be caused by "packets" of sand transgressing the intertidal zones from either adjacent parts of the coast, or the sub-tidal zone. Textural variation of the low-tide muds may not be simply a function of seasonal change.

The transitional mudflat, introduced earlier during description of the high-tide muds is genetically and textually an extension of the low-tide mudflats. All facies occur in a topographic framework related to tidal inundations. Therefore the relative widths of the transitional mudflat, the mangrove fringe and the low-tide mudflats, are a function of the slope of the intertidal apron. Furthermore, the boundary between the
transitional mudflat and the high-tide mudflat is poorly defined in the absence of a chenier ridge. However, in the presence of cheniers, it is clear that the high-tide mudflat occurs landward of the chenier whilst the transitional mudflat occurs seaward of the chenier. Finally, it will be shown in the next section that the modern chenier ridges are constructed on the transitional mudflat.

STRANDLINE ENVIRONMENT, CHENIER PLAIN.

Shelly chenier sands

The term "chenier" was discussed in detail in Chapter 1. Chenier ridge sediments can be gravel (Koster, 1955), sand and shell (Greensmith and Tucker; 1968, and Byrne et al., 1959), or entirely shell and shell fragments (Brouwer, 1953). Whatever their composition, the ridges are always markedly coarser than the base on which they rest.

Discussion of Carpentaria cheniers is divided into a description of modern or actively forming ridges, and relict cheniers. Such division is necessary because certain processes, especially carbonate leaching and recrystallization, continue after the strandline becomes "perched" or isolated from the modern coast.

Active cheniers are separated from the sea by a fringe of mangroves, except near river mouths. Figure 3-12, showed an active chenier over 2000 meters landward of the mangroves 5km west of Wildhorse Creek. Incipient shell ridges, such as in the center of Figure 3-12, develop on the transitional mudflat between the mangroves and cheniers, but constructional processes continue on the larger chenier to the landward. Only after these incipient ridges become continuous does the landward chenier become relict and inactive.

Adjacent to river mouths, such as the Norman or Bynoe rivers, chenier ridges adjoin the low-tide flat (Figure 3-19). Sediments in these
Figure 3-18. A section cut through low-tide muds shows the alternate layering of coarse and fine sediments characteristic of this facies. The alternation is caused by annual variation of sediment type being deposited on the low-tide flat. Scale divisions are 10cm.

Figure 3-19. A sandy chenier west of the mouth of the Norman River adjoins the low-tide flat. Sand for the formation of this chenier is provided by wave sorting of sediment in the ebb-tidal delta of the Norman River, just to the right of this aerial view.
ridges are dominantly terrigenous sand (up to 70-80%). Sandy cheniers may laterally grade to a shelly facies away from tidal inlets. However, the general rule holds that sandy cheniers form adjacent to inlets and are separate from the mangrove fringes. Regardless of where the ridges form, their crests are up to 3m above the high-tide flat and approximately 6m above prediction datum. Length varies from a few hundred meters to tens of kilometers whilst width ranges from 50m to 150m.

Initial colonization on cheniers is by of *Casuarina* sp. *Casuarina* suffers a high mortality rate even among mature trees, when high energy wave events cut into the beach and expose the roots. Mechanical damage or removal of the tree itself is also critical in the decimation of pioneer communities during cyclones.

Relict cheniers of Holocene age are distributed across the high-tide flat between Karumba and Burketown (Figure 3-2). Radiocarbon ages of these landforms will be discussed in Chapter 5. The best developed ridges are in the eastern portion of the plain, except where the Norman, Flinders and Bynoe rivers have cut eroded older ridges so that some relict strandlines may only be defined by the connecting of ridge remnants. These remnants are isolated shelly mounds perched on the mudflat. They are 100-300 meters long, and rarely more than 50 meters wide. Aeolian silt and clay deposits, flank the ridge margins.

A well developed series of cheniers remains preserved between the Flinders and Bynoe rivers (Figure 3-2). These ridges are 5-10km in length, 50-250m wide and rise 3-4m above the high tide flat. They are composed of whole or broken molluscs with varying amounts of silt and sand. Bioclastic content can be as high as 75% by weight, but field estimates by volume from localities at Port Norman rise to 95% (Figure 3-20).

The best information on lithology, geometry and sedimentary
structures is available from scour cuts through relict cheniers such as that shown in Figure 3-21. The crest of the ridge is 5.0 meters above mean low water. This section is cut in a relict ridge approximately 500m from the modern beach in the interfluve area between the Norman and Bynoe rivers. Soil development is seen in the upper 20-25cm on top of which non-halophytic grasses and some upland shrubs grow, notably Parkinsonia sp. Below this soil zone is 90-100cm of composite bedded sand and shell layers. The shell assemblage consists of Anadara sp., Mactra sp., Turritella terebra, Trisidos yongei, Patro australis, Murex coppingen, Gaet lessoni and Tellina sp. (Figure 3-22). It shows the nearly horizontal bedding typical of crestal portions of the chenier. Individual beds are sometimes continuous for tens of meters and appear to be draped over the lower portion of the ridge, showing seaward dips of 3-5 deg. on the seaward flank and steeper dips to 10 deg. with crossbedding on the landward side.

Below the layered sand and shell zone is 65-75cm of finely bedded shelly sands with steeply dipping landward crossbedding (15-20 deg.). The top of this cross-bedded zone is approximately 3.6m above mean low water, which is equivalent to the level of maximum wave runup during the winter season. This is the zone which is transitional to the low tide mud facies which appears in Figure 3-20 at the base of the scale. Below this transitional zone are the finely interbedded sandy silts and clays with flaser and wavy bedding typical of the low-tide muds which extend landward to form the transitional mudflat.

Cheniers which formed 2-3km away from estuarine inlets are less sandy than that shown in Figure 3-21 and more typically appear like the ridge in Figure 3-20. Landward dipping shell beds (5-8 deg) of shells characterize the back section of the ridge. Crossbedding and continuity of individual beds are difficult to observe in the very shelly section,
composite bedded sand and shell

cross bedded shelly sands

low-tide muds

crest

soil development

Figure 3-20. A shallow section through a chenier at Port Norman illustrates the high shell content typical of most cheniers.

Figure 3-21. Wet season erosion has breached a relict chenier near the mouth of the Norman River providing information on internal structure. See text for detailed discussion. Scale divisions are 10cm.
partly due to the large size range of the shells and the lack of any distinctive imbrication. Dominant species are *Anadara* and *Mactra*.

Carbonate mobilization is present as evidenced by the chalky white coating on all shells older than the modern beach. Some ridges show a layer of calcite cemented sand and shell 1 to 2 cm thick at their bases overlying the clays. Ridges on the inner portion of the plain have well developed soil profiles 0.5-0.8 m thick, from which most carbonate has been leached.

Dense thickets of *Cryptostegia grandiflora* are mixed with upland trees and shrubs, dominantly *Atalaya hemiglauca*, *Grevillea striata*, *Parkinsonia sp.*, *Eucalyptus pruinosa*, and *Eucalyptus dichromaphloia*. Upland species appear to survive because of the increase in height of ridge crests away from the sea. The crests of the most landward ridges are 6.8-7.5 m above prediction datum whilst their bases are 5.0 m to 5.3 m above datum, an important observation to be considered in later discussions.

Strandline and bare mudflat environments dominate the overall morphologic appearance of the chenier plain. Deposition in these environments is closely related to tidal inundation, annual sea level change and catastrophic high energy events. Figure 3-23 summarizes the topographic relationships between these adjacent environments and tidal inundation. Topography shown in Figure 3-23 is a synthesis of over 500 elevations observed on the chenier plain. All of these observations were related to a permanent benchmark at Karumba which was in turn related to tidal observations. Facies occurrence is closely controlled by the frequency of inundation. It is apparent from Figure 3-23 that the base of modern cheniers is clearly defined by a facies change which occurs around the median value of 3.0 m above mean low water. The surface of the high-tide flat averages 3.3 m with a standard deviation of 0.1 for 68
Figure 3-23. Modern depositional environments bear a definable relationship to prediction datum and mean sea level. Surveyed site data in the Karumba transect are summarized here.
observations. Elevation of the base of the aeolian deposits is closely controlled by the level of the flat on which they develop: the high-tide flat. Elevations of the dune crests is probably limited by the competence of the wind to accrete upward rather than outward. The elevations of facies changes noted near or on the modern coast are central to the discussion of relative sea level change to be discussed in Chapter 6.

NEARSHORE ENVIRONMENT, CHENIER PLAIN.

The nearshore environment consists of ebb-tidal deltas and a subtidal zone (Figure 3-1). These two depositional environments produce their respective facies ebb-tidal delta sands and subtidal muds (Table 3-1).

Ebb-tidal deltas

Ebb-tidal deltas are composed of shelly and silty sands occurring at the mouth of every estuarine channel of the chenier plain. The ebb-tidal deltas are of various shapes, but nearly all have a central distributory channel from which smaller channels branch. Figure 3-24 shows the outline of tidal deltas at the mouths of the Norman, Flinders and Bynoe rivers and Wildhorse Creek, of which the Flinders exhibits the best developed delta.

A characteristic intertidal levee extends seaward from the low-tide mudflats adjacent to major estuarine inlets. These levees are 1.0 - 1.5m in height and slope away from the main channel for 250-500m. Between 750 and 2000m in length, they are frequently broken by smaller distributary channels near their seaward margins (Norman River on Figure 3-24). Occasionally, as in the case of Wildhorse Creek, the distal distributary channels become so sub-divided that no main channel is present. This
Figure 3-24. Ebb-tidal deltas occur at the mouth of all major rivers and inlets. This LANDSAT image (Band 7, far infra-red) taken in December 1972, shows tidal delta morphology for the Norman, Bynoe and Flinders rivers. The transitional mudflat landward of the mangrove fringe shows evidence of inundation as a result of wet season rise in mean sea level.
morphology is somewhat in contrast with typical ebb-tidal delta and inlet morphology common to sandy coasts. For instance, Hayes et al. (1973) present an ebb-tidal delta model which contains the components ebb channel, channel margin bars, terminal lobe, and swash platforms. The Carpentaria ebb-tidal deltas lack a well developed terminal lobe and the swash platform is usually incised by lateral distributaries. In the Hayes model, such channels are flood produced, whereas on the chenier coast, these appear to be ebb produced. The arcuate swash bars, so common to the type of delta studied by Hayes are totally absent in the southern Gulf where wave modification appears to be minimal. Hayes (1973) suggests that ebb-tidal delta morphology is dependent on the time-velocity asymmetry of tidal currents. His model is developed in areas where the estuaries have a large tidal storage and tidal exchange occurs in a well-developed bi-directional manner through the tidal inlet. In addition, such estuaries tend to experience a net infilling of sediment by growth of their flood tidal deltas. In contrast, the Carpentaria inlets exhibit a net seaward flux of sediment especially during the dominantly uni-directional flow of major rivers during the wet season. The resulting ebb-tidal delta form observed on the chenier plain is similar to that described by Wright (1977) as a "friction dominated effluent." Although Gulf ebb-tidal deltas exhibit middle ground bars which are more subdivided than those of Wright's "type B" model, the general morphology and sediment distribution are analogous.

Tidal delta lithologic characteristics also show considerable internal variability. Composite layering of alternate fine and coarse bedded material similar to the flaser and wavy bedding of the intertidal muds is characteristic of these deposits. Figure 3-26 shows the type of interlayering common to this facies. Sand and shell layers may be 10 cm to 30 cm thick and mud layers vary from 5 cm to 15 cm. Reineck and Singh
I. term such conditions as "coarsely interlayered" and explain how this may happen. Their example which is most applicable to the Gulf situation involves occasional sand transport in an environment dominated by mud deposition. Observations of sedimentary processes in the Gulf of Carpentaria show that sand is deposited under wave-dominated conditions and mud is deposited under tide-dominated conditions. This alternation of texture according to dominant conditions is not strictly a wet/dry season phenomenon, as winnowing and redeposition of sand and shell is common during the short period waves typical of late winter and early spring.

Figure 3-25 summarizes the textural characteristics of ebb-tidal delta sands. These sands show marginally better sorting than tidal channel sands and the presence of more rounded grains suggests mixing in this environment. Calcium carbonate can rise to 80% in the shelly layers and drop to less 1% in mud layers.

Live shellfish communities have not been located on intertidal portions of the ebb tidal deltas. Whole or broken shells present in the sediment show evidence of considerable abrasion and therefore may have been transported some distance from their habitat. Species determination is difficult on broken shells, but whole shells consist of Architectonia sp., Solen vagina, Trisidos yongei, Murex cuppeningen and Turetella terebia.

Subtidal muds

Subtidal sediments which are adjacent to the chenier plain are dominantly silt and clay. The facies is referred to as subtidal muds and is present on a rather featureless gently dipping sea floor with gradients less than 1:4000 (few depths in excess of 20m are within 100km of the coast). The distribution of subtidal muds and their relationship
Figure 3-25. Ebb-tidal delta sands have a lithology which results from the mixing of broken shell with medium grained sands.

Figure 3-26. Sand and shell layers are typical of the lithology of ebb-tidal delta sands. Coarse layers alternate with mud layers due to varying wave conditions during the year.
to offshore sands was discussed by Rhodes (1978). Bottom sediments in this portion of the Gulf were shown to be extremely muddy, with offshore sand only found in the central and northern Gulf. This subtidal facies will be discussed later in a broader context than the chenier plain.

DUNE ENVIRONMENT, CHENIER PLAIN.

Dunes composed of fine silt and clay are present in various environments throughout the world. Price and Kornickef (1961) reported clay and silt dunes derived from lagoonal sediments on the Texas coast. Australian inland playa lakes with their associated ancient and modern lunettes and clay dunes have been discussed in detail by Bowler (1973). Inland and coastal sebkas are known to host a significant windblown facies (Kinsman, 1969; Glennie, 1970; and Reineck and Singh, 1973).

Aeolian silt and clay on the chenier plain comes from the mobilization by wind of the high-tide muds and therefore some lithologic similarities exist between the aeolian silt and clay and its parent facies. It is the surficial "skin" of the high-tide flats which is mobilized. This skin contains a high salt content so that such dunes, may consist of as much as 25% salt. Two thirds of this salt is NaCl and the remainder is soluble and insoluble sulphates. The organic content is less than 1.0% whilst calcium carbonate is either absent or present as finely broken pieces of shell. Figure 3-27 shows that in samples selected for sub-micron size analysis, up to 50% of the material may be in this fraction. However, size distribution varies as a function of source. Dunes adjacent to the high-tide flats near L Creek (Figure 3-2) contain over 50% fine sand due to the fluvial contributions to the mudflat facies in this area. Closer to the coast, along the same transect, near the mouth of Wildhorse Creek (Figure 3-2), the dunes are predominantly silt and clay. Similarly, near the mouth of the Norman
Figure 3-27. A lithologic summary of the dune facies shows that salts, silt and clay are major components of aeolian deposits on the chenier plains. Furthermore, active dunes near the sandy upland surface reflect the availability of fine sand in this area, whereas, dunes at the seaward edge of the plain are dominantly clay. An X-radiograph of a core taken from a stabilized dune demonstrates the homogeneous nature of dune sediments after several seasons of alternate wetting and drying.
River (Figure 3-2), dunes contain over 50% clay where they are actively forming behind the modern beach (Figure 3-28). The dunes at this locality are typical of active dune formation behind chenier ridges. They develop most extensively on both landward and seaward sides of the cheniers, forming a hummocky vegetated flank covered entirely with a grassland of Enneapogon arenicola. These undulating dune fields become sediment sinks after stabilization by vegetation. The salt-free foothold offered by this supratidal position, encourages rapid growth of vegetation which is effective in trapping further sediment during the next dry season (Figure 3-29).

Individual dunes develop in one season to heights of 1.5m above the high-tide flat. Topographic levelling along the Karumba transect (Figure 3-2) showed that the undulating dune fields were almost as high as the ridge crests. In the interfluve areas between the Norman, Bynoe and Flinders rivers, the dunes are found on both the seaward and landward sides of the relict cheniers Figure 3-30. These aeolian flanks may be up to 1500 meters wide and as long as the cheniers.

The older dunes have a structureless homogeneity as shown in the X-radiograph in Figure 3-27 and after several seasons of alternate wetting and drying the surface becomes indurated. Salts, including gypsum, and carbonate are apparently leached from these silts and clays during the first wet season following deposition.

SUMMARY

The study of modern environments and their resulting facies provides more than just a basis for the interpretation of the Holocene stratigraphic record. Some environments on the chenier plain do not produce a significant depositional product but are important in understanding the morphology and development of other environments. The
Figure 3-28. Actively developing clay and silt dunes near the mouth of the Norman River are building on the landward flank of an active modern chenier. The Gulf is to the left behind the vehicle.

Figure 3-29. Woody and herbaceous plants provide a trap for rapid accretion of clay and silt dunes across the entire chenier plains. These dunes are forming 50 kilometers from the sea, near the upland surface. Southeast winds from the left background of the photo are deflating high tide flats seen in the distance.
entire chenier plain remains a site of dynamic processes despite seaward progradation of the active coast and apparent "stranding" of some features.

It is apparent that of the high-tide flats form an environment of net non-deposition in equilibrium with tidal and aeolian processes. This maintenance of equilibrium is key to the present morphology of the chenier plain. Without it, silt and clay dunes would be less well developed, the annual change of sea level would inundate less of the plain, and storage of wet season waters on the high-tide flat would be considerably reduced. Figure 3-31 summarizes the processes essential to the maintenance of the high-tide flats in their present state of equilibrium.

Certain other important factors must be incorporated into a summary of chenier plain environments. These observations are key to interpretations made in later chapters, especially those dealing with relative sea level and episodic progradation.

a. The estuarine system is markedly different in the wet and dry seasons. During the wet, ebb flow dominates with both suspended and bedload transport being important. Dry season estuarine processes are limited to tidal flushing of main channels and tidal creeks with further distribution of the suspended load.

b. The high-tide flats are in equilibrium with modern processes. The flats are maintained at uniform elevation from the sea to the upland surface by alternate inundation involving deposition and erosion during the wet and deflation during the dry.

c. Chenier deposits are built by wave processes during the high sea level associated with the wet. The base of this facies is sharply defined where it overlies low-tide muds. This facies change is an important sea level indicator.
Figure 3-30. Dunes have developed on the landward and seaward flanks of this chenier. Southeast winds during the winter are modified by a land and sea breeze system to permit aeolian deposition on both sides of sediment traps such as cheniers or tidal creek mangrove fringes.

Figure 3-31. A summary of the processes by which the high-tide flats are maintained at an equilibrium condition with respect to tidal deposition and aeolian deflation.
d. Mangrove fringes do not produce a lasting facies beyond the life of the ecologically defined environment. Mangroves are present only where geomorphological conditions permit their growth and these plants exercise little self determination over their own physical environment.

e. Tidal deltas are only created by ebb-tidal flows. They show seasonal variation in geometry and lithology. They are repositories for the small amount of terrigenous sands introduced by rivers until wave processes gradually redistribute these sands into the subtidal, intertidal and strandline environments.

f. Clay and silt dunes are important sediment sinks of fine material deflated from the high-tide flats. In addition, they add breadth and mass to relict chenier ridges, preventing complete reworking of these features during wet inundation of the flats.

BEACH-RIDGE PLAIN ENVIRONMENTS AND FACIES

Most of the environment/facies relationships outlined in Table 3-1 can be applied to the beach-ridge plain. However the content of Table 3-1 requires some minor modification before it can conveniently encompass all the needs of environment/facies description on the beach-ridge plain. Chapter 2 outlined the general geomorphology of the beach-ridge plain and its regional geologic setting. It is this regional setting and its geomorphic characteristics which necessitates revisions to the previous environment/facies organization.

The upland behind the beach ridge plain is mantled with overlapping and interfingering Cainozoic alluvial fans which because of the high average annual rainfall (Figure 2-1) contribute a massive amount of quartzose sand to the bedload of the rivers and streams which flowing across it. The result at the coast is a mixed terrigenous-carbonate
system in which terrigenous muds and sands are mixed with shells to produce a prograding coast quite different from the chenier plain in the southern Gulf. Wave-induced sorting is further facilitated by nearshore gradients of approximately 1:2000 which contrast with those of 1:4000 seaward of the chenier plain. Consequently there is a narrow subtidal mud zone with coarse offshore sands occurring close to shore, especially near large river mouths.

Mangroves are rarely present outside of the estuarine environment. Within the tidal rivers and creeks they form forests flanking the channels and contribute to the development of a mangrove peat facies. These forests help stabilize channel banks and their presences precludes the formation of an overbank mud facies such as found on the chenier plain. The unvegetated high-tide mudflats which are dominant on the morphology of the chenier plain are less extensive on the beach-ridge plain. An additional facies, the supratidal silts and clays, is used to describe the vegetated coastal flats which lie slightly above the high-tide mudflats. Table 3-3 summarizes comparisons between environment and facies relationship on the chenier and beach-ridge plains. Figure 3-32 indicates the spatial relationship of the environments whilst Figure 3-33 shows the locations of samples and photos discussed in this section.
Figure 3-32. Modern environments on the beach-ridge plain. Facies occurrence is summarized in this generalized view of the beach-ridge plain.
Figure 3-33. Sample and photograph locations are shown in this map of the coast between the Kendall River and the Mitchell Rivers.
Table 3-3

<table>
<thead>
<tr>
<th>ENVIRONMENT</th>
<th>FACIES</th>
<th>ENVIRONMENT</th>
<th>FACIES</th>
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<tr>
<td></td>
<td>CHENIER PLAIN</td>
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<td>BEACH-RIDGE PLAIN</td>
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<tr>
<td>Fluvial</td>
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<td>upland channel sands</td>
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<tr>
<td></td>
<td>tidal point-bar muds</td>
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<td>tidal point-bar muds</td>
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<td></td>
<td>tidal overbank muds</td>
<td></td>
<td>mangrove peats</td>
</tr>
<tr>
<td>Mudflat</td>
<td>hightide muds</td>
<td></td>
<td>supratidal silts and</td>
</tr>
<tr>
<td></td>
<td>intertidal organic muds</td>
<td></td>
<td>clays</td>
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<td></td>
<td>low-tide muds</td>
<td></td>
<td>high-tide muds</td>
</tr>
<tr>
<td></td>
<td></td>
<td></td>
<td>low-tide muds</td>
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<tr>
<td>Strandline</td>
<td>shelly chenier sands</td>
<td></td>
<td>beach-ridge sands</td>
</tr>
<tr>
<td>Nearsheore</td>
<td>ebb-tidal delta sands</td>
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<td>ebb-tidal delta sands</td>
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<tr>
<td></td>
<td>substidal muds</td>
<td></td>
<td>substidal muds</td>
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<td></td>
<td></td>
<td></td>
<td>offshore sands</td>
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<tr>
<td>Dune</td>
<td>aeolian silts and clays</td>
<td></td>
<td>aeolian silts and clays</td>
</tr>
</tbody>
</table>

FLUVIAL ENVIRONMENTS, BEACH-RIDGE PLAIN

Rivers and streams crossing the beach-ridge plain are smaller and more numerous than those crossing the chenier plain. These rivers drain an area in which sandstone lithologies are dominant, although the headwaters of the larger rivers such as the Archer and Mitchell originate in the granitic uplands of Cape York Peninsula, some within 100km of the east coast of the Cape. Even rivers as small as the Holroyd or Edward collect water and sediment as far as 150-180km from the coast where the continental divide is 160-200m above sea level producing an average stream gradient of 1:1000.

Upland river channels of major rivers such as the Archer and Mitchell are straight, whereas medium-sized and smaller rivers like the Edward, Coleman and Alice have anastomosing channels. The ephemeral nature of fluvial flow permits inspection of the channels during the dry season. Large sandwaves of 10-100m length and with 0.5 - 1.0m amplitudes cover the channel bottom. The main channels are incised 5-15m below the
surface of the alluvial fans. Wyaba Beds, a Pliocene or older sandstone and conglomerate, frequently outcrops on the floor of the main channels. The lower reaches of the fluvial system, within 10km of the estuarine system, are characterized by channels choked with sand, and become a series of poorly connected freshwater ponds during the dry season.

Upland channel sands

Upland channel sands were sampled at 15 localities (Figure 33). They are dominated by poorly rounded to angular quartz grains showing highly pitted irregular surfaces under binocular microscope examination. Some grains show iron staining which passes through fractures to the interior of the grain. This red coloring is most noticeable in channels draining remnants of the laterized Aurukun Surface. The gravel fraction consists of large quartz grains to 5mm, laterized rock fragments and pisolithic ironstone. Calcium carbonate is absent. Organic material is quite variable, usually less than 2% and usually consisting of wood or leaf litter. Figure 3-34 presents certain lithologic characteristics of upland channel sands. Most samples are 80% or more sand, except for 3 samples with significant gravels. These samples were collected from tributaries of the Archer system near the Weipa Plateau where large fragments of laterized rock or pisolithic gravel are common in the channel deposits. Streams and rivers which flow off the Weipa Plateau show increased gravel fraction due to their steeper gradients. Size frequency distribution of upland channel sands will be discussed in relation to other sand facies later in this chapter.

Point-bar deposits on the uplands are of the same coarse to medium sands which characterize channel deposits. The fluvial system frequently flows for long periods of the wet at or above bank full stages (Chapter 2) thereby mixing channel and point bar sands. Flood flows produce levee
FLUVIAL SANDS

100% quartz sand and rock fragments

Average Composition

Size Fractions

GRAVEL

SAND

SILT

Figure 3-34. Fluvial sands are composed of a poorly sorted mixture of sand and gravel. All samples shown here were collected in channel bottoms and several kilometers upstream from the limit of tidal effects.
and overbank deposits in the savannah woodlands adjacent to the river channels. Such deposits overlap with the soils developed on the surface of the older Quaternary deposits. Sheets of fine- to medium sand are found migrating across the interfluve areas during the wet season, frequently extending onto the mudflat surface at the juncture of the uplands and the beach ridge plain.

Neither upland point-bar sands or overbank sands were characterized lithologically. They appear to be very similar to channel sands. Furthermore, the indirect route taken by overbank sands to the Gulf means that they are probably not as significant as channel sands in terms of sediment input.

ESTUARINE ENVIRONMENTS, BEACH-RIDGE PLAIN

The estuarine systems of the beach ridge plain are much smaller when compared to their counterparts on the chenier plain. Chapman Creek, 3km south of the Edward River settlement has approximately 15km of tidal channel from its mouth to its tidal limit. Estuaries on the beach-ridge plain are approximately the size of tidal creeks on the chenier plain. Despite a smaller size, the hydrologic behavior is similar. The flows are bi-directional by tidal action in the dry season, but uni-directional in the wet season. The narrow breadth of the beach ridge plain (5-8km) reduces the tidal prism of the estuaries. In the wet season, tidal channels are little more than conduits in which fluvial sands and clays are transported across the coastal plain and added to the marine system.

Tidal Channel Sands

Tidal channel sands are a minor facies in both the modern system and the Holocene stratigraphic record. The exception to this condition is found in the Mitchell River delta, 50km south of Edward River, and the
Archer River estuary, 130km north of Edward River (Figure 3-33). The Mitchell has a lobate delta with several distributary channels whereas the Archer empties into the sea through an estuary, partly closed by Holocene barriers. Both exhibit perennial flow. Their estuarine systems merit separate detailed study and were not sampled as part of this investigation.

Chapman Creek, south of Edward River, was selected as a typical medium sized estuarine system contributing to the marine plain. Qualitative description of its channel sand confirmed the "conduit" role of estuaries on this part of the coast. At the extreme tidal limit, coarse to medium upland sand is moving seaward in a narrow creek, 6m wide and only 2m deep at low tide. Binocular examination shows similar tidal channel sands from this point to within 2km of the mouth of the estuary. Occasionally, lag deposits of laterite pebbles, mud balls and wood are found in the crossover sections of the channel between meander bends. Near the mouth, the channel depth exceeds 3m at low tide and channel sands are mantled with a thick layer of blue-green clay which may be tidal accumulation during the bi-directional flows of the dry season.

Tidal point-bar muds.

Despite the presence of medium to coarse sand in the channel, point-bar and bank deposits are muddy, except near the upper tidal limit, where point-bar deposits show an interlayering of mud and sand. Point-bar deposits in the lower estuarine portion are almost identical to that facies on the chenier plain. They fine seaward, being composed of 80-90% mud near the channel mouth. Organic content remains low (at 2% or less) despite the proximity of mangrove forests.
Mangrove peats

The mangrove forests growing on top of the point bar deposits and along the margins of the estuarine channels contribute to the formation of mangrove peat facies. Peat development proceeds slowly; even before compaction, this facies rarely exceeds 20 cm in thickness. However, it is a distinctive marker in both the modern and stratigraphic context. Organic content ranges from 15 to 60% with the terrigenous fraction composed of fine quartz silt mixed with clay. The peat is fibrous with horizontal lamination. Mangrove pollen is dominant, (G. Singh, pers. comm., 1976) and relict deposits show the similar species to the modern environments.

A mangrove forest adjacent to Christmas Creek was selected as typical of the forests which flank the tidal creeks. Organic deposits varied from 2 cm to 15 cm in thickness. The width of the forest ranged from several meters along the straight sections of the channel to approximately 50 meters or more on point bars, or where the mangroves extend into swales between ridges. The peat facies is continuous along the channel margin except where the creek has incised a steep cut bank into the relict beach-ridge deposits. Peat extends to the channel mouth where a deposit with several remnants of mangrove stumps in growth position outcrops landward the ebb tidal delta. Table 3-4 presents the mangrove species list from this forest.
Figure 3-35. Beach-ridge plain facies occur at various levels with respect to prediction datum and mean sea level.
Table 3-4

<table>
<thead>
<tr>
<th>Plant Species</th>
</tr>
</thead>
<tbody>
<tr>
<td>Xylocarpus australasicus Ridl.</td>
</tr>
<tr>
<td>Exocoecaria agallocha L.</td>
</tr>
<tr>
<td>Aegiieras corniculatum (Stickm.) Blanco</td>
</tr>
<tr>
<td>Thespesia populnea (L.) Sol ex Corr.</td>
</tr>
<tr>
<td>Avicennia sp.</td>
</tr>
<tr>
<td>Lumnitzera racemosa Willd.</td>
</tr>
<tr>
<td>Croton insularis Baill.</td>
</tr>
<tr>
<td>Rhizophora stylosa Griff</td>
</tr>
<tr>
<td>Bruguiera exaristata Ding Hou</td>
</tr>
<tr>
<td>Aegialitis annulata R.Br</td>
</tr>
<tr>
<td>Ceriops tagal (Perr.) C.B. Rob.</td>
</tr>
<tr>
<td>Hakea pedunculata F. Muell.</td>
</tr>
</tbody>
</table>

MUDFLAT ENVIRONMENTS, BEACH-RIDGE PLAIN

Topographic zonation of mudflats is better defined in the beach ridge environment than in the chenier plain. There is little ambiguity in the recognition of three distinctive facies by either topographic position or lithologic characteristics. The occurrence of these three facies, supratidal silts and clays, high-tide muds, and low tide muds are related diagrammatically in Figure 3-35.

Supratidal silts and clays

Silts and clays compose the facies of the supratidal vegetated flat. The surface of this flat is supratidal in the traditional sense because it is rarely inundated even by the higher astronomical tides of the wet season. Severe upland flooding in combination with meteorological damming on the coast is required for inundation of these flats. Figure 3-36 shows wet season flooding on the south arm of Breakfast Creek, 5km north of the Edward River settlement. The surface in the upper left portion of the photo is the supratidal flat. Tide levels approximately 3.5m above prediction datum are necessary for prolonged flooding of this surface.

The surface of the flat is vegetated, largely by a dense grass mat of Enneapogon arenicola. The sediment is a mixture of silt and clay in
varying amounts which is clearly undergoing soil development. Dolomite nodules (identified by XRD), 1-2cm in diameter (Figure 3-43) are present in the upper 50-70cm of the facies. Dating of these clasts from a site near Edward River yields a date of 4150±80 years BP (ANU 1897), which must be considered as a minimum for the age of the carbonate. Excavation in the flat yields slightly degraded shells of *Telescopium telescopium*, a gastropod usually associated with intertidal mudflats and mangrove swamps.

The supratidal flats do not appear to be an environment of active deposition and the lower high-tide flats appear to be eroding into the supratidal flats. The 30-50cm scarp pictured in Figure 3-37 separates the supratidal flat from the high-tide flat. Coleman *et al.* (1966) identified a similar scarp separating alluvial and coastal flats from unvegetated tidal flats on the eastern coast of North Queensland. They suggested that erosion occurred at the base of the scarp and was promoted by the salt crystal wedging and frequent leaching by tidal waters at this juncture.

High-tide muds

High-tide flats on the beach-ridge plain occur only between the Pleistocene strandline and the oldest Holocene ridge. The flats are limited in extent, usually associated with margins of a tidal creek or river which crosses the beach ridge plain. The flat is of uniform elevation over its breadth, with as little as 5cm change in 1000 meters. Irregularly shaped-mudflats such as shown in Figure 3-38, up to 3km wide, can interconnect for several kilometers along tidal creeks.

The high-tide flat is scoured by floodwater during the wet season in a manner similar to the chenier plain. Headward creek erosion is less marked and drainage channels from the flat appear more stable, with well
Figure 3-36. Wet season inundation, (February 1976) of the beach-ridge plain shows supratidal mudflats flanking the Pleistocene beach ridge in the background of the photo. High-tide flats in the foreground have been flooded. Mangrove forests define the trace of the Breakfast Creek.

Figure 3-37. An erosional scarp separates supratidal flats from high-tide flats on the beach-ridge plain. Nodules of soil carbonate lie in the foreground as a lag deposit after removal of clay and silt by wet season tidal flooding of the high-tide flats.
developed channel bank mangrove communities and little or no evidence of
bank erosion.

Lithologically, the facies deposited on the high-tide flats of the
beach-ridge plain is similar to that on the chenier plain. Thick algal
mats of *Lyngbya aestuarii* such as shown in Figure 3-39 develop during the
wet season but are dried out and broken during the dry season. Brackish
water remains ponded on the flats after heavy rains and internal drainage
permits evaporative formation of salt crusts.

Similar to the chenier plain, the high-tide flats on the beach-ridge
plain appear to be an environment of non-deposition, which gain and lose
only small amounts of sediment during the wet and dry seasons. The flats
are slowly expanding in size via scarp retreat into the supratidal
facies. Thicker algal coverings combined with a longer wet season than
experienced by the chenier plain precludes deflation and removal of the
high-tide muds by the wind. Surface elevation is maintained by tidal
flooding and scour during the wet season.

**Low-tide muds**

The facies deposited on the low-tide mudflats exhibits considerable
seasonal variation in physical characteristics. These mudflats are
exposed during the wet season only by spring low tides, whereas they are
inundated during the dry season only by spring high tides. Low-tide muds
are a viscous mixture of mud and water during the wet season. This
viscous mixture shown in Figure 3-40 is a large lenticular body with a
sharp boundary against the beach face. The facies extends seaward and
merges with the subtidal muds. Augustinus (1978) noted that a similar
semi-liquid mud on the coast of Surinam was effective in damping wave
action. Wave damping characteristics of the low-tide muds were observed
near Edward River in February 1976, in the middle of the wet season and
Figure 3-38. An aerial view looking seaward over Chapman Creek during the dry season (September, 1975) shows the irregular shape of high-tide flats flanking the tidal creek. Remnants of supratidal flats vegetated by *Enneapogon arenicola* may be seen in the foreground.

Figure 3-39. Algal mats of *Lyngbya aestuarii* grow on the high-tide flats during the wet season. They are effective in reducing the rate by which this mudflat is dried and deflated during the dry season.
during the peak of fine-grained input to the Gulf.

During the dry season, spring high tides reach only the base of the beach face, therefore low-tide muds are exposed to subaerial drying for 80% of the time during the months of June through September. The surface becomes heavily cracked and in some cases indurated as shown in Figure 3-41. Spring high tides rip up this cracked, indurated surface, forming mud clasts such as those in Figure 3-42. During these high tides, moderate wave action deposit pelecypod shells, especially articulated Anadara, on the low-tide flat. These shells along with the mud clasts are readily incorporated in beach deposits.

Shallow sections in the low-tide flat during the dry season, reveal alternation of fine and coarse layers similar to those of the low-tide flat on the chenier plain. However, interbedding is more massive with individual layers of 5-15cm thickness continuous over several meters. There is less evidence of flaser or wavy bedding with little or no preservation of crossbedding or individual laminae in the coarse layers. Figure 3-43 shows the poor sorting present in the coarse shelly layers. Clay bands are generally thinner than the shelly sand layers. The viscous mud of Figure 3-40 is eroded and/or compacted to the thin layers of Figure 3-43 by the end of the dry season.

The mud layers of the low-tide mudflats represent shoreward extension of the subtidal muds. The lithologic characteristics of these muds are the same, and therefore will be discussed together in the later section. The shelly sand layers are also lithologically the same as the facies found in the beach ridges, a similarity to be expected after observing the migration of coarse sediments across the low-tide flat enroute to deposition on the beach.

Lens shaped bodies of viscous mud in the intertidal zone appear to be common on tropical coasts. Nossin (1962, p.299) in his description of
Figure 3-40. A viscous mixture of mud and water is deposited on the low-tide mudflat during the wet season.

Figure 3-41. The mud shown in Figure 3-40 becomes dried and heavily cracked during dry season exposure. Lower mean sea levels in this season reduce the frequency of tidal inundations of the low-tide flat.
Figure 3-42. Mud balls and molluscs are deposited on the low-tide flat during spring high tides of the dry season.

Figure 3-43. Low-tide muds are typically interbedded with massive layers of mud and shelly sand. This alternation is a result of the seasonal variation in sediment type available for deposition on the low-tide flat.
coastal sediments on the east coast of the Malay Peninsula said:

around the outlets of the rivers the clayey bed load settles in characteristic homogenous clay layers which may be 30cm thick, and which alternate with equally homogenous sand layers of marine origin. The deposition of the clay layers takes place rapidly: a clay sheet of about 20cm thickness was observed to settle in a matter of weeks.

Reineck and Singh (1973) explained that sequences of coarsely interlayered sediments are common to mixed intertidal flats, but that their mode of origin is not yet clearly understood. They suggest that one of the possible modes of origin is the occasional deposition of sand in an environment where mud normally dominates. Such an explanation is highly analogous to the processes on the low-tide flats near Edward River. Large quantities of mud is introduced along with sand during the wet season. The mud moves alongshore in a viscous mass until deposited from suspension. During the dry season, short period wind waves are effective in winnowing sand and shell from shallow subtidal sediments and ebb-tidal deltas for transport onto the low-tide flat enroute to the beach.

STRANDLINE ENVIRONMENT, BEACH-RIDGE PLAIN

Beach-ridge sands

Strandline deposits on the beach ridge plain are termed beach-ridge sands. Modern ridges are built to levels of 4m above prediction datum, whereas relict ridges have been measured near Edward River and Christmas Creek at 6.5m above datum. Relict ridge lithology and morphology will be discussed in a section separate from modern ridges.

Modern ridges vary from narrow berms 25 meters across to formidable ridges over 100 meters wide. Individual ridges are continuous for 10km and become discontinuous only when cut by an inlet or eroded during realignment of the coast.
The terrigenous fraction of beach-ridge sands is fine to medium quartz sand (Figure 3-44) with individual grains being angular to shard-like. The sand fraction is negatively skewed when the carbonate is removed, in contrast to the fluvial sands which are positively skewed. Molluscs comprise 5-70% of the sediment, although estimates of shells by volume in field section can be 95%. Pelycypod shells tend to be deposited with concave surfaces down and are frequently nested two to three shells within another. The shells and shell fragments show limited abrasion, frequently with muscle scars still fresh. Intact articulated Anadara antiquata are common on the lower beach face after moderate wave activity. Figure 3-45 shows other dominant species in addition to Anadara.

Beach-ridge sediments rest on sandy muds of the low-tide flat. The boundary between shelly sands and sandy muds is abrupt (Figure 3-46). Figure 3-48 presents data from a section excavated through a young beach-ridge near the Edward River settlement. This excavation permitted inspection of beach ridge structure from the vegetated crest of the ridge to the facies boundary with the low-tide muds. The lowest portions of the ridge show horizontal bedding with a change to crossbedded and seaward dipping beds at the top of the lower one-third of the ridge. The upper two-thirds of the ridge contains both seaward and landward dips with steepest landward dips of 25-30 degrees located at the back of the ridge. Beach/berm bedding carries through to the top of the ridge with the uppermost beds conforming to the shape of the ridge as if the uppermost sediments were draped over the crest.

The internal structures of prograded beach ridges on the tropical coasts, Mexico, have been described by Psuty (1966) and Curray et.al, (1969). Psuty described a beach-ridge plain in Tabasco where internal structures are similar to those found in the Gulf. The Tabasco ridges
Figure 3-44. Beach-ridge sands are fine to medium sands with a variable component of whole and broken shell. The samples showed a negative skewness when the shell fraction in removed.
Figure 3-45: Shell assemblages found in the beach-ridge sands are somewhat different from those found in chenier sands. A common assemblage is shown here in order of dominance. a. Mactro sp., b. Murex coppingen, c. Anadara antiquata, d. Turritella terebra, e. Polinices sp., f. Trisidos yongei and g. Melos umbilicatus.
Figure 3-46. The boundary between low-tide muds and beach-ridge sands is well defined. Shown here just above the 50cm scale, it occurs within a definable topographic range in modern ridges.

Figure 3-47. Beach ridges south of Christmas Creek are marked by well-developed *Meleleuca* sp. forests on their flanks and are separated by fresh water swamps which remain wet throughout most dry seasons.
consist of horizontally laminated beds at the base of the ridge which grade upward into cross-bedded units. This is in close accordance to the structure shown in Figure 3-48. However, Psuty commented that Tabascan ridges rest on an "erosional unconformity", whereas ridges in the Gulf of Carpentaria are marked by a facies changes at their base, a change which occurs at a predictable topographic level above prediction datum. Curray et al. (1969) investigated beach ridges of the coast of Nayarit, Mexico. Stratigraphic findings of the Nayarit study will be compared with stratigraphy of the Carpentaria study in Chapter 4. However, it is interesting to note that Curray et al. found that the beach-ridge sands showed fair to poor stratification and little lamination. In addition, the Nayarit ridges have an aeolian capping which is virtually absent in the Gulf of Carpentaria.

Relict beach ridges are distinctively different from modern beach ridges. Organic production on top of the older ridges is low, and only ridges older than 3000 years show organic discoloration of the upper 1.0m of soil. Soil development is slow, but preceded by rapid mobilization of carbonate downward. Aragonitic molluscan shells are dissolved if they are above the groundwater table and recrystallized if they are at or near the water table. Anadara and Turritella shells, from ridges near Edward River, which were analyzed for calcite/aragonite content, showed 100% recrystallization of aragonite to calcite within 3500 years. Leaching of carbonate proceeds roughly as a function of age. Ridges from which shells date as 5000-6000 years old, have 1.5 - 2.0 meters of leached quartz sand overlying highly recrystallized beach-ridge rock, (see Chapter 4).

Casuarina is the first tree or shrub to be established on the strandline. However, upland species grow on the ridges in behind the Casuarina fringe and develop into "vine forests" of trees and shrubs.
Figure 3-48. A cross section of a ridge excavated near Edward River shows horizontal bedding in the lower portions of the section. The upper two-thirds of the section contains both seaward and landward dipping beds with occasional charcoal. Note the landward dipping beds overlapping the seaward dipping unit at the far right of the section. The underlying seaward dipping unit is part of the adjacent beach ridge to the east. Vertical scale is not shown relative to datum.
The components of these thickets are summarized in Table 3-5. These species are limited to the crests and flanks of the ridges, alternating with open swamps in the swales, and are effective in defining the ridge morphology from the air.

**TABLE 3-5**

<table>
<thead>
<tr>
<th>Species</th>
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<tbody>
<tr>
<td>Brevnia sp</td>
</tr>
<tr>
<td>Gyrocarpus americanus</td>
</tr>
<tr>
<td>Melaleuca dealbata</td>
</tr>
<tr>
<td>Pleurostylia opposita</td>
</tr>
<tr>
<td>Pouteria sericea</td>
</tr>
<tr>
<td>Securinega melanthesioides</td>
</tr>
<tr>
<td>Syzygium suborbiculare</td>
</tr>
<tr>
<td>Terminalia subacrontera</td>
</tr>
<tr>
<td>Thryptomene oligandra</td>
</tr>
</tbody>
</table>

Swales between the beach ridges vary from subdued troughs 10m wide to 100-200m wide swamps. Sediments in the swales are silty sands with varying amounts of organic matter. Wider swales, such as those near Edward River have a muddy substrate due to deposition of silt and clay from wet season flood waters. In these situations the swales may remain unvegetated if they develop hypersaline conditions due to evaporation during the dry season.

More usually, the narrow swales between the heavily vegetated relict ridges contain a freshwater swamp which retain ponded fresh water during the dry. Such a swamp is visible in the lower right corner of Figure 3-47. These freshwater swamps develop a distinctive vegetation consisting of the low fan palm (*Livistona* sp.) and cane grass thickets. Heavy organic muds approximately 0.3m thick overly 1.5 - 2.0m of clayey sand in these swales. Sediments grade downward into clays similar to those beneath the beach-ridges. Where the swales are connected to estuarine channels, the frequent incursion of saline waters promotes growth of mangrove forests with accompanying peats.
Swale sediments are not regarded as a separate facies, merely an extension of beach-ridge sands which interconnect adjacent ridges. Frequently ridges are so close that swales are all but absent such as the structure shown in Figure 3-48 where seaward dipping beds from the previous ridge are overlapped by crossbedded material of the next ridge.

Discussions in this section have emphasized the seasonal variation of inundation, especially with respect to such facies as the low-tide muds, which barely qualify as intertidal during the dry season. The lower sea levels associated with this time of the year cause the beach environment to be undisturbed by wave activity for approximately 5 months. The high-tide flats are identical to their chenier plain analogues with respect to frequency of inundation, remaining intertidal only for the wet season and supratidal for the dry season. The actual supratidal vegetated flat, which is unique to the beach ridge plain, is clearly beyond reach of astronomical spring tides except during the wet season (Figure 3-35).

NEARSHORE ENVIRONMENTS, BEACH-RIDGE PLAIN

Nearshore environments can be sub-divided similarly to the chenier plain. Ebb-tidal deltas develop at the mouth of all estuarine inlets along the beach ridge coast. Sub-tidal nearshore sediments characterized by an "inshore" mud zone which is a subtidal extension of low-tide muds and an "offshore" sand zone. The term "offshore" is used in a relative sense because water depths over these sands range from 16-30 meters. These three sub-environments will be discussed separately in the sections which follow.

Ebb-tidal delta sands

Large rivers such as the Kendall possess a crescentic shaped
intertidal lobe of sand across their mouths (Figure 3-49). Subtidal and intertidal swash bars are located either side of the terminal lobe. Such a morphology is in accordance with Hayes et al. (1973) description of ebb-tidal deltas on barrier island coasts. Few ebb-tidal deltas on the beach ridge plain present the birdsfoot type of channel pattern which was evident on the tidal deltas of the chenier plain.

The form of the ebb tidal deltas on the beach-ridge coast is quite variable. This is a result of seasonal variation in sediment supply and wave regime. Short-term changes of an ebb tidal delta over a period of 20 years are presented in Figure 3-50. This tidal delta is at the mouth of Moonkan Creek, 8km north of the Edward River settlement. Photographs taken in 1957 showed a levee type morphology with slight indication of a subtidal terminal lobe. A "hammerhead" shaped swash bar was connected to the shore and a dominant southward longshore drift is suggested by the spit building from the north side of the channel. By 1969, the southern swash bar had lost its connection to the beach, a bar oriented normal to the shore blocked the main channel and the spit developed from the north side had grown further south, further deflecting the mouth of the inlet. Photography from 1977 showed a well-developed terminal lobe, bifurcation of the inlet channel into two branches with the south branch being the larger, and well developed intertidal swash bars on flanking the terminal lobe. The spit from the northern side had been shortened and replaced by a swash bar.

The dominant southward deflection of the inlet during the last twenty years is not in agreement with the northerly oriented relict recurved spits shown in Figure 3-51. The north side of the channel shows undercutting of old beach ridge sediments along the lower two kilometers of channel. It is unlikely that inlet migration is caused by channel meandering because the meander pattern in the lower channel is quite
Figure 3-49. A crescentic shaped middle ground is common to the ebb-tidal deltas on the beach-ridge plain. This bar, at the mouth of the Kendall River, is 400-500m long, consisting of a mixture of medium quartz sand and broken shell.
unlike that landward of the Holocene ridges. Estuarine channels through Holocene beach-ridge plains at other locations north and south of Edward River are all unusually straight with a distinct northerly deflection.

Intertidal portions of ebb-tidal deltas, especially the terminal lobes, are usually smooth gently convex surfaces. However, where channels cut across the swash platforms, ebb-oriented megaripples with 1.0 - 1.5m ripple length and 0.2 - 0.3m amplitude are associated with ebb flows during the wet. Subtidal swash bars which are concentric to the lobate bulge of the entire delta have amplitudes of 0.5 - 0.75m and show a bifurcating enechelon form, (Figure 3-51).

A wide variation in grain size is typical of ebb-tidal delta sands. Terrigenous quartz sands are mixed with broken and whole shells producing a variable lithology as seen in Figure 3-52. Calcium carbonate may reach 75% by weight in sandy shell layers which may contain as much as 15-20% comminuted shell in the silt size range. The composite layering formed in the ebb-tidal deltas is similar to that found in the low-tide flats, except that sand and shell predominates on the tidal deltas. The mode of origin for this interlayering is probably similar.

Live molluscan communities were not noted on the intertidal portions of tidal deltas. At the mouth of Moonkan Creek, Mactra sp. form a detrital assemblage in the southern swash bar. Similar findings are common along the beach-ridge coast, with bioclasts providing up to 95% of the volume for spits and marginal swash bars connected to the beach. Local residents insist that edible molluscs, apparently Mactra, may be collected from intertidal areas near tidal inlets at some times of the year. It is assumed that the shell assemblages which litter the beaches originate from intertidal and subtidal communities adjacent to the ebb-tidal deltas.
Figure 3-50. An ebb-tidal delta at the mouth of Moonkan Creek has exhibited extensive variation in form and extent over the past 20 years.

Figure 3-51. An aerial oblique of the Moonkan Creek ebb-tidal delta also shows a northerly deflection of the inlet as the coast prograded in the upper Holocene.
Figure 3-52. A lithologic summary of ebb-tidal delta sands shows the high content of shell material in this facies. The terrigenous fraction is medium to coarse, poorly sorted quartz sand.
Subtidal muds

Rhodes (1978) considered the subtidal and intertidal muds as one facies, deposited in the "inshore mud zone". There is little reason to separate discussion of the intertidal and subtidal muds except that the intertidal zone was sampled and described using subaerial techniques. The subtidal zone has been sampled on a more regional basis as shown in Figure 3-53, indicating a continuous zone of subtidal muds fronting the coastline from Weipa to the southern end of the Gulf. Shallow water sediments (2-6m depths) contain up to 90% mud and there is a general coarsening of this facies seaward until sediments of 80% sand are encountered in depths of 20-22 meters. These muds are grayish or blue-green and viscous or plastic when sampled. Textural qualities of subtidal muds are summarized in Figure 3-54. Calcium carbonate levels vary from 10% to over 50% in the mud fraction. Highest amounts of fine carbonate were found in the northern portions of the Gulf, especially in the protected embayment associated with Albatross Bay off Weipa (Figure 3-53).

Offshore sands

A zone of clean offshore sands called the offshore sand facies consisting of relict beach and fluvial sands is seaward of the 80% sand contour in Figure 3-53. These offshore sands vary from rounded to angular. Calcium carbonate contents range from 5 to 25% and size analysis of the carbonate free fraction shows a positively skewed suite of samples ranging from medium to coarse sands. They frequently contain black or brown clasts of secondary calcium carbonate probably derived from reworked beach rock (Figure 3-55). The quartz grains show a strong red staining in some areas, similar to that noted in fluvial sands from upland areas adjacent to laterized surfaces. Many samples contain a
Figure 3-53. The subtidal sediments of the nearshore zone show a trend toward coarsening offshore. Percent sand in the bulk sample is contoured at 10%, 50% and 80% levels. Sample locations are marked by solid dots.
Figure 3-54. Subtidal muds show a considerable variation in grain size. Although not indicated here, some samples may contain up to 50% calcium carbonate at silt-size or smaller.
OFFSHORE SANDS

Average Composition

calcium carbonate 36.3%
terrigenous material mostly quartz
organic < 0.5%

Size Fractions

GRAVEL

SAND

SILT

Figure 3-55. Offshore sands are well sorted, much better sorted than either fluvial or beach sands. They contain high levels of calcium carbonate as both broken shell and calcite-cemented clasts.
mixture of very immature grains with mature well rounded grains, suggesting a considerable mixing of sediments from different sources.

Poor correlations exist between either depth and calcium carbonate content, or depth and sand content. There appears to be more justification for a correlation between sand content and downdrift proximity to mouths of major rivers (Figure 3-53).

Linear ridges of coarse sandy material are common in the eastern Gulf (Rhodes, 1978). Their occurrence at uniform depths in the 12-20 meter zone suggests correlation ridges at similar depths. Some ridges showed evidence of cementation and they are considered as relict shoreline features subjected to sub-aerial cementation during lower sea levels. Their age and full extent remains to be established.

AEOLIAN ENVIRONMENTS, BEACH-RIDGE PLAIN

Aeolian facies are notable by their absence from the beach-ridge plain. Only small silt and clay dunes (0.2 - 0.3m high) develop on the high-tide flats next to tidal creeks. Very few of these showed any accretion during the two seasons of field work. A higher rainfall and longer wet season precludes drying of the mudflats to the extent as noted on the chenier plain. The unvegetated flats are less extensive and their long dimensions are usually transverse to the dry season winds. Salt efflorescence is not as active and the binding effect of the algal mats is more effective than on the chenier plain.

Between the Mitchell and Holroyd rivers, aeolian capping of modern ridges is absent. Capping on relict ridges may be difficult to confirm because leaching of carbonate destroys lamination and bedding. Development of dunes elsewhere along the west coast of Cape York Peninsula is ephemeral and infrequent. Most incipient foredunes are destroyed by high energy waves during the wet season. Smart (1977) did
not report dunes on the beach ridges of Cape Keer Weer. However, north
of the mouth of the Archer River, where the Gulf washes directly against
the Weipa Plateau, small active foredunes are located at the heads of
embayments. In addition, a reconnaissance to the mouth of the Wenlock
River, north of Weipa, showed active foredunes approximately 1.0 - 2.0m
high near the Mapoon peninsula. These isolated occurrences do not alter
the generalization that aeolian facies are not significant on the
beach-ridge plain.

SUMMARY, BEACH-RIDGE PLAIN

Following the previous discussion of the Carpentaria chenier plain,
several important factors were included in a summary of chenier plain
environments. It was stated that such observations were essential to the
interpretations to be made in later chapters, especially in the
discussions of relative sea level and episodic progradation. A parallel
set of observations is presented for the beach-ridge plain.
a. The typical estuarine systems of the beach-ridge plain is smaller
but similar to that of the chenier plain. Although the northern
rivers are less ephemeral than those at the southern end of the
Gulf, it is still the high rates of discharge during the wet season
which introduce the sand and the mud in a quantity sufficient for
beach-ridge progradation. The estuarine systems in both cases
carries this sediment load across the prograded plain and introduces
it to the marine systems.
b. The high-tide flats on the beach-ridge plain are erosional surfaces
which are subjected to both down wasting and lateral extension into
the supratidal flat. Tidal creeks across both the high-tide and
supratidal flats show headward erosion. Aeolian deflation of the
high-tide flats on the beach-ridge plain is insignificant.
c. Beach ridges are built by wave processes during the higher sea levels associated with the wet season. Waves during the dry season barely reach the base of the beach face. The base of the beach-ridge facies where it rests on low-tide muds is well defined. This observation is critical to the study of relative sea levels during the upper Holocene.

d. Mangrove forests are absent on the open coast, but are well developed on creek and river margins. At these sites they produce a mangrove peat facies which appears to be preserved stratigraphically.

e. Tidal deltas occur only in ebb-tidal forms, due to the net seaward transport of upland sediments and the narrow shape of most estuaries. Such deltas are well developed at the mouths of large rivers where their exposure to wave processes permits continued sorting and redistribution of the upland sediments.

f. Aeolian facies are virtually absent on the beach-ridge plain, and where developed, they do not contribute greatly to the morphostratigraphic record of the prograded coast.

BIOCLASTS, CLAYS, and GRAIN SIZE IN MODERN ENVIRONMENTS

Sediments of depositional environments discussed in this chapter are composed of 3 major constituents: bioclastic calcium carbonate (shells), terrigenous clays and terrigenous sands. It has long been the objective of researchers working in modern depositional environments to characterize environments by analytical means with only minor assistance of morphology and primary sedimentary structures. Discussion of modern environments in the Gulf of Carpentaria is concluded by evaluating the possibility of establishing distinctive characteristics for some environments based on detailed analysis of the three major constituents.
Shell content

The ubiquitous occurrence of skeletal molluscan fragments suggest that zonation of sediments could be possible by examining fossil assemblages. However, the identification of shell assemblages is only possible from the material found in the strandlines, macrofossil examination is therefore limited value. It has been shown previously in this chapter that live molluscan recovered from bottom samples agree in neither species or number with death assemblages. Environmental discrimination by fossil assemblage appears to be of limited use in the Gulf of Carpentaria. Examples of this limited use have been shown in the case of assemblages found in chenier ridges landward of mangrove fringes. Table 3-2 enumerated an assemblage of shells which were found in cheniers formed behind mangrove swamps. Furthermore, comparison of shells in Figure 3-23 with those in Figure 3-45 indicate that assemblages are slightly different for these two types of coast.

The possibility of micro-fossil analysis of environments exists. Although study of bottom samples from the Gulf is proceeding at the Australian Museum, Sydney, discussion of this work is beyond the scope of this thesis.

Clay mineralogy

Clay mineralogy in the marine environment was reviewed by Grim (1968) who discussed major findings to that time. Grim concluded:

... all types of clay minerals have been identified in present-day marine sediments. The occurrence of halloysite is not unequivocal, ... Illite and chlorite are abundant components, with kaolinite commonly abundant near shore in localized areas... In general, the clay-mineral composition is likely to reflect climatic conditions in the source. It is not certain, on the basis of available data, if there is a correlation between type of

Porrenga (1967) found that distribution of montmorillonite and kaolinite in sediments of the Niger delta, probably resulted from differential sedimentation of the larger kaolinite particles near shore. Porrenga reported a similar distribution and explanation for clay mineralogy adjacent to the Orinoco delta. He did not report finding halloysite in any of his study areas (Niger and Orinoco deltas and Sarawak shelf). Its occurrence as metahalloysite in Carpentaria sediments (Table 3-6) was therefore unexpected.

More recently Gibbs (1977) has shown that clay mineral segregation offshore of the Amazon River is entirely attributable to variation in particle size of montmorillonite, kaolinite and mica. Gibbs was able to fully dismiss chemical alteration as a cause of clay mineral zonation. He reported that differential flocculation, although present, was not significant in causing the smaller montmorillomite particles to be deposited seaward of the larger kaolinite and mica.

A cursory inspection of clays from various modern environments in the Gulf of Carpentaria was conducted using X-ray diffraction and SEM techniques outlined in Appendix A. Results of this effort are summarized in Table 3-6. The obvious outcome of the attempt to "fingerprint" with clay mineralogy validates the caution of Grim (1968). Kaolinite, montmorillinite, illite and chlorite occur in a variety of environments. There appears to be no differential precipitation of clays with regard to particle size. Montmorillinite, kaolinite and illite occur in various amounts in both low-tide flat and subtidal facies. There is little evidence to suggest that the chlorite from the low-tide flats is authigenic. Electron photomicrographs showed no evidence of needle or fibrous forms, only highly weathered plates, scales and fleecy sheets. However nearly ubiquitous occurrence of metahalloysite in non-marine
<table>
<thead>
<tr>
<th>ENVIRONMENT</th>
<th>mixed layer</th>
<th>montmorillinite</th>
<th>montmorillinite</th>
<th>metahalloysite</th>
<th>kaolinite</th>
<th>chlorite</th>
<th>illite</th>
<th>amorphous clay</th>
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<tr>
<td>(6 samples from diverse area)</td>
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Table 3-6

CLAYS
clays is in marked contrast to its absence in marine sediments. This
distribution may be due to either differential floculation/precipitation
or diagenetic changes.

The variety of source materials and the shallow nature of the Gulf
may preclude clay mineral zonation in the terms of either Porrenna (1967)
or Gibbs (1977). Variation in source material is the most plausible
explanation for clay mineral distribution noted in Table 3-6. These
results do not support Smart's (1977) assertion that the presence of
illite in a carbonaceous clay of late Pleistocene age found in cores from
the bottom of the Gulf of Carpentaria must necessarily result from
authigenic alteration of montmorillonite. There appears sufficient
variability in clay mineralogy in upland, fluvial, nearshore and offshore
sediments to account for any mixture of clays from Table 3-6.

Lithologies in the drainage basin of the Gulf vary from sandstones
to siltstones, all having been subjected to various extents of tropical
weathering. The products of this intense weathering are sufficiently
degraded that they enter the fluvial/marine system with little potential
for further alteration. SEM photographs illustrated the highly degraded
crystalline forms representative of tropical weathering products.
Excepting geographic zonation of clays by source areas, there is nothing
in results to date to support the premise that a given suite of clay
minerals can be a unique environmental indicator in the Gulf of
Carpentaria.

Sand fraction analysis

Three principal modern environments of deposition in the Gulf
contain significant amounts of sand, sufficient to warrant
categorization of this fraction separately from silt and clay. Fluvial
and beach-ridge sands associated with the beach-ridge plain and marine
sands of the offshore sand zone (Rhodes 1978) provided potential for comparison on the basis of size-frequency distribution.

Friedman (1961, 1962, 1967) established the rationale for such comparisons. A number of workers since that time have met with varied success. Valia and Cameron (1977) have reviewed most of the more recent studies. They applied the skewness measure to unconsolidated Miocene sediments in central Delaware, U.S.A., in an attempt to confirm paleoenvironmental interpretations. They too, met with limited success, partly because they used graphic measures of Folk and Ward (1957) despite the fact that Chappell (1967) made it clear that moment measures should be applied, especially when only one parameter is used as a paleoenvironmental indicator. In addition, Valia and Cameron were forced to draw their comparisons with modern analogues from beyond the basin in which they were working.

The initial objective of size frequency analysis of sand fractions in the Gulf of Carpentaria was to differentiate between fluvial and beach sediments. Sand samples already summarized in Figure 3-34 and 3-44 were subjected to size frequency analysis by moment measures after sizing with a settling tube (Appendix A).

Despite later efforts by other workers to improve on techniques of environmental distinction by size-frequency distribution, Friedman's (1967) approach remains the most direct strategy with a minimum of numerical manipulations. His use of moment measures plotted against one another facilitates rapid determination of success or failure for these techniques in any given study area. Results from the Gulf of Carpentaria suggest that if any distinction is possible, it will appear simply by plotting one of the first three movements against another. Further manipulations merely enhance the separation of domains. Therefore, scatter plots were prepared using Friedman's (1967) suggestion of: mean
vs. standard deviation, skewness vs. standard deviation, mean cubed deviation vs. standard deviation, and mean cubed deviation vs cubed deviation (Figure 3-56). Visual separation of sands into discreet domain of beach and fluvial sands was possible on all plots, but to neither the extent or configuration of Friedman's (1967) results. Friedman suggested that with few exceptions, beach sands are negatively skewed, whilst river sands have a fine-grained tail which gives them a positive skewness. In addition, Friedman (1961) stated that river sands could have either positive or negative skewness if a significant fraction (5%) of the grains are larger than 1.00 phi. However, Carpentaria river sands plotted in Figure 3-56 average 57% of their grains larger than 1.00 phi, but the result is a predictable negative skewness with respect to beach sands which are usually positively skewed.

In brief, fluvial sands entering the Gulf of Carpentaria are medium to coarse, moderately well to poorly sorted with a predominant negative skewness. Modern beach sands show similar sorting, but are considerably finer with their means in the fine to medium range, with a positive skewness. Although Ly (1978) found roundness to be a discriminant in sands on the New South Wales coast, this parameter does not appear to be of use in the Gulf. Sands in the beach facies are often so fine that they approach coarse silt. Fluvial sands, however are coarse, therefore the degree of rounding may be more a function of grain size and less a matter of environment. Finally, mineralogy appears to be of little use because both fluvial and beach sands are dominantly quartz.

During the discussion of strandline environments in the beach-ridge plain, a separate section was devoted to the discussion of relict ridges as different from modern ridges. It was concluded that relict ridges, although having crestal elevations higher than modern ridges and usually lacking carbonate in their upper sections, were very similar to modern
Figure 3-56. Using strategies presented by Friedman (1967) scatter plots of beach and fluvial sands were created. These plots are presented here as a. mean vs. standard derivation, b. skewness vs. standard deviation, c. mean cubed deviation vs. standard deviation, and, d. mean cubed deviation vs. cubed deviation. The best separation occurs in plots a and b with fluvial sands being medium to coarse, poor to moderately sorted and positively skewed. Beach sands are fine to medium, with sorting similar to fluvial sands, but negatively skewed. Solid lines drawn on each graph separate beach and fluvial domains.
ridges. Size frequency analysis of 15 sands from relict ridges in the Edward River area confirms this similarity at the textural level. Figure 3-57 shows relict ridge sands overplotted on modern ridge sands in a mean vs. standard deviation format. The close accordance of the modern and relict sands suggests the value of such simple comparisons. The clear differentiation of both relict and modern beach sands from modern river sands suggests that source and composition of beach sands has remained relatively consistent throughout the late Holocene.

It was hoped that foregoing success in comparing relict to modern beach sands could be extended to the offshore sands described by Rhodes (1978). These sediments consist of coarse, well sorted variably skewed sands. Offshore sands may be found within 5 km of the coast and in less than 12 meters of water, especially adjacent to major rivers such as the Archer, Kendall, Edward, Mitchell, Staaten, and Gilbert (Figure 3-53). Bulk samples from the sea bottom of this zone are usually more than 80% sand with the sand fraction coarser than beach sands but approximately the same as fluvial sands. However, the presence of both primary bioclastic carbonate and secondary carbonate militates against a simple fluvial origin for these sands.

When plotted into simple scatter plots such as in Figure 3-58, there is a clearcut separation of offshore sands into two domains. The finer group of sands were recovered from the transition zone of the southern Gulf (between the 50% and 80% sand contours of Figure 3-54). The coarser group were collected from stations seaward of the 80% sand contour in the central and northern area. The fine group show size distribution similar to modern beach sands and the coarse group shows good overlap with fluvial sands. Bottom sediments of the Gulf may be something other than relict equivalents of modern environments which have been transgressed by rising Holocene seas. The bottom of the Gulf is also an active modern
Figure 3-57. Relict beach ridge sands are compared against modern beach-ridge sands in this plot of mean vs. standard deviation. The domains are reasonably conformable suggesting that the sorting and reworking of sands prior to deposition in beach ridges has been similar throughout the upper Holocene.
Figure 3-58. Offshore sands are split into two discrete domains by comparison of mean grain size against standard deviation. The finer group of samples represent bottom sediments in the southern Gulf whilst the coarse group were collected from the northern Gulf.
environment where modification of previously deposited sediments is continuing to an appreciable extent.

Table 3-7 presents a summary of size-frequency parameters attributable to modern fluvial, modern beach, relict beach, and offshore sands. Such a table is a rather glib abridgement of the scatter plots presented in previous figures. However, it constitutes a manageable and reasonably accurate characterization of these sand facies, to be used in later discussions of stratigraphic findings.

Table 3-7

<table>
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<tr>
<th>STATISTIC</th>
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<td>fluvial sands</td>
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<td>relict beach sands</td>
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<td>(15)</td>
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<td>(20)</td>
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<tr>
<td>mean</td>
<td>.79</td>
<td>2.36</td>
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<td>median</td>
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<td>.48 to 1.13</td>
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<td>skewness</td>
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</table>

SUMMARY

Table 3-8 is a summary of important observations made of depositional environments on both types of coast. It is unnecessary to reiterate all principal findings here. The importance of this chapter as a whole rests with the identification and characterization of fourteen distinctive facies, each of which may be described in terms of three parameters: geometry, lithology and sedimentary structures. Table 3-8 performs a review of modern depositional environments and their resulting facies. Of the three parameters used in Table 3-8, it is geometry and lithology which will prove most valuable in the presentation of the next chapter dealing with morphostratigraphy.
<table>
<thead>
<tr>
<th>Fluvial</th>
<th>Upland channel sands</th>
<th>variable from a thin gravel lag deposit to sands several m thick</th>
<th>medium to coarse quartz sand pisolithic ironstone laterite gravel 2-3 cm diameter</th>
<th>poorly laminated, occasional sand waves and scour mega-ripples</th>
<th>ephemeral channels 50-500 m wide choked with sands and gravels 5-10 m thick</th>
<th>medium to coarse quartz sand with minor pisolithic ironstone laterite gravel</th>
<th>poorly laminated sand waves 1.0-1.5 m amplitude</th>
</tr>
</thead>
<tbody>
<tr>
<td>Estuarine</td>
<td>Tidal channel sands</td>
<td>variable as above usually overlying irregular bedrock floor</td>
<td>90-95% medium quartz sand 4-6% rock fragments 1-4% pisolithic ironstone increased CaCO3 seaward</td>
<td>ebb-oriented lunate megaripples wavelengths 1.0-1.5 m amplitudes 0.2 m with larger in thalweg</td>
<td>not observed</td>
<td>same as fluvial sands</td>
<td>not observed</td>
</tr>
<tr>
<td></td>
<td>Tidal channel point-bar sands</td>
<td>5-10 m thick interfinger horizontally with high tide muds and underlying deposits</td>
<td>homogeneous dark brown muds, median 9.0 phi 0.5-1.5% organic 2.0-6.0% CaCO3</td>
<td>moderately bioturbated rapid lateral alumping and burrow infilling</td>
<td>not observed</td>
<td>similar to chenier plain</td>
<td>not observed</td>
</tr>
<tr>
<td></td>
<td>Tidal overbank muds</td>
<td>1.0 m thick lobate 100-200 m wide wedging onto mudflat</td>
<td>fine sand and silt light brown 1-2% organic, 2.0-4.0% CaCO3 as fine comminuted shell</td>
<td>finely laminated with occasional thin bands fine sand 3-5 cm thick</td>
<td>not observed</td>
<td>absent on beach-ridge plain</td>
<td>not observed</td>
</tr>
<tr>
<td></td>
<td>Mangrove peats</td>
<td>not developed on chenier plain</td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>Mudflat</td>
<td>Supratidal silts and clays</td>
<td>absent on chenier plain</td>
<td></td>
<td></td>
<td></td>
<td>cracked and oxidized light brown clays and silts with CaCO3 nodules 1-2 cm diameter</td>
<td>absent probably due to alternate wetting and drying and illuviation</td>
</tr>
<tr>
<td></td>
<td>High-tide muds</td>
<td>approx 1-2 m thick continuous for up to 50 km grade downward to underlying low-tide muds</td>
<td>deeply cracked and oxidized brown muds 90% clay 9% sol salts 4-7% insol salts 2-3% organic less than 1.0% CaCO3</td>
<td>surface heavily scoured and reworked fine hor laminate disappearing downward due to efflorescence, wetting and drying</td>
<td>variable thickness exceeding 1.0 m continuous over km, scarped adjacent to high-tide muds</td>
<td>similar to chenier plain</td>
<td>similar to chenier plain</td>
</tr>
<tr>
<td></td>
<td>Intertidal organic muds</td>
<td>vegetation defined zone 100-300 m wide continuous up to 20 km</td>
<td>fine silt and clay with 2-3% organic, subsurface textural parameters similar to low-tide muds</td>
<td>gastropod tracks</td>
<td></td>
<td>absent on beach ridge plain</td>
<td></td>
</tr>
<tr>
<td></td>
<td>Low-tide muds</td>
<td>1-2 m thick</td>
<td>light to dark brown muds fine silty sands</td>
<td>flaser and wavy bedding with fine sand and clay</td>
<td>1.0-1.5 m thick lenticular mud layers same as subtidal muds sand layers same as</td>
<td>massive composite bedding of shelly sand and fine mud</td>
<td></td>
</tr>
</tbody>
</table>

**Table 3-8**
<table>
<thead>
<tr>
<th>Location</th>
<th>Description</th>
<th>Characteristics</th>
</tr>
</thead>
<tbody>
<tr>
<td>High-tide</td>
<td>Approx 1-2m thick continuous for up to 50km grade downward to underlying low-tide muds</td>
<td>Deeply cracked and oxidized brown muds av 50% clay, 9% sol salts, 4-7% insol salts, 2-3% organic less than 1.0% CaCO3. Surface heavily scoured and reworked fine hor laminae disappearing downward due to efflorescence, wetting and drying.</td>
</tr>
<tr>
<td>Intertidal</td>
<td>Vegetation defined zone 100-300m wide continuous up to 20km</td>
<td>Fine silt and clay with 2-5% organic, subsurface textural parameters similar to low-tide muds. Gastropod tracks.</td>
</tr>
<tr>
<td>Low-tide</td>
<td>1-2m thick prismatic interfingering seaward into subtidal muds continuous for 100km or more</td>
<td>Light to dark brown muds, fine silty sands to silty clays 40% submicron. Flaser and wavy bedding with fine sand and clay symmetrical and asymmetrical ripple marks. 1.0-1.5 m thick lenticular interfingering with subtidal muds. Mud layers same as subtidal muds sand layers same as strandline sands.</td>
</tr>
<tr>
<td>Strandline</td>
<td>Chenier/ beach ridge sands 4-5m thick 50-150m wide continuous up to 20 km</td>
<td>Up to 75% molluscan shell with variable silt sand and pisolithic gravel. Horizontal bedding mixed with steep crossbedding near back of ridge gentle seaward at front of ridge. 3-4m thick 25-100m wide continuous up to 10km. 5-70% molluscan shells mixed with fine to medium quartz sand. Gently dipping seaward beds at front and top horizontal at base grade upward to crossbed at center and back.</td>
</tr>
<tr>
<td>Nearshore</td>
<td>Ebb-tidal delta sands lobate complexes 2-3m thick 500-2000m wide 1000m long</td>
<td>Medium shelly quartz sands with CaCO3 variable 1-80% massive composite bedding of shell and sand layers occasional scour mega-ripple. Crescentic shaped lobe 2-3m thick 100-200m wide 750m long. Medium to coarse sand with shell occasional mud CaCO3 variable at 15-75% composite layering dominated by 0.1-0.2m thick sandy shell layers and thin mud bands. Characteristics similar for both coastal types.</td>
</tr>
<tr>
<td>Subtidal</td>
<td>Prism shaped bodies up to 8m thick width variable to 10 km continuous along entire coast</td>
<td>Blue-green muds with variable sand, silt and clay median size 9.0 phi CaCO3 varies 10-20%. Not observed but probably highly burrowed.</td>
</tr>
<tr>
<td>Offshore</td>
<td>Absent in southern Gulf except for transitional zone containing 50-80% sand</td>
<td>Thickness unknown probably several meters continuous over central and northern Gulf. Coarse mixed terrigenous, primary and secondary carbonate sand quartz grains heavily stained with iron oxide. Not observed.</td>
</tr>
<tr>
<td>Dune</td>
<td>Aeolian silts and clays 2-3m thick dune fields may be 1-2km wide and 10 km long</td>
<td>Light brown silt and clay with some sand soluble salts up to 25% organics less than 1%. Usually homogeneous with some inherited micro-structure. Facies not well developed on beach-ridge plain.</td>
</tr>
</tbody>
</table>
INTRODUCTION

Morphostratigraphic concepts

The purpose of this chapter is to establish the surficial and subsurface relationships of the record left by migration during the middle and upper Holocene of the depositional environments discussed in Chapter 3. The concept of morphostratigraphy based on surface form is combined with stratigraphy based on lithology to describe the Holocene coastal record. This chapter provides the framework to which an age structure will be assigned in Chapter 5.

The formal terminology of surficial geology is usually inadequate to thoroughly deal with Holocene coastal formations and needs extension by other mapping techniques. Frye and Willman (1960, p.7) proposed the use of the morphostratigraphic unit for the analysis of sequences of surficial deposits. They said:

a morphostratigraphic unit is defined as comprising a body of rock that is identified primarily from the surface form it displays; it may or may not be distinctive lithologically from contiguous units; it may or may not transgress time throughout its extent.

Although Frye and Willman proposed the term for use in mapping Pleistocene glacial deposits, other authors have applied it to a variety of surficial deposits. That morphostratigraphic principles can be readily applied to Quaternary coastal deposits is demonstrated by Thom (1967) and Mixon and Pilkey (1976).

However, there remain certain inadequacies even in the Frye and Willman strategy when working with stratigraphic data obtained by drilling. The comments of Krumbein and Sloss (1963) concerning
lithofacies are particularly relevant in this context.

The information from sub-surface sources is overwhelmingly weighted toward the physical data of lithostratigraphy rather than toward the paleontologic data of bio- and time stratigraphy..., subsurface stratigraphy tends to emphasize tangible units based on lithology. Facies concepts developed under this emphasis stress observable lithologic variations rather than the more subjective and abstract interpretation of environment (Krumbein and Sloss, 1963, p. 327).

Chapter 3 described the environments of deposition and their resulting facies of the chenier and beach-ridge plains. Associated with these facies is a distinctive geometry, a portion of which is surficial form. Furthermore, each facies may be described by a number of lithologic parameters, thereby providing a lithofacies associated with each environment. In this study of Carpentaria coastal sediments, distinctive geometry to include surficial form is combined with lithofacies characteristics to define a morphostratigraphic unit. Therefore, each facies presented in Chapter 3 may occur as a morphostratigraphic unit of the chenier plain. A discussion of the surficial expression of these units precedes the description of their subsurface occurrence and characteristics. Such a separation of map and stratigraphic data divides the chapter into six principle sections:

a. regional morphostratigraphy of the chenier plain supplemented by a detailed description of one area.
b. presentation of stratigraphic data from drill hole transects across the chenier plain.
c. discussion of chenier plain morphostratigraphy.
d. regional morphostratigraphy of the beach-ridge plain.
e. presentation of stratigraphic data from drill hole transects across the beach-ridge plain.
f. discussion of beach-ridge plain morphostratigraphy.
Carpentaria paleosol

By association of a modern facies with a morphostratigraphic unit, the entire Holocene sedimentary sequence of the prograded coastal plain is easily described. However, beneath this sequence of prograded units is an unconformity separating Holocene sediments from upper Pleistocene sediments. A distinctive marker of unconformity is the Carpentaria paleosol.

The Carpentaria paleosol is lithologically defined by characteristics which make it an effective marker in all drill holes. This fine, highly compacted mud contains carbonate nodules, pisolitic ironstone pebbles, and coarse quartz sand. The unit varies in color from a dark brown beneath the chenier plain to a mottled light or yellow brown beneath the beach-ridge plain. The carbonate nodules are dolomitic, 1-2cm in diameter and a C-14 date from these nodules beneath the modern beach near Karumba yielded an age older than 28,000 yrs B.P. (ANU 1895). Pisolitic pebbles in the soil are small, often less than 5mm and occasionally no larger than coarse sand. Quartz grains vary in size from medium to coarse sand with a highly pitted and iron-stained appearance typical of quartz grains from modern soil profiles. The upper 10-20cm of this zone is largely (70-95%) silt and clay with up to 40% of the clay in the submicron size range. Further north, beneath the Edward River beach-ridge plain, this soil becomes sandier, with an attendant decrease in clay material, a change to be expected considering the dominance of coarse sandy sediments in the uplands of this region.
REGIONAL MORPHOSTRATIGRAPHY - CHENIER PLAIN

Mapping units

A small scale morphostratigraphic map has been constructed for the chenier plains between Karumba and the Leichhardt River (Figure 4-1). For ease of discussion this map has been divided into five sectors from east to west. Shown on the map are six major mapping units which define the morphostratigraphy of the chenier plain at a regional scale. Pre-Holocene rocks ranging in age from mid-Quaternary to lower Cretaceous underlie the upland surface in Figure 4-1. Pleistocene strandline deposits for which Doutch (1976) has inferred an age of 110-120,000 yrs B.P., were described in Chapter 2 and are distributed in a discontinuous fashion around the Gulf at or near the coastal plain/upland juncture. An alluvial facies, alluvial silts and sands combines the meander belt point-bar facies and flood plain deposits (Ingram, 1973). The other three units, which are Holocene, have geometries and lithologies which have been described in Chapter 3. Shelly chenier sands, high-tide muds, and aeolian silts and clays are time-transgressive units, with the oldest parts of the unit occurring at the inland edge of the chenier plain.

Sector 1 (Figure 4-2)

North of the Norman River a complex of chenier ridges extends from the modern coast to approximately 8km inland. High-tide muds and aeolian silts and clays surround this complex and extend inland almost 20km adjoining Pleistocene strandline deposits and alluvial silts and clays at the edge of the upland surface. In this area, Pleistocene beach sediments are unconsolidated, and are composed of highly leached medium to coarse sands. The ridges are less than 0.5m high, therefore lacking topographic expression and their presence is distinguishable only by a
Figure 4-1. A regional morphostratigraphic map for between Karumba and Burketown shows the location of in presented in the following figures.
The morphostratigraphic map for the chenier plain shows the location of individual sectors among figures.
Figure 4-2. Morphostratigraphic map of sector 1, chenier plain.
marked lineation of vegetation along their trend.

Sector 2 (Figure 4-3)

South and west of the Norman River is the area of the chenier plain most extensively studied. Extending across the Bynoe River toward the Flinders River is a sequence of well preserved chenier ridges and remnants of ridges flanked by aeolian silts and clays. Correlation of groups of cheniers across the Norman/Bynoe and Bynoe/Flinders interfluves may be done visually by inspection of their general trends. Correlation of cheniers by radiocarbon ages will be discussed in the next chapter. These cheniers demonstrate a recurved form next to the major rivers suggesting that the Norman, Bynoe and Flinders rivers have altered course little during the upper Holocene.

At Red Bluff and adjacent to the upland surface south of Russell Creek, Holocene cheniers are welded directly against pre-Holocene deposits, usually Pleistocene strandline material. The sediments of these cheniers show the expected increase in terrigenous components, especially pisolitic ironstone pebbles and laterite gravels.

In the Bynoe/Flinders interfluve area, the upland surface occurs 25-30km from the active coast and high-tide muds and aeolian silts and clays extend over this entire distance. The landward chenier ridges have been reduced to remnants aligned in a subtle lineation east-west across the back of the chenier plain. Erosion by floodwaters of oldest cheniers appears to typify the central portion of the chenier plain.
Figure 4-3. Morphostratigraphic map of sector 2, chenier plain.
Sector 3 (Figure 4-4)

Between the Flinders River and Morning Inlet, the younger cheniers exhibit best preservation where they are flanked by aeolian silts and clays. Some cheniers are continuous for more than 10km, although near the mouth of Wildhorse Creek, even the younger ridges are discontinuous. Whether the chenier ridges in this area were discontinuous originally, or were later eroded by floodwaters during the wet seasons cannot be established. As shown in Chapter 3, there is evidence from modern processes to support both possibilities.

In this central sector, aeolian silts and clays have formed into large dune fields such as the one southwest of Wildhorse Creek. A similar complex is located west of Morning Inlet. These dune fields exhibit drainage patterns superimposed in their primary ridge and swale pattern. Dune crests trend in a northwesterly-southeasterly direction, a trend not frequently observed in modern active clay and silt dunes. The highly eroded nature of the dunes suggest these landforms to be of greater age than the oldest dunes flanking chenier ridges (perhaps pre-Holocene). A number of excavations in the dune field adjacent to the Morning Inlet failed to provide datable materials, therefore their age remains indeterminate.

In the vicinity of Wildhorse Creek, young active dunes are being deposited contiguous with older eroded ones, but with a differing crest trend. Active dunes in this and other areas of the chenier plain exhibit a multi-directional trend along their crests with lunate and crescentic forms common. A non-systematic orientation of crest and swale trends results in the irregular undulatory surface exhibited by young dune fields flanking chenier ridges.
Figure 4-4. Morphostratigraphic map of sector 3, chenier plain
Sector 4 (Figure 4-5)

West of Morning Inlet where both Holocene and Pleistocene strandlines are welded parallel to the upland surface, numerous water holes such as Manrika Lake and Rocky Waterhole occupy a relict lagoonal position landward of the Pleistocene strandline. Other bodies of fresh water, such as Bird Lake fill similar depressions between the Holocene and Pleistocene ridges. Seaward of this area the debris strewn transitional mudflat is widest. This mudflat was described in Chapter 3 as having a gentle seaward slope and being the surface on which cheniers are developed. The transitional mudflat is 10km wide near Middle Point but narrows to less than a kilometer near Gore Point. Except for its seaward slope, it is morphologically and lithologically the same as the low-tide tide mudflat.

Sector 5 (Figure 4-6)

Beyond Disaster Inlet, the chenier plain broadens and becomes similar to that in the east. A series of evenly spaced chenier ridges flank the mouth of the Leichardt River, extending westward in a morphology similar to the eastern portion of the plain. Mangrove fringes are more intermittent west of Gore Point, frequently interrupted by sandy low-tide flats and active cheniers, probably a result of increased terrigenous sand supply near the outlet of the Leichardt River.

Norman/Bynoe interfluve

Figure 4-7 is a morphostratigraphic map prepared for a limited portion the Norman/Bynoe interfluve area. At this larger scale, differentiation of eight Holocene morphostratigraphic units is possible

(subtidal muds, ebb-tidal sands, low-tide muds, mangrove swamps,
Figure 4-5. Morphostratigraphic map of sector 4, chenier plain.
Figure 4-6. Morphostratigraphic map of sector 5, chenier plain.
Figure 4.7. A detailed morphostratigraphic map for the Norman/Bynoe interfluve shows eight Holocene and two pre-Holocene units. Note the use of the unit *mangrove swamps* to map both the intertidal organic muds and the tidal point-bar muds.
SUBTIDAL MUDS

EBB-TIDAL DELTA SANDS

LOW-TIDE MUDS

MANGROVE SWAMPS

HIGH-TIDE MUDS

TIDAL OVERBANK MUDS

MEOCANE SILTS AND CLAYS

CHENIER SANDS

PLEISTOCENE BEACH RIDGES

UPLAND SURFACE

5 km
high-tide muds, tidal overbank muds, aeolian silts and clays and shelly chenier sands). At such a large scale, tidal point-bar muds and intertidal organic muds appear to coincide with mangroves. The vegetation defined map unit, mangrove swamps, represents both of these facies.

STRATIGRAPHIC DESCRIPTION—CHENIER PLAIN

Three transects were placed across the chenier plain between Karumba and Burketown. The localities of these transects are shown in Figure 4-8. Two transects; Karumba transect and Pandanus Yard transect, were drilled and topographically surveyed, but the Wildhorse Creek transect was sampled only for dating purposes. The two drill transects will be utilized in this section to establish morphstratigraphic sections of the chenier plain.

Figures 4-9 and 4-11 present the stratigraphy of Karumba and Pandanus Yard transects respectively. The top of each drill hole is positioned at its surveyed position above tidal datum, therefore facies boundaries are located in their correct topographic position with respect to prediction datum. Topographic positions of certain facies boundaries are essential to the discussion of sea level change (Chapter 6).

Karumba transect

The Karumba transect extends from the edge of the upland surface south of Russell Creek, north on the Norman/Bynoe interfluve, and reaches the sea near the mouth of the Bynoe River. It is 15.2km long, crossing 3 major groups of cheniers and several smaller remnants. Drill sites were usually placed on chenier ridges or high-tide flats, except 5229, thereby omitting aeolian silts and clays from most of the drill logs. Holes on
Figure 4-8. Three transects were used to characterize the morphostratigraphy of the chenier plain.
cheniers illustrate the distinct facies boundary between shelly chenier sands and low-tide muds, a boundary to be utilized as a paleo-sea level indicator in later discussion.

Located on the oldest Holocene beach ridge, drill hole 5217 (Figure 4-9) contains a thin Holocene record. Holocene sediments are laterally and vertically close to iron-rich laterized sandstones at this site. Pisolitic ironstone and laterite gravel is therefore present in all facies. The 0.7m thick sandy layer directly overlying the laterite pallid zone may be an alluvial equivalent of the Carpentaria paleosol rather than a younger transgressive facies as indicated. This possibility is a recurring problem in which the Carpentaria paleosol is either absent or poorly developed and Holocene marine sediments are in abrupt contact with laterized bedrock. The differentiation is rather inconsequential.

Stratigraphy from hole 5229 is ambiguous. Located approximately 2.4km north of hole 5217 (Figure 4-9), the presence of a sandy mud below the aeolian silts and clays offers some interpretative problems. Shown here as high-muds, the parent material may have been either low-tide or subtidal muds modified by the downwasting which has occurred at the back of the chenier plain. Lower in the stratigraphy, a clear pre-Holocene boundary is difficult to establish, the shelly sands and gravels are best accepted as the Holocene basal transgressive unit.

Data to the seaward of hole 5229 is far more straightforward. The stratigraphic equivalents of the subtidal muds and low-tide muds are readily identifiable. The boundary between shelly chenier sands and low-tide muds which is important as a sea level indicator is present in all drill hole which pass through chenier ridges, whereas, the veneer of high-tide muds is present everywhere beneath the surface of the high-tide
Figure 4-9. A stratigraphic summary of the Karumba transect shows drill hole locations.
The presence of a clearly defined soil with carbonate nodules, pisolitic ironstone pebbles and a brown clay matrix in all other holes excepting 5227A simplifies location of the basal Holocene deposits.

The most seaward drill hole, 5221, provided valuable information on the facies underlying a modern chenier. Beneath the shelly chenier sands is an abrupt facies change to low-tide muds. These muds show a high degree of preservation of their original color and texture. It is evidence from such sites which indicates that cheniers, once developed to some critical height and width on the transitional mudflat, do not roll back over high-tide flats. From the bottom of drill hole 5221, calcium carbonate nodules from the Carpentaria paleosol provided a minimum radiocarbon age of 28,000 yrs BP (ANU 1895). The certainty of contamination in such a sample suggests that the real age is closer to 42,000 yrs BP (see Chapter 5).

Drill hole 5227 (Figure 4-9) is taken as representative of chenier plain stratigraphy. Figure 4-10 illustrates down-hole variation of mud, calcium carbonate, organic, and sulfate content, characteristics which were distinctive in some modern facies. Percent mud gives a general impression of grain size, percent calcium carbonate is directly related to the presence of either bioclastic or soil carbonate, percent organics may suggest the presence of mangroves, and sulfate suggests the presence of gypsum which is usually associated with environments subjected to subaerial efflorescence. Differentiation of aeolian silts and clays from underlying muds cannot be made from parameters displayed in this figure. Distinction of these facies was made at the drill site on the basis of apparent homogeneity of aeolian silts and clays compared to the alternating layers of silt and clay found in the lower unit. In this case, the aeolian facies overlies high-tide muds. These muds had
Figure 4-10. The drill log of 5227 is presented as typical of chenier plain stratigraphy. Down-hole variation of mud, calcium carbonate, organic material and gypsum content is presented here. Dark squares indicate actual sample position.
opportunity to develop on a parent material consisting largely of low-tide muds prior to the formation of the overlying clay dune.

From drill hole data, low-tide muds appear to have a gradational contact with subtidal muds. The distinctive characteristic between the facies is one of sedimentary structure, with the subtidal muds showing little or no layering in contrast to the layered low-tide muds. A color change is usually associated with the low-tide / subtidal boundary. Subtidal sediments appear gray and less oxidized than the brownish colors of the overlying low-tide muds. Other characteristics such as calcium carbonate, organic and sulfate content are of little assistance in separating these two units.

A coarsening downward trend is typical of subtidal muds. There is also a trend toward increased shell content downward in this facies. A decrease in mud fraction with depth is in agreement with the generalized trend for modern Gulf bottom sediments to coarsen with increased water depth and distance from shore (Chapter 3). The subtle variation in organic content in subtidal muds is atypical when compared with other drill logs and probably not significant. The decrease in organic material in the weathered pre-Holocene sediments is to be expected.

The sharp increase in sulfate just above the contact with the Carpentaria paleosol is recurrent in all drill logs where such analysis was performed. The most plausible explanation is upward movement of sulfates from gypsum formed sub-aerially in the pre-Holocene soil. Analysis of samples at and below the buried the soil in this and other drill holes show only trace amounts of sulfates.

The boundary of the subtidal muds with the Carpentaria paleosol is always accompanied by a textural and color change. Mud content increases
sharply as shown in Figure 4-10 and decreases abruptly below the upper 10-20 cm of the soil. Calcium carbonate is usually absent except as dolomitic soil carbonate. A color change from light gray or blue gray to a dark or reddish brown is the single most distinctive indicator of this unconformity. The Carpentria paleosol is invariably more compact than the overlying marine sediments, a change readily noted on site by the drill operator.

Pandanus Yard transects.

The Pandanus Yard transects are located in the western portion of the chenier plain, near Disaster Inlet. (Figure 4-5). The easternmost transect is an 8 km section across Holocene and Pleistocene features. A short 2 km section to the west was used to establish topographic relationships between Pleistocene beach rock capped with aeolianite and the Holocene strandline sequence. Figure 4-11 presents a compilation of data from these two transects.

Holocene coastal morphology at Pandanus Yard, is a composite of chenier plain and beach-ridge plain types (Figure 4-12). The Pandanus Yard transect was selected to provide information on this composite coastal progradation. It offers little information on chenier plain stratigraphy additional to the Karumba transect. However, the change from a beach-ridge to a chenier plain may be correlative with episodic and/or sea level influenced progradation noted elsewhere on the Carpentaria coastal plain.

Drill hole 5243B was drilled on the Pleistocene portion of the transect (Figure 4-11). Pleistocene beach-rock was found within 2 m of the surface. The overlying 2 m of highly leached sandy material is probably beach sediment of Pleistocene age, especially with its obvious
Figure 4-11. A stratigraphic summary of the Pandanus yard transects projects the data from short transect near Bird Lake onto the major transect which includes beach ridges cheniers and mudflats. (see Figure 4-8)
Figure 4-12. The coastal plain near Pandanus Yard is a composite of beach-ridge and chenier plain morphology. Locations of the Pandanus Yard transect and the nearby supplementary transect at Bird Lake are shown on this vertical air photo. Approximate scale: 1cm = 800m.
lack of shell material.

Seaward of the Pleistocene section is a 100m-wide series of Holocene beach ridges separated by shallow swales of less than 1.5m amplitude, with at least eight distinctive ridge crests appearing in the topographic survey. Stratigraphy beneath this section of the transect is not as readily interpreted as the chenier plain portion to the seaward. Drill hole 5243A (Figure 4-11) contains shelly sands overlying a cemented and recrystallized shell hash which in turn overlies compact dewatered, slightly oxidized clays. The carbonate enriched sands are probably subtidal sand or nearshore transgressive sediments deposited over either transgressed lagoonal muds or a muddy pre-Holocene surface as the marine transgression slowed and reached its highest level. In drill hole 5242, the base of shelly beach-ridge sands is clearly indicated. However, underlying these sands are interbedded silts and clays grading downward into a more homogeneous gray clay. Such a sequence again supports the view that low-tide muds remain preserved beneath chenier ridges where they are protected from sub-aerial exposure. The general mode of accretion on this portion of the transect appears to have been initially vertical and later lateral as a prograded beach-ridge plain.

Sediments beneath the chenier plain section of the transect show stratigraphy similar to that of the Karumba transect. High-tide muds appear as a thin layer over low-tide muds, suggesting that the lower facies may have provided a parent material from which the overlying facies has developed by subaerial exposure and tidal inundation. The seaward chenier shows a well defined boundary with low-tide muds at its base. At the bottom of the section, subtidal muds blanket a pre-Holocene surface with no evidence of a basal transgressive sand.
Morphologic descriptions of other chenier plains have been extensive, but a universally applicable model is not available. The work of Gould and MacFarlan (1959) is probably the best example of chenier plain description. However, block diagrams in their paper are difficult to read because different units are used on the surface from those used in vertical section. Cook and Mayo (1977) presented what is probably an oversimplified landform map of the Broad Sound chenier plain. It is likely that some of their map units, especially the "supratidal mudflat and coastal grassland" unit could have been further subdivided. Augustinus (1978) described the Surinam coast detailing the sedimentologic and ecologic aspects of the coastal plain. He presented maps and photos of local features in combination with cross sectional diagrams illustrating environmental relationships and soil development. However, regional geomorphology of the coastal plain was not presented in a map view. The data presented here show the Carpentaria situation to be morphostratigraphically unlike Louisiana and Surinam but with some similarity to the Broad Sound, Queensland, marine plain. The dominance of extensive unvegetated flats of uniform elevation suggests morphologic comparisons with arid or semi-arid deltaic coasts such as Cambridge Gulf, Australia (Thom et al., 1975) or the Gulf of California (Thompson, 1968).

Comparison with examples from temperate coasts is implicitly difficult. Temperate coasts, whether they are deltaic or estuarine, are usually characterized by marsh development. Marsh development results in de-activation of high-tide and supratidal areas by upward organic accretion. Conversely, coasts with high evaporation rates and impeded drainage support the development of hypersaline soil conditions which
preclude vegetational development on large areas of mudflat. Hypersaline soil conditions on a portion of the Surinam coast are discussed by Augustinus (1978). He found that the pronounced annual dry season raises soil salinities to 82ppt during periods of limited drainage, seriously affecting vegetation.

Comparable high soil salinities are probably most responsible for the maintenance of extensive vegetation free zones on the high-tide flats of the Carpentaria chenier plain. Furthermore, considering the physical effect of the long dry season on high-tide mudflats in the Gulf, it is likely that such conditions are the most critical environmental factors responsible for chenier plain morphology in the Gulf of Carpentaria.

Stratigraphic models of chenier plain development (Curray, 1969; Hoyt, 1969, and Reineck and Singh, 1973) are based on a limited number of examples, with southwestern Louisiana providing the overwhelming amount of influence (Gould and MacFarlan, 1959; Price, 1955;). Implicit in the definition of a chenier plain is the requirement for coarse strandline materials to overlie a fine-grained facies with this facies boundary at or near mean sea level. This underlying fine-grained facies has been attributed to various environments of deposition (Chapter 1). Principle among the work reported in Chapter 1 was the report by Gould and MacFarlan (1959) who showed some Louisiana cheniers to be overlying Gulf bottom or nearshore deposits. Coleman (1966) revealed that Pecan Island, a Louisiana chenier, was underlain by marsh deposits at its west end and nearshore marine deposits at its east end. Cook and Polach (1973) showed Broad Sound cheniers to be resting on mangrove deposits 1-2m thick which overlie intertidal muds, a situation analogous to that reported by Jennings and Coventry (1973) for coarse strandline deposits in the Fitzroy estuary.
Stratigraphic findings from the Gulf of Carpentaria correlate closely with descriptions of the Louisiana cheniers which rest on Gulf of Mexico nearshore sediments. Carpentaria cheniers are underlain by a lithologically defined facies which matches precisely the parameters established for modern low-tide muds. Landward movement and overwash of chenier deposits onto high-tide muds appear to be an exception rather than the rule. Cheniers in the Gulf of Carpentaria appear to accrete vertically in situ from a base of low-tide muds and once established remain stable. This finding is central to further interpretations of Holocene relative sea level and modes of deposition on the Carpentaria chenier plain.

REGIONAL MORPHOSTRATIGRAPHY- BEACH-RIDGE PLAIN

Mapping units

A small scale map (1:250,000) has been constructed for the beach-ridge plain between the Kendall and Mitchell Rivers (Figure 4-13). This region has been subdivided into 4 sectors from north to south. In comparison with the chenier plain, there are slightly different mapping units used at the regional scale. The upland surface consists of several Holocene and pre-Holocene alluvial units. Grimes (1977) named the surface of these units according to a physiographic scheme shown in Figure 2-9. Major land surfaces are the Holroyd Plain, the Mottle Plain and Mitchell Fan consisting of channel, levee and floodplain deposits. In some cases, as illustrated in Chapter 3, upland alluvium has breached the oldest Holocene strandline and has been deposited over Holocene coastal sediments. Usually, there is a well defined juncture between Holocene coastal deposits and this alluvial facies, permitting the combination of all alluvial sediments into one mapping unit, the upland surface. This condition is in contrast to the chenier plain where
Figure 4-13: A regional morphologic plain between the Mitchell and Kendalls Creek sectors presented in individual sections. Indicate the locations of Christman Creek transects.
Figure 4-13. A regional morphostratigraphic map for the beach-ridge plain between the Mitchell and Kendall rivers shows the locations of individual sectors presented in the following figures. Section lines indicate the locations of Christmas Creek and Edward River stratigraphic transects.
distinction between older lithified rock units and younger alluvium required the use of two map units.

**Pleistocene strandline deposits**, which are correlative with the cemented calcarenite facies found adjacent to the chenier plain, occur in adjacent to the beach-ridge plain as leached, mottled, brown and yellow quartzose sands. Although Smart (1976) found these sands to be sub-horizontally bedded at Cape Keer-Weer, excavations east of Edward River settlement showed them to be apparently structureless in section. Pleistocene beach ridges are frequently forested with upland species. They consist of two or three crests, which together may be 1.5km wide. Loss of height by leaching of calcium carbonate during the late Pleistocene and throughout the Holocene causes original crestal elevation of ridges to be indeterminate. If the percentage of bioclastic material in modern ridges is any indicator of that in the Pleistocene ridges, then a 40-60% loss in volume could be assumed. The best indicator of mean sea level during the formation of these ridges might therefore be derived from stratigraphic data on their bases. However, drilling results show that the leached beach-ridge sands overly sandy alluvial deposits which exhibit textural qualities similar to the marine sands. Pleistocene ridge bases, even when seen in section at tidal creek banks are difficult to identify.

**High-tide muds, supratidal silts and clays** and **beach-ridge sands** are all facies of Holocene age discussed in Chapter 3. Their application as morphostratigraphic mapping units is in accordance with principles presented at the beginning of this chapter by which environment, facies, and morphostratigraphy are related. The same sedimentary parameters are therefore applicable to these units when used in the morphostratigraphic context.
Sector 1 (Figure 4-14)

It is the northern portion of the map area, adjacent to the Holroyd and Kendall rivers, where the alluvial upland facies has extensively breached the Pleistocene strandline. Alluvial material has been deposited contiguous with Holocene coastal deposits where the Pleistocene strandline exhibits a recurved form at the Kendall River. The presence of alluvium over Holocene coastal deposits in this area is probably attributed to the large sandy bedload transported by the Kendall and its tributary the Holroyd. During overbank stages, sandy material is deposited beyond the channel by processes described in Chapter 3.

Except for the beach ridges at the mouth of the Kendall, which are part of the Cape Keer-Weer complex, the coastal plain in the northern portion is dominated by supratidal silts and clays with limited erosion next to tidal creeks producing small areas of high-tide muds. Average coastal plain width is 3km and high-tide flats flank the tidal creeks, occasionally extending through a breach in the Pleistocene strandline onto the upland surface. Too small to be shown in here, the active coast is comprised of a single sandy beach-ridge which protects the mud facies to the landward from erosion. Well-cemented Holocene beach-ridge rock is present along the active coast, frequently showing evidence of erosion and undercutting.

Sector 2 (Figure 4-15)

Near Christmas Creek, a Holocene beach-ridge plain is deposited against the Pleistocene strandline. Individual ridge crests on the Holocene plain exceed 50 in number. The complex is almost 6km in width, but discrete ridge crests are difficult to count, especially in the oldest sections of the plain. Many crests are vegetation defined and
Figure 4-14. Morphostratigraphic map of sector 1, beach-ridge plain.
Figure 4-15. Morphostratigraphic map of sector 2, beach-ridge plain.
only visible on aerial photographs with topographic variation of older crests less than 0.5m from swale to crest.

Despite the apparent conformable relationships of ridges at the widest part of the plain, there is at least one lithologically defined discontinuity which subdivides the plain into two sets of ridges. These sets show distinctive lithologic and radiocarbon dating characteristics. The radiocarbon date findings will be discussed in Chapter 5.

Lithologic qualities which assist in the separation of ridge sets include soil development, the presence or absence of shell in the upper 0.3m of the ridge, and the sharp well-defined crests of younger ridges in contrast to the subdued, rounded crests of older ridges. These parameters, when supplemented by morphologic discontinuities at the distal (north and south) ends of the complex give evidence of episodes of coastal accretion separated by either non-accretion or erosion. Some observations suggest that conformable ridge crest orientation does not preclude the existence of either diastems or erosional disconformities in the history of the prograded plain. Given the appropriate physical conditions at the onset of the next progradational period, it is apparent that accretion may proceed with conformable orientation to a record considerably older.

Christmas Creek, which cuts across the northern end of beach-ridge complex is flanked by both high-tide and supratidal flats. The high-tide flats extend across the Holocene portion of the plain, through a gap in the Pleistocene ridges and onto the upland surface. The lower channel of Christmas Creek presents a 1km crosssection through some of the younger ridges before being deflected 5km south by the seaward ridges. It is this 1km channel section which provides some of the best descriptive
information concerning internal stratification of relict beach ridges (Chapter 3).

Sector 3 (Figure 4-16)

South of the beach-ridge complex, in the vicinity of the north arm of Balaurgah Creek, the supratidal flats extend landward of the Pleistocene strandline, suggesting that parts of this morphostratigraphic unit may be older than mid-Holocene. Since the supratidal surface is mapped largely by the flat grasslands which cover it, and can only be confirmed as a sedimentary facies by auger drilling, there may be some ambiguity in areas such as this one where auger holes are lacking. That the most landward portions of this unit existed as mudflats or lagoonal environments for the Pleistocene seas whilst the seaward portions were tidal flats during the mid-Holocene is a plausible history for this landform. Topographic surveys showed that although local relief is considerable (0.4-0.5m) overall seaward slope is gentle with slopes of angle 1:1000 - 1:2000.

High-tide mudflats near the confluence of Balaurgah Creek and Edward River are separated from the supratidal flats by a scarp described in Chapter 3. These high-tide flats remain inundated during a portion of the wet season, and although the two rivers have separate mouths, they are joined by several tidal channels across these flats. Tidal creeks which extend south on the coastal plain between the Holocene and Pleistocene ridges exhibit headward gully erosion into the supratidal flats, an observation which reinforces the active erosional nature attributed to the scarp separating the two mudflat surfaces.
Figure 4-16. Morphostratigraphic map of sector 3, beach-ridge plain.
Sector 4 (Figure 4-17)

South of the Edward River, and across the channels of the Moonkan and Chapman Creek, another beach-ridge plain extends toward the margins of the Mitchell River delta. Near the Coleman River, this beach-ridge plain broadens into a chenier plain which continues across the distributaries of the Mitchell. The Edward River settlement is located on the beach-ridge plain between the Moonkan and Chapman creeks and the location of the settlement is approximately coincident with a well-defined oblique intersection of two ridge sets. The relationship of ridges on both sides of this discontinuity was the subject of both stratigraphic and radiometric investigations, matters to be discussed later in this chapter and Chapter 5 respectively.

The Holocene ridges between the Moonkan and Chapman creeks are separated from the Pleistocene strandline by both supratidal and high-tide flats. This area provided many of the examples used to describe modern environments which were presented in Chapter 3. The scarp which evident in this area and headward gully erosion on both creek systems is extending the high-tide flats inland through the breaches in the Pleistocene ridges and south toward Malaman Creek.

This area presents the most expansive section of high-tide flats between the Kendall and Mitchell rivers and it is the only locality where evidence of minor aeolian processes may be observed. Incipient clay and silt dunes less than 0.3m high are noted against the supratidal scarp near Chapman Creek. Tidal creeks show small aeolian accumulations along their banks where vegetation provides a baffle for sediment collection and where they are flanked by broad areas of high-tide flat to the southeast. However, none of these occurrences are significant to warrant the addition of an aeolian facies to the morphostratigraphic map.
Figure 4-17. Morphostratigraphic map of sector 4, beach-ridge plain.
At Malaman Creek, the beach ridge plain begins to broaden, and at the Coleman River it becomes a chenier plain in response to sediment input from the Mitchell River (Figure 4-18). The chenier plain associated with the Mitchell delta is a local occurrence of this morphology, with beach-ridge plains present south of the Mitchell delta. This delta requires descriptive and analytical methods somewhat different from the beach-ridge plain to the north. There is evidence of several additional facies, probably unique to this sort of deltaic system, which necessitate exclusion of this delta from the present investigation.

STRATIGRAPHIC DESCRIPTION—BEACH-RIDGE PLAIN

Two transects were placed across the beach-ridge plain between the Kendall and Mitchell Rivers (Figure 4-13). Both the Christmas Creek and Edward River transect were drilled and surveyed, with topographic data linked to a tide gauge on Chapman Creek. The Chapman Creek tide records, discussed in Chapter 2, provided a twelve month record which established the amount of annual sea level variation at this locality.

Figures 4-19 and 4-20 present the stratigraphy for Christmas Creek and Edward River transects respectively. Stratigraphic facies are identical to those used on the chenier plain excepting supratidal silts and clays which occupy a position vacated by aeolian silts and clays, a facies not significant on the beach-ridge plain. To facilitate later discussions of age structure and relative sea level, drill holes are placed in their topographically corrected position with respect to the modern datum.

Christmas Creek transect.

This transect is located across the widest part of a beach ridge.
Figure 4-18. South of Malaman Creek, the coastal plain broadens and the beach-ridges trend into chenier ridges resting on a muddy substrate. Ridges in the left foreground are composed of mud with small amounts of whole shell. The Coleman river is out of view under the aircraft.
complex between Christmas and Balaurgah Creeks. (Figure 4-19). The Holocene sequence is approximately 4300m wide with an additional 1000m of coastal plain comprised of Pleistocene beach ridges. The transect was selected for its apparent continuous record of Holocene progradation.

Proceeding from landward to seaward, drill hole 5297B is located on the most seaward Pleistocene ridge. It served to characterize both the Pleistocene beach ridge facies and the underlying alluvium. Separating this Pleistocene ridge from the oldest Holocene ridge is a narrow swale in which drill hole 5297A was sited. The surface of this swale consists of a dark brown clay with pisolithic ironstone overlying a fine yellow quartzose sand, similar to sands of the Pleistocene ridges. The clay is interpreted as a facies equivalent of the supratidal clays and silts which form the expansive plains between the Holocene strandline sediments suggests that the swale materials could be either Pleistocene or Holocene in age, a point which supports the assertion that many of the coastal plain facies are time-transgressive.

Drill hole 5297 is located on the oldest Holocene ridge. Although lacking datable shell material, the ridge is certainly of Holocene age. Its morphology, lithology and contiguous alignment with the Holocene complex provides evidence of its relative age. Identification of subtidal muds below the low-tide muds is difficult. However, the well defined boundary at the base of the beach facies provides important data for later discussion of mid-Holocene sea level.

Drill hole 5296C, 500m to the seaward of 5297 provides the first continuous section containing all the fine-grained Holocene facies. The top of the hole shows supratidal silts and clays overlying low-tide muds, with almost two meters of subtidal muds separating these from the Carpentaria paleosol. Radiometric dating from the low-tide and subtidal
Figure 4-19. A stratigraphic summary of the Christmas Creek transect shows drill hole locations.
Facies confirms the Holocene age of these sediments.

Four drill holes to the west of 5296C are sited on ridges with a gently undulating form composed of leached carbonate-free sands in the upper two meters. Calcium carbonate occurs below this leached zone in a calcite form, cementing shell hash into material similar to the beach-ridge rock described by Thom (1969). Only shells which are well below the seasonal oscillations of the groundwater table retain an aragonitic mineralogy. The upper sands of these ridges vary from red and yellow medium quartz sands in drill hole 5296A to white fine quartz sand in 5295 and 5296 becoming yellow and slightly coarser again in 5294B. Below the beach-ridge sands is the critical facies boundary with low-tide muds, in all cases identifiable by the lithologic contrast with overlying coarse material.

Drill hole 5294B is adjacent to a lithologic boundary which is defined by the absence of shell fragments in surficial sands to the landward and an abundance of shell in these sands to the seaward. Comparison of stratigraphic data between holes 5294A and 5294B confirms the lithologic contrast which is surficially indicated. Sands from the upper 1.5m portion of 5294B are carbonate-free, whilst sands from the upper portion of 5294A are extremely shelly, sufficient to be described as shell hash. Beach-ridge rock cementation by low-magnesian calcite is present in both holes near the groundwater table, but consolidation in the seaward ridges is not to the extent noted in older ridges to the landward. Red oxidized sands are present below the water table and near the facies boundary with low-tide muds in both holes.

A sequence similar to 5294A is repeated in the adjacent drill hole, 5294, to the seaward. Sandy shell hash overlies slightly recrystallized shell which grades downward into red oxidized sands and yellow shell
hash. Drill hole 5293A, sited in a water filled swale an additional 200m seaward confirmed that a sequence correlative with that underlying the ridges also underlies the swales. This swale is floored with a thin layer of organic debris which grades into a low-tide mud facies within 1.0m of the surface.

The three remaining drill holes to the seaward, 5293, 5292, and 5291 (see Figure 4-19), show a stratigraphic sequence which typifies the Holocene facies important to coastal progradation: beach-ridge sands, low-tide muds and subtidal muds. The beach-ridge sands are progressively less cemented by low-magnesian calcite in the younger ridges, with no recrystallization observed in hole 5291, drilled on the modern beach. Boundaries between facies, especially the sandy beach-ridge sediments and the muddy low-tide flat deposits are unambiguous in the younger ridges, permitting assessment of the topographic variation in this boundary across the entire prograded plain.

A facies absent in this section is the basal transgressive sand, which appears only occasionally in the stratigraphy of the Edward River transect, 35km to the south. Basal transgressive sands occur in a limited manner beneath the beach-ridge plain, similar to their occurrence beneath the chenier plain. Their presence is usually accompanied by evidence suggesting a pause in transgression accompanied by an upbuilding of sandy material out of the subtidal and intertidal zone into a strandline position, a situation most common beneath the oldest Holocene ridges.

Edward River transect.

This transect is located through the township of Edward River (Figure 4-20). It is positioned across a morphologically obvious
discontinuity which may correlate with the shelly/non-shelly discontinuity described on the Christmas Creek transect. Landward of the Holocene beach-ridge complex and flanked by supratidal flats, is the most extensive high-tide mudflat between the Kendall and Mitchell Rivers. The transect is 7000m long, 4000m of which is Holocene ridges with the balance consisting of supratidal flats, high-tide flats and a single Pleistocene beach ridge.

At the landward edge of the supratidal flats, (Figure 4-20), near the Pleistocene beach ridge, a silt and clay facies overlies sandy material in drill hole 5288. This sandy facies is apparently a nearshore equivalent of the Pleistocene beach ridges. The silt and clay facies near the surface may be a soil horizon developed on the sandy material since the Last Interglacial (120,000 B.P.). The possibility of later modification by mid-Holocene tidal processes during deposition of the westernmost parts of this facies cannot be eliminated. Supratidal silts and clays must then be considered a time-transgressive facies, portions of which may date from the upper Pleistocene. Oldest portions of this facies may be equivalent to the Carpentaria paleosol.

Seaward of the Pleistocene beach-ridge, drill holes 5288, 5263, and 5262A were drilled on the supratidal and high-tide flats, in an effort to establish the relationship between Holocene ridges, the flats and the Pleistocene strandline deposits. Unfortunately, nowhere between the Holocene and Pleistocene ridges can the Carpentaria paleosol be identified at the surface by lithologic parameters equivalent to its stratigraphic occurrence. As shown in Chapter 3, only the upper portion of the supratidal flat near drill hole 5262A provided datable materials consisting of carbonate nodules and gastropod shells. The carbonate nodules provided a minimum mid-Holocene age of 4150 ± 80 (ANU 1897).
Figure 4-20. A stratigraphic summary of the Edward River transect shows drill hole locations.
Elsewhere, datable materials in the form of either organic or shell material are absent. Therefore, age correlations east of drill hole 5262A are unclear.

Drill hole 5262 (Figure 4-20) was placed on the oldest Holocene ridge and its log is different from the others on this transect. The uppermost three meters of the ridge consists of fine quartzose sand which changes from brown at the surface to yellow downward. Datable shell was recovered from near the groundwater table, but shell material higher in the section is virtually absent. A boundary with underlying silts and clays occurs lower in the section than elsewhere on the beach-ridge plain. The evidence that the ridge began building upward prior to the mid-Holocene transgressive maximum is strong, especially in the light of a radiocarbon date obtained from shell in this ridge (Chapter 5).

A group of three drill holes to the seaward present a stratigraphy similar to that of the Christmas Creek transect. These three holes (5260B, 5260A and 5260) are located across a series of beach ridges which appear to be morphologically and lithologically contiguous. The underlying stratigraphy is also readily interpreted with the exception of some ambiguity at the base of the beach-ridge sand facies in hole 5260A. This ambiguity may be due to poor recovery at the field site. Data from adjacent holes is used to clarify the position of this boundary.

Drill holes 5260 and 5259A straddle the morphologic discontinuity defined by the oblique intersection of ridge sets. Stratigraphic data reinforces surficial observations concerning the relative differences of sediments across the discontinuity. Shell is markedly absent from sands landward of the boundary, whereas shell hash is prevalent on the seaward side of the boundary. A lower topographic position for the base of beach-ridge sands seaward of the discontinuity suggests that this feature
may represent something more than just a minor diastem in beach-ridge accretion (see Chapter 5).

The remaining three holes to the seaward (5259A, 5259 and 5258) exhibit a lithologic trend toward the younger ridges which is repetitive of the sequence noted at Christmas Creek. Evidence of recrystallization decreases seaward. The facies boundary below beach-ridge sands becomes easier to distinguish whilst the lithologic characteristics of facies beneath the modern beach ridge (hole 5258) are almost identical to the modern equivalents occurring in the intertidal and subtidal zones.

Drill hole 5259 (Figure 4-20) was selected for downhole variation in certain lithologic parameters (Figure 4-21). These parameters included mud content, calcium carbonate, organic content and sulfate analysis. Mud content in the samples shows the expected increase across the facies boundary between beach sands and low-tide muds. In contrast with the chenier plain, there is no marked increase in mud content at the boundary with the Carpentaria paleosol. Calcium carbonate indirectly is a measure of shell content, except where dolomitic soil nodules are present in the Carpentaria paleosol. Low shell content is indicated for the low-tide muds, an observation confirmed in the modern environment where fine-grained terrigenous sediments tend to dominate. An increase in shell content in the upper subtidal muds is noted beneath both the chenier plain and the beach-ridge plain with a decrease just above the Carpentaria paleosol. Relict soil carbonate in the Carpentaria paleosol beneath the beach-ridge plain increases calcium carbonate content in that portion of the log.

Organic content shows a more significant variation than beneath the chenier plain. The upper portions of the subtidal muds give visual suggestion of organics, which is confirmed by laboratory analysis.
Figure 4-21. The drill log of 5259 is presented as typical of beach-ridge plain stratigraphy. Downhole variations of mud, calcium carbonate, organic material and gypsum content shown here. Dark squares indicate the actual sample positions.
However, this 1.0% level organics is minor in comparison with levels in some other Gulf sediments, especially considering that mangrove peat facies may contain 7-15% organics by weight.

Similar to the findings beneath the chenier plain, sulfate content remains concentrated in the subtidal muds, although in this case, near the top of the facies. Highest sulfate levels beneath the chenier plain occurred just above the Carpentaria paleosol. Analysis presented in Chapter 3 showing only trace amounts of sulfate in modern subtidal sediments suggests that its presence here may be due to post-depositional mobilization. Replacement of sea water by fresh groundwater would accompany the regression of the coast. This influx of ground water into previous marine sediments accompanied by the seasonal oscillation of water table resulting from wet and dry seasons may promote sulfate movement from a lower facies such as the Carpentaria paleosol.

DISCUSSION—BEACH-RIDGE PLAIN

General

Beach-ridge plains in other parts of the world have attracted attention largely because of the promise of two types of information from their physiography. First crest heights of ridges on beach-ridge plains have been alleged to be directly related to mean sea level during the time of their formation. Second, discontinuities in the plan view of beach-ridge plain have been attributed to episodic changes in sediment supply and nearshore bathymetry, although it has been shown earlier in this chapter that uniform ridge alignment does not necessarily indicate time-continuous progradation. The discussion of beach-ridge plains in Chapter 1, especially the section dealing with genesis, reviewed the various opinions of the use of beach ridges in these two contexts. It is
now useful to return to that review in the light of findings from the Carpentaria coastal plain.

Similar to chenier plains, beach-ridge plains have been traditionally described by their physiography, with only a limited number of workers able to explore the underlying stratigraphy. Special attention has been devoted to the analysis of ridge orientation, spacing and topography for reasons already described. Less emphasis has been placed on the other landform units found on a beach-ridge plain which are often contiguous with the ridges — such as tidal flats.

Early work by Johnson (1919) together with more recent reviews such as Davies (1957) on ridge genesis encouraged researchers to focus their attention on the physiographic aspects of ridges and swales of prograded coastal plains. Nossin (1962, 1965a and b), in Malaysia and Psuty (1966) in Mexico presented detailed maps of the discontinuous relationships between ridge sets whilst Curray et al. (1969) paid full attention to all environments and resulting facies on the Nayarit, Mexico, coastal plain. More recent summaries of coastal landforms such as Bird (1972) and Davies (1972) have neglected to elaborate upon beach ridges in the context of their relationship to adjacent environments of deposition.

Within Australia, Ly (1976), Thom et al. (1978) and Thom (1978) have studied beach-ridge plains on the New South Wales coast where morphostratigraphic evidence derived from form and orientation is less important than the lithologic and radiometric data from stratigraphic sources. In addition, only a limited number of eastern Australian prograded coasts have a similarity with the Carpentaria beach-ridge plain. Thom et al. (1978) consider the eastern Australian examples to be various morphostratigraphic sub-divisions of bay-barriers, permitting comparison of Carpentaria beach-ridge plains only with the "prograded
barrier" type 1a of Thom et al. (1978). These eastern Australian examples are closer to a Galveston Island model (Bernard et al., 1962) than the broad prograded beach-ridge plains of the Gulf of Carpentaria. Only Smart (1976), has studied such a coast in his work at Cape Keer Weer, an area immediately north of the Kendall River. Unfortunately, Smart's stratigraphy does not appear entirely correlative with stratigraphy from this study. Inspection of the drill logs which support his findings, indicate that the disparities are lithologic and not interpretative.

Despite the diverse extent to which other beach-ridge plains have been described, and the lack of a universally applicable depositional model, the morphostratigraphy in some cases, and certainly the morphology in most cases, can be shown to be analogous to the Carpentaria situation. Although the paucity of precise topographic control in other areas may be falsely reassuring, the Carpentaria beach-ridge plains appear similar to examples which have different climatic regimes, a similarity not accorded to chenier plains. Where description of structure or stratigraphy has been more complete as in the case of Psuty (1966) and Curray et al. (1969) the comparisons on a morphostratigraphic level appear favorable. Beach-ridge plains, wherever they occur, may have common genetic properties accompanied by identical or near-identical response to variations in such parameters as sediment supply, nearshore bathymetry and wave climate (see Chapter 7).

Comparison is made with only a few examples. The works of Bernard et al. (1962), Curray et al. (1967), Thom et al. (1978) and Smart (1976) provide the basis for the discussion which follows.

Galveston Island example

Galveston Island is to the study of coastal sand barriers as the
Louisiana coastal plain is to the study of chenier plains. This barrier island which is approximately 45km long and 3-6km wide is comprised of beach-ridges with an aeolian cap. Bernard et al. (1962) defined five distinctive facies in their model of barrier development. In the Galveston sequence, dune sands mantle a foreshore deposit which is up to 3.0m thick. Below this facies, is a subtidal shoreface deposit extending from 3 to 9m below low water. A transition zone extends from 9 to 12m below datum whilst shelf muds occur from 12m to greater depths. The entire Holocene sequence overlies a Pleistocene surface.

Comparison of the Galveston Island sequence with Carpentaria data is possible if depth zones are ignored. the Carpentaria beach-ridges have no aeolian cap, but beach-ridge sands are analogous to the Galveston foreshore deposits. Low-tide muds in the Gulf of Carpentaria could be comparable to a combined facies of shoreface and transition zone sediments on Galveston Island. However, lower shoreface deposits on Galveston Island contain both finely laminated bedding and cross bedding. Cross bedding is rare in low-tide or subtidal facies in the Gulf where either laminated or massive interlayed bedding predominates. Subtidal muds in the Gulf have an analogue in the shelf muds seaward of Galveston Island. However, in waters deeper than 20m in the Gulf, subtidal muds transition to offshore sands, a facies for which Bernard et al. did not describe an analogue.

Nayarit example

Curray et al. (1967) found a stratigraphic sequence beneath the Nayarit plain which strongly resembles that of the Carpentaria system. They identified five stratigraphic facies: littoral sands, alluvium, lagoon and marsh, shelf muds and silty sand. The major difference from
Carpentaria findings is the inclusion by Curray et al. of beach, intertidal and nearshore sands into one lithofacies. The Carpentaria environment required the distinction of two facies (beach-ridge sands and low-tide muds) for this portion of the sequence.

Some of the Nayarit beach ridges, overlie inner shelf facies muds, especially the Novillero transect. These shelf muds are stratigraphically comparable to the subtidal muds found in the Gulf of Carpentaria.

Lagoon and marsh facies are not present in the Gulf for reasons already explained in the section on chenier plains. Perhaps, an appropriate comparison exists between the supratidal silts and clays of Edward River transect and the Nayarit lagoonal facies.

Transgressive sands are present beneath the Nayarit plain, but absent in the Gulf of Carpentaria although a distinctive pre-transgressive alluvium underlies both plains. The absence of a lagoonal or marsh-facies formed in advance of the transgressing Holocene seas in the Gulf of Carpentaria is most responsible for the obvious differences noted at the base of the Holocene sequence.

Despite these rather marked contrasts, the two plains are comparable. Curray et al. make use of the term "regressive sands" in their discussion of the relationship between accretion and sediment influx. They outline factors responsible for progradation of the Nayarit plain, including change in relative sea level and climate, factors which will be shown in Chapter 7 to be applicable to the Carpentaria example.

New South Wales example

Thom et al. (1978) present examples of prograded Holocene coastal
plains which are geographically closer to the Gulf of Carpentaria but less similar than the Nayarit example. It has been shown previously that only one type of prograded sand barrier from Thom et al. is morphologically analogous to the Carpentaria coast. Some stratigraphic comparisons are possible with this barrier type, the "bay barrier type 1a".

The New South Wales example was described stratigraphically in terms of seven facies (Thom et al. 1978, Figure 4). Although these beach-ridges have a marked facies change at their basal contact with nearshore shelly sands, no distinction is made between intertidal and sub-tidal deposits as in the Gulf of Carpentaria. A transgressive facies providing radiocarbon ages of 8000-9000 yrs B.P. and an estuarine clay is present lower in the eastern Australia section, facies which have been shown as absent in the Gulf of Carpentaria.

Sections across New South Wales barriers demonstrate the inherent lithologic and morphologic differences between the two prograded coasts. Sediment source and supply, nearshore and offshore morphology and wave regime are all factors which affect the lithology and distribution of sediments available for accretion. These factors are superimposed on a pre-Holocene topography on the New South Wales coast which has considerably steeper slopes (1:50-1:100) than the gentle pre-Holocene topography beneath Gulf sediments (1:1000-1:4000).

Regressive facies on the New South Wales coast include a sandy nearshore unit and its overlying beach-ridge sands with aeolian caps. These beach-ridge sands are comprised of terrigenous sand mixed with shell, but do not contain as much shell as the Gulf beach ridges. The New South Wales coast shows no evidence past or present of a mud influx equivalent to that in the Gulf of Carpentaria, an influx which is
essential to regression on the latter coast.

Cape Keer Weer example

Smart (1976) provides the remaining stratigraphic example useful in this discussion. However, the most disappointing aspect of Smart's findings at Cape Keer Weer, is the apparent dissimilarity between these and findings beneath Christmas Creek and Edward River transects. Smart employs four stratigraphic facies to describe the Cape Keer Weer sections. He has been able to lithologically sub-divide beach ridge sand facies into discrete episodes of accretion. His lithologic distinctions are confirmed by a limited radiocarbon age structure to be discussed in the next chapter.

In brief, Smart noted four episodes of Holocene ridge-building, each with an identifiable sub-facies separated by morphologic discontinuities. Beach-ridge sands are distributed across an alluvial surface with a gentle seaward gradient, typical of the Gulf pre-Holocene topography. Smart found the oldest Holocene ridge set resting on a clay facies, an arrangement which is superseded by the second subfacies which rests directly on pre-Holocene alluvium, resulting in a stratigraphy comparable to the barrier island model of Le Blanc (1972). The two younger ridge sets rest on clayey sand, somewhat like cheniers, but exhibiting near-continuous deposition during the latter part of the Holocene. The boundary between these-beach ridge sands and the underlying silts and clays is not as consistent or uniform as reported beneath transects at Christmas Creek or Edward River.

The lack of agreement between neighboring stratigraphic sections is best explained by the possible anomalous nature of Cape Keer Weer. The Cape is a prominent bulge on the coast approximately 60km south of the
present mouth of the Archer River. It exhibits an arcuate delta-like form which is not in agreement with present discharge of terrigenous sediment into the Gulf. The Archer, one of Cape York's major rivers empties into an estuary partly closed by a Holocene barrier far to the north. The Kendall River spills sediment into the Gulf just south of the Cape and it is possible that northward drift during a part of the year moves some of this input toward Cape Keer Weer. The Pleistocene and Holocene strandlines at the Cape exhibit a similar, but not conformable bulges to the seaward. Some of the Pleistocene ridges show a recurved pattern at breaks in their trend. these breaks carry only minor consequent streams at present.

The Holocene geology of Cape Keer Weer may be highly influenced by morphology inherited from Pleistocene or older topography. The possibility of an ancestral Archer distributing sediment across Cape Keer Weer and forming a large over-flattened fluvial plain to the southwest needs to be considered. In addition, existing structural and tectonic evidence does not indicate a basement monocline in this area, but a structural "high" similar to the Vrylia structure (Smart et al., 1978) could be present. This condition occurs further north and should not be completely dismissed until more sub-surface data is available. Whatever the explanation for the present morphology of Cape Keer Weer, it may not provide a appropriate generalized stratigraphy of coastal plains on the Gulf of Carpentaria.

Summary

Certain commonalities are apparent after review of the previous examples. Most relevant to the interpretation of Carpentaria stratigraphy is the finding by other workers that morphologic
discontinuities on the surface, unless accompanied by a distinctive lithologic contrast, may not be traceable below the surface. As a corollary to this, interpretative difficulties may be overcome by detailed characterization of the modern facies before stratigraphic work. For example, where a modern intertidal or subtidal facies exists as distinct from beach-ridge sediments, this distinction may be used to identify the facies boundaries in the stratigraphy. Finally, the observation that facies boundaries in the Gulf of Carpentaria, whether they are beach-intertidal or intertidal-subtidal, exist in a definable topographic framework with respect to sea level offers the promise of an upper Holocene sea level history. The existence of such boundaries in coastal stratigraphy containing radiocarbon datable material makes the study of the Gulf of Carpentaria sequence attractive.
INTRODUCTION

The Holocene chronology derived from radiocarbon dates of marine shells from both chenier and beach ridge plains is presented in stages:

a. review of rationale and problems associated with radiocarbon (C-14) dating of marine carbonates.
b. presentation of methods for grouping of dates when independent evidence of episodic events is present.
c. identification of the age structure of chenier and beach-ridge plains.
d. presentation of average rates of progradation and comparison with other such data.
e. identification of episodic events on the chenier and beach-ridge plains and some correlations between them.

RADIOCARBON DATING PROGRAM AND PROBLEMS.

General

Radiocarbon dating, applied to Holocene coastal sites since the late 1950's, is subject to a variety of problems which affect interpretation of the results. Given that the investigator correctly understands the relationships between dated material and stratigraphy of a particular site, C-14 results must be interpreted in the light of error sources. As reviewed by Mangerud (1972), Polach (1976) and Chappell et al. (1974) these 4 sources are as follows:

1. Statistical errors are associated with laboratory measurement of C-14, arising in counting of the sample itself, of the modern reference standard and of the counting machine "background." These errors are expressed in the standard error stated with every C-14 age estimate. Minimization of these errors is the responsibility of the C-14 laboratory. It must be noted here that the exact value of the C-14 half-life is disputed; however, ages are conventionally calculated by
the Libby half-life despite the fact that Polach (1976) suggested that the Libby half-life may be too low. Furthermore, calculations using the conventional Libby figure permits comparison with other determinations performed at other laboratories.

(2) Isotopic fractionation, during growth or deposition of carbon-bearing substances, can cause small differences in initial C-14 content from that to be expected in any given environment or type of sample. These differences are systematic. Correction is a routine task at the ANU Radiocarbon Lab and is based on mass-spectrometric measurement of the concentration of C-13 in each species of sample (Polach, 1976).

(3) Reservoir or environmental effect is the name given to the conditions where C-14 concentration varies between contemporaneous samples from different environments arising because of C-14 variations between different carbon "pools" or reservoirs. For example, ocean surface water contains only about 90% of the atmospheric C-14 level, causing modern shellfish to have an apparent C-14 age of approximately 400 years (Gillespie and Polach, 1980). Estimation of the correction factor due to reservoir effect is discussed below.

(4) Contamination of relict marine shells by younger carbon with a higher C-14 activity is a problem frequently encountered in dating of marine carbonates, especially those which may have been subaerially exposed or in contact with fresh water after deposition. The most common mode of contamination involves recrystallization of aragonitic carbonate into calcite with an attendant external carbon exchange (Chappell and Polach, 1972). This error source will be discussed more fully later.

Sources of error including those major ones summarized above have been discussed by Polach (1976). A summary of Polach's findings are reproduced in Table 5-1 and each error source is discussed in relation to its effect on the radiocarbon observations.
<table>
<thead>
<tr>
<th>Sources of error</th>
<th>Effect on Age Determination</th>
<th>How to Minimize</th>
</tr>
</thead>
<tbody>
<tr>
<td>1. Precision of Age Determination</td>
<td>Statistical, Typically 1% dC-14 or less</td>
<td>Big samples, optimal counting times</td>
</tr>
<tr>
<td>2. Inherent</td>
<td>&quot;Libby&quot; H/2 probably too low</td>
<td>Multiply ages BP x 1.030</td>
</tr>
<tr>
<td>(i) C-14 Half-life</td>
<td>410 years is common</td>
<td>Mass-spectrometry on portion of sample</td>
</tr>
<tr>
<td>(ii) C-13/C-12 Fractionation</td>
<td>550 years is possible</td>
<td>International cross-checking of secondary standards</td>
</tr>
<tr>
<td>(iii) C-14 Modern Std</td>
<td>&lt; 80 years</td>
<td>Tree-ring calibration; otherwise interpret in terms of radiometric time scale</td>
</tr>
<tr>
<td>(iv) Variation in past C-14 production rates</td>
<td>Nil to 800 years; beyond 10,000 B.P. not determined</td>
<td>Interpretation of results choice of contemporary standards</td>
</tr>
<tr>
<td>(v) Distribution of C-14 in nature</td>
<td>Surface ocean, latitudinal dependence -400 to -750 years. Deep ocean -1800 years</td>
<td>Interpretation of results</td>
</tr>
<tr>
<td>(vi) Changes of C-14 concentration in the atmosphere</td>
<td>Industrial effect ca. -2.5% and Atom Bomb effect, ca. +160% in atmosphere</td>
<td>Interpretation of results choice of material chemical/physical purification</td>
</tr>
<tr>
<td>3. Contamination</td>
<td>Nil to 3000 years up to 15,000 BP &gt;10,000 possibly beyond 25,000 BP</td>
<td>Choice of material and interpretation of results</td>
</tr>
<tr>
<td>4. Age of Material</td>
<td>&lt;10 years to &gt;1000 years</td>
<td>Interpretation of results</td>
</tr>
<tr>
<td>5. Association of Sample and Event</td>
<td>Indeterminate</td>
<td>Care in interpretation, interdisciplinary collaboration</td>
</tr>
<tr>
<td>6. Human</td>
<td>Indeterminate</td>
<td></td>
</tr>
<tr>
<td>7. Interpretation of Results</td>
<td>Indeterminate</td>
<td></td>
</tr>
</tbody>
</table>

(from Polach, 1975)
Estimation of reservoir or environmental effect.

That the carbon-14 activity of marine shells is different from terrestrial wood, was an observation made early in the development of radiocarbon dating techniques (Craig, 1954). Further work showed that this difference was also present in the C-14 activity of ocean surface waters (Craig, 1957). The term "apparent age" of marine shells developed from the work of several authors to include Berger et al. (1964) and Taylor and Berger (1967). Rafter et al. (1972) went on to demonstrate the variability of such apparent ages, especially in areas of restricted circulation or high latitudes, a phenomenon further expanded by the works of Mangerud (1972) and Mangerud and Gulliksen (1975). Currently the term "reservoir effect" is accepted in reference to the constant by which apparent radiocarbon ages of marine carbonates must be corrected to give a true radiocarbon age.

Conventional radiocarbon ages require a correction for the reservoir effect (Gillespie and Polach, 1976 and Thom et al., 1978). A correction factor of 410 years for New Zealand coastal waters was prescribed by Rafter et al., (1972). This was later refined by Gillespie and Polach (1980) to be 450 ± 35 years. However, the correction which is indicated for southern and eastern coasts of Australia may not be valid in the anomalously shallow, warm, well-mixed waters of the Gulf of Carpentaria. In order to establish the correction for this region, three "modern" shell samples were dated. Two of the samples, ANU 1828 (Anadara) and ANU 2092 (Telescopium) were collected "live" (from 15m of water) in 1903 offshore of Mapoon, Qld, approximately 80km north of Weipa (Figure 5-1). Although they are museum specimens (Australian Musium Nos. C14195 and C14280), notes kept by the collector, Hedley, indicate that the specimens were live at the time of collection. Supplementing these findings, is a
Figure 5-1. Locations are shown for sites where dating of Holocene geologic and palynologic histories has occurred in northeastern Australia.
determination (ANU 2099) on pelecypods (Volachlamys sp.) collected live from 25 meter depth offshore of Edward River in 1978. Determinations on these modern shells are presented in Table 5-2.

Table 5-2

<table>
<thead>
<tr>
<th>Sample Number</th>
<th>Type of Shell</th>
<th>C-14 Results</th>
</tr>
</thead>
<tbody>
<tr>
<td>ANU 1828</td>
<td>Anadara (1903)</td>
<td>590 ± 60 yrs B.P. 330 ± 60 % Modern</td>
</tr>
<tr>
<td>ANU 2092</td>
<td>Telescopium tegesopium (1903)</td>
<td>119.7 ±0.9%</td>
</tr>
<tr>
<td>ANU 2099</td>
<td>Volachlamys (1978)</td>
<td></td>
</tr>
</tbody>
</table>

Polach (pers. comm., 1978) applied methods of combining radiocarbon determinations as outlined in Polach (1969, 1972) and stated that ANU 1828 and 2092 combined to give an error weighted mean of 460 ±60 years. Polach also states that the live 1978 collection compares favorably with determinations on other "modern" live specimens presented in Gillispie and Polach (1976). Until additional data are obtained, it appears that the correction factor in use in eastern Australia can be applied to shells collected from the Gulf of Carpentaria. All ages reported in this work are corrected by the factor described. Furthermore, when comparisons are made with fraction-corrected dates reported by authors who have not adjusted for reservoir effect, such as Smart (1976), then a correction is applied to the other dates before comparison.

Contamination due to post-depositional changes.

One of the most common modes of contamination results from post-depositional recrystallization of marine carbonates. Chappell and Polach (1972) recognized two modes of aragonite recrystallization in corals and Tridacna shells. One of these modes involves subtle coarsening of the aragonite fibers from which the shell is built, without inversion to calcite. By C-14 measurement of shells older than 50,000
years B.P., Chappell and Polach were able to show that this mode of recrystallization proceeds in a virtually closed system, which excludes exchange with modern carbon. The other mode, by which aragonite is altered to sparry calcite, showed radiocarbon determinations which suggested recrystallization proceeded in an open system which allowed exchange with modern carbon. Therefore, it can be inferred that the presence of sparry calcite may affect the reliability of the radiocarbon date. Thom (1973), Bloom et al. (1974) and Chappell et al. (1974) further discuss this problem of contamination by recrystallization. Thom et al. (1978) indicate that the most reliable dates from the New South Wales coast have been obtained on shell hash from below the ground water table where exchange with modern carbon has been minimal. These authors advocate the use and reporting of the $^{18}O/^{16}O$ ratio as a validation of the radiocarbon results. According to Thom et al., the principle of the $^{18}O/^{16}O$ ratio is based on the fact that meteoric waters are more depleted in $^{18}O$ than ocean water. They suggest an equilibrated $^{18}O/^{16}O$ ratio equal to $-1.9 \pm 1.2\%$ w.r.t. PDB for marine shells on the east coast of Australia. Any marked negative departure from this value suggests ionic exchange during recrystallization and therefore increases the likelihood that contamination by modern carbon has occurred.

Given the possibility of major dating errors due to contamination by modern carbon in highly recrystallized shells, it was necessary to devote several age determinations to evaluating the level of possible error. Highly recrystallized shells may retain an aragonitic inner fraction which may be obtained by mechanical removal of the outer calcite fraction. Dating of these two fractions provided one estimate of possible range of errors. Down-section dating in drill holes through relict ridges on the beach-ridge plains provided another estimate, especially when vertical age structure of older ridges are compared with
very young ridges in which the shells appear unaltered. Finally, stable isotope ratios were determined on a number of samples and in combination with calcite/aragonite relationships used to indicate the reliability of the date. Discussion of these three tests of age reliability follows:

1. ANU lab samples 1740A and 1740B were collected from the most inland Holocene chenier ridge on the Karumba transect (Figure 5-1). The shells were recovered from a pit dug adjacent to drill hole 5217 (Figure 4-4) located on a "perched" chenier ridge. Sample 1740A, the inner fraction of Anadara sp. shells, comprised of 4% calcite (96% aragonite) yielded an age of 5540 ± 95 yrs B.P. Sample 1740C, the outer fraction, comprised of 24% calcite, (76% aragonite) provided an age of 5330 ± 95 yrs B.P.. Application of the Z statistic of Polach (1972) to these two means indicated they are not significantly different despite the higher level of calcite in the outer fraction (ANU 1740B).

2. The Christmas Creek transect (Chapter 4) is placed across a broad beach-ridge plain north of Edward River (Figure 5-1). The relict ridges exhibit increased carbonate recrystallization with increased age. Drill hole 5294 (Figure 4-19) penetrated a relict ridge approximately 1200 meters from the modern coastline. Granular, sparry calcite was encountered within 1.5m of the surface and graded to loose shell hash below the water table. Low-tide muds occur as well defined facies 1.0m thick beneath the beach-ridge sands. A radiocarbon determination on highly recrystallized granular carbonate (75% calcite) from the beach-ridge facies provided an age of 2680 ±75 yrs B.P. (ANU 1736). Shells (4% calcite) from the low-tide mud below the sand facies yielded an age of 2770 ±80 yrs B.P. (ANU 1737). Such results suggest that recrystallization may not be accompanied by contamination to the extent
that C-14 ages are significantly altered.

The results described above prompted further investigation of the age structure of individual ridges, particularly a younger ridge where recrystallization has not proceeded to any appreciable extent. Such a ridge was selected near the Edward River settlement (Figure 5-1) and cut open by backhoe excavation. It provided the internal structure details previously discussed and presented in Figure 3-48. One shell sample was selected from the ridge crest and another from below its base, in the low-tide muds. Both samples were bivalves of the Anadara sp. and when analyzed by X-ray diffraction showed 100% and 98% aragonite for the upper and lower samples respectively. The upper sample provided an age of 240 ± 85 yrs B.P. (ANU 1899) whilst the lower sample dated at 160 ± 80 yrs B.P. (ANU 1898), determinations which statistically are the same age. It appears therefore that dates from the relict sediments in drill hole 5294 are not adventitious and that the age of a ridge from base to crest is probably contemporaneous in terms of radiocarbon resolution. Furthermore, these data suggest that deposition of low-tide muds immediately beneath the base of a ridge may also be contemporaneous with the construction of the ridge.

3. Table 5-3 summarizes data associated with the 55 radiocarbon observations from which the Carpentaria age structure is developed. Although a replicate of most shell samples was subjected to XRD analysis to determine carbonate mineralogy, not all samples could be submitted for measurement of C-13/C-12 and O-18/O-16 ratios.

The use of stable carbon isotope measurements (C-13/C-12) in evaluating the reliability of a date, is accepted by Thom et al. (1978). However, they state that the value of stable oxygen isotope ratios
<table>
<thead>
<tr>
<th>ANU CODE</th>
<th>δ¹³C%</th>
<th>δ¹⁸O‰</th>
<th>CALCITE %</th>
<th>CONVENTIONAL AGE BP</th>
<th>ENVIRONMENT CORRECTED AGE BP</th>
<th>MATERIAL</th>
</tr>
</thead>
<tbody>
<tr>
<td>1690</td>
<td>-2.2</td>
<td>-2.0</td>
<td>40</td>
<td>6400 ± 90</td>
<td>5950 ± 95</td>
<td>shell hash</td>
</tr>
<tr>
<td>1691</td>
<td>0.0</td>
<td>(E)</td>
<td>60</td>
<td>5830 ±100</td>
<td>5380 ±105</td>
<td>shell-predom Mactra</td>
</tr>
<tr>
<td>1728</td>
<td>-0.2</td>
<td>(M)</td>
<td>38</td>
<td>1920 ±110</td>
<td>1470 ±125</td>
<td>shell hash</td>
</tr>
<tr>
<td>1729</td>
<td>-1.1</td>
<td>(M)</td>
<td>6</td>
<td>6160 ±180</td>
<td>5710 ±185</td>
<td>shell hash</td>
</tr>
<tr>
<td>1730</td>
<td>0.0</td>
<td>(E)</td>
<td>59</td>
<td>5590 ±250</td>
<td>5140 ±250</td>
<td>shell hash</td>
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<tr>
<td>1732</td>
<td>0.0</td>
<td>(E)</td>
<td>ND</td>
<td>5370 ±60</td>
<td>4920 ±70</td>
<td>shell hash</td>
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<td>1733</td>
<td>-0.4</td>
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<td>15</td>
<td>3570 ±120</td>
<td>3120 ±125</td>
<td>shell hash</td>
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<tr>
<td>1736</td>
<td>-2.9</td>
<td>(M)</td>
<td>100</td>
<td>3610 ±70</td>
<td>3160 ±80</td>
<td>shell hash</td>
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<tr>
<td>1735</td>
<td>0.0</td>
<td>(M)</td>
<td>75</td>
<td>3130 ±65</td>
<td>2660 ±75</td>
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<td>1736</td>
<td>0.0</td>
<td>(M)</td>
<td>4</td>
<td>3220 ±70</td>
<td>2770 ±80</td>
<td>shell hash</td>
</tr>
<tr>
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<td>NA</td>
<td>6450 ±100</td>
<td>NA</td>
<td>peat 34% organic</td>
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<tr>
<td>1739A</td>
<td>-4.1</td>
<td>(M)</td>
<td>95</td>
<td>6060 ±90</td>
<td>5610 ±95</td>
<td>shell Anadara and Turritella</td>
</tr>
<tr>
<td>1739B</td>
<td>-5.0</td>
<td>(E)</td>
<td>100</td>
<td>5530 ±90</td>
<td>NA</td>
<td>calcite matrix</td>
</tr>
<tr>
<td>1740A</td>
<td>-1.1</td>
<td>(M)</td>
<td>4</td>
<td>5990 ±90</td>
<td>5540 ±95</td>
<td>shell Anadara, inner fraction</td>
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<tr>
<td>1740C</td>
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<td>24</td>
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<td>5330 ±95</td>
<td>shell Anadara, outer fraction</td>
</tr>
<tr>
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<td>34</td>
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<td>3390 ±145</td>
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<td>630 ±70</td>
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<td>8</td>
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<td>3110 ±80</td>
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</tr>
<tr>
<td>1827</td>
<td>-0.7</td>
<td>(M)</td>
<td>10</td>
<td>2520 ±60</td>
<td>2070 ±70</td>
<td>shell Mactra</td>
</tr>
<tr>
<td>1869</td>
<td>0.0</td>
<td>(E)</td>
<td>100</td>
<td>34850 ±2000</td>
<td>NA</td>
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</tr>
<tr>
<td>1895</td>
<td>-5.0</td>
<td>(E)</td>
<td>ND</td>
<td>Background</td>
<td>NA</td>
<td>carbonate nodules</td>
</tr>
<tr>
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<td>0.0</td>
<td>(E)</td>
<td>17</td>
<td>5900 ±90</td>
<td>5450 ±95</td>
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<tr>
<td>1897</td>
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<td>2</td>
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</tr>
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<td>(E)</td>
<td>2</td>
<td>610 ±70</td>
<td>160 ±80</td>
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</tr>
<tr>
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<td>(E)</td>
<td>0</td>
<td>650 ±80</td>
<td>240 ±85</td>
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</tr>
<tr>
<td>1900</td>
<td>0.0</td>
<td>(E)</td>
<td>3</td>
<td>2290 ±70</td>
<td>1840 ±80</td>
<td>shell Anadara and Mactra</td>
</tr>
<tr>
<td>1920</td>
<td>0.0</td>
<td>(E)</td>
<td>4</td>
<td>6030 ±90</td>
<td>5580 ±95</td>
<td>shell Anadara</td>
</tr>
<tr>
<td>1921</td>
<td>0.0</td>
<td>(E)</td>
<td>2</td>
<td>2250 ±80</td>
<td>1800 ±85</td>
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</tr>
<tr>
<td>1922</td>
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<td>(E)</td>
<td>2</td>
<td>4430 ±90</td>
<td>3980 ±95</td>
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</tr>
<tr>
<td>1923</td>
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<td>(E)</td>
<td>4</td>
<td>1150 ±70</td>
<td>700 ±80</td>
<td>shell Anadara and Mactra</td>
</tr>
<tr>
<td>1924</td>
<td>0.0</td>
<td>(E)</td>
<td>2</td>
<td>530 ±65</td>
<td>80 ±75</td>
<td>shell Mactra</td>
</tr>
<tr>
<td>1925</td>
<td>0.0</td>
<td>(E)</td>
<td>4</td>
<td>4660 ±90</td>
<td>4210 ±95</td>
<td>shell Anadara and Mactra</td>
</tr>
<tr>
<td>1926</td>
<td>0.0</td>
<td>(E)</td>
<td>4</td>
<td>3780 ±90</td>
<td>3330 ±95</td>
<td>shell Anadara and Anadara</td>
</tr>
<tr>
<td>1927</td>
<td>0.0</td>
<td>(E)</td>
<td>4</td>
<td>1770 ±70</td>
<td>1320 ±80</td>
<td>shell Anadara</td>
</tr>
<tr>
<td>1928</td>
<td>0.0</td>
<td>(E)</td>
<td>3</td>
<td>2440 ±65</td>
<td>1950 ±75</td>
<td>shell Mactra</td>
</tr>
<tr>
<td>1929</td>
<td>0.0</td>
<td>(E)</td>
<td>0</td>
<td>2200 ±85</td>
<td>1750 ±90</td>
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<tr>
<td>1997</td>
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<td>0</td>
<td>680 ±70</td>
<td>230 ±80</td>
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</tr>
<tr>
<td>1998</td>
<td>0.0</td>
<td>(E)</td>
<td>15</td>
<td>550 ±80</td>
<td>100 ±85</td>
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<tr>
<td>1999</td>
<td>0.0</td>
<td>(E)</td>
<td>96</td>
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<td>50 ±80</td>
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<tr>
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<td>(E)</td>
<td>74</td>
<td>5540 ±160</td>
<td>5090 ±145</td>
<td>shell Mactra and Anadara</td>
</tr>
<tr>
<td>2037</td>
<td>0.0</td>
<td>(E)</td>
<td>3</td>
<td>1240 ±70</td>
<td>790 ±80</td>
<td>shell Anadara</td>
</tr>
<tr>
<td>2038</td>
<td>0.0</td>
<td>(E)</td>
<td>3</td>
<td>2900 ±80</td>
<td>2050 ±85</td>
<td>shell Anadara</td>
</tr>
<tr>
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<td>0.0</td>
<td>(E)</td>
<td>100</td>
<td>3750 ±80</td>
<td>3300 ±85</td>
<td>shell hash</td>
</tr>
<tr>
<td>2060</td>
<td>0.0</td>
<td>(E)</td>
<td>100</td>
<td>3170 ±110</td>
<td>2720 ±115</td>
<td>shell hash</td>
</tr>
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<td>(E)</td>
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<td>2210 ±70</td>
<td>1760 ±80</td>
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<td>3</td>
<td>4310 ±80</td>
<td>3860 ±90</td>
<td>shell Mactra</td>
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<tr>
<td>2093</td>
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<td>(E)</td>
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<td>1670 ±90</td>
<td>NA</td>
<td>peat 7% organic</td>
</tr>
<tr>
<td>2100</td>
<td>0.0</td>
<td>(E)</td>
<td>ND</td>
<td>6000 ±100</td>
<td>5550 ±105</td>
<td>shell Anadara and Mactra</td>
</tr>
<tr>
<td>2101</td>
<td>0.0</td>
<td>(E)</td>
<td>ND</td>
<td>5760 ±110</td>
<td>5310 ±115</td>
<td>shell Anadara and Mactra</td>
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<tr>
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<td>(E)</td>
<td>ND</td>
<td>3430 ±100</td>
<td>298 ±105</td>
<td>shell hash</td>
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<tr>
<td>2103</td>
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<td>(E)</td>
<td>ND</td>
<td>1880 ±90</td>
<td>1430 ±95</td>
<td>shell Anadara</td>
</tr>
</tbody>
</table>

1) δ¹³C/δ¹²C and δ¹⁸O/δ¹⁶O ratio depletion (-) or enrichment (+) w.r.t. PDB: M = Measured ±0.2‰; E = Estimated ±2.0‰
2) XRD Determinations of recrystallisation from aragonite, except ANU 1897, soil carb. nodule which were 2% calcite and 98% dolomite. ND = Not determined; NA = Not Applicable
3) Conventional radiocarbon age BP, based on δ¹³C corrected δ¹⁴C depletion w.r.t. 0.95 NBS Ox, and 5568 ± half-life; † = Not acceptable on stratigraphic evidence.
4) Conventional age BP after substraction of oceanic reservoir effect constant of 450 ±35 y. The Environment corrected age BP must be used when relating shell ages to peat or wood ages.
(0-18/0-16) are less widely appreciated, reporting these ratios for the first time in conjunction with a date list. The rationale for the determination of oxygen ratios was described briefly in a previous section, and fuller explanation of such measurements is undertaken by Thom et al. (1978).

Inspection of Table 5-3 shows that a number of carbonate samples have been subjected to stable oxygen and carbon isotope measurement. From this data, a graph such as that shown in Figure 5-2 may be constructed which shows 0-18/0-16 and C-13/C-12 ratios compared to the degree of calcite recrystallization. Figure 5-2 indicates that 3 samples (ANU 1734, 1735 and 1739A) exhibit abnormally negative 0-18/0-16 values. Similarly these samples show depressed C-13/C-12 ratios accompanied by high percentage of calcite replacement of original aragonite.

There is a subtle trend shown by Figure 5-2 to suggest that samples containing a high percentage of calcite which also show unusually negative values for stable carbon and oxygen isotope ratios may have been contaminated by younger carbon during recrystallization. However, not all samples with a high calcite content are so affected, for example ANU 1736, has stable isotope ratios which suggest that calcite recrystallization proceeded without contamination. Such contamination, positively identified should invalidate a C-14 determination, or at least indicate a particular age as minimum for that observation. Given the apparent correlation from Figure 5-2 which shows very high calcite levels (95-100%) as correlative with depressed stable isotope ratios, it is likely that ages on some other samples with calcite levels in the 95-100% range on which stable isotope measurements were not obtained, should be identified as questionable. These samples include ANU 1739B, 1869, 1999, 2059 and 2060 in addition to the three samples graphed in Figure 5-2.
Figure 5-2. A scatter plot of carbon and oxygen depletion ratios compared to percent calcite in the sample suggests that high levels of calcite are related to unusually negative stable isotope ratios.
GROUPING OF RADIOCARBON OBSERVATIONS

Before the presentation of the age structure for the Carpentaria coastal plain in either plan or section, further statistical assessment is necessary. Since the acceptance of radiocarbon dating for determining the age of late Quaternary events, it has been common to combine, group or "cluster" radiocarbon ages when they appear to date the same event. When the "event" length is shorter than the resolution permitted by the laboratory counting error (1 standard error) it is common to test the significance of grouped ages by some sort of a statistic, such as the Z statistic presented by Polach (1972) and combine these determinations into an "error weighted" mean (Polach, 1969). However, when the "event" length exceeds the time in radiocarbon years associated with the laboratory counting error, two other methods, both graphical, have been employed. A bar graph display where multiple dates are grouped according to their respective ages and graphed through a range of 1 standard error about their mean (some workers insists on 2 standard errors) permits what is loosely called the "eye-balling" technique. If the determinations "look" like they fall into a group, then they are grouped accordingly. Alternatively, a rank-order display, such as that in Thom et al. (1978, Figure 3) also permits visual grouping of determinations. These graphical approaches are subjective. It is better to use numerical techniques which provide (1) a statistical confidence level of separation between groups, and (2) a group mean and standard deviation based on the assumption that all determinations in the group are taken from a normal distribution around some unknown mean.

(1974). These techniques were originally intended for application to events where sampling variance is not present and the "event" is brief in terms of radiocarbon resolution. The application of such techniques is sufficiently critical to the discussion of Carpentaria coastal deposition to warrant summary of the numerical methods.

When sampling error is present, one can take the true age of a dated sample to be related to the observed value by:

\[ a = \theta + r + e + f + g \]

where:  
- \( a \) = reported C-14 age of sample  
- \( \theta \) = true radiocarbon age  
- \( r \) = error associated with sampling  
- \( e \) = laboratory counting error  
- \( f \) = calibration error  
- \( g \) = sunspot error where suspected

It must be assumed that any reported value \( a \) is normally distributed about some unknown mean \( \theta \) with a variance \( \sigma^2 + E^2 + F^2 + G^2 \) due to the errors shown above. It is also assumed that the sampling error \( r \) is normally distributed around the true radiocarbon age \( \theta \) with variance \( \sigma^2 \) and that all other errors are independent of the sampling error and of each other where:

\[ e_i \sim N(0, E_i^2), f \sim N(0, F_i^2), \text{ and } g_i \sim N(0, G_i^2) \]

The null hypothesis is therefore:

\[ H_0 : \theta_1 = \ldots = \theta_n \]

The alternative hypothesis is that:

the group of observations belong to two groups with unknown means \( \theta_1, \theta_2 \) and sampling variances \( \sigma_1^2, \sigma_2^2 \) respectively. Since the sampling variances \( \sigma_1^2 \) and \( \sigma_2^2 \) cannot be assumed to be equal because sampling variability
cannot be considered constant and taking into consideration the additional complexities due to counting error, calibration error and sunspot error, the determinations of a test statistic is not as direct as in a simple analysis of variance (ANOVA).

The following simultaneous equations allow us to determine the Maximum Likelihood estimates \( (\hat{\theta}, \hat{\sigma}^2) \) of \( \theta \) and \( \sigma^2 \), the unknown group mean and variance.

\[
\hat{\theta} = \left( \frac{\sum A_i}{\mid \sigma^2 \mid + S_i^2} \right) \left( \frac{\sum 1}{\mid \sigma^2 \mid + S_i^2} \right)
\]

\[
\sum_{i=1}^{n} \frac{1}{\mid \sigma^2 \mid + S_i^2} = \sum_{i=1}^{n} \frac{(A_i - \hat{\theta})^2}{\mid \sigma^2 \mid + S_i^2}^2
\]

these equations are solved iteratively to produce the estimates \( \hat{\theta} \) and \( \hat{\sigma}^2 \).

To deal with the conditions outlined in the alternative hypothesis, Wilson (1978, personal communication) recommends the following method:

First find \( \hat{\theta} \) and \( \hat{\sigma}^2 \) (Maximum Likelihood estimates of the entire group). Consider then the ordered observations.

\( (A_1, S_1^2), \ldots, (A_n, S_n^2) \)

where: \( A_i \) represents the \( i \)th observation

and: \( S_i^2 = E_i^2 + F_i^2 + G_i^2 \)

Next a search is made for the maximum of twice the logarithm of the likelihood ratio \( \Lambda \), at some point \( k \) which splits the ordered values into two groups with different means and sampling variances. Determination of this maximum \( \Lambda \) value requires a test split to be performed all \( k \) \( (2 \rightarrow n-1) \).
According to Wilson, the \( A \) value at any split \( (k), (k+1) \) is:

\[
A = \sum_{1}^{n} \left( A(i) - \hat{\theta} \right)^{2} / \left( S(i)^{2} + \hat{\sigma}^{2} \right) - \sum_{1}^{k} \left( A(i) - \hat{\theta}_{1} \right)^{2} / \left( S(i)^{2} + \sigma_{1}^{2} \right)
\]

\[
\sum_{k+1}^{n} \left( A(i) - \hat{\theta}_{2} \right)^{2} / \left( S(i)^{2} + \hat{\sigma}_{2}^{2} \right) + \sum_{1}^{n} \ln \left( S(i)^{2} + \hat{\sigma}_{2}^{2} \right)
\]

where: \((\hat{\theta}, \hat{\sigma}^{2})\) = Maximum Likelihood estimates of mean and sampling variance for all observations

\((\hat{\theta}_1, \hat{\sigma}_1^2)\) = Maximum Likelihood estimates of mean and sampling variance for observations \((1 \rightarrow k)\)

\((\hat{\theta}_2, \hat{\sigma}_2^2)\) = Maximum Likelihood estimates of mean and sampling variance for observations \((k+1 \rightarrow n)\)

The above expression for the test value \( A \) is an extension of Wilson and Ward (in press, Equation 8) in which differences in sampling variability within the test groups is considered. The assumption is made that \( r \) (sampling error) is normally distributed over the entire group being tested i.e., under the null hypothesis:

\[
\begin{align*}
A(1), \ldots, A(n) \\
\leftarrow r \rightarrow \\
N(\hat{\theta}, \hat{\sigma}^2 + E_j^2 + F_j^2) \quad j=1, \ldots, n
\end{align*}
\]
whereas for any arbitrary split:

\[
A(1), \ldots, A(k), A(k+1), \ldots, A(n)
\]

\[
N(\hat{\theta}_1, \sigma_1^2 + E_{j_1}^2 + F_{j_1}^2)
\]

\[
N(\hat{\theta}_2, \sigma_2^2 + E_{j_2}^2 + F_{j_2}^2)
\]

\[
j_1 = 1, \ldots, k
\]

\[
j_2 = k+1, \ldots, n
\]

and taking $G_i^2 = 0$

Once the test ratio $A$ is maximized at some value $k(A_{\max})$, this site is identified as the best split within the group under consideration. It remains for the investigator to decide the validity of the null hypothesis. S.R. Wilson (personal communication, 1978) advised that values for the test ratio at the 95\% confidence level are approximately as shown in Table 5-4.

| TABLE 5-4 |
|---|---|---|---|
| (observations) | 4 | 10 | 20 | 40 |
| (test value) | $A = 18$ | 24 | 33 | 55 |

Once any split is performed on a group, the process can be repeated on the subgroups until splitting proceeds into subgroups which are no longer statistically significant.

The result is a hierarchal arrangement of groups of radiocarbon determinations. This hierarchal splitting frequently confirms that which is visually apparent from graphical presentations. However, the computation of a group mean with an associated sampling standard deviation facilitates the comparison of grouped ages in one locality with grouped ages in another area. The estimate of the sampling standard deviation is actually an underestimation of the true sampling standard
deviation value because it is calculated after the variability due to other sources is taken in account. However, in the case of Carpentaria date, this error due to other sources is minimized by omitting $F^2$ and $G^2$ from the calculations. Consideration of variance due to calibration error between radiocarbon and sideral years ($F^2$) is unnecessary because that conversion is not made. The variance due to sunspot error ($G^2$) is considered as negligible for the Carpentaria study.

AGE STRUCTURES

The age structures of Carpentaria coastal sediments are based on ages of shells deposited in beach or chenier ridges and fine-grained muds, mainly the subtidal mud facies. Unless otherwise stated, it must be assumed that the age of the shell indicates the age of the depositional facies. Although, it is apparent from discussions in Chapter 3, that shell material is reworked on both the chenier and beach-ridge plains, the dating of whole shells whenever possible probably reduces the likelihood of error in relating the date to the depositional event.

The presentation of age structures here follows the form of discussion in Chapter 4. First ages are presented in map plan, and second in stratigraphic sections. All ages reported on the following figures are corrected for reservoir or environmental effect.

Chenier plain ages.

The age structure of the chenier plain between Karumba and Disaster Inlet is shown in plan by Figure 5-3, whilst Figures 5-4 and 5-5 show age determinations for stratigraphic sections of the Karumba and Pandanus Yard transects. Cheniers are easily dated because of their high shell
Figure 5-3. The chenier plain between Karumba and Disaster Inlet is dated with 24 radiocarbon determinations. Ages shown here are corrected for reservoir effect (environmental correction) and are presented with standard errors and laboratory numbers.
It is noted with CARPENTARIA

UNDIFFERENTIATED COASTAL AND FLUVIAL SEDIMENTS (HOLOCENE AND PLEISTOCENE)

CHENER RIDGES (HOLOCENE)

BEACH RIDGE (PLEISTOCENE)

UPLAND SURFACE (TERTIARY AND OLDER)

AGES SHOWN IN RADIOCARBON YEARS BP (ENVIRONMENTALLY CORRECTED)

NUMBER IN PARENTHESES INDICATES ANU LAB NO
content; in contrast, the paucity of shell in fine-grained facies such as high-tide muds, low-tide muds and subtidal muds frustrates precise age determination of these units. Four major outcomes of chenier plain dating are made obvious in Figure 5-3.

a. In most instances, samples from different sites along the axis of any given chenier ridge yield concordant dates.

b. "Anomalous" radiocarbon observations contradicted by adjacent dates are absent from the chenier plain.

c. Strandline deposition appears to have been initiated at or near 5500-5800 yrs B.P. with the oldest shoreline extending in a smooth accurate trace across the more irregular pre-Holocene landforms.

d. Ages appear to be distributed in groups across the chenier plains with similar distributions shown on more than one transect. (Statistical evaluation of such grouping will be performed later in this chapter).

Ages obtained from sections are shown in Figure 5-4 and 5-5. Only the Karumba transect provides information additional to that already shown in Figure 5-3. Besides providing important age-topographic data relevant to the sea level discussion of Chapter 6, four major observations may be gained from the age structure of the Karumba transect.

a. Time lines (see Figure 5-4), based on ages such as ANU 1746, 1743 and 1744 suggest an isochron which has a slope of approximately 1:1200. This gradient is of similar magnitude to modern offshore gradients which approximate 1:4000. Comparison of the age of ANU 1743 with the overlying determination, ANU 1928, suggests that on the chenier plain, deposition of a chenier ridge may not be contemporaneous with the underlying mud facies.
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Shelly chenier sands

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High-tide muds

0

Low-tidemuds

lsml Subtidal muds

-3.0
~BACKGROUND

[!] Transgressive sands

(1895)

-4.0

Age Structure
KARUMBA TRANSECT
Ages shown in radiocarbon years B.P.
Number in parentheses indicates A,NU Lab No.

0
0

Carpentaria paleosol
Pleistocene beach-ridge sands

illustrates
Figure 5-4. The age structure the Karumba
transect
age
for the
flat-lying
time
lines,
a distinctive pre-Holocene
Carpentaria paleosol and vertical accretion rates for the Holocene
sediments of 1-2mm per year.
N

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Figure 5-5. Limited dating of the Pandanus Yard transect shows the seaward chenier to be very young and indicates that the mudflat to the landward has prograded during the last 2000 years.
b. The Carpentaria paleosol which underlies the Holocene sequence contains abundant carbonate nodules; these yield minimum pre-Holocene ages such as ANU 1895 (background) which indicate this buried soil developed before the onset of the Postglacial Marine Transgression.

c. Downhole dating of the sections shows vertical accretion to be slow in contrast to rapid horizontal progradation; this is illustrated by C-14 dates ANU 1928, 1743 and 1744 which indicate vertical accretion rates in the 1-2mm/year range.

Beach-ridge plain ages.

Radiocarbon observations from the beach-ridge plain may be presented in a similar manner. Figures 5-6 and 5-7 presents the details of C-14 distributions in plan for Christmas Creek and Edward River areas. The following summarizes the date patterns.

a. Beach-ridge plains apparently prograded rapidly following the rise of sea level to its postglacial maximum approximately 6000 yrs ago, producing a large number of ridges during the period 5000-6000 B.P. The younger boundary of this episode of rapid progradation will be discussed in the context of more rigorous statistical evaluations later.

b. One anomalous result (ANU 2060) appears on the beach-ridge plain. The unreliable nature of both 2060 and 2059 has already been highlighted in the earlier discussions of calcite recrystallization and stable isotope ratios.

c. The time gap across obvious morphologic discontinuities such as the shelly/ non-shelly boundary on the Christmas Creek transect and the obliquely intersecting beach-ridges on the Edward River transect
Figure 5-6. A broken line indicates the shelly/non-shelly discontinuity of the Christmas Creek transect. Age differences across the boundary are not as great as suggested by the different lithologies.
Figure 5-7. On the Edward River transect, a well defined morphologic boundary separates mid-Holocene ages from upper Holocene ages. High calcite levels in samples used for date numbers 2059 and 2060 require these ages to be regarded as unreliable. Note that ANU 1847 is a determination on soil carbonate nodules from supratidal flats east of the Holocene beach ridge complex.
is not as great as expected. Either the dates (ANU 1734, 2059, 2060) which lie to the landward of the discontinuities are unreliable as already suggested, or, it is possible that these discontinuities are lithologic and morphologic rather than chronologic. There are insufficient data available to resolve a choice between these options.

Figures 5-8 and 5-9 presents the distribution of C-14 ages along the two beach-ridge sections. In addition to providing age and topographic information useful to the discussion of sea level, the following 3 points are important.

a. It is difficult to characterize time lines below the strandline because the lack of datable material on the lower facies. However, the observation that ridges are contemporaneous from their crests to below their bases on the low-tide muds has been presented earlier in this chapter. This suggests that the age structure of a beach-ridge plain more accurately dates regressive events than the age structure of a chenier plain where ages are determined only on chenier ridges.

b. Downhole dating of ridges suggests that the vertical age structure of ridges is homogeneous from base to crest and that ridge building may occur in a single or closely-spaced events. As shown in an earlier section, the findings from drillholes 5258 and 5294 support this view.

c. Rates of progradation have varied since the initiation of ridge-building in the mid-Holocene. The highest rates were observed during the initial episode following 6000 yrs B.P. (see later), and the lowest rates occurred during the formation of the younger portions of the plains. This change could be due to either the increased water depth into which the coast prograded or lower rates of sediment supply.
Figure 5-8. The age structure of the Christmas Creek transect shows good agreement in downhole dating such as drillhole 5294.
Figure 5-9. Edward River stratigraphy shows a young phase of beach-ridge progradation from such dates as ANU 1898 and 1899. Both of these dates are projected onto the section from a nearby excavation into the same beach ridge.
RATES OF PROGRADATION—SOME COMPARISONS

The accuracy of progradation rate estimates is affected by the radiometric dating errors, and more importantly, by the fact that ridge-building is probably associated with groups of storms perhaps involving a temporary transgression of the highwater shoreline by an unknown amount, especially in the case of the chenier plains. Consequently, progradation rates based on chenier or beach-ridge ages can only be expressed as averages over intervals during which several ridges are built. Rates of volumetric addition, to be accurate, need to be defined by the amount of material between isochrons on a crosssection. (see Figure 5-4). Since isochrons in section are poorly defined in the Gulf, and cannot be extended offshore without further field data, volumetric estimates are not presented. Furthermore, the age of a chenier may not indicate the age of the mudflat on which it rests (see Figure 5-4). Therefore, the average rates for the Gulf of Carpentaria presented in Table 5-5 are considered first approximations. Calculated rates for the beach-ridge plain are probably more accurate, but no adjustment can be made for diastems or other periods of non-deposition and erosion. Therefore such estimates must be considered minimum values as they are for other areas (see Table 5-5).

Comparison with eastern Australian examples is favorable except for the initial period of progradation noted at Moruya, New South Wales (Table 5-5). Thom (pers. comm., 1979) notes that some anomalously high rates during the mid-Holocene are to be expected as the transgression slowed and wave action attempted to restore a nearshore equilibrium profile by moving relict material landward. This also appears to be the case of the Carpentaria beach-ridge plain where, despite the absence of a basal transgressive sand, it is obvious that rapid progradation occurred
<table>
<thead>
<tr>
<th>Locality</th>
<th>distance between points (meters)</th>
<th>age range (C-14 years)</th>
<th>duration</th>
<th>progradation rate m/yr</th>
</tr>
</thead>
<tbody>
<tr>
<td>Karumba</td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>total</td>
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<td></td>
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<td>3800-2000</td>
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</tr>
<tr>
<td>Pandanus Yax-1</td>
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<td></td>
<td></td>
</tr>
<tr>
<td>total</td>
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<td>5400-0</td>
<td>5400</td>
<td>0.80</td>
</tr>
<tr>
<td></td>
<td>600</td>
<td>5400-2000</td>
<td>3400</td>
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at the maximum of the transgression, probably due to an abundance of relict material in the nearshore zone.

Comparisons with the chenier plain are difficult because formation of cheniers does not imply progradation. Conversely, the outbuilding of mudflats appears to result from increases in fluvially supplied terrigenous material rather than an oversupply of elastic material in transgressed sediments, a process which will be explained further in Chapter 7. There is, however, a slight increase in the sand and gravel content of oldest cheniers, but this may be partly due to a relative loss of shell material by leaching.

Comparison with rates of progradation from areas outside Australia shows that only the chenier plain at Karumba exhibits rates as high as the Nayarit, Scheveningen and Bay of Plenty examples (Table 5-5). However, the seemingly lower rates for the Gulf of Carpentaria, may be misleading in volumetric terms. Although offshore sediment geometry is unknown, it is apparent from the limited offshore lithologic data presented in Chapter 3 that deposition of fine-grained terrigenous material may occur 10-20km from the modern shoreline. It is likely, that coincident with shoreline progradation, there is widespread nearshore-offshore accretion which maintains the overflattened offshore profile during seaward progradation. This is in contrast to nearshore conditions such as those found in eastern Australia where steep nearshore gradients cause similarly steep time-lines in the stratigraphy with the shoreline prograding into increasing depths since the mid-Holocene.

EPISODES OF PROGRADATION AND CORRELATIONS

The episodes

Perhaps more important than average rates of progradation, is the
Figures 5-10a and b. Using numerical strategies outlined earlier, C-14 observations from the chenier and beach-ridge plains may be divided into discrete groups.
study of the occurrence of episodic progradation. The large number of dates used to support the age structure presented in this Chapter permits numerical analysis of the tendency for the dates to fall into discrete groups or episodes of strandline formation. This section evaluates this tendency toward grouping by the numerical strategy outlined previously. Although fine detail is lost by such a technique, it is necessary when comparisons are being made between the evolution of Carpentaria coastal plains and other episodes of climatic and/or sea level variations claimed for other parts of the world. The development of one chenier or beach-ridge is too adventitious and there are too many of such events, rendering subjective comparison invalid.

In the case of the chenier plain, all strandline dates between Karumba and Disaster Inlet are pooled into one list on the premise that a regional response is suggested from qualitative inspection of the dates. A similar justification is offered for the pooling of all beach-ridge dates from Edward River and Christmas Creek areas.

Grouping of radiocarbon determinations from the chenier plain are presented in Figure 5-10A. Four distinct episodes (stages) of chenier formation with no overlap at two standard deviations are indicated. Stage CR1 (chenier ridge) began at or near the time when sea level reached its postglacial maximum. This brief period of strandline formation continued for approximately 350 years until a stage of mudflat development, CM1 (chenier mudflat), approximately 750 years long separated it from Stage CR2. Stage CR2 was a longer event of approximately 1600 years duration, centered about 3700 B.P. Another hiatus in strandline formation separates this period from Stage CR3 a brief 350 years period centered about 1900 B.P. Finally, there is evidence for a young or modern Stage CR4 which may have begun at 1200
B.P., and is presently active. There is evidence that some cheniers are in the process of being deposited by modern processes (see Chapters 3 and 4).

Strandline development on the beach-ridge plain also followed an episodic history. Figure 5-10B shows the numerical evaluation of radiocarbon dates extracted from the beach ridge plain. Strandline development proceeded for approximately 1200 years (Stage BR1) after the postglacial maximum. No beach ridges appear to have formed in the period 4700 B.P. to 3500 B.P. Stage BR2 was of moderate duration (1000 years), producing beach ridges from 3500 to 2500 B.P. A brief break in deposition separates this episode from Stage BR3 centered around 1400 B.P. Finally a young or modern period of deposition appears to have been active since 400 B.P.(BR4).

Correlations

At first inspection it appears that the four respective depositional episodes form the chenier and beach-ridge plains are distinctly "out of phase". Strandline deposition on the two types of coast appear to be synchronous only during the period following the termination of the Postglacial Marine Transgression. Even in this case, Stage CR1 is considerably shorter than Stage BR1.

It was established in Chapter 1 that traditional models of deposition for chenier ridge formation require a relative decrease in sediment supply during the actual time of ridge development. This is especially true for those researchers who have suggested that cheniers are built from coarse material, either sand or shell, which is sorted out of the fine-grained nearshore facies or in some cases a marsh facies. Therefore, chenier ridge formation indicates either vertical accretion or
erosion instead of progradation. Conversely, the formation of a ridge in a beach-ridge plain does imply progradation, because it is by the addition of ridges rather than mudflats that this type of coast progrades.

Figure 5-11 suggests the relevant relationship between chenier plain and beach-ridge plain age structures. In this figure, total width of coastal plain progradations is shown with respect to radiocarbon years B.P. From this form of presentation, it can be suggested that progradation of the chenier plain occurs by deposition of mudflats at some time between the times of maximum ridge formation. These periods of mudflat progradation might be expected to be approximately synchronous with beach-ridge development if a regional forcing function was responsible for increases in sediment supply. The lack of total agreement for episodic progradation (Figure 5-11) may be due to two possible shortcomings in data analysis.

1. The episodes of mudflat progradation may be incorrectly inferred because they are defined indirectly by demonstrating when cheniers formed. This requires the assumption that most mudflats prograded only when cheniers were not developing, an assumption which may not be entirely correct. (See further discussion in Chapter 7.)

2. The small numbers of dates which define each episode of chenier and beach-ridge formation (Figure 5-11) may not fully describe the normal distribution which is assumed for the age of the depositional events. It is possible that larger selections of strandline ages from both plains might refine the chronologic alignment of the two areas.
Figure 5-11. Chenier ridge and beach-ridge age structures do not appear synchronous because they are formed by very different conditions of sediment supply. However, it is more likely to expect a synchronous relationship between mudflat progradation on the chenier plain and beach-ridge progradation on the beach-ridge plain.
CHAPTER 6
RELATIVE SEA LEVEL MID- Holocene TO THE PRESENT

In this chapter, morphologic and stratigraphic data presented in previous chapters are reviewed and used to develop Holocene relative sea level curves for the chenier and beach-ridge plains. These progradational sequences extend back to about 6000 yrs B.P. (Chapter 5) and evidence for earlier relative sea level changes has not been uncovered in this study.

SEA LEVEL INDICATORS

Various indicators have been used in other parts of the world to study paleo-sea levels. Many of these indicators are found in chenier and beach-ridge plain environments, although one of the most frequently utilized, the boundary between marine and non-marine peats, is unavailable in an environment such as the Gulf of Carpentaria. Ridge crest heights, coral platform elevations, wave-cut scarps, relict mud or supra-tidal flats have been used with varying degrees of success on all types of coasts and with the exception of coral platforms, all are available in the Gulf of Carpentaria.

A sea level indicator must bear a definable relationship to mean sea level. This relationship may be a direct one such as exhibited by some attached organisms which live within a narrowly defined range within the tidal curve. Alternatively, the relationship may be an indirect one in which the indicator is related to some tidal inundation level, such as wave peak level. The reliability of all sea level indicators, especially those of the latter category is variable.

Strandline crests have been one of the most frequently used indicators of sea level change during the upper Holocene. Johnson (1919) stated with considerable assurance that analysis of beach ridges such as those of Dungeness would yield a sea level history. Davies (1961)
postulated that Tasmanian beach-ridge elevations probably indicated a sea level trend during the upper Holocene. Height of storm ridges in conjunction with evidence of high spring tide benches and "perched" tidal-stream flats were used by Schofield (1960) to indicate Holocene sea level fluctuations in the Firth of Thames, New Zealand. However, all studies which use the height of a feature, constructed by infrequent high energy events, to reconstruct a sea level datum are using ambiguous primary data.

Where it can shown that high energy events are actually constructive, such as areas where beach face stratification extends to the crest the highest ridges, then these events will also vary in energy levels. If a frequency-magnitude relationship such as that shown in Figure 6-1A is considered, it is apparent that events of relatively higher energy occur less frequently, but may be as important to the constructional record as the more common moderate energy events. The six beach ridges shown in Figure 6-1B might be products of the six events plotted in Figure 6-1A. Each of these events has a different energy level which contributes to the construction of a beach ridge of some particular height.

Furthermore, height of ridge construction is influenced by factors other than the explicit correlation with wave energy level which is outlined above. Sediment availability in the nearshore zone and the amount of progradation by moderate energy events since the previous high energy event affect not only ridge height, but ridge spacing. Excavations in Carpentaria beach ridges show not only considerable variability in lithology but evidence that compound ridge crests and closely spaced berms may form when constructional high energy events closely follow one another. Ridge height and form bears a multi-variable relationship with several nearshore environmental parameters of
Figure 6-1. Application of the frequency-magnitude concept to beach-ridge construction indicates that ridge crest height is directly related to magnitude of the ridge-building event.
which relative sea level is probably less diagnostic than the factors mentioned above.

A more cautious approach to the value of ridge crest elevations is proposed by Tanner and Stapor (1972) who show that ridge height and the slope angle of the seaward face of the beach ridge are closely related to wave height at the time of ridge building.

Tanner and Stapor show that this relationship takes the form of:

\[
HW = \frac{HR}{0.6 \ln(\sin a) + 4.3}
\]

where:  
HW = wave height  
HR = height of ridge crest  
a = slope angle of beach face

Tanner and Stapor further develop their argument by stating:

Wave energy levels obviously do not remain constant. However, berms and beach ridges are built, if they are built at all, by relatively energetic waves: the small bermlets which adorn some beaches when wave activity is low will be destroyed quickly when activity picks up again. That is, there is a range "maximum wave energy" which stays pretty much the same year in and year out and it is waves which fall in this range which build the features discussed here (Tanner and Stapor, 1972, p.397).

The value of Tanner and Stapor's model probably lies in its ability to predict wave heights associated with the construction of specific ridges. It is less effective as a means to establish long term (upper Holocene) fluctuations in sea level as suggested by the two authors in a later paper (Stapor and Tanner, 1977). Use of the term "maximum wave energy" implies some sort of finite cutoff of the right hand portion of the frequency - magnitude curve in Figure 6-1. There is no evidence to justify such an arbitrary upper limit to the magnitude of ridge building events in the Gulf of Carpentaria.
The importance of frequency-magnitude relationships is borne out by evidence from Pacific islands and the Great Barrier Reef (McLean et al., 1978) and the east coast of Australia (Thom and Chappell, 1978) which suggests that the construction of storm berms or "perched" ridges may proceed during high energy events with a recurrence interval of tens to hundreds of years. There are further implications (Thom, 1978) suggesting that past climatic fluctuations during the late Holocene may have varied the degree of storminess and therefore altered the storm magnitude distribution.

Wave-cut scarps may provide indication of paleo-sea levels, but their use in this context must be tempered by limitations similar to those of beach-ridge crests. The use of erosional features to infer sea level necessitates some confidence in the relationship between storm wave heights and sea level. If the resolution sought is within the range of temporarily elevated sea levels which are common during storms, then wave-cut features are to be regarded with caution. The issue of magnitude-frequency relationship plays a critical part in the reliability of such features as wave-cut scarps, only in cases where the expected paleo-sea level lies well above the infrequent storm surge should wave-cut features be admitted as qualitative evidence. Even then, "high energy windows" of a time variable nature such as described by Neumann (1971) must be considered as possible explanations of such features.

Mudflats of either an intertidal or supratidal nature have a direct relationship with tidal amplitude and therefore a secondary relationship to mean sea level, provided tidal amplitude can be shown to be constant. Proof of constant tidal amplitude requires two tide defined horizons whose separation through time remains unchanged. Availability of such proof is rare, and a single horizon, usually related to mean high water or high water spring are often substituted.
Deactivation of mudflats by upward accretion during tidal inundation is a frequent end result of tidal accumulation. Further deposition of terrigenous material from fluvial sources may further raise the levels of relict tidal flats, especially those adjacent to upland surfaces. This possibility must be considered when using perched mudflats as paleo-sea level indicators.

Sea level curves are best inferred from a rigorously evaluated data base. Evidence for higher than present sea levels which is based on isolated single-event type data is to be regarded conservatively. Only a record in which systematic replication of the same form of evidence from several independent sites should be considered as a basis of a relative sea level curve. Even then, care should be taken to establish the "envelope of uncertainty" for the particular indicator whether it is a depth zoned biologic evidence or wave and tide constructed morphologic features.

THE GULF OF CARPENTARIA RECORD

Previous chapters, especially Chapters 4 and 5, have alluded frequently to the value of a particular facies boundary as a relative sea level indicator. This facies boundary is at the base of strandline sediments and separates the beach ridge sands from the low tide muds. That this boundary bears a consistent relationship to tidal datum and therefore to mean sea level is established by observations at eroded ridges, tidal creek exposures, drill hole logs and excavations. Figure 6-2 shows the "modern" topographic relationships of the strandline/low tide mud boundary for the chenier and beach-ridge plains; observations are based on 6 and 7 carefully surveyed sites on the chenier and beach-ridge plains, respectively. These measurements were taken on the
Figure 6-2. Modern and relict features on both the chenier and beach-ridge plain exist in a topographic framework with respect to tidal datum. These observations are particularly important in the interpretation of relative sea level indicators.
strandlines which appear to be presently active; however, radiocarbon determinations on shells from these strandlines place the youngest on the chenier plain at 650 yrs B.P. and on the beach ridge plain at 150-240 yrs B.P. Although younger strandlines exist in the central and western portions of the chenier plain, they are too remote from the Karumba benchmarked survey grid to be useful.

Only a limited number of radiocarbon determinations were used to construct the Holocene sea level relationships which follow. Ages with an identifiable relationship to the tidal datum via the strandline/low-tide mud boundary are plotted as a two standard deviation range either side of the mean. In addition, their vertical placement is defined by an "envelope of uncertainty" which is taken as twice the vertical range of modern observations for the following reason. Since any occurrence of the diagnostic facies boundary for strandlines older than modern may have developed at either the lowest or highest possible occurrence of this boundary (as well as anywhere in between), the uncertainty amounts to the modern range taken both above and below the observed level.

Relative sea level on the chenier plain

Relative sea level as preserved in the chenier plain from mid-Holocene to present is shown in Figure 6-3. Sea level reached its present position sometime prior to 6000 yrs B.P. After a mid-Holocene maximum at approximately 5500 yrs B.P., there was a stillstand followed by a rather rapid relative sea level fall of approximately 2.4 meters starting about 4000 yrs B.P. Five radiocarbon dated chenier or chenier complexes are used for input to this curve and all dates presented are confirmed by at least two additional determinations on the same or a correlative strandline.
Figure 6-3. Relative sea level curves such as this for the chenier plain must indicate the envelope of error associated with the age of a sample as well as the uncertainty of its modern relationship to sea level. Relative sea level on the chenier plain appears to have fallen 2.4m since 4000 yrs B.P.
Relative sea level for the last 500 years shown in Figure 6-3 is unknown as the youngest chenier is approximately 500 years old. However, refinement in this portion of the curve is not necessary to indicate an unequivocal relative fall in sea level since the mid-Holocene in the southern Gulf of Carpentaria.

Relative sea level on the beach-ridge plain

Data from the Holroyd transects are not as smooth as the chenier plain observations. Levels and ages of the facies boundary from Edward River are presented in Figure 6-4 and from Christmas Creek in Figure 6-5. At Edward River the data shows that relative sea level reached its present position prior to 6000 yrs B.P. and continued to rise approximately 0.8m above present level until 5500 yrs B.P. Following a stillstand between 5500 and 5000 yrs B.P., a rapid fall brought relative sea level to its present level or somewhat below. The uncertainty introduced by the topographic positions of ANU 2059, 2102, and 2103 (Figure 6-4) will be discussed later. Data from Christmas Creek indicate a 1.5m stand above present level at approximately 5500 yrs B.P. However, this highstand was immediately followed by a rapid relative fall until after 5000 yrs B.P. Between 5000 yrs B.P. and the present, there appears to have been a slow fall in sea level. The possible oscillation at 3000 yrs B.P. may be discounted because the relative levels of ANU 1734 and 1736 lie within the 0.4m range of the diagnostic boundary.

Edward River and Christmas Creek transects should yield comparable results. They have similar marine exposures, the same tide range and are separated by only 60km. Figure 6-6 shows a composite envelope constructed by overlaying the observations from these neighboring transects. The summation of the two beach-ridge curves suggests a high stand of sea approximately 1.5m above present level 5500 years BP.
Figure 6-4. Sea level data from the Christmas Creek transect indicate a 1.5m fall during the last 5000 radiocarbon years.
Figure 6-5. Sea level data from the Edward River transect indicate a relative fall in sea level followed by a rise during the upper Holocene.
Figure 6-6. A rigorous evaluation of beach-ridge plain sea level data requires that observations from both transects be pooled. An inner envelope of confidence indicates a 1.5m fall in sea level for this portion of the Gulf during the last 5500 radiocarbon years compared to 2.4m on the chenier plain in the southern Gulf.
Similarly to the chenier plain, this high stand was followed by a general fall in relative sea level until present. If the two anomalously low observations from Edward River (ANU 2102 and ANU 2103) are omitted, there appears no evidence for a slight regression at approximately 1400 yrs B.P.

There are several reasons for the anomalous appearances of the beach-ridge plain curves with respect to that presented for the southern Gulf. Firstly the Edward River transect was logged at 1.5m (5 ft) auger intervals. The low positions of ANU 2102 and ANU 2103 could be due to errors of drill hole logging. ANU 2102 is contradicted by ANU 2059, both of which are of similar age, therefore ANU 2102 is allowed to remain outside the envelope constructed in Figure 6-4. The inclusion of ANU 2103 is probably questionable but not fully contradicted by other data. Secondly, stratigraphic descriptions in Chapter 4 showed that the location of the facies boundary beneath the beach-ridge plain is not always as closely indicated as on the chenier plain. This boundary may be difficult to identify if sandy lenses are present in the low tide facies. The critical parameters involved in its vertical definition is the highest occurrence of interbedded clay, silt and sand, similar to that found on the low tide flat. Study of the modern equivalent for this facies showed that sandy lenses may occur on the low tide flats, therefore causing the sand-mud boundary to occur at a lower level. Although the horizontal bedding characteristic of this environment can remove some of this ambiguity if the stratigraphy is studied in section, bedding cannot be determined from auger returns. Ambiguity is least when the sand-mud facies boundary is distinct in textural terms.

The Christmas Creek transect (Figure 6-5) was logged at 0.75m (2.5ft) intervals to overcome some of the difficulties described above.
There appear to be fewer anomalous relationships in this more closely studied stratigraphy. The absence of a radiometrically young modern strandline leaves some question as to the form of the curve in the upper Holocene. However, the proximity (30km) of the Christmas Creek transect to Edward River permits extension of modern observations to the former site because tidal range, coastal morphology and stratigraphy are similar.

The most conservative view based on the envelopes of uncertainty clearly establishes two points: (1) a relative higher than present sea level occurred during the mid-Holocene on both the beach ridge and chenier plains, and (2) this change in relative sea level appears different in magnitude for the two types of coast, showing 2.4m of change on the chenier plain and 1.5 meters of change on the beach - ridge plains. Figure 6-7 summarizes the relative sea level findings for the two areas. Several possible explanations for these records need to discussed.

INTERPRETATION OF THE RECORDS

Interpretations of strandlines existing at higher than present levels may be a function of one or more of the following six factors of sea level change.

1. a data misinterpretation
2. neo-tectonic movement
3. global eustatic sea level change
4. variation in tidal amplitude if it can be shown that the sea level indicator is related to either high or low water levels.
5. variation in storminess if the sea level indicator is related to storm wave energy rather than mean sea level.
6. hydro-isostatically induced emergence.
Figure 6-7. A summary and comparison of chenier and beach-ridge plain data shows a greater amount of sea level fall for the chenier plain.
Data misinterpretation

Although field methods are outlined in Appendix A, a brief review of some of the precautions taken to ensure data quality may confirm that the relative sea level changes presented here are "real". Despite the possibility of imprecise drill hole logging at Edward River, such is not the case on either the Karumba or Christmas Creek transects. The Karumba transect provided data from sections cut across the ridges. Both primary structures and lithology confirmed the location of the facies change under these conditions. It has already been shown that the Christmas Creek transect was logged at twice the resolution of the Edward River transect. The effects of field error on interpretations from these latter two transects is well understood.

Survey techniques ensured a 2-4cm accuracy at closure for all transects with the error distributed along the transect. At Karumba, the presence of a stable benchmark to which previous tide records have been related ensure accurate positioning with respect to the tidal datum. At Edward River, one year's tidal records during 1976 validate the approximate tidal datum used for the Edward River and Christmas Creek transects. Finally, despite certain inconsistencies in the Edward River transect, three independent localities exhibit the same general trend of sea level change. It is this general trend which is certainly "real" and which merits further consideration.

Tectonics

Neotectonics not associated with postglacial hydro-isostasy could be invoked to explain not only the net relative sea level change, but also the differential change between the two localities. If tectonics is the explanation, then evidence of greater differential uplift might be expected in the topography of the last interglacial strandlines. These
strandlines, probably formed 120,000 yrs B.P., are ubiquitous in their occurrence around the Gulf. Although they vary in lithologic character, especially with regard to depletion of carbonate, they form a general reference to which Holocene strandlines may be compared. Figure 6-8 is a summary of the topographic relationships between the Pleistocene, Holocene, and modern levels.

It is difficult to fix the precise level to which Pleistocene strandlines were built because of their loss of calcium carbonate due to leaching. In addition, Pleistocene beach ridges rest on a sandy alluvium landward of the Holocene beach-ridge plain. The boundary between highly leached marine sands and medium to well-sorted alluvial sands is not readily identified. Therefore comparison of ridge crest levels is the only possible means by which to examine levels of the Pleistocene strandline.

Assuming one meter loss due to leaching on the beach-ridge plain in a strandline facies which is now up to 3 meters thick, the consistency for Pleistocene-Holocene datums from widely separated areas is remarkably good. From Figures 6-2 and 6-8 it can be calculated that tectonic movement since deposition of the 120,000 yrs B.P. strandline may be insignificant. This conclusion is developed as follows:

Compare the observed height of Pleistocene cheniers with Pleistocene beach ridges:

Pleistocene strandline (chenier plain) max crest height= 10.0m
or 8.2m above mean sea level (see Figure 6-2)

Pleistocene strandline (beach-ridge plain) max crest height= 6.0m +1.0m
or 5.6m above mean sea level

\[
\frac{8.2}{5.6} = 1.46 \quad \text{(ratio 1)}
\]
Figure 6-8. A comparison of Pleistocene, Holocene and modern levels is used to dismiss uniform tectonic change as an explanation for relative sea level change. See text for calculations.
Whereas, assuming uniform tectonic uplift, one would expect the following ratio:

\[
\frac{\text{observed relative sea level change (6000yrs) chenier plain} = 2.4m}{\text{observed relative sea level change (6000yrs) beach-ridge plain} = 1.5m}
\]

and adjusting for average thickness of strandline facies, where:

- average chenier thickness = 2.8m
- average beach ridge thickness = 2.6m

and considering continuous tectonic movement from 120,000yrs B.P. to the present, the expected height ratio of Pleistocene (last interglacial) ridge crests at the two locations should be:

\[
\frac{2.5m + 2.8m \times \frac{120,000yrs}{6,000yrs}}{1.5m + 2.6m} = 25.9 \quad \text{(ratio 2, i.e., more than 10 times greater than observed height ratio of approximately 1.5, see ratio 1 above)}
\]

Although the amount of material lost by deflation and leaching of calcium carbonate is unknown, comparison of the two ratios above suggests that simple uniform tectonic movement cannot be the explanation for the observed differences in relative sea level change. There may be errors due to loss of material as well as variations in mean sea level and associated survey errors. However, none of these sources of error can account for the magnitude of difference between the ratios 1 and 2.

It is also unlikely that some non-uniform tectonic arching could explain the emergent record. Grimes and Doutch (1978) found no evidence for major post-Tertiary tectonic activity in the Karumba Basin. Although tectonism may have continued into the Holocene in the uplands, the major Cainozoic alluvial fans of the Karumba Basin exhibit no dislocation. Uniform tectonics throughout the later Quaternary can be omitted as an explanation for the observed relative sea level curve for reasons shown above. Oscillatory or intermittent neotectonics need not be considered unless all other possibilities are eliminated.
Eustasy

Differential movement between Edward River areas and Karumba areas in the Gulf of Carpentaria cannot be explained in eustatic terms. However, sea level fall indicated in both areas should be examined in a eustatic context since Holocene high sea levels have been inferred in other parts of the world. Eustatic changes, those changes in sea level of a global nature which are observable as synchronous over wide areas, are difficult to identify, as Holocene glacio and hydro-isostatic movements may have affected the entire globe (Clark et al., 1978). A true eustatic curve, from which isostatic influences have been removed may only be achieved by calculation of the isostatic factor. Early work on the calculation of this factor proceeded in areas of large glacio-isostatic changes adjacent to the Fennoscandian and Laurentide ice caps (Morner, 1971). Other workers have preferred to avoid the complications of glacio-isostasy and have sought "synthesized" eustatic curves from widely separated data sites (see Fairbridge 1962 and 1976), whilst others have advanced the notion of "stable" areas which have experienced a minimum of isostatic movement (Bloom, 1970).

It is apparent from even a cursory review of sea level studies that two distinctive divisions have existed in the overall controversy. Newman (1968) described these divisions:

the North American-Gulf Coast School... who find that sea level has risen to its present position only during the last millenia, and the Indo-Pacific school... who believe that sea level rose to its present level some 5000-6000 years ago and has fluctuated within 3 meters of its present level, including one or more conspicuous stands at +2 to 3 meters. (Newman, 1968 p152).

The development for so-called "synthesized" sea level curves by workers such as Fairbridge (1961, 1976) is frustrated by two major hurdles. (1) The possible error inherent in assembling observations from widely scattered sites where stratigraphic interpretations, levelling and
dating accuracy may be in doubt greatly detracts from the value of such a curve. (2) Even areas formerly labelled as "stable" are now viewed as having been subject to variable hydro-isostatic loading during and after the Holocene transgression.

Clearly different from the "synthesized" curves of Fairbridge are the single-site studies in which data from one area is developed into a generalized curve for a region. Such studies include those of Curray (1960), Bloom (1969, 1970), Morner (1971), Scholl et al. (1969), Schofield (1960, 1977) and Tooley (1974).

Curray (1960) presented a "single site" curve from radiocarbon dates on marine shells recovered from subsurface facies in the northwest Gulf of Mexico. For each type of shell a modern depth range was determined and the assumption was made that transportation since death of the organism had not occurred. Curray's first approximation for the history of the transgression in this area was presented as irregular curve constructed through the C-14 ages and the depth ranges of the organisms. Although Curray was probably correct in assuming stability for this coast, a broader envelope of error would have better described the time-depth form of the Postglacial Marine Transgression.

Bloom (1969, 1970) noted that a serious shortcoming of the search for a "stable" coast is that these sites have previously been located on continental coasts where isostatic factors cannot be ignored. His data from Micronesian islands of Truk, Ponape and Kusaie, which are far from continental coasts, were purported to provide a depth gauge type of record which represented only the volumetric change of the oceans. Bloom's eustatic curve is included with others in Figure 6-9, however it should not be expected that the Gulf of Carpentaria should show a sea level record similar to the Micronesia curve. The Gulf of Carpentaria is entirely located on the continental shelf and coupled on three sides to a
Figure 6-9. A summary of several well known sea level curves shows lack of agreement with respect to both geography and form. The oscillatory Fairbridge curve is in marked contrast to the more linear Bloom curve.
continental shoreline. Such an environment will not provide the simple eustatic dipstick type record which Bloom sought in the Pacific islands.

Southern Scandanavia provides a record of shoreline uplift which is due to both isostatic and eustatic factors. Morner (1969, 1970) claimed that the isostatic factors could be removed once the axis of subsidence was located. He further stated that uplifted areas such as southern Scandanavia may provide the most suitable location for such work. Morner claimed that the Bloom (1969) curve was not a true eustatic curve because the Micronesia data only represented "ocean volume changes" rather than time eustatic behavior. Morner defined true eustatic changes as:

the final changes of ocean level after the influence by the possible hydro-isostatical movements of the ocean floor as well as other isostatic changes in the world affecting ocean level (Morner, 1970, p. 69)

Morner offered as tentative proof that eustatic oscillations are real, the fact that there is some agreement between his curve and Fairbridge (1961). However, such comparisons may not be justified in the light of Morner's own theory that eustatic data can only be gained after removal of all isostatic factors, a practice which Fairbridge does not apply to his synthesized curves. In a later work, Morner (1976) suggested that global variation have occurred in the form of geoid changes. These geoid changes appear to be related to activity at the core/mantle interface. According to Morner, the shape of the earth has not remained stable through the Quaternary and therefore eustatic "average" sea level curves are meaningless.

There is similarity between the Bloom (1970) curve and that of Scholl et al. (1969). The Florida curve of Scholl et al. is based on a generalized trend exhibited by organic deposits of both lagoonal and estuarine peats on the southeastern Florida coast. Comparison of the Micronesian and Florida curves in Figure 6-9 indicates that the two
curves demonstrate approximately parallel submergence trends for the last 3000 radiocarbon years. Scholl et al. are among the few workers who advocate the use of an envelope of uncertainty which in their case is an error of 1.0 m above and below the trend line. Whether or not the Scholl et al. submergence curve reflects only eustatic factors without a hydro-isostatic component remains to be established. If the Micronesia curve from Bloom (1969) represents an ocean volume curve which is very similar to a eustatic curve, then southwest Florida must represent a stable continental margin apparently immune to hydro-isostatic effects in order to permit such concordance of the two records.

A study such as that presented by Tooley (1974) may be set apart from most previous efforts because of its rigorous methodology and well controlled age structure. Tooley notes nine and perhaps ten oscillations of relative sea level by analysis of transgressive and regressive sequences in marsh drill holes. The author considers that, although the frequency and direction of the oscillations are in part eustatic, the magnitude may be of local character.

Tooley presented a relative sea level curve (Figure 6-9) based on dated biogenic sediments whose relationship to former environments of deposition, and therefore sea level, were established by pollen and diatom analysis. His dated sites are estuarine marshes in northwest England and north Wales from which stratigraphic data were gained by drillholes and excavations. Sediments varied from marine sand and gravels to undifferentiated organic material from upper saltmarsh facies. The quality of both the stratigraphic description and detailed topographic analysis is unequivocal.

However, several matters detract from this apparently rigorous analysis. Firstly, some of the radiocarbon defined stages are rather brief, specifically Lytham IV - 175 years and Lytham IV - 180 years. The
identification of periods of such length by radiometric means is perhaps an over extension of the present state of the art. Secondly, Tooley weakens his argument by suggesting that the lack of correlation between his findings and the findings of others, including the curve of Fairbridge (1961), may be due to insufficient radiocarbon determinations, sediment consolidation and compaction and similar local factors in other areas, factors which could just as easily invalidate details in Tooley's interpretations. Finally, consideration must be given to the possibility that many of the estuarine sites from which Tooley recovered his data have been subject to tidal range variation as the estuary in-filled during the late Holocene.

Schofield's (1977) report of atoll deposits in the Gilbert and Ellis Islands provided additional insight regarding possible relative sea levels in the western Pacific. Schofield probably over extends his latest interpretations by suggesting that Gilbert island data are correlative with transgressive peaks from the Firth of Thames chenier plain. Schofield admits that the Gilbert Island data lack "second order" regressions as seen in the Firth of Thames record. However, he suggests that the New Zealand and Gilbert Island data are correlative, largely due to the apparent synchronous "fit" of the peaks. Since Schofield's (1960) primary data have been argued to be suspect because of frequency-magnitude relationships of ridge building events, less confidence should be placed in the details of these two curves, whilst accepting that they both indicate a higher than present sea level for the respective areas. The real importance of the Schofield (1977) curve from the Gilbert Islands is that it unequivocally contradicts Bloom's (1970) interpretations from the CARMARSEL expedition. Schofield's major contention is that peat layers, as used by Bloom, on the Caroline and Marshall islands may not be reliable indicators of relative sea level
change.

Some of the curves presented in Figure 6-9 are flawed in that their authors do not present the envelopes of uncertainty which are implicit in radiometric determinations even if the facies-datum relationships are absolute. If the same liberties were taken with data from the Gulf of Carpentaria, three curves (Figure 6-10) could result, all differing and all distant approximations of the true relative sea level history. At no time since the mid-Holocene has relative sea level in the Gulf returned to the height achieved at the end of the Flandrian transgression. Although the curve from Christmas Creek may give circumstantial evidence for a minor transgression between 2500 and 3000 yrs B.P., it has been shown that the facies boundary associated with these observations could have been drawn from anywhere within the envelope of uncertainty. Other curves, which show continued submergence (Scholl et al., 1969 and Bloom, 1970), are in marked contrast to the Gulf record which exhibits continued emergence.

However, the best reason for not relating the Gulf of Carpentaria data to any eustatic model from elsewhere in the world is the apparent lack of such a model which "fits" the Gulf data. Perhaps the resolution of the Carpentaria data is insufficient to reveal the eustatic fluctuations claimed elsewhere. Perhaps an epicontinental sea with an annual change of sea level produces a record which masks such oscillations. Whatever the explanation, there appears no evidence to suggest that the eustatic changes claimed by other workers are applicable to the Carpentaria curves.

Tidal range variation

Holocene variation in tidal range and associated sedimentation has been shown to have occurred in the Ord River, W. Australia, (Wright et
Figure 6-10. In this example, Gulf of Carpentaria sea level data are presented without envelopes of uncertainty. Such presentations, frequently used by earlier workers, are unacceptable because excessive confidence is placed in single observations.
al., 1972), the Bay of Fundy (Grant, 1970) and Broad Sound, Qld (Cook and Mayo, 1977). That such variations could have affected the relative sea level record exhibited in the Gulf of Carpentaria must be considered. Temporal variation of tidal range during the upper Holocene has been claimed as the cause of locally observed relative changes in sea level at the three localities mentioned above. All these examples are either enclosed or semi-enclosed bodies of water which makes them only marginally comparable to the Gulf. There are no studies of modern epicontinental seas showing such changes.

Grant (1970) found evidence that the Bay of Fundy has been experiencing continued submergence in both the historical and geologic past. This body of water, which has a modern tidal range of 15m, is a long narrow arm of the sea connected to the Gulf of Maine. Under present conditions, tidal amplitude increases from the mouth toward the head of the bay, an observation to be expected considering the geometry of the basin. However, historical records and archeological evidence less than 300 years old shows high tide levels have increased in relative elevation by 0.8m - 1.0m during this period. Radiocarbon dating of submerged fresh water peats showed this submergence to be operative for the last 5000 years. The selection of a geobotanical defined level (fresh - salt water boundary) called "Higher High Water" is similar to the environmental indicators utilized by Tooley (1974) in British marsh stratigraphy.

After removal of submergence from hydro-isostatic, glacio-isostatic and eustatic sources, Grant found a component which was still unexplained. He attributed this to tidal variation in the Gulf of Maine during the Flandrian transgression. The Gulf of Maine is separated from the North Atlantic by a narrow channel adjacent to Georges Bank. Prior to 8000 yrs B.P. a narrow threshold width precluded tidal amplification. However, by 6000 yrs B.P. threshold width was increasing as shown by
Grant (1970, Figure 14) and by 4000 yrs B.P. the "upstream" tidal range was increasing exponentially.

The Ord River - Cambridge Gulf example (Wright et al., 1972) revealed evidence of emerged tidal flats at the mid-Holocene limit of tidal influence. According to these authors, Holocene infilling of the embayment has caused shallowing of the upper tidal reaches increasing frictional dissipation. Prior to infilling, a longer tidal wavelength in conjunction with funnelling effects caused a larger tidal amplitude at the point where elevated deposits were observed.

The Broad Sound example is based on data gained from "in situ" mangrove deposits buried beneath supratidal flats. This estuary, which is a macro-tidal environment on the middle North Queensland coast was extensively studied by Cook and Mayo (1977) who reported a clearcut trend to increased elevation of mangrove deposits with decreasing age. The authors infer that this trend suggests an increase of approximately 1.0m in tidal range during the last 6000 years. This change is attributed to an increase in the V-shaped configuration of the estuary during the last 5000 years.

It is interesting to note that claimed effects from similar causes are in opposition when comparing Broad Sound with Cambridge Gulf. Cook and Mayo argued that an increase in the V-shaped nature of the estuary has increased tidal amplitude whilst Wright et al. show that sediment infilling of a V-shaped estuary has decreased the amplitude. Cook and Mayo have also shown that concomitant with the increase in the V-shape of Broad Sound there has been an increase in subtidal shoaling. Comparison of these two examples demonstrates that tidal range variation may occur from different factors under various estuarine geometries and dimensions.

Review of the three previous examples provides little progress toward explaining Carpentaria emergence with tidal range variation. The
Gulf has no restricted threshold analogous to that between the North Atlantic and the Gulf of Maine. Although the Wessel Rise, which separates the Gulf from the Arafura Sea, was probably transgressed in the late Pleistocene if our present understanding of the postglacial sea level rise is near correct, there appears to have been no time variable threshold width to regulate tidal exchange with the Arafura Sea. After flooding of the Gulf was complete, a shallow epicontinental sea existed with a form almost the same as present in which overflattened nearshore gradients contributed to frictional dissipation from the start. Although the precise thickness of Holocene deposits since 6000 yrs B.P. in the subtidal zone can only be inferred from their thickness at the shore and their approximate offshore limit, it is unlikely that deposition of this thin veneer of material over the pre-Holocene surface could alter the tidal range in a manner shown by the Gulf of Carpentaria record.

Previous chapters have shown that the Carpentaria shore prograded as a smooth arcuate coastline with few estuarine inlets. The emergence which is noted is found in the strandline record which was formed at the mouth of what few inlets existed. Therefore, no analogy with variation of tidal amplitude due to estuarine infilling can be shown. The conclusion can only be that if variation of tidal amplitude has occurred in the Gulf, it has resulted from some process other than shown in the examples outlined above. Certainly secular variation of tidal amplitude is possible but the proof of such a variation cannot be obtained from data presently available.

Variation in storminess

Variation in storminess has been considered in Chapter 5 as one of the causes of episodic progradation. Because of the association of storms with elevated sea levels it is appropriate to reconsider this
factor during a discussion of relative sea level change. Previous chapters have shown the Gulf of Carpentaria to be subject to a pronounced annual sea level change of 1-2 meters which is a result of dominant northwest winds during the summer months. Superimposed on this higher sea level during the summer are frequent storm surges associated with cyclonic disturbances. These storms have been shown to elevate Gulf levels near the shore at least 2.0 meters above tidal predictions.

A critical "missing link" in this discussion is the absence of short term observations of strandline response to cyclonic events. A general aerial survey of several hundred kilometers of Gulf shoreline after cyclone Ted in December 1976, accompanied by ground inspection of a modern chenier adjacent to Karumba showed an overall erosive effect on the beach-face and berm crest to the extent that well established Casuarina trees were either toppled or removed by the storm waves. Although some cheniers showed minor landward transport of material by overtopping waves, others showed massive seaward movement of sediment apparently by the ebb currents flowing off the flooded coastal plains following the storm surge.

Catastrophic events of the above nature may occur early in the wet season before the major influx of terrigenous sediment to the nearshore zone. Nearshore response under these conditions will usually permit high levels of energy to reach the strandline causing erosion as observed after cyclone Ted. However, a cyclone in the month of February finds different nearshore conditions. Although Gulf water levels are still raised, there is a viscous slurry of mud, 1.0 - 1.5m thick, in the subtidal zone which is effective in damping wave energy, a phenomenon also noted by Augustinus (1978) on the Surinam coast. Within this viscous mixture there is also a coarse fraction, both bioclastic and terrigenous, which is winnowed and starts its progression toward
strandline deposition. Advice from apparently reliable sources such as residents of Edward River indicates that the beach may actually gain material on beach face and berm crest during such high energy events. However, beach profiles installed in 1975 and resurveyed in 1976 failed to show any clearcut trend toward erosion or deposition on either cheniers or beach ridge locations during the period of observation.

Therefore the response of Gulf strandlines to high energy events is speculative. It does appear that storms may cause deposition as well as erosion depending on their timing. However there is no evidence to show that they predictably cause deposition at higher levels, especially deposition of fine-grained facies which is the indicator feature used in this discussion. Chapter 3 showed that contact between fine low tide muds and the coarse strandline facies is a sharply defined one. Special, rare conditions such as an incompletely welded ridge causing a trough behind a berm must occur for mud facies to be deposited within beach sands.

Reason to disqualify variations of storminess from any role in apparent relative sea level change is therefore contained within an understanding of the modern environment. The modern system appears to adjust to storms and exhibit erosion or deposition as a function of nearshore conditions. Storm surges which double present tidal amplitude are an annual event. Modern observations do not provide an example of "perched" strandline deposition as a result of this storminess. Increased frequency of storms would not effect levels of deposition of key facies provided the present monsoonal system persists.

Hydro-isostasy

There can be little doubt, given the present understanding of upper crust rheology, that hydro-isostatic warping of the earth's crust is as
"real" as glacio-isostatic relaxation. Qualitative descriptions of such responses were put forward as early as Daly (1925), whilst Bloom (1967) and Morner (1971) argued that removal of the hydro-isostatic effect might be useful in determining true eustatic behavior of the oceans. Walcott (1972) showed that differential response to hydro-isostatic loading may proceed in areas remote from glacio-isostatic unloading, and can account for higher than present relative sea levels during the last 6000 years without eustatic changes.

Two problems in such computations are the lack of quantitative values available for the viscous and/or elastic parameters of the earth to loading and unloading, as well as establishment of a realistic layered earth model. Additional computational difficulties arise when dealing with sloping continental shelves of irregular geometry flooded by a time-varying sea level load. Considering these handicaps, it is a strong point in favor of hydro-isostatic theory that Chappell (1974) predicted deformations which closely match the observed Holocene and modern warping in eastern North America and in the Baltic region. Further confirmation of the theory can be gained from more recent results of Clark et al. (1978) whose predictions for several other regions resemble observed data. This latter work is an effort to improve on models for visco-elastic behaviour originally selected by Cathles (1975) and later used by Peltier and Andrews (1976) to identify variable response in different geographic areas. Clark et al. identify six zones of differential response to postglacial melting and provide the best preliminary world wide syntheses of hydro-isostatic predictions to date (Figure 6-11).

Of particular relevance to the discussion is Zone VI (Clark et al.) which consists of continental shorelines with emerged strandlines. The authors are able to draw favorable comparison with Fairbridge's (1976)
data from Recife, Brazil, whilst predicting the emerged beaches of the Indian Ocean (Stoddart, 1971) as well as emergence of the Australian coasts. Australia lies in Zone V, an area of oceanic emergence which according to the authors experienced 1.5-2.0m of sea level fall in the last 5000 years. Although Clark et al. suggest that iterative solution toward a more constrained viscosity profile may refine the "fit" of the global model, for solution of regional problems, it would be more appropriate to consider the layered earth model of Chappell (1974) as shown in Figure 6-12.

Chappell (pers.comm., 1978) suggests that a numerical model for the northern Australian shelf is certainly possible, but difficult owing to the lateral transformation of continental to oceanic lithosphere - asthenosphere conditions at the continental margin. Irregular geometry of the North Australian shelf makes computational methods used by Chappell (1974) inapplicable. Chappell (1974) computes the responses of circular or infinite parallel basin margins to transgressions and regressions using Hankel and Fourier transforms to manipulate the stress-strain equations. These methods cannot be applied directly to irregular geometries. However, this view does preclude a qualitative assessment of the role of isostasy in this area.

To invoke hydro-isostasy as an explanation of emergent Gulf shorelines requires consideration of two aspects. (1) Why is there differential response on shorelines of the same epicontinental sea? (2) What has been the time-varying position of the hingeline or isopleth of zero relative change which separates the emergent sections from the submergent ones? The analysis by Chappell (1974) reproduced in Figure 6-13 indicates that different shelf geometries will respond in different manners to the same hydro-isostatic load. Furthermore, Chappell has shown that for a shelf similar to that on which the Gulf of Carpentaria is
Figure 6-11. Using computational techniques based on visco-elastic behaviour of the earth's crust during post-glacial melting, Clark et al. (1978) predict six zones of differential response. Zone 6, continental shorelines is lightly shaded.

Figure 6-12. Chappell (1974) suggested this layered earth model when calculating the hydro-isostatic response at regional rather than global scales.
situated, the zero isopleth probably moved landward during the upper Holocene. That the Karumba chenier plain and the Holroyd beach ridges remained on the the uplift (continental) side of the zero isopleth during the upper Holocene appears likely from Figure 6-14. Concepts of hydro-isostasy, especially those which account for shelf loading, permit mass transfer from beneath the loaded basins and shelves to beneath the adjacent continents. Chappell (1974) has already suggested that eastern Australia, particularly the Great Barrier Reef province may exhibit a hinge line of time-variable location. He notes that Holocene reef crests on the outer reef appear to be presently growing upward with no evidence of emergence higher than present levels. This condition is in contrast to slight emergence shown by dead reef part way across the shelf and considerable emergence of up to 1.0-1.5m closer to shore (Thom and Chappell, 1978).

Elsewhere in northeastern Australia, slight emergence is visible on aerial photographs of a chenier plain at Princess Charlotte Bay, an east coast embayment which is at equal latitude with Edward River. Data are non-existent for the remainder of the Cape York Peninsula as well as the Torres Strait region. The mouth of the Jardine shows local uplift of coastal features in excess of that further south in the Gulf. However, this emergence as well as that near Port Musgrave and Weipa has not been quantified and could represent local neotectonics of a nonhydro-isostatic nature. Data from the western side of the Gulf and in Arnhem land are also non-existent so only limited generalizations can be made about the possible vertical movements of this portion of Australia.

Considering hydro-isostatic loading as a simple qualitative model which ignores the irregular geometry of the respective shelves as well as Papua New Guinea, Figure 6-14 suggests the possible sources of loading to the Gulf of Carpentaria. Whatever the form and rate which can be
Figure 6-13. The principle of differing hydro-isostatic response to upper Holocene loading of continental shelves of different geometries helps explain varying sea level curves at widely separated locations (from Chappell, 1974).
ascribed to the "real" Postglacial Marine Transgression, it is certain that water encroached onto the Northern Australian shelf from several directions. The trace of the present zero isopleth is certainly more complex than this figure suggests. However, this diagram exhibits the difficulties of modelling such a complex shelf, especially where empirical data are absent or sparse. Answers to the two questions posed at the beginning of this section are suggested from Figure 5-14. The zero isopleth may remain within the Gulf or intersect a portion of the western shore. The differential in emergence is readily explained by the respective locations of Holroyd and Karumba sites from the two sources of loading.

Clark et al. (1978) concluded their discussion of postglacial sea levels with a broad generalization:

Although agreement with observations is not perfect, the predictions of this model suggest that sea levels from all over the world may be explained without invoking any change in eustatic sea level during the past 5000 years (p. 285).

It may now be appropriate to re-evaluate previous findings which used eustasy to explain shorelines of higher than present level, for example, observations of Schofield (1960) at the Firth of Thames, New Zealand, might be best explained in hydro-isotatic terms once the original data was contained within an envelope of uncertainty to remove the "noise". However, every such case requires re-evaluation not only of the physical field evidence in order to establish the envelope of uncertainty, but also determination of quantitative estimates of likely isostatic histories. This latter problem is at present unsolved as shown by the above discussion of Gulf of Carpentaria isostasy.
Figure 6-14. A qualitative model of hydro-isostatic loading on the Northern Australian shelf shows that flooding from the Arafura and Coral seas would explain the Carpentaria observations.
CONCLUSIONS

This chapter has reviewed the use of various indicators in the study of postglacial sea level. Many of these indicators, and especially those associated with high-energy constructional events are suspect because frequency-magnitude relationships of these events are difficult to quantify. It has been shown that the most dependable indicators are those which bear a direct finite relationship to mean sea level rather than tidal amplitude. The facies boundary beneath Gulf strandline facies is a reliable indicator which appears to be unresponsive to infrequent storms, whilst exhibiting a defineable relationship to mean sea level.

The Gulf of Carpentaria record shows that relative sea level reached its present position sometime prior to 6000 yrs B.P. This observation is in accord with the results reported by Thom and Chappell (1975) whose envelope of rising relative sea level intersects present sea level at or near 6000yrs B.P. Figure 2 in Thom and Chappell (1975) contains a number of radiocarbon dates which indicate that relative sea level elsewhere in Australia, particularly Victoria, might have reached or exceeded present sea level earlier than 6000yrs B.P. However, more recent findings from sites in eastern Australia, specifically those reported by Thom et al. (1978), further validate 6000yrs B.P. as the time when sea level reached or exceeded present level along most stable portions of Australia's coast.

There is unambiguous evidence on both the chenier plain and the beach-ridge plain to indicate a higher than present level by 5500 yrs B.P. Relative sea level was 2.4m and 1.5m above present on the chenier and beach-ridge plain respectively during this highstand. A relative fall in sea level taking a slightly different form and rate in the two areas brought sea level to its present position not later than 500 yrs B.P. in both areas.
Six possible explanations for this sea level history have been considered. Of these six, misinterpretation of primary data and neo-tectonic movement can be readily discounted. The form and timing of eustatic changes advocated by other authors do not appear to "fit" the Carpentaria data, although it is possible that annual sea level changes in the Gulf have masked minor eustatic fluctuations. Variations in tidal amplitude and storminess do not appear viable explanations, but for different reasons. Secular variation in tidal amplitude is difficult to document without the use of two tide related horizons, and such indicators are not available in the Gulf record. Variation in storminess, although considered as an important factor affecting periodic progradation, appears to have little effect on the topographic position of the diagnostic facies boundary used in this study. Evidence weighs heavily in favor of a hydro-isostatic factor to explain most of the sea level changes observed in the Gulf record for the last 6000 years. This conclusion is reached not by default of the other factors, but through careful evaluation of field evidence in comparison with findings from other areas where hydro-isostatic response can be proven. Hydro-isostasy has been shown to be unequivocally responsible for much of the relative sea level change during the upper Holocene on other continental coasts. This mechanism of relative sea level change is the only one which adequately explains in qualitative terms the Carpentaria record. Although quantitative comparison of a predictive model with primary data is possible for complicated shelf geometries such as the Gulf of Carpentaria, there remain unsolved problems concerning realistic estimates of crustal behavior to water loading of such irregular continental margins.
INTRODUCTION

In this final chapter, four sections, each utilizing material presented in previous chapters, will reconstruct the Holocene history and mechanisms of coastal progradation on the chenier and beach-ridge plains in the Gulf of Carpentaria. First, modes of deposition of the chenier and beach-ridge plains will be summarized. Second, a discussion of the periodic progradation demonstrated in Chapter 5, will be resumed, with some suggestions concerning a trigger mechanism. Third, the chapter will present an upper Holocene geologic history of the Gulf of Carpentaria. The chapter will conclude with a comparison of the Carpentaria depositional history with certain evidence for environmental change outside the study area.

DEVELOPMENT OF COASTAL PLAINS

Chenier plain - general

The chenier plain is composed of two distinctly different facies types: (1) muddy facies composed mostly of terrigenous silts and clays; and (2) sandy facies composed of terrigenous sand and skeletal carbonate clasts. These two facies types are deposited by different modes of deposition and in some cases under greatly differing environmental conditions.

Chenier plain - muddy facies

Subtidal muds are the "foundation blocks" of coastal progradation on the chenier plain. At present, these muds are derived from the upland and transported to the Gulf by a fluvial system during the annual wet season. Some fine sediment resides briefly in the estuarine portions of the system where muds are deposited as point-bar or overbank facies. However, the majority of the suspended material which enters the
estuaries is carried directly to the sea during periods of high discharge. Once at the mouth of estuaries, fine terrigenous material is distributed by waves and wave-induced currents. Chapter 3 demonstrated that some fine-grained sediments are found in ebb tidal deltas on the chenier plain; however, the dominance of sand in such deltas suggests that the mud fraction, especially clay, remains suspended and is distributed more widely in the nearshore zone before it is deposited as low-tide and subtidal mud facies. The dispersal of fine-grained sediments between their source areas in the uplands and the depositional environments on the chenier plain is summarized in Figure 7-1.

As the subtidal mud facies accretes upward into the tidal zone, there is a slight coarsening of grainsize as shown by the lithology of the low-tide muds which contain alternating layers of mud and sand. The low-tide muds then provide the base on which thin intertidal organic muds are deposited during the adventitious development of mangrove fringes on the open coast. Alternatively, chenier sands may accumulate over these low tide muds, producing a marked lithologic change. The deposition of overlying chenier sands, which are largely bioclastic, is subject to a special set of environmental conditions and will be discussed later.

Upward accretion of mudflats on the open coast restricts mangrove colonization to a narrow tidal inundation-defined band along the coast. As accretion continues, intertidal organic flats behind the mangroves gradually reach the level of spring high tides during the dry season. When this level is reached, the flats become too saline for most plants in the dry season and are covered only with algal mats in the wet season. Dessication of the algal-covered surface during the latter part of the dry season facilitates aeolian deflation of the surface, permitting an equilibrium surface to be maintained at a critical elevation. The upward accretion of the intertidal organic flats to become high-tide flats
CLAY DUNES
CHENIER RIDGES
OVERBANK DEPOSITS

CLAY FLUVIALLY TRANSPORTED SILT (SEDIMENTS FROM UPLANDS)
SAND WITH MUD FRACTION DOMINANT

UPLAND SURFACE
HIGH-TIDE MUDFLAT

Figure 7-1. The transport and dispersal of sediments on the chenier plain is initially dominated by fluvial factors. Sorting by wave processes occurs after terrigenous sediment enters the nearshore zone from estuarine channels. Mud remains in suspension, being redistributed along the coast and offshore.
illustrates the passive rather than active role of mangroves on this coast which is in contrast to their suggested role as landbuilders elsewhere (see Chapter 5). Although mangrove fringes may assist in sediment collection, their presence is very much controlled by sedimentary factors, which may change to those conditions inhospitable for mangrove growth or reproduction, causing massive die-back.

The high-tide mud facies is the uppermost marine deposited fine-grained facies on the chenier plain. This facies overlies the low-tide muds and on the older portions of the chenier plain, is probably derived from parent material previously deposited as subtidal muds. Chapter 3 demonstrated that scour and deflation processes alternate with seasonal deposition to maintain the high-tide flat at a consistent level with no regional tilt over tens of kilometers. Accretion on these flats results from both tidal and estuarine overbank processes with supplementation from upland overland flow. Ebb tidal scour and upland run-off processes move sediment seaward across the high-tide flats during the wet season. It appears that this scoured material is introduced directly from the mudflats into the nearshore zone and contributes to accretion in the intertidal and subtidal zones.

The development of clay dunes results from dry season deflation of the high-tide flats. Silt and clay, bound together by salt and algal fibers, are removed from the surface of the flat and piled into dunes by winds moving across this unvegetated surface. The dunes are sediment sinks in which silt and clay become stabilized by vegetation and are no longer available for redistribution over the plain, unless eroded by the migration of a tidal channel. Aeolian deflation and deposition appears to proceed independently of coastal progradation, with seasonally lowered sea levels permitting the drying of the flats which result in the mobilization of the surface sediments. A subsequent rise in sea level
during the next wet season allows further sedimentation and erosion of the high-tide flats, facilitating the long term maintenance of the mudflat surface at a regionally uniform level.

Chenier plain - sandy facies

Chenier ridges are composed of sands with up to 75% marine bioclastic carbonate. The balance consists of various terrigenous components. Bioclastic carbonate is a mixture of pelycope and gastropod detritus. The terrigenous fraction is fine quartz sand and silt although the oldest Holocene ridges contain pisolithic ironstone gravels because of their proximity to the upland surface and the probable addition of material swept landward during the Postglacial Marine Transgression. Chapter 3 demonstrated that the coarse terrigenous fraction is introduced as a fluvial bedload to the estuarine system during the wet season. However, once in the estuaries, tidal currents, assisted by high fluvial discharge during the wet season causes a net seaward movement of the coarse fraction. The estuaries basically serve as conduits for the upland sands and have no coarse estuarine facies other than channel bottom or lag deposits.

A sharp decrease of ebb current velocities at the mouth of most estuaries facilitates the development of ebb-tidal deltas whilst the form of the deltas depends in many cases on wave modification. At such deltas and in the adjacent shallow subtidal zones, wave action mixes terrigenous sands with marine shell material, moving this mixture away from river mouths and redistributing it in the shallow subtidal and intertidal zones. Elsewhere along the coast, under conditions which promote chenier development, wave action sorts a coarse fraction which is largely shelly material from the subtidal muds transporting this material shoreward into chenier ridges.
Terrigenous sand and marine shell material are predominantly redistributed during periods of reduced terrigenous input and greater wave activity. Chenier formation on a large scale takes place during periods when wave winnowing and onshore transport of coarse sediment is more effective than subtidal mud accumulation.

An example of coarse fraction sorting and redistribution may be observed on a small scale during the annual cycle of wet and dry seasons. This annual increase and decrease in sediment supply to the inshore zone of the Gulf which accompanies the alternation of wet and dry seasons is a short term model of the probable response to variations in sediment supply occurring over periods of 100-1000 years. Deposition of unsorted mud proceeds rapidly in the nearshore zone during high fluvial input of the wet season when cyclone induced waves are absent. When cyclones develop high wave energies (wave heights greater than 1.5m) recently deposited cheniers which are not stabilized by clay dunes or vegetation may become flattened on their crests or occasionally pushed back over the mudflats. However, field observations indicate that there is also a massive onshore movement of coarse material associated with some high energy events. Moderate to low wave energy (wave heights less than 1.5m) of longer duration occur during the dry season at a time of decreased sediment influx. These conditions result in removal and sorting of the coarse fraction from nearshore muds with a migration of sandy material onto the low-tide flats. The end of the dry season is marked by an increase in onshore winds due to the approaching monsoons. This increase in wave activity is not accompanied by increased sediment flux because the upland rivers have not yet responded to the early rainfalls. Therefore, sediment sorting proceeds at a high rate just prior to the onset of the wet season. There is onshore movement of shell and terrigenous sand producing the coarse layers which alternate with fine
layers in the low-tide muds. In addition, this late dry season sorting of sediment places coarse materials where they may be swept into chenier ridges during the high wave energy events of the wet season.

Once in the intertidal zone, whole and broken shells move as individual particles. They often pass through the mangroves with little addition to the intertidal organic muds, becoming incorporated into chenier deposits landward of the mangrove fringe. Finely comminuted shell and quartz sand moves landward in thin sheets or broad, low relief ridges. Movement of these sheets of coarse sandy material into mangrove communities helps promote die-back described in Chapter 3.

After the shell and terrigenous sand is deposited in a chenier ridge, there is little further transport of this material. Roll back of established cheniers over high-tide mudflats is limited, perhaps due to the stabilization of cheniers by the clay and silt dunes which quickly develop on their landward sides.

Chenier plain - progradation and sea level change

The effect of relative sea level change on a prograding coast may be studied on a chenier plain such as the Carpentaria example. In Chapter 6, on the basis of elevations of the low-tide mud/chenier-sand interface, it was established that the Carpentaria chenier plain has experienced a relative sea level fall of approximately 2.4 meters during the last 5500 radiocarbon years. The major question which may be posed from this observation is:

How did the prograding plain respond to such a change in relative sea level?

The study of modern depositional environments presented in Chapter 3 suggests that the level of the plain is continually adjusted by wet season inundation and sediment deposition in conjunction with dessication
and aeolian deflation during the dry season. Such a combination of conditions produces a bare mudflat which is algal covered and waterlogged half the year and hypersaline during the remainder. It appears that during the last 5000 years high-tide flats of the chenier plain have been deflated at their landward margin as progradation has progressed and sea level fallen.

Whether or not progradation in this region is independent of relative sea level fall is a question less easily answered. An answer may be sought by observing the effect of the annual rise in sea level associated with the wet season. Upward accretion during the high sea level of the wet season dominates during this brief 3-4 month period. It could be assumed that a similar response at a different rate would result during a transgression lasting tens to hundreds of years. All other parameters remaining constant, once the high-tide flats adjusted to the new sea level, progradation would continue at the seaward edge of the flat. However, nowhere on the eastern Gulf shoreline is there evidence for progradation during transgression suggesting that the Postglacial Marine Transgression exceeded the maximum upward accretion rate of the coastal plain. Since all subsequent relative sea level change in the Gulf has been regressive, the response of this coast to transgression remains undefined unless examples from outside the Gulf of Carpentaria are utilized. Several other coasts have experienced either a relative stillstand or slow rise in sea level during the upper Holocene and these coasts have exhibited progradation. Some of these examples (Mississippi, Nayarit, and Tabasco) are shown in Table 7-1 which indicates that coasts may prograde under a variety of conditions including submergence when sediment supply is sufficient.
Table 7-1

<table>
<thead>
<tr>
<th>Gulf of Carpentaria</th>
<th>Karumba</th>
<th>Edward R.</th>
<th>Mississippi</th>
<th>Nayarit</th>
<th>Tabasco</th>
<th>New South Wales</th>
<th>Surinam</th>
<th>Schvenengen (Galveston Island model)</th>
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</thead>
<tbody>
<tr>
<td>Wave energy</td>
<td>Low</td>
<td>Low</td>
<td>Low</td>
<td>low- moderate</td>
<td>low- moderate</td>
<td>moderate-high</td>
<td>low- moderate</td>
<td>moderate-high</td>
</tr>
<tr>
<td>Mean spring tide</td>
<td>3.0 - 3.5 m</td>
<td>2.5 - 2.8 m</td>
<td>0.5 - 0.75 m</td>
<td>1.0 - 1.25 m</td>
<td>0.4 - 0.5 m</td>
<td>1.5 - 2.0 m</td>
<td>2.8 m</td>
<td>2-3.0 m</td>
</tr>
<tr>
<td>Sediment type(s)</td>
<td>mud/shell</td>
<td>mud/shell</td>
<td>mud/shelly sand</td>
<td>sand</td>
<td>mud/shelly sand</td>
<td>sand</td>
<td>mud/shell</td>
<td>sand/mud</td>
</tr>
<tr>
<td>Sediment source</td>
<td>fluvial, seasonally acid from mudstone and silicic uplands</td>
<td>fluvial, mixture of silt/clay/sand at delta</td>
<td>fluvial, fluvial, and fluvial sedimentary uplands</td>
<td>fluvial, continental, relict transgressed, with fluvial supply</td>
<td>fluvial, continental, relict transgressed, with fluvial supply</td>
<td>fluvial, continental, relict transgressed, with fluvial supply</td>
<td></td>
<td></td>
</tr>
<tr>
<td>Coastal morphology</td>
<td>chenier plain</td>
<td>beach-ridge plain</td>
<td>beach-ridge plain</td>
<td>beach-ridge plain</td>
<td>sand barriers with and without beach ridges and dunes</td>
<td>Cheshire</td>
<td>prograded barrier and dune</td>
<td>prograded barrier islands</td>
</tr>
<tr>
<td>Holocene sea level</td>
<td>relative fall since 5500 years B.P.</td>
<td>relative fall since 5500 years B.P.</td>
<td>relative rise</td>
<td>slow relative rise</td>
<td>relative rise</td>
<td>stillstand since 6000 years B.P.</td>
<td>relative stillstand</td>
<td>slow relative rise to present</td>
</tr>
<tr>
<td>Climatic change</td>
<td>yes - probably precip. related</td>
<td>yes - probably precip. related</td>
<td>indeterminate but shifting level of discharge most important</td>
<td>yes - effective runoff and coastal winds</td>
<td>undetermined but evidence exists for changing level of fluvial discharge</td>
<td>yes - probably regional variation in storms and extreme conditions</td>
<td>indeterminate but some evidence exists for varying fluvial discharge</td>
<td>yes - inferred from N. Europe paleobotany record</td>
</tr>
<tr>
<td>Annual fluctuation of sea level</td>
<td>yes - 1-2 m</td>
<td>yes - 1-2 m</td>
<td>yes - due to storm surges</td>
<td>not reported</td>
<td>yes - &lt; 1 m</td>
<td>no</td>
<td>yes - due to onshore winds</td>
<td>no</td>
</tr>
<tr>
<td>Offshore gradient</td>
<td>1:4000</td>
<td>1:2000</td>
<td>1:1000</td>
<td>moderate</td>
<td>moderate</td>
<td>steep</td>
<td>moderate</td>
<td>moderate-steep</td>
</tr>
</tbody>
</table>
The chenier plain - summary and sediment budget

Traditional explanations of chenier plain development (Chapter 1) indicate that cheniers form when sediment flux decreases whilst mudflats prograde when sediment flux increases in the form of unsorted muds. Chapter 4 reinforced this interpretation in the context of findings from the Gulf of Carpentaria. Furthermore it has been shown that fine unsorted sediment enters the Gulf in large quantities and that the morphostratigraphy and age structure of the coastal plain established such sediment influx as operating in discrete episodes during the last 5000 years. Episodes of either mudflat progradation or chenier building may be coupled to episodes of maximum or minimum sediment flux respectively. A graph of such oscillations in sediment supply is shown in Figure 7-2. The block diagrams which accompany the graph illustrate the sedimentary response to various levels of terrigenous input. In addition the figure shows that maximum and minimum terrigenous sediment flux are merely end members on a continuum. Both cheniers and mudflats form with conditions in between these limits, a consideration which is important when evaluating the statistically grouped episodes of chenier and beach ridge development which were presented in Chapter 5 and will reviewed further in this chapter.

Beach ridge plain - general

The beach-ridge plain, although narrower, and at some locations, less continuous than the chenier plain, also exhibits discontinuities which suggest discrete episodes of ridge-building activity. These discontinuities are shown only in the sandy facies, a subaerial prism of sand several meters thick which overlies the muddy facies. Muddy facies on the beach-ridge plain are composed of fine material similar in lithology to the muds of the chenier plain. Sandy beach-ridge facies
Figure 7-2. Progradation of the chenier plain is related to relative rates of sediment flux. This flux varies through time producing alternating chenier and mudflat modes.
differ from the chenier facies, by being composed of mostly terrigenous quartz sands with a subsidiary bioclastic component.

Beach-ridge plain - muddy facies

Clay and silt are introduced by the fluvial system to the estuarine portion of the beach-ridge plain. Estuaries serve as conduits (Chapter 3) carrying terrigenous material across the coastal plain and into the nearshore system of the Gulf. Most estuarine channels are only a few kilometers in length and depositional environments are much more limited in extent and number than on the chenier plain (Chapter 3).

Subtidal and low-tide muds are facies essential to coastal progradation on the beach-ridge plain. Chapter 3 demonstrated that rapid rates of sedimentation have been observed in the subtidal and low-tide muds during the wet season. The low-tide muds were shown to be in part an intertidal extension of the subtidal muds and a sharp lithologic boundary separates beach-ridge sands from the underlying muds.

Other mud facies, and especially high-tide muds are less important to the process of coastal progradation on the beach-ridge plain than on the chenier plain. High-tide muds occur mainly between the Holocene and Pleistocene beach ridges and may in part represent a lagoonal facies. Aeolian processes were shown to be unimportant on this portion of the Carpentaria coast, with foredune sand dunes and high-tide flat clay dunes being poorly developed.

Beach-ridge plain: sandy facies

Terrigenous sands on the beach-ridge plain are introduced to the nearshore system along a path similar to the muds. Fluvial processes carry sands into the estuarine system in which they move as bedload of the tidal channels toward the Gulf. At the mouth of estuaries, ebb-tidal
deltas are as important to the distribution of sands as they are on the chenier plain. The terrigenous fraction of tidal delta sands is coarser than the similar fraction of modern beach-ridge sands. This observation, as well as the presence of coarse well-sorted sands offshore of major rivers suggests that there is a separation of grain size which occurs in the tidal deltas. Ebb tidal deltas have a medium grain size which is between the size of coarser fluvial sands and the finer beach sands. However, offshore sands in depths of 20 meters or more and in the shallow subtidal areas immediately seaward of the major rivers are the coarsest and best sorted sands.

Identification of major ridge forming mollusc shells (Chapter 3) indicates that these organisms live in the shallow subtidal zone. They exhibit little evidence of wear or abrasion when found on the beach face, indicating distance of transport has been minimal, although the soft muddy character of subtidal muds may minimize abrasion during onshore transport. Bioclastic contribution to beach-ridge formation is variable (5-70% shells). There are several modern examples which show that the beach-ridge plain may prograde with minimal bioclastic contribution, especially near the mouths of large rivers.

Beach ridge accretion in the Gulf of Carpentaria appears to occur as successive draping of laminae over an initially flat lying base of fine sand on the low-tide flat. In Chapter 3, it was shown that such lamination of Carpentaria ridges is similar to the internal structure of beach ridges reported by Psuty (1966) in the Tabasco, Mexico. The major difference between the Carpentaria and Tabascan ridges is the dominance of seaward dipping beds in the upper portions of the Carpentaria ridges whilst the Tabascan ridges exhibited mostly crossbedding in this portion of their structure. Beach ridges in Nayarit, Mexico as described by Curray et al (1969) show little lamination but have aeolian caps, a
feature which is not developed on the Carpentaria ridges. The dominance of seaward dipping beds in the Gulf is also in strong contrast to the description of beach sections given by Jennings and Coventry (1973) for the barriers in the Fitzroy Estuary, W. Australia. According to Jennings and Coventry, these beaches are dominated by topslope and backslope (landward dipping) beds. The disparity between gravelly Fitzroy and sandy Carpentaria ridges may be a result of magnitude of the depositional event required to transport the two different types of strandline material. The Fitzroy ridges appear to be products of moderate to high energy events which deposit stratified gravel on the top and backslope of existing ridges, whereas Carpentaria ridges appear to develop during the entire spectrum of energy levels. Carpentaria ridges exhibit deposition whenever there is sufficient wave runup to transport sandy material as far as the ridge crest.

Discussion of modern beach-ridge formation in Chapter 3 suggested that after horizontally bedded sands developed on the low-tide flat, the ridge continued to build upward for approximately one-third of its height with crossbedded and seaward dipping beds. This portion of the ridge structure is probably emplaced as a series of low amplitude ridges migrating across the low-tide flat and onto the flat-lying basal sands. Such ridge migration can be observed during the annual dry season when sediment supply is reduced and limited winnowing dominates for 2-3 months. The upper two-thirds of the modern ridge consist of seaward and landward dipping beds, some of which are continuous over the crest of the ridge. These draped beds are interpreted as deposits related to moderate wave energy levels which overtop the ridge crest but do not cause erosion on the front. Unconformities noted in the ridge section probably result from high wave energy events which erode the seaward portion of the ridge, but which may infill the swale with overwash sediment. Observation of
beach behavior during cyclone Ted in November 1976 suggests that even during such catastrophic events accompanied by raised water levels, there appears to be a dominant landward movement of beach sands. This includes removal of sand from the seaward portions of the ridge.

A major feature of Carpentaria beach-ridge plains when viewed in plan is the apparent stability of inlets throughout most of the progradational history of the plain. This stability is attributed to the formation of calcite cemented beach-ridge rock. Relict ridges older than 4000 yrs B.P. are leached of bioclastic carbonate except near their bases, making it difficult to compare the original shell content of relict ridges with modern ridges. However, the limited examples where original lithology is preserved in mid-Holocene beach-ridge rock suggests that these older ridges contained as much if not more shell than modern ridges.

Beach-ridge plain: progradation and sea level change

The beach-ridge plain and the chenier plain have experienced similar relative sea level histories during the last 6000 years. Data presented in Chapter 5 showed that whilst the chenier plain was subjected to a 2.4 meter fall in relative sea level. The beach-ridge plain experienced a fall of 1.5 meters over this same time period. Furthermore, discussion of modes of progradation associated with the chenier plain indicated that the relative dominance of chenier or mudflat deposition is regulated by sediment supply and is not dependent on relative sea level fall for rapid progradation. A parallel response to relative increases in sediment supply has been demonstrated for the beach-ridge plain. The argument used to support the concept that the chenier plain progrades independently of sea level change may be applied equally well to the beach-ridge plain.
Beach-ridge progradation is most responsive to sediment supply. Provided sufficient terrigenous input is maintained, any temporary coastal erosion caused by a small transgression is followed by further progradation, once the depositional environments have adjusted to the higher mean sea level. Since progradation on both plains appears to be governed by relative changes of terrigenous sediment input, it seems appropriate to consider relative sea level change as a factor which may accentuate or retard progradation rates, whilst remaining preserved in the stratigraphy of the prograded plains particularly the basal mud facies.

Beach-ridge plain: summary and sediment budget

The mode of coastal progradation on the beach-ridge plain has remained relatively uniform during the last 6000 years. The important contrast with the chenier plain is that the beach-ridge plain requires both terrigenous mud as well as terrigenous sand to prograde whilst the chenier plain requires only mud. This condition is certainly an effect of upland lithology of respective source areas. The beach-ridge plain is supplied by rivers which originate in the granitic highlands of eastern Cape York and which flow across sandy upland surfaces toward the Gulf. The terrigenous mud and sand components which reach the Gulf are in relatively equal quantities providing a basis for terrigenous beachface progradation to which bioclastic carbonate is added in varying quantities.

The present annual cycle of nearshore and beach processes may be used as a model for longer term fluctuation in sediment supply. Presently, at the end of the wet season, there is an oversupply of terrigenous material in the nearshore zone, a condition which initially promotes accretion on the low-tide flat and later provides material for
winnowing of a coarse fraction. Onshore movement of this coarse fraction continues during the dry season, therefore allowing shoreward movement of sand and concomitant beach face accretion to proceed after the terrigenous input of the wet season has ceased.

Similarly, a long term decrease in terrigenous supply would be followed by initial winnowing and shoreward movement of the coarse sandy material over the muddy low-tide flat. However, if the relative level of terrigenous input continues to decline, the long term result is a diastem with respect to beach-ridge development, a condition which appears several times in the age structure of the plain. Figure 7-3 summarizes this sequence of responses. It should be noted that the proposed mode of response to fluctuations in sediment supply to the beach-ridge plain differs considerably from the chenier plain. Oversupply of terrigenous material on the beach-ridge plain promotes strandline development whilst a similar oversupply on the chenier plain results in mudflat accretion (Figure 7-2). This difference in response modes has already been illustrated by comparison of the age structure of the two plains in Chapter 5 where it was shown that strandline formation on the beach-ridge plain appears contemporaneous to mudflat development on the chenier plain.

Modes of deposition - a summary

Chapter 1 reviewed the findings of other workers concerning the genesis of both chenier and beach-ridge plains. Despite divergence of opinion concerning the exact process by which chenier plains prograde, there was general agreement that some variation in the supply of fine-grained sediment determines whether a mudflat or a ridge is constructed. Theories of beach-ridge genesis are spread along a spectrum between the high energy event and the frequency-magnitude hypothesis,
with no general consensus on beach-ridge development.

Study of the Gulf of Carpentaria prograded coastal plains has shown that there is a predictable response to variation in sediment supply on both types of coast (Figure 7-3). Modes of response may differ greatly. In the case of the chenier plain, the formation of strandlines is associated with a relative decrease in terrigenous supply. Furthermore, the level of sediment supply has been postulated as variable throughout the upper Holocene and this variation may have a forcing function which is external to the marine system. Figure 7-2 demonstrated that during periods of abundant sediment supply, the construction of strandlines (chenier ridges) is actually suppressed by this oversupply of terrigenous material. Such an oversupply inhibits the sorting and winnowing processes which are responsible for the deposition of coarse-grained facies which comprise chenier ridges.

The Carpentaria example also suggests that traditional theories of beach-ridge genesis which are wave energy dependent, may be incorrect in their emphasis. The frequency of ridge construction and the total width of the prograded plain are primarily regulated by the abundance of material for ridge construction. Although the geometry of the individual ridges may be at least modified by the wave energy parameter, this factor should be considered as secondary to the major one of sediment supply. Response to increased sediment supply is opposite to that of the chenier plain. Increased sediment to the beach-ridge plain permits the construction of strandlines (beach ridges) whilst a decrease causes cessation of progradation and ridge building.

Finally, the role of relative sea level fall is probably less important than either of the above two factors provided any such change occurs over the time span measured in the Gulf of Carpentaria. Evidence from both the chenier and beach-ridge plain suggests that relative sea
Figure 7-3. The beach-ridge plain progrades by a different mode from the chenier plain. In this diagram, coastal plain response is compared under conditions of sediment abundance and scarcity.
level change has accentuated progradation in both cases, but has not been a primary cause.

There have been frequent attempts during the last decade, in the fields of coastal geomorphology/geology, to categorize coastlines of the world on the basis of physical parameters such as present climate, wave energy, tidal range, recent change in sea level, coastal morphology and sediment type. When the rationale of such categorizations are considered in the light of the Carpentaria model for coastal progradation, it becomes apparent that only a limited number of factors may actually have an effect on coastal depositional morphology. Table 7-1 presents a review of several other examples of prograded coasts in comparison to the Gulf of Carpentaria. This review demonstrates: (1) coasts prograde independently of several physical parameters which include tidal range, wave energy, and relative sea level change; (2) modes of coastal progradation appear regulated by sediment type, sediment source and offshore gradient; and (3) climatic change in some form appears to have affected progradation on many of these coasts but its role remains difficult to determine.

Such factors as tidal range, level of wave energy and relative sea level change appear as minor contributors to the processes of coastal progradation. It cannot be denied that wave dominated coasts incur sediment transport systems which may differ from those of a tide dominated area. However, although the mode of delivery to the shore face may differ on these two types of coast, progradation may occur regardless of the tidal or wave regime. Earlier portions of this Chapter discounted the Holocene sea level history of the Gulf of Carpentaria as contributing greatly to the mechanics of progradation. Comparison with the other examples of prograded coasts confirms that whether the coast is submergent or emergent is less important than the factors described below.
The factors which are most important in regulating coastal progradation appear to be sediment type, sediment source, and offshore gradient. Sandy shelf sediments when associated with steep offshore gradients have resulted in prograding sand barriers of which most eastern Australian barriers are a good example. In the presence of abundant fluvially supplied sediments and low offshore gradients, coastal types subdivide into chenier plains and beach-ridge plains with chenier plains linked to a dominant input of fluvial mud and the presence of some shells and/or sand for chenier formation.

The role of climatic change is a parameter which appears highly important in regulating the rate and episodic nature of coastal progradation. Although this parameter may not govern morphology, it may initiate and/or interrupt progradation. Unfortunately, climatic change can only be inferred from proxy data gathered in the coastal zone, and more often from continental sites, with the resulting model from such data conceived to fit the observations rather than being tested by the observations.

PERIODICITY IN PROGRADATION

There is a diachronous relationship between periods of chenier and beach-ridge formation. It was demonstrated in Chapter 5 that this relationship could be explained in terms of two differing sets of conditions necessary for the development of ridges and cheniers. A further test of this explanation may be performed by examining the age structure of another radiocarbon dated beach-ridge plain in the Gulf of Carpentaria.

Smart (1976) presented 25 radiocarbon observations on regressive ridge facies from Cape Keer Weer. Inspection of these 25 determinations,
which have been corrected for reservoir effect (Table 7-2), shows a range over the 4 episodes of progradation determined for the beach ridge plain at further south Edward River. However, Smart noted that limited datable material is available from the oldest Holocene beach ridges (QHM4 in his nomenclature) and presented only two dates older than 4000 years (SU 183 and SU 201A). Since it is probably inappropriate to establish a grouping based on two dates of such age, these observations were removed from the list prior to application of splitting techniques (Chapter 5). The resulting splits, and associated episodes of ridge formation are shown in Figure 7-4. Comparison between the age structure of two independently studied beach-ridge plains suggests synchronous response over 120 kilometers of coast.

Unfortunately, no other chenier plain on the Gulf of Carpentaria has been subjected to detailed age determination. However, Cook and Mayo (1977) present an age structure for a chenier plain adjacent to the Broad Sound estuary, approximately 120 km north of Rockhampton, Qld and 500 km south of the Carpentaria study area (Figure 5-1). Dates from the chenier ridges at Broad Sound were adjusted for reservoir effect, then laboratory replicates were discarded, resulting in the list presented in Table 7-3. It is apparent from inspection of this table that ANU-907 requires omission for the same reason the two oldest dates from Smart (1976) were disqualified. The resulting 12 determinations show a grouping which is somewhat comparable to Carpentaria results (Figure 7-4). The two episodes in the 1500-2700 B.P. time range could be an artifact of the limited number of observations (two) on which the younger of these two groups is based. This younger group may therefore be an incompletely sampled extension of either a neighboring group or some independent group. Despite such disparity, the building of cheniers during a modern episode and an episode centered about 3500 B.P. offer favorable
Table 7-2

<table>
<thead>
<tr>
<th>Lab No.</th>
<th>Beach Ridge Group</th>
<th>Corrected Age B.P. (reservoir effect)</th>
</tr>
</thead>
<tbody>
<tr>
<td>SU 197A</td>
<td>Qhm3</td>
<td>20 90</td>
</tr>
<tr>
<td>NSW 157</td>
<td>Qhm3</td>
<td>50 90</td>
</tr>
<tr>
<td>SU 198A</td>
<td>Qhm4</td>
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</tr>
<tr>
<td>SU 184</td>
<td>Qhm4</td>
<td>370 80</td>
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<td>Qhm4</td>
<td>460 80</td>
</tr>
<tr>
<td>SU 198B</td>
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<td>530 80</td>
</tr>
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(modified from Smart, 1976)

Table 7-3

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(modified from Cook and Mayo, 1977)
Figure 7.1. Radiocarbon data from Cape Keer Meer and Broad Sound are subjected to the same grouping techniques as the Carpinteria results. Cape Keer Meer prograded in a manner synchronous to the beach-ridge plan further south. Comparison with Broad Sound suggests limited agreement with Carpinteria results.
comparison with similar Carpentaria episodes. This may suggest a regional response throughout North Queensland to some trigger mechanism. Even if the tenuous Broad Sound comparison is omitted, there is a sound case for synchronous episodes of terrigenous supply to the Gulf of Carpentaria separated by periods of relative sediment scarcity.

Certain prograded coasts such as those of Louisiana, the Gulf of California, the Nayarit and Tabasco plains, Mexico, and perhaps even the Niger delta, experience variable sediment supply by diversion of deltaic channels or by alteration of longshore drift patterns. However, an arcuate coast adjacent to an epicontinental sea such as the Gulf of Carpentaria does not permit variation of sediment supply by merely altering discharge location or longshore drift pattern. Sediment enters the Gulf via discrete estuarine channels which are 20-40km apart (Figure 1-2). Each major estuary is an extension of an upland river system, many of which are ephemeral, but all of which respond to a regional set of climatic conditions. It has been shown above that this variation in supply during some stages affects widely separated areas concurrently (Figure 7-4). Variation must therefore occur in the primary source of terrigenous material to the coast - the fluvial system.

It has been argued in Chapter 4, and to a greater extent in this chapter, that coastal progradation has proceeded over a wide geographic area of North Queensland in a somewhat synchronous manner. If stage CM3 on the chenier plain and BR3 on the beach-ridge plain (Figure 7-5) are taken as the most recent examples of coastal progradation by two differing modes at two separate locations, then some cause for this regional response must be sought.

Stage CM3 on the chenier plain was an episode of mudflat progradation which implies an abundant supply of muds from the uplands. Mudflat development proceeded because the supply of material to the
nearshore zone exceeded the capacity of waves and currents to sort it into a fine and coarse fraction (see discussion of Todd, 1968 in Chapter 1).

Stage BR3 on the beach-ridge plain was an episode of beach-ridge progradation which implies that sediment supply from the uplands was in the form of mixed mud and sand sufficient for long-term accretion of many ridges.

Both of these stages were followed by periods of reduced sediment supply. On the chenier plain, chenier ridges were developed by the sorting and redistribution of a coarse fraction by wave activity. On the beach-ridge plain, the age structure suggests a brief halt in beach-ridge development (Chapter 5 and Figure 7-5).

Variations of sediment input by fluvial systems such as those described above may be achieved by several mechanisms, most of them associated with altering either rainfall or the watershed response to rainfall. In the latter category are responses to agricultural land use, man-made diversions or sediment traps, or deforestation by fire. However, episodic sediment supply has been shown to be not only pre-agricultural, but pre-historical as well. Although evidence is available (Tindale, 1978) to show that man occupied Carpentaria watersheds during this period, it is highly unlikely that Aboriginal man significantly altered the sediment supply even by extensive burning off.

The remaining option appears limited to some sort of secular variation in precipitation. Such variations fall into two broad categories:

a. change of net annual precipitation
b. changed intensity of precipitation without significant net annual increase.
Although Simpson and Doutch (1977) found Carpentaria rivers "underfit" and Grimes and Doutch (1978) postulated widespread periods of aridity followed by pluvials throughout the Karumba basin during the Quaternary, there is even stronger evidence of a biogeographic nature to suggest fluctuation in rainfall continued into the Holocene in North Queensland. Using palynologic evidence from lake sediment cores, Kershaw (1970, 1971, and 1975) clearly showed that variations occurred in rainfall on the Atherton Tablelands (Figure 5-1) during the upper Pleistocene and lower Holocene. However, complications are inherent in extending Atherton Tableland models (which contain both a summer and winter rainfall component) across North Queensland to the Karumba Basin which receives only a summer (wet season) rainfall. Such a limitation does not preclude the development of a hypothesis concerning Holocene changes in Carpentaria sediment supply.

The unique geomorphic conditions of the Gulf which include its conservation of sediment within a semi-enclosed basin provide a system sensitive to small but sustained changes in rainfall. The seasonal aridity of the uplands further amplifies geomorphic response to variability in precipitation as pointed out by Kershaw and Nix (1975).

Variability in precipitation is most obvious in arid and semi-arid areas where intrinsic environmental instability leads to variations in rates and kinds of sediment accumulation and where deposition is punctuated by periods of erosion. (p.1)

Rainfall intensity is one means by which precipitation may vary without causing a feedback factor on vegetation. Carson and Kirkby (1972) qualitatively argued the importance of rainfall intensity with respect to geomorphic processes. In their discussion of surface water erosion, they say:

The distribution of rainfall intensity is also an important factor in determining the rate of soil erosion. Intense rain contains more large drops which are more effective in compacting and
dispersing the soil, and intense rain is also more likely to exceed the infiltration capacity of the soil and so produce Horton overland flow (Carson and Kirkby, 1972, p 212).

Early studies on soil erosion considered agricultural aspects which included factors such as slope, vegetation, soil type and land use. Musgrave (1947) was among the first to include a rainfall intensity factor in his soil loss equation. This equation which later became known as the Musgrave equation was widely accepted for use in watershed conservation studies. The significance of the Musgrave equation is that erosion was directly related to maximum 2-year recurrent 30-minute rainfall raised to the 1.75 exponent, thereby demonstrating the importance which Musgrave accorded the intensity factor. Wischmeier and Smith (1965) developed an explicit empirically based predictive model called the "universal" soil-loss equation which remains accepted by many soil scientists. Of particular interest is the comment by Wischmeier and Smith that:

The data showed that a rainfall factor used to estimate overall annual soil loss must include the cumulative effects of the many moderate-size storms as well as the effects of the occasional very severe ones. (Wischmeier and Smith, 1965, p.3).

These authors show that when other factors are held constant storm soil losses are directly proportional to the product of the storm's total kinetic energy and its maximum 30-minute intensity. Annual measures of the rainfall factor are achieved by summing computed storm values throughout the year.

This brief review of soil-loss equations suggests that both net annual rainfall as well as intensity play an important role in introducing sediment to the fluvial system. Previous soil-loss studies have been calibrated to strict agricultural situations and therefore are not calibrated to the Carpentaria catchment. Consideration of the effects of variation in rainfall intensity on the Carpentaria watershed parallels discussion in Chapter 6 that the recurrence interval of
moderate and especially high energy events play an important role in
governing rates of change and the character of the morphostratigraphic
record. Although the present context is different from the example cited
for beach-ridge deposition, it is apparent that any alteration of
frequency magnitude relationships which persist for periods of 300-1000
years may grossly alter rates of landform modification.

Although either increased precipitation and/or storm energy offers a
mechanism by which sediment input may be increased, there remains some
uncertainty concerning the primary cause of such variations, especially
variations of the episodic type noted in the Gulf sediments. To explain
similar episodic activity noted in beach-ridge progradation and sand dune
activity of the New South Wales coast, Thom (1978) has postulated a
secular variation in storminess for the Tasman Sea associated with
increased extra-tropical cyclogenesis. That such increased cyclogenesis
could extend to the tropical areas in Northern Australia is a possibility
which will considered in the final section of this chapter.

GEOLOGIC HISTORY

Postglacial Marine Transgression

the Postglacial Marine Transgression flooded the Gulf of Carpentaria
over the Wessel Rise, a bathymetric high which separates the Gulf from
the Arafura sea. Both the form and rate of this transgression into the
Gulf of Carpentaria is largely unknown except for data presented by
Phipps (1970) and Smart (1977). Phipps presents evidence from bottom
cores containing datable material and fossil assemblages suggesting that
the Gulf of Carpentaria was a brackish or non-marine continental
environment prior to complete flooding. Furthermore, the model
synthesized by Phipps shows that the floor of the Gulf may have risen at
approximately the same rate as the Post Glacial Marine Transgression until approximately 6000 years B.P. at which time it was depressed and flooded. Smart uses the data presented by Phipps in combination with information gathered by mineral exploration programs to present a slightly different view. In Smart's model, the Gulf flooded by a transgressing sea similar to the curve presented by Bloom et al (1974) and Thom and Chappell (1975) but with the possibility of minor regressions superimposed on the trace of the major transgression. Despite these uncertainties concerning the upper Pleistocene-lower Holocene sea level record, there is unequivocal evidence to show that relative sea level reached and passed above present level prior to 5500±300 years B.P. in the eastern and southern Gulf.

The surface transgressed by the rising seas shows evidence of having a well developed soil zone with carbonate nodules, pisolithic ironstone and highly pitted quartz grains in a compact dark brown clay matrix. This soil, called the Carpentaria paleosol in this study, underlies the Holocene marine deposits and in the absence of a basal transgressive unit, is used as a distinctive stratigraphic marker indicating the Holocene-Pleistocene unconformity. The paucity of coarse clastic materials in the original Carpentaria Soil in conjunction with its semi-indurated character may explain the absence of a basal transgressive sand beneath the chenier and beach ridge plains. Where sand deposits were present on the pre-Holocene surface, it appears that winnowing processes left coarse lag deposits. Sands in water depths of 20m or more probably lost their fine fraction during the transgression whereas shallower deposits continued to be winnowed under modern conditions.

It has been shown in the previous section of this chapter that certain progradational processes appear to have operated in an episodic
manner during the middle to upper Holocene. Although these episodes are generally correlative between the chenier and beach-ridge plains, there is not complete agreement in all episodes at a regional level (Figure 7-5). This may be due to either incomplete sampling which yields an age structure not reflective of the true history, or some other explanation such as a lag factor between the northern and southern portions of the study area.

It is not the purpose of this section to further examine possible explanations for a diachronous relationship. The lack of complete agreement at a regional scale is offered as a justification for separating discussion of the Holocene geologic histories of the two plains. This separation, although still permitting regional comparison, facilitates discussion of episodic progradation, without confronting some of the interpretative problems which may only be due to incomplete information.

Chenier plain.

6000 to 5500 years B.P.

Coastal progradation was initiated by mudflat development on the chenier plain sometime at or near 6000 years B.P. (Figure 7-5). This stage of progradation is difficult to date because of the lack of carbonate material in datable quantities in the oldest mudflats. Mud deposition in-filled the irregular embayments formed by the drowning of the pre-Holocene topography. The formation of wave cut scarps into some laterized headlands may have occurred at this stage. However, the possibility that these features are inherited from a Pleistocene interglacial sea level cannot be ignored.
Figure 7-5. Correlation of progradational stages on the chenier and beach-ridge plain indicate nearly synchronous response during some episodes. Modelling strategies require a large number of normally distributed observations. The data on which the Carpentaria response model is based may be insufficient to fully describe the upper Holocene climatic history.
5500 to 4500 years B.P. -(Stages CR1 and CM1)

The transgression slowed or completely halted by the beginning of this stage. Relative sea level reached a high of 2.4 meters above present began falling toward the end of the stage, perhaps in response to hydro-isostatic loading of the continental shelf (Figure 6-3). A shoreline of chenier ridges extended east-west along the prograded flats, forming a smooth arcuate shoreline broken only by eight major rivers (Figure 4-1). Adjacent to the mouths of the larger rivers, recurved spits marked the limit of the transgression into the drowned valley topography. These landward curving, mid Holocene shorelines are especially noticeable adjacent to the major rivers such as the Kendall, Mitchell, Norman and Leichhardt (Figures 4-1 and 4-8). The construction of cheniers during this period, which appears to have been one of rapid terrigenous mud input, was probably more an effect of coarse clastic material swept shoreward by the earlier transgression and less a result of the more conventional winnowing processes postulated in the model for this type of coast. Oldest cheniers have a high sand and gravel content with considerable pisolitic ironstone. Material of this type is likely to be reworked landward by the transgression. These cheniers are narrower and less continuous than younger strandlines, suggesting that strandline construction was not dominant during this period when compared to the 4800 meters of mudflat progradation shown on the Karumba transect during stage CM1.

4500 to 2900 years B.P. (Stage CR2)

Chenier development dominated during this period with a limited progradation of mudflats. Apparently little fine terrigenous material was available for mudflat progradation, whilst winnowing processes and biogenic carbonate production continued. The limited mudflat
progradation which is expressed by the narrow separation of chenier
ridges near the Karumba transect (Figure 5-1) was probably enhanced by a
continuing relative fall in sea level. In contrast with the earlier
mudflat progradation stage CM1, only 1700 meters of mudflat on the
Karumba transect, can be attributed to this period (Figure 5-8).

2900 to 2000 years B.P. (Stage CM2)

The chenier plain underwent further progradation of mudflats in this
stage with a marked absence of chenier formation. Mudflat progradation
totalled 4200 meters on the Karumba transect. Such rapid progradation
without the formation of cheniers may have been somewhat accelerated by
the 0.7m relative fall in sea level during this stage.

2000 to 1700 years B.P. (Stage CR3)

A brief lapse in the availability of fine sediment during this period
was marked by the winnowing of the coarse fraction from the prograded mud
and the associated development of cheniers. Chenier building appears to
have ended abruptly with the return to mudflat progradation by 1700 years
B.P.

1700 to -1300 years B.P. (Stage CM3)

A continuing relative fall in sea level in combination with a high
rate of terrigenous input accentuated progradation rates during this
third stage of mudflat development on the chenier plain. Over 6000
meters of mudflat appear to have developed on the chenier plain at this
stage (Figure 5-4). At the close of Stage CM3, relative sea level was
approximately 0.5 meters above present.
1300 years B.P. to present (Stages CR4 and CM4)

Winnowing processes appear to have dominated for this final stage, suggesting a decrease in the availability of fine sediment. However, the presence of approximately 500 meters of prograded flat seaward of modern cheniers may indicate the trend toward another stage of mudflat progradation. Relative sea level reached its present level at or prior to 600 70 years B.P.. The form of the relative sea level curve between then and the present is somewhat in question due to the lack of very young cheniers on the Karumba transect.

Beach-ridge plain

6000 to 4800 years B.P. (Stage BR1)

As the Postglacial Marine Transgression reached its maximum on the chenier plain, immediate ridge development and progradation occurred. Relative sea level was 1.5 meters above present (Figure 6-6), remaining stable until shortly after 5000 years B.P.. Progradation took the form of closely spaced beach ridges which in some cases were welded against the Last Interglacial strandline. Sand supply from both offshore and upland sources permitted rapid outbuilding of the coast with a total of 3200 meters of progradation on the Christmas Creek and Edward River transects.

4800 to 3400 years B.P.

The beach-ridge plain experienced a period of nondeposition between Stages BR1 and BR2. This observation is in accordance with a relative decline in the amount of sediment available for mudflat progradation in the chenier plain during approximately the same period (Figure 7-6). It was demonstrated in Chapter 5 that such periods of non-deposition were not always represented by morphologic discontinuities, assuming that the
Age determinations correctly describe the age structure of the plain. Further inspection of Figure 5-5 indicates that the shell/non-shelly boundary on the Christmas Creek transect is close to the inferred position of the diastem, which might be associated with this non-deposition. Furthermore, if ANU 2060 is anomalously old, there is an obvious plan discontinuity of approximately this age across the Edward River transect. Figure 6-6 shows that sea level continued to fall during this period, a trend which can only be inferred from the dates on beach ridges older or younger than this age. It is clear from Figure 6-6 that by 3400 years B.P., relative sea level on the beach-ridge plain had fallen to within 0.8 m of its present position, approximately one-half of the total upper Holocene change.

3400 to 2450 years B.P. (Stage BR2)

Beach-ridge progradation resumed shortly after 3400 years B.P. and continued for almost 1000 years. Total progradation along both transects sum to over 2000 meters for Stage BR2. If the postulated regional response to increased terrigenous input is viable, then it is likely that Stage BR2 correlates with CM2 on the beach chenier plain.

2450 to 2300 years B.P.

This is a brief period of non-deposition on the beach-ridge plain, inferred from a marked statistical split between radiocarbon observations which belong respectively to Stages BR2 and BR3. Beach-ridge morphologies appear contiguous during this portion of their age structure. A diastem at this time in the development of the beach-ridge plains is equivocal without further dating of ridges assumed to be older and younger than this period.
2300 to 500 years B.P. (Stage BR3)

During the 1750 year time span of Stage BR3, the two beach-ridge plain transects exhibited approximately 2500 meters of progradation. Relative sea level was continuing to fall and probably reached present level before the end of this stage. Certainly, falling relative sea level may have accentuated progradation rates, but if the Carpentaria watershed was responding at a regional level to increased runoff, then stage BR3 may correlate with Cm3.

550 years B.P. to present

The last 500 years of geologic history on the beach-ridge plain is shrouded with an uncertainty similar to that attached to the chenier plain during the same time span. Grouping of radiocarbon observations in Figure 7-6 suggests that a short diastem may have preceded the present period of progradation. It is possible that the recent progradation on the beach-ridge plain may be associated with the same relative increase in terrigenous material which has been observed on the chenier plain.

CLIMATIC HISTORY

The study of Quaternary geologic history has frequently been extended to reconstructions of climatic history and this extension has traditionally been through the study of glacial deposits. Beyond the areas directly affected by Pleistocene glaciation and glacial outwash, other geologic processes have fluctuated in response to Quaternary climatic change. For example, the effect of climatic change on fluvial processes is implicit in the generalization that cooler periods in some parts of the world were accompanied by wetter conditions, although the precise response of streams to such climatic change especially in dry regions is the subject of varying opinions (see Flint, 1971). It can be
further stated that certain geomorphic systems, especially coastal environments, are extremely sensitive indicators of variations in climatic conditions. Considerable uncertainty shrouds both the trigger mechanism for such variation as well as the degree to which such variations are coupled on a global scale. These uncertainties are amplified when dealing with the nearshore and coastal systems, where separation of cause and effect are difficult due to feedback factors. In an example already cited, Thom (1978) has suggested a relationship between degree of storminess, coastal progradation and transgressive dune development on the New South Wales coast, Australia. However, the problem with such a hypothesis is that the separation of regional response from local factors is difficult. Verification of a regional model such as Thom is proposing can only be accomplished by extensive chronologic control at widely separated sites which may be shown to be dominantly influenced by regional factors. All such inferences, whatever the scale, use proxy data to characterize past climates. The inference of climatic information from proxy data is suspect because of several limitations inherent in generalizing from a small number of observations. Webster and Streten (1978) when reviewing evidence for recent climatic change, stated:

Any single observation should therefore be considered as a statistic that defies full interpretation unless considered with other members of the statistical ensemble (Page 280).

Only after the same or parallel type of climatic response is observed at several widely separated localities, is it appropriate to postulate a model. Table 7-4 presents some middle to upper Holocene climatic events, described by other workers (Dodson, 1974a and b; Hope, 1974; Thom, in prep.; Burrows, 1975; and Denton and Karlen, 1973) which may have some expression in the geologic record of the Gulf of Carpentaria.
The water level records of Lakes Keilambete and Leake are interpreted by Dodson (1974 a and b) as suggesting periods when rainfall was probably higher than present. However, ambiguity exists in the use of such parameters as water level or salinity from enclosed basins where evaporation and the factors that affect it play a role as important as rainfall. Furthermore, Dodson infers 10,000 years of climatic history based on 4 radiocarbon observations. Even with ample dating, it is often difficult to identify the climatic event with the material being dated. If the time indicator is an organic deposit (peat or wood) there is difficulty in unequivocally linking the datable material to the event.

Hope (1974) recorded vegetation evidence from a pollen record at Wilson's Promontory, Victoria which suggested a moist climate around 5700 yrs B.P., a trend in general agreement with the findings of Dodson.

Thom (unpublished data) attempts to link phases of transgressive dune building to periods of increased storminess in the Tasman Sea. Thom has shown that dune building periods appear out of phase with periods of beach-ridge accretion, a condition to be expected if it can be shown that ridges and mostly constructed during periods of less storminess.

The expansion of alpine glaciers in New Zealand is discussed by Burrows (1975) whilst the generalized response of several glaciers in North America and North Europe is presented by Denton and Karlen (1973) Admittedly, the relationship between glacial expansion which implies a cooling trend and continental or global pluvial phases may be debatable. However the findings of both Burrows and Denton and Karlen are included to demonstrate both the potential and the difficulty of such broad scale correlation.

The examples of apparent climatic change described above are also compared in Table 7-4 with the phases of beach-ridge progradation demonstrated for the beach-ridge plain. The beach-ridge plain is used as
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<td>SE AUSTRALIAN SAND BARRIERS</td>
<td>NEW ZEALAND GLACIERS</td>
<td>NORTHERN HEMISPHERE GLACIERS</td>
<td>GULF OF CARPENTARIA BEACH-RIDGE PLAINS</td>
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| 7000 - | WET | MOIST | INCREASED STORMINESS | EXPANSION | INCREASED SEDIMENT FLUX |
| 6000 - | WET | MOIST | INCREASED STORMINESS | EXPANSION | INCREASED SEDIMENT FLUX |
| 5000 - | WET | MOIST | INCREASED STORMINESS | EXPANSION | INCREASED SEDIMENT FLUX |
| 4000 - | WET | MOIST | INCREASED STORMINESS | EXPANSION | INCREASED SEDIMENT FLUX |
| 3000 - | WET | MOIST | INCREASED STORMINESS | EXPANSION | INCREASED SEDIMENT FLUX |
| 2000 - | WET | MOIST | INCREASED STORMINESS | EXPANSION | INCREASED SEDIMENT FLUX |
| 1000 - | WET | MOIST | INCREASED STORMINESS | EXPANSION | INCREASED SEDIMENT FLUX |
| PRESENT- | WET | MOIST | INCREASED STORMINESS | EXPANSION | INCREASED SEDIMENT FLUX |
the primary site of comparison for two reasons: (1) the episodes of beach-ridge progradation are directly dated by shells included in the prograded sediments, whereas on the chenier plain, the dated material is from a winnowed deposit formed during periods of progradation, and (2) the episodic nature of the beach-ridge plain sequence has been validated by comparison with the dates reported in Smart (1976) for Cape Keer Weer.

Initial progradation on the beach-ridge plain occurred during the period 6000-4800 years B.P. This episode of deposition followed closely the completion of the Postglacial Marine Transgression and is synchronous with a wetter climatic phase in most of southern Australia. The initial phase of increased cyclogenesis proposed by Thom (unpublished data) came at the close of the pluvial indicated by lake levels (Dodson, 1974 a and b) and palynologic evidence (Hope, 1974). If increased cyclogenesis could be shown to have occurred simultaneously in both tropical and temperate areas of eastern Australia, then a mid-Holocene pluvial associated with such storminess may have contributed to the mid-Holocene episode of coastal progradation by increasing sediment supply from North Queensland to the Gulf of Carpentaria. On a global scale, glacial advances in the Northern Hemisphere described Denton and Karlen (1973) appear to be in phase with this initial progradation whereas the record described by Burrows (1975) for New Zealand glaciers does not suggest similar glacial activity. However, it has already been suggested that pluvials, especially in Australia, may not necessarily have been characterized by cooler conditions.

Following the mid-Holocene, there is little evidence for synchronicity between beach-ridge progradation and wetter conditions elsewhere in Australia until after 3000 yrs B.P. However, comparison of progradation phase BR2 (Figure 7-4) may be made with the glacial readvance shown by Denton and Karlen between 2400 and 3300 yrs B.P. This
requires the assumption that Northern Hemisphere high latitude cooler periods were wetter in northern Australia. Furthermore, these activities appear to coincide with the latter stages of New Zealand alpine glacier expansion (Burrows 1975).

Given the limited chronologic control of Dodson (1974b) and Hope (1974), there is general agreement in a wetter trend after 2000 yrs B.P. as shown by lake levels and the vegetation record. This trend may also have global implications when compared with the North American glacial record. However, New Zealand glaciers again did not respond in a synchronous manner whilst the progradation episode on the beach-ridge plan was a long one, spanning a period from 2300 yrs B.P. to approximately 600 yrs B.P., with most progradation centered around 1400 yrs B.P.

A brief comparison of upper Holocene climatic change such as shown above, illustrates many of the difficulties arising when attempting to correlate widely separated sites from which proxy climatic information has been gained in different ways. There are distinct overlaps or even total lack of agreement when similar responses over wide areas might be postulated. Such disparities may be attributed to three causes.

1. Coupled response factors in climatic change have been incorrectly characterized, for instance pluvials might be not only cooler and wetter but warmer and wetter.

2. There may be lags in the response of some environments to climatic change, and these lags are sufficient to explain the varying chronologies at different locations.

3. Small scale climatic fluctuations may occur on a regional or hemispherical basis but in a manner such that global response is not synchronous. Any attempt to correlate world-wide activity might result in comparisons of events which are either not the same event or inherently out-of-phase.
requires the assumption that Northern Hemisphere high latitude cooler periods were wetter in northern Australia. Furthermore, these activities appear to coincide with the latter stages of New Zealand alpine glacier expansion (Burrows 1975).

Given the limited chronologic control of Dodson (1974b) and Hope (1974), there is general agreement in a wetter trend after 2000 yrs B.P. as shown by lake levels and the vegetation record. This trend may also have global implications when compared with the North American glacial record. However, New Zealand glaciers again did not respond in a synchronous manner whilst the progradation episode on the beach-ridge plan was a long one, spanning a period from 2300 yrs B.P. to approximately 600 yrs B.P., with most progradation centered around 1400 yrs B.P.

A brief comparison of upper Holocene climatic change such as shown above, illustrates many of the difficulties arising when attempting to correlate widely separated sites from which proxy climatic information has been gained in different ways. There are distinct overlaps or even total lack of agreement when similar responses over wide areas might be postulated. Such disparities may be attributed to three causes.

1. Coupled response factors in climatic change have been incorrectly characterized, for instance pluvials might be not only cooler and wetter but warmer and wetter.

2. There may be lags in the response of some environments to climatic change, and these lags are sufficient to explain the varying chronologies at different locations.

3. Small scale climatic fluctuations may occur on a regional or hemispherical basis but in a manner such that global response is not synchronous. Any attempt to correlate world-wide activity might result in comparisons of events which are either not the same event or inherently out-of-phase.
Geologic histories can be as useful as vegetation histories in demonstrating climatic change provided cause and effect relationships are fully understood. The testing of hypotheses concerning climatic change in the Holocene may proceed in arid or semi-arid areas where vegetation histories are absent or poorly developed. Arid and semi-arid areas have been shown to be especially sensitive to fluctuations in rainfall which may affect the fluvial system. Where the final repository for fluvially transported material is an epicontinental sea, there is considerable potential for interpreting the nearshore and coastal depositional record in terms of climatic change provided other factors such as relative sea level may be identified and their effects isolated in the record.

The inference of global correlations from such records is considered to be premature because of the present lack of understanding concerning trigger and feedback mechanisms for both coastal deposition and climatic change. Furthermore, the lack of knowledge concerning factors controlling deposition, non-deposition and erosion of the coastal zone is accentuated by the limited number of Holocene geologic histories which are supported by a detailed radiocarbon age structure.

The Gulf of Carpentaria has provided approximately 6000 years of geologic history in which cause and effect appear to be separable. The resulting "filtered" record supports a model which may be tested further by independent data from other geographic areas which may have been subject to regional variations in climate.
RESEARCH TECHNIQUES

The purpose of this section is to explain the manner and in some cases the rationale by which data from the Gulf of Carpentaria was collected and analyzed. This discussion is organized into three categories of activities. Although separated here for ease of presentation, many of these activities were concurrent or overlapping (see Table A-1). Categories of activities were:

a. fieldwork
b. analysis of environmental data
c. sedimentologic and geochemical analysis

FIELDWORK

Access to the Gulf of Carpentaria was described in Chapter 2. The distinct wet and dry seasonality of the area dominates all fieldwork and limits wet season data collection to those activities which can be carried out by boat or aircraft. Data for this research were collected during two lengthy dry season (5-7 months) and two wet season (2 months) field trips. Table A-1 summarizes the work carried out during these 16 months in the field.

ANALYSIS OF ENVIRONMENTAL DATA

Environmental data are of a physical and biologic nature and describe the large scale setting of the chenier and beach-ridge plains. Such data are distinct from the parameters which were used to specify sedimentary facies.

Meteorologic records are limited in the Gulf. A geographic gap in the collection of rainfall and wind data exists between Weipa and Karumba. An attempt was made to fill this gap by installing a wind speed
**ACTIVITY**
- Coastal reconnaissance
- Drilling
- Geomorphic mapping
- Sampling and descriptions of modern coastal environments
- Topographic surveying
- Tidal data acquisition
- Aerial photography
- Offshore sampling

**DURATION**
- June-August 1975
- August-October 1975
- August-November 1975
- August-November 1975
- February 1976
- July-December 1976
- September-October 1976
- Install November 1975
  Remove December 1976
- Intermittent during above periods
- March 1976
  March 1977

**VESSELS AND EQUIPMENT USED**
- Aircraft, small boat and 4 x 4 vehicle
- Gemco 210b (trailer mounted) and 4 x 4 vehicles
- Aircraft and 4 x 4 vehicles
- Small boat, float plane, and 4 x 4 vehicle
- Sokkisha automatic level
- Weathermeasure Model F552 water level recorder
- Hasselblad 500CM with 70 mm magazines
- RV Kalinda (20m)
  RV Sprightly (43m)

**AREA STUDIED**
- Weipa south to Karumba and west to Burketown
- Chapman Creek near Edward River
  Entire eastern shore of Films most frequently used
  Northern Gulf and selected sites along eastern shore
  Northern Gulf

**COMMENTS**
- Required for selection of detailed study areas
- Samples collected from 12 cm diameter auger
- Chenier and beach-ridge plains
- Chenier and beach-ridge plains and selected sites along eastern shore
- Chapman Creek near Edward River
  Rectilinear chart recorder on 7 day mechanical clock.
  Float protected in stilling well.
- Sampling done with 30x30 cm grab
and direction recorder on the coast at Edward River. However, the humid environment caused recurring malfunction of the paper advance system such that events could not be identified by date. It was therefore necessary to rely on more widely spaced data from sources referenced in Chapter 2. Discontinuous rainfall data are kept by stations and missions on Cape York peninsula, but the most reliable data are from Weipa and Normanton.

Records of tidal behavior in the Gulf of Carpentaria are insufficient at even the standard ports of Weipa and Karumba. Furthermore, it appeared that both human and natural events, for which the area is noted, precluded the accurate recovery of some of the previous tide gauge benchmarks. For example, when tidal data was requested from a previous officer in charge of the CSIRO Karumba field station, he supplied the data with these remarks:

The main problem seems to be that part of the jetty on which the machine was mounted subsided over the period of observation. The heights given in the attached documents were applicable when the instrument was installed but it is unknown for how long these values were valid. Probably there had been some subsidence prior to our arrival in July 1963 and we were aware of further subsidence during summer cyclones during 1963-1964 and 1964-1965. (...) There was a painted board gauge fixed to an adjacent pile and every time the paper was changed the base on the paper was supposed to be set so that the pen agreed in height with that observed on the board. The board would have subsided with the gauge mounting. There was another check in the form of a steel rod with welded pegs at one foot intervals. I think this value was independent of the jetty and was supposed to serve as a check against subsidence of the gauge and the board. I am unaware of any observations recorded and was told by the officer deputizing in my absence that he hammered this steel gauge into the river bed to match the wooden one. I recall him stating "about six inches". He was not noted for sobriety. (Ian S. R. Munro, written comm. 1978)

Tidal data were acquired specifically for this study at Edward River. This effort was successful and the annual change in sea level observed at this location was discussed in Chapter 3. The harmonic analysis of the data was performed by personnel at CSIRO Fisheries and Oceanography Laboratory, Cronulla, New South Wales.
Topographic survey data were tied to tidal datums at both Karumba and Edward River. At Karumba, the benchmark was S.P.M. 601 in the center of the township near the former seaplane apron. This benchmark was checked for validity by comparison with tidal records and a supplementary benchmark maintained by CSIRO Karumba field station. At Edward River, a local benchmark grid was established with respect to gauge datum on the tide gauge at Chapman Creek. This temporary grid was tied to Division of National Mapping and U. S. Air Force triangulation monuments in the township. Levelling data were reduced in the field and closing errors up to 2.0cm over 1000m were considered acceptable. Larger errors required resurvey over the relevant portions of the transect.

Plant collections from the field were preserved for herbarium identification. Strandline communities from the Edward River area were identified and curated by the CSIRO Division of Forest Research, Atherton, Queensland. Mangrove collections were identified and curated by CSIRO Black Mountain Herbarium, Canberra.

Shell assemblages were identified by the Department of Malacology, Australian Museum, Sydney. Molluscs which were not already in the museum collection have been retained in Sydney.

SEDIMENTOLOGIC AND GEOCHEMICAL ANALYSIS

With the exception of some shipboard work on the R/V Sprightly, most of the sedimentologic and geochemical analyses were performed in the laboratories of various departments of the Australian National University. Shipboard estimation of sand and mud content was performed on grab samples collected from the Gulf. This method entails a volumetric measurement of the sample before it is wet sieved on a 62.5 micron sieve. The initial volume is usually 100ml, and measurement of the material remaining on the sieve after wet sieving provides a percent
sand in the bulk sample. Crosschecking of ten replicate samples in a soils laboratory by dry weight methods found the volumetric method accurate to 5%.

Granulometric analysis was completed by settling tube, soil hydrometer and centrifuge techniques for the respective sand, mud and submicron fractions. Table A-2 summarizes the measurement of these size fractions. Grain size statistics were calculated by the method of moments using software written by the author, but based on statistical strategies outlined by Folk (1965). Scatter and probability plots were drawn by a drum plotter interfaced to the ANU DEC-10 computer using graphics routines developed by the author and based on Calcomp subroutines.

X-ray diffraction techniques were required to study carbonate and clay mineralogy. The XRD facility, located in the Geology Department, ANU, consisted of a Siemens Type F (Omega) flat bed goniometer linked to a Siemens scintillation counter and powered by a Philips 2kw power source. Output was produced on a Philips flat bed recorder. Pretreatment of the samples and interpretation of the traces were done in accordance with techniques set forth by Bloss (1971). Identification of minerals was based on the ASTM (American Society for Testing Materials) tables of interplanar spacings.

Geochemical analyses were conducted in the geochemistry facility of the Department of Biogeography and Geomorphology. Most of this work was completed by or performed under the supervision of the staff geochemist. Four major analyses were performed: organic carbon, sulfates (soluble and insoluble), soluble chlorides and total carbonate. Methods used in these determinations are based on accepted published strategies. These strategies and the appropriate references are outlined below.

- organic carbon. This analysis proceeded by the net oxidation
<table>
<thead>
<tr>
<th>SIZE FRACTION</th>
<th>TECHNIQUE</th>
<th>PRETREATMENT/SUMMARY</th>
<th>FACILITY USED</th>
<th>REFERENCE</th>
</tr>
</thead>
<tbody>
<tr>
<td>Sand</td>
<td>Settling tube</td>
<td>Wet sieve to remove mud; organic removal with H₂O₂, carbonate removal with dilu HCl, dried and split to 2.0 grams</td>
<td>Bureau of Mineral Resources settling tube</td>
<td>Zeigler et al. 1960</td>
</tr>
<tr>
<td>Mud silt and clay to 4 microns</td>
<td>Soil hydrometer</td>
<td>Dispersal by drink mixer in distilled water with Calgon and NaOH (62.5g sediment in 1250 ml)</td>
<td>Soils lab, Dept of Bio- Folk, 1965</td>
<td>Mayo, 1974</td>
</tr>
<tr>
<td>Submicron clays</td>
<td>Centrifuge and dry weight methods</td>
<td>Remove 2 micron and smaller replicate from hydrometer settling cylinder, centrifuge removal of 1, 0.5, 0.25, 0.13, 0.06 micron sizes, evap to dryness and weigh.</td>
<td>As above</td>
<td>Steele and Bradfield, 1934</td>
</tr>
</tbody>
</table>
technique outlined by Schollenberger (1945). After measurement of the total carbon, it was assumed that this carbon is all due to organic material in the sediment and is therefore reported as total organics.

b. total sulfates. After a hot acid leach with 2N hydrochloric acid, titration with barium chloride produced barium sulfate. When soluble sulfate analysis was desired, distilled water was substituted for the hot acid and the titration was repeated. In reporting sulfates present in stratigraphic samples, it was assumed that those found were due to the presence of gypsum, the most ubiquitous sulfate in modern environments. Therefore, conversion was made to indicate total gypsum by weight in the bulk sample.

c. soluble chlorides. Silver nitrate titration according to the Mohr method provided total concentration of these salts. This analysis is only meaningful in a limited number of supratidal facies such as aeolian silts and clays which are above the ground water table during the dry season.

d. total carbonate. A small replicate of the bulk sample was measured by a volumetric calcimeter. The volume of gas given off during acid leach was compared to a known standard (see Presley, 1975) it was assumed that all carbonate present was due to either calcite, aragonite or dolomite although small amounts of ferrous iron appeared in some samples. XRD techniques were employed to differentiate the carbonate minerals, especially calcite and aragonite.
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