Journal of Coastal Research | 11 | 2 | 454–477 | Fort Lauderdale, Florida | Spring 1995

Barrier Stratigraphy on the Macrotidal Central Queensland Coastline, Australia

Gerhard Masselink† and Guilherme Lessa‡

†Centre of Water Resource
Department of Civil Engineering
University of Western Australia
Nedlands 6009, Western Australia

‡UFPR. Centro de Estudos do Mar
Av Beira Mar
Pontal do Sul
Paranaguá (PR) 83255-000 Brazil

ABSTRACT

MASSELINK, G. and LESSA, G., 1995. Barrier stratigraphy on the macrotidal central Queensland coastline, Australia. *Journal of Coastal Research*, 11(2), 454–477. Fort Lauderdale (Florida), ISSN 0749-0208.

The morphostratigraphy of two macrotidal, low energy, sandy barrier systems on the central Queensland coastline (Australia) were investigated. The barriers contain transgressive and regressive sequences and are underlain by a continuous layer of basal transgressive mud, which is exposed at approximately 10 m water depth. The thickness of the macrotidal barrier sequences is 8–12 m, and the limited vertical extent of the barriers is ascribed to the low wave energy level (wave height < 0.6 m) and the gentle gradient of the continental shelf (1/1,000) off the central Queensland coastline. Offshore sediment and seismic data indicate an inner shelf depleted of Holocene sediments, with a pre-Holocene substratum outcropping in many places. The stratigraphy of both barrier systems suggests that (a) sea level was about 2 m higher than present at the maximum postglacial marine transgression (PMT) and (b) a tidal window may have occurred close to PMT.

ADDITIONAL INDEX WORDS: *Barrier stratigraphy, macrotidal, Holocene sea-level history, Australia, central Queensland.*

INTRODUCTION

Coastal sand barrier systems are a dominant feature along the eastern Australian coastline and usually consist of a sandy barrier complex containing nearshore, beach, dune and washover deposits, a backbarrier lagoon and a tidal inlet. They have been extensively described for the southeast microtidal Australian coast where they are usually composed of a transgressive and a regressive sequence (THOM *et al.*, 1978 1981, 1992). Transgressive barrier sequences generally develop under conditions of rising sea-level (*e.g.*, KRAFT, 1971) whereas regressive sequences usually form under stable or falling sea-level conditions in association with abudant sediment supply (*e.g.*, BERNARD *et al.*, 1962; THOM *et al.*, 1981). Adequate sediment yield may even allow for progradational sequences to develop under conditions of slowly rising sea level (BEETS *et al.*, 1992; HILL and FITZGERALD, 1992).

Barrier sequences vary from one coastal region to another in response to different combinations of environmental conditions, including sea-level history, the width and gradient of the continental shelf, incident wave energy level, tidal range, sediment availability and process regimes in the Late Quaternary. Moderately steep offshore slopes (< 1°) and high wave energy levels favour the formation of extensive barrier sequences, and thicknesses may range from 15 m for the Californian coastline (HOWARD and REINECK, 1979 in MCCUBBIN, 1982), to up to 25 m for some southeastern Australian barrier systems (THOM *et al.*, 1981). Conversely, in low wave energy environments with gentle continental shelf gradients, such as the Gulf of Mexico, the thickness of the barrier sequence does not exceed 15 m (BERNARD *et al.*, 1962). To date, the effect of tidal range on the character of the barrier stratigraphy has not been investigated, although DAVIS and HAYES (1984) proposed an upper tide range limit for barrier formation suggesting that barriers do not occur in macrotidal environments (tide range > 4 m).

However, more than 30 barrier systems can be found along the low energy, macrotidal coastline of central Queensland, Australia (see Figure 1 for location study area). These macrotidal barrier systems are generally backed by a partially infilled estuary which is connected to the open sea by a tidal inlet. Although the environmental conditions in central Queensland are not conducive to barrier formation (low wave energy, gentle shelf gradient and macrotidal ranges), the central

---

94018 received 6 June 1994; accepted in revision 15 August 1994.

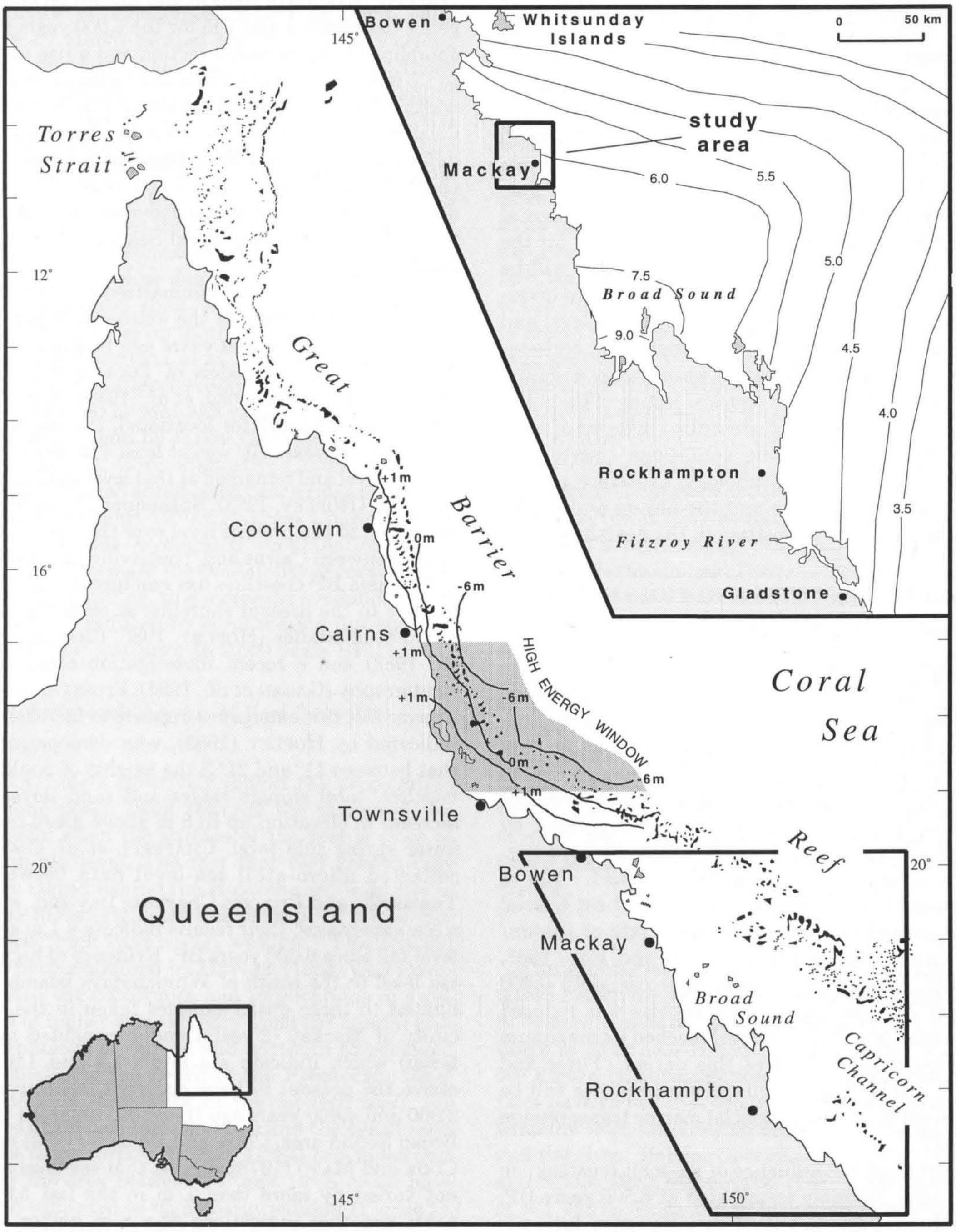

Figure 1. Location of study area, average maximum tidal ranges around Mackay (BPA, 1979), location of HOPLEY's (1984) high wave energy window and the differential warping due to isostatic rebound after HOPLEY (1983).

Queensland barriers fit in the classification of southeast Australian barrier systems (Thom *et al.,* 1978) and fall in the category of prograded barriers.

The aim of this paper is to provide information regarding the morphology, stratigraphy, and Holocene evolution of two macrotidal, prograded barrier systems on the central Queensland coastline. Since the Holocene evolution of these barriers is strongly dependent on the environmental conditions, the Holocene sea-level history for the central to northern Queensland coastline will be discussed first. This will be followed by an investigation of the relationship between sea level and the elevation of the Great Barrier Reef, addressing its effect on the incident wave energy and tidal range. In addition, the physical setting of the study area will be extensively described in terms of present day oceanographic conditions (waves and tides), coastal geomorphology, shoreface profile, pre-Holocene surface and the nature and extent of the Holocene deposition in the nearshore area.

SEA LEVEL, WAVES AND TIDES HISTORY

Quaternary Sea-level History

During the Last Interglacial maximum, the sea level reached +5 m relative to present sea level in southeastern Australia, as indicated by evidence from Pleistocene barrier systems in New South Wales (Marshall and Thom, 1976). In northeastern Australia, a Pleistocene sea level of +4 m relative to present level is suggested by Hopley (1982) based on fragmented inland barriers in Queensland. Sea level reached −120 m around 18,000 years BP during the Last Glacial and started to rise at an average rate of 13 mm/yr until 7,700 years BP (Thom and Roy, 1983, 1985). Between 7,700 and approximately 6,400 years BP, the rate of sea-level rise was reduced to 4 mm/yr and the sea level reached its maximum elevation at the end of this period (Thom and Roy, 1983). The postglacial sea-level rise will be referred to as the postglacial marine transgression (PMT).

Although the influence of ice-melt (eustasy) on sea level probably terminated at 6,500 years BP, further relative sea-level variation may have occurred due to isostatic movements caused by the water load on the continental shelf. Thom and Chappell (1978) applied a hydro-isostatic model to the central and northern sections of the Great Barrier Reef under the assumption that the ice-melt contribution had completely ceased by 7,000 years BP. Results of their model suggest an emergence between 0.5 and 1 m for the 6,000 years BP shoreline in the northern section and a rise of up to 1.5 m in the central section due to the increased width of the continental shelf at this location. Chappell *et al.* (1982) calculated hydro-isostatic deformations for northern Queensland and the Gulf of Carpentaria since 5,500 years BP and demonstrated that positive movements of more than a meter along the Coral Sea shoreline may have occurred.

Several field studies (summarised by Hopley, 1982, 1983) demonstrate the existence of higher sea levels close to 5,000 years ago in support of the hydro-isostatic models of Thom and Chappell (1978) and Chappell *et al.* (1982). North of Cairns (see Figure 1 for locations), the sea level around 5,800 years BP was at least 1 m above its present level and remained at that level until 3,000 years BP (Hopley, 1983). Subsequently, sea level decreased to the present level over the next 1,000 years. Between Cairns and Townsville, the 5,000–6,000 years BP coastline has emerged 1 to 1.5 m relative to the present shoreline as evidenced by micro-atoll studies (Hopley, 1983; Chappell *et al.,* 1983) and a recent investigation of barrier stratigraphy (Gagan *et al.,* 1994). From Cairns to Townsville, this emergence appears to increase as indicated by Hopley (1983), who demonstrates that between 11° and 21° S the heights of boulder beaches, coral shingle ridges and sand terraces increase in elevation up to 6 m above mean high water spring tide level. Chappell *et al.* (1983) collected micro-atoll sea-level data between Townsville and Princess Charlotte Bay and, with a few exceptions, their results indicate a 1 m sea-level fall since 6,000 years BP. Evidence of higher sea level to the south of Whitsundays Islands is limited to three dated samples taken in the vicinity of Mackay (2 reefal and 1 cemented material) which indicate sea levels 1.2 and 1.6 m above the present to have occurred respectively 3,900 and 4,600 years ago (Hopley, 1983). In the Broad Sound area, Cook and Polach (1973) and Cook and Mayo (1978) suggest that sea level has not varied by more than 1 m in the last 5,000 years and that indications of a 6 m higher sea level at the eastern side of the estuary appears to be related to neotectonic movements along a fault line.

In general there are strong indications of higher mid-Holocene sea levels for the coastline between

Cairns and Rockhampton, although there are areas, such as the Whitsunday Islands, without any evidence of emergence. Based on field data, HOPLEY (1983) plotted the differential warping over the last 5,000 years along the northern Queensland coast (Figure 1). The results are in good agreement with the CHAPPELL *et al.* (1982) model, showing positive deviations of more than 1 m along the coastline and negative adjustments of up to 6 m on the outer shelf. South of 20° S the +1 m contour trends away from the continent suggesting an increased emergence of the 6,000 years BP shoreline due to the increased width of the continental shelf. A smooth descent of the relative sea level up to present time is commonly accepted (CHAPPELL *et al.*, 1983), although a sudden fall of sea level around 3,000 years BP has been inferred by MCLEAN *et al.* (1978) in northern Queensland and by FLOOD (1983) in southeastern Queensland. Coincidentally, CHAPPELL *et al.* (1983) reported accelerated beach ridge formation after 2,500 years BP in Princess Charlotte Bay. Although this accelerated progradation is explained in terms of the availability of coral shingle for ridge building, this increased availability may have been induced by coral erosion due to a fall in sea level just prior to 2,500 years BP.

Morphodynamic Aspects of Reef Growth

Waves

Coral reefs effectively inhibit the propagation of long ocean swell towards the mainland, especially when the surface of the reef is close to mean sea level. The fact that reef growth may have lagged behind the Holocene sea-level rise in some areas originated the concept of a mid-Holocene "high wave energy window." This concept has been examined by HOPLEY (1984) who investigated the occurrence and implications of a period of higher wave activity upon the coastal geomorphology of central and northern Queensland by analysing the depth of the Pleistocene limestone (the framework of the modern reef), the rates of vertical reef accretion and the age of reefs close to the surface. HOPLEY (1984) concluded that only the area between 17° and 19° S could have been subjected to high energy window (Figure 1).

The likelihood of a high energy window is primarily dependent on the depth of the Pleistocene substratum. As mentioned above, average rates for sea-level rise in eastern Australia were 13 mm/yr until 7,700 years BP and 4 mm/yr from 7,700 to 6,400 years BP (THOM and ROY, 1983, 1985). Estimated vertical reef growth rates are 7 to 8 mm/yr for framework construction, but as high as 16 mm/yr for branching corals (HOPLEY and KINSEY, 1988). Consequently, vertical reef growth during the mid-Holocene may have lagged 4 m behind sea level every thousand years until 7,700 years BP, after which the rate of sea-level rise was reduced. The deeper the Pleistocene surface relative to sea level, the earlier it is flooded and the more time is allowed to increase the difference in elevation between the sea level and the reef surface. As pointed out by HOPLEY (1984), the depth of the Pleistocene substratum below sea level between 17° and 19° is always larger than 10 m, or even greater than 20 m, whereas north of 17° and south of 19° the Pleistocene occurs generally at depths of 6 to 12 m and is occasionally emerged. The southernmost reef included in HOPLEY'S (1984) work is Redbill Reef, just north of Mackay, but there is some evidence that the southern sector of the Great Barrier Reef may also have experienced a high energy window. Using seismic profiles, MARSHALL and DAVIES (1984) show that the depth of the Pleistocene surface ranged from 7 to 23 m between 23° and 24° S. In addition, the seismic study of HARVEY (1986) yields Pleistocene basement depths ranging from 8.5 to 23 m, with an average of 13 m. The depths of the Pleistocene surface reported by MARSHALL and DAVIES (1984) and HARVEY (1986) are similar to the depths found in the high energy window section of HOPLEY (1984), and the possibility of a high energy window for the central Queensland coastline (between Mackay and Rockhampton) cannot be excluded.

Tides

The central section of the Great Barrier Reef is well known for its large tidal ranges, the highest in eastern Australia, with maximum spring tidal range of 10 m near Broad Sound (EASTON, 1970; see Figure 1). Investigations of these anomalous tides have shown that the large tidal ranges result from the tidal canalisation between the mainland and the Great Barrier Reef (EASTON, 1970; BODE and STARK, 1983; MIDDLETON *et al.*, 1984; GRIFFIN *et al.*, 1987). Since the propagation of the tidal wave across the barrier reef is largely inhibited due to friction, it travels into the region via gaps in the northern and southern reefs. Thus, there are two tidal waves travelling in opposite directions with phases which tend to reinforce each other in the central part of the basin, leading to a quasi-resonant condition of the semi-diurnal tidal components (BODE and STARK, 1983; STARK *et al.*, 1984; GRIFFIN *et al.*, 1987).

It becomes appealing to think of the effect that deeper reefal surfaces may have had on the propagation of the tide and the resulting tidal ranges. Hypothetically, larger reefal depths would allow for unimpeded tidal flow across the reef which would inhibit the formation of the standing wave pattern between the reefs and the mainland. Consequently, the tidal range under such conditions would be reduced. Analogous to HOPLEY'S (1984) high energy window, we may define a "tidal window" for the central Queensland coastline during which the tidal impedance across the central Great Barrier Reef was reduced, resulting in reduced tidal ranges. Such a tidal window would have occurred simultaneously with the high energy window in the mid-Holocene.

The likelihood of smaller tidal ranges depends upon the reef's "transparency" to tidal flow. MIDDLETON *et al.* (1984) provide an analytical investigation of the relative amount of tidal energy which can be transmitted across the reef, assuming an ocean with uniform depth separated from the continental shelf by a reef platform. MIDDLETON *et al.* (1984) calculated a transmission coefficient (T) across the reefs based on the depth of the shelf, the depth of the ocean, and the width and depth of the reef. If the transmission coefficient T is between 1 and 2, the tide is transmitted across the reef with some amplification, but if T is much smaller than 1, the tidal surface displacement across the reef is negligibly small. The present value for T is estimated at about 0.13 (MIDDLETON *et al.*, 1984).

Since it is expected that the transparency of the reef, parameterised by T, varied over the Holocene period, an exercise was undertaken to investigate the possible variation of T over time using the model of MIDDLETON *et al.* (1984). The sea-level curve of THOM and ROY (1983) was used to represent the sea-level history. No hydro-isostatic effects were accounted for, since the Great Barrier Reef is located in the vicinity of the 0 m contour (Figure 1). The calculation starts at 9,000 years BP with sea level at −20 m (the depth of the Pleistocene limestone) and rising at a rate of 13 mm/yr. HOPLEY (1984) and HOPLEY and KINSEY (1988) show that some reefs reached the present position as late as 3,500 years ago, and it will be assumed that on average the reefs arrived at the surface around 4,500 years BP. Vertical reef growth rates are estimated at 8 mm/yr for water depths over 5 m but as low as 4 mm/yr for water depths of 2 m (HOPLEY and KINSEY, 1988). Following the above assumptions, it is apparent that the reefs could not have grown significantly during the first millenia after inundation at 9,000 years BP, otherwise they would have surfaced much earlier than 4,500 years BP. This possibility has also been addressed by DAVIES and MARSHALL (1979), who suggest that reef growth would have difficulties in colonising the antecedent surface just after submergence due to high turbidity.

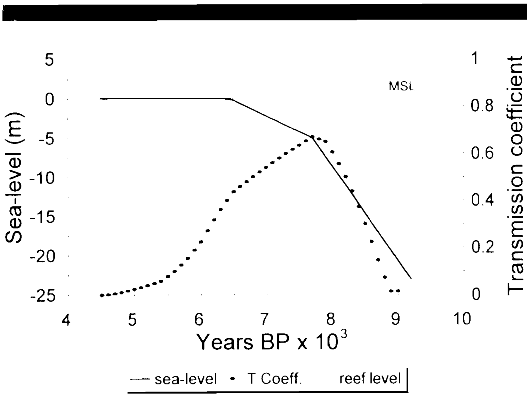

Figure 2. Holocene history of sea level, reef surface and tidal transmission coefficient, T.

Figure 2 shows the evolution of sea level, reef depth and T from 20,000 to 4,500 years BP according to the model of MIDDLETON *et al.* (1984) and the above assumptions. It is observed that until 7,700 years BP, the reef surface increasingly lags behind the rising sea level, mainly due to its later growth initiation. Vertical reef growth catches up with the rate of sea-level rise after 7,700 BP and the reef reaches the surface around 4,500 BP. The transmission coefficient T progressively increases until it reaches a value of 0.7 at 7,700 years BP, the time of maximum reefal depth. After 7,700 years BP, T decreases towards 0 as the reefs become shallower. A reduced tidal impedance across the reef could have occurred close to 6,500 BP as implied by the maximum value of T around this time, and it may be suggested that smaller tidal ranges might have been felt during mid-Holocene along this part of the Queensland coast.

Unfortunately the transmission coefficient T is not a quantitative indicator of tide range, and it is difficult to assess the absolute reduction of the tide ranges during the probable tidal window using the model of MIDDLETON *et al.* (1984). STARK *et al.* (1984) shed some light into the subject by numerically simulating the M_2 tidal constituent in the barrier reef. It is reminded that the resonance in the area affects mainly the semi-diurnal frequency, and the M_2 component accounts for roughly 50% of the total tidal range in the area. A model run was designed with the complete removal of the reef elements, creating a continuous line between the shelf break and the mainland. According to the model, the amplitude of M_2 undergoes a reduction of 25% and in the Mackay region, this results in a decrease of the M_2 amplitude from 1.6 m to 1.3 m. The model of STARK *et al.* (1984) suggests that maximum tidal ranges around Mackay in the absence of the Great Barrier Reef would not be larger than 5.0 m, in comparison to a present maximum tide range of 6.8 m. It is noted that the width of the continental shelf plays the dominant role in the process of tidal shoaling. Even without the Great Barrier Reef and its framework, the central Queensland coastline would experience macrotidal conditions.

Summarising, a wave and tide window, *i.e.* a period with higher waves and smaller tides, respectively, may have been present on the central Queensland coastline from 6,500 to 5,500 years ago.

THE CENTRAL QUEENSLAND COASTLINE

General Introduction

The study area in the vicinity of Mackay, on the central Queensland coast, covers around 60 km of coastline extending from Hay Point to Cape Hillsborough (Figure 3). The Pioneer River is the principal supplier of sediments to the coast (JONES, 1987) with an estimated rate of sand and gravel yield of 40,000 m^3/yr (GOURLAY and HACKER, 1986). The Great Barrier Reef is located adjacent to the central Queensland coastline and is separated from the mainland by a relatively wide (200–400 km) and shallow (< 80 m) continental shelf. Waves are predominantly locally generated by southeasterly winds and have wave heights off Mackay ranging from 0.6 to 1.0 m with periods between 5 and 7 sec (BPA, 1986). The tides are semi-diurnal and may be classified as macrotidal with mean spring tide ranges decreasing from 5.5 m at Hay Point to 4.5 m at Cape Hillsborough. Cyclones influence the area regularly with an average frequency of 5 per decade passing within 320 km of the coast and generating waves of 3 to 4 meters in height (BPA, 1986). The majority of the storm surges recorded to date have less than 0.6 m, although values as high as 3.65 m have been experienced (GOURLAY and HACKER, 1986). Net sediment transport direction in the inner shelf and nearshore is predominantly northwards along the whole macrotidal coast, between Fitzroy River and Bowen (DEPARTMENT OF HARBOURS AND MARINE, 1981; BPA, 1983; LESSA, 1993). The longshore sediment transport rate in the nearshore area off Harbour Beach and Black's Beach is estimated at 25,000 m^3/yr (DEPARTMENT OF HARBOURS AND MARINE, 1981; BPA, 1983).

Coastal Geomorphology

The style of Holocene coastal accretion varies markedly across the region shown in Figure 3. South of Hay Point several small beach ridge barrier systems, bounded by headlands, are present. These systems are relatively narrow and the coarse beach sediments contain large quantities of shelly material (> 10% carbonate). Since the net nearshore and inner shelf sediment transport is likely to be directed northwards (LESSA, 1993), these barrier systems cannot be supplied by the Pioneer River (Figure 3). As there are no major rivers entering the sea south of the study area, the inner continental shelf is the most probable sand source for these barrier systems. One of the barrier systems described in this study, the Louisa Barrier, is located within this area.

Between Sandringham Bay and the mouth of the Pioneer River, a narrow coastal fringe of estuarine backbarrier deposits and small beach/foredune ridges occurs on the landward margin of an extensive intertidal/subtidal terrace up to 3 km wide extending to between −3 and −5 m below mean sea level (see profile A in Figure 3). The terrace becomes less defined to the north and between the mouth of the Pioneer River and Slade Point the profile is generally steep and concave upward (profile B, Figure 3), with sometimes more complex topography arising from offshore rocks and undulating sand banks (profile C, Figure 3). A large barrier system is present between the river and Slade Point, reaching a width of 2 km. Behind the barrier lies an extensive infilled estuary and MARTIN (in JONES, 1987) reports the presence of Pleistocene beach ridges behind these Holocene estuarine deposits.

To the north of Pioneer River, until Shoal Point,

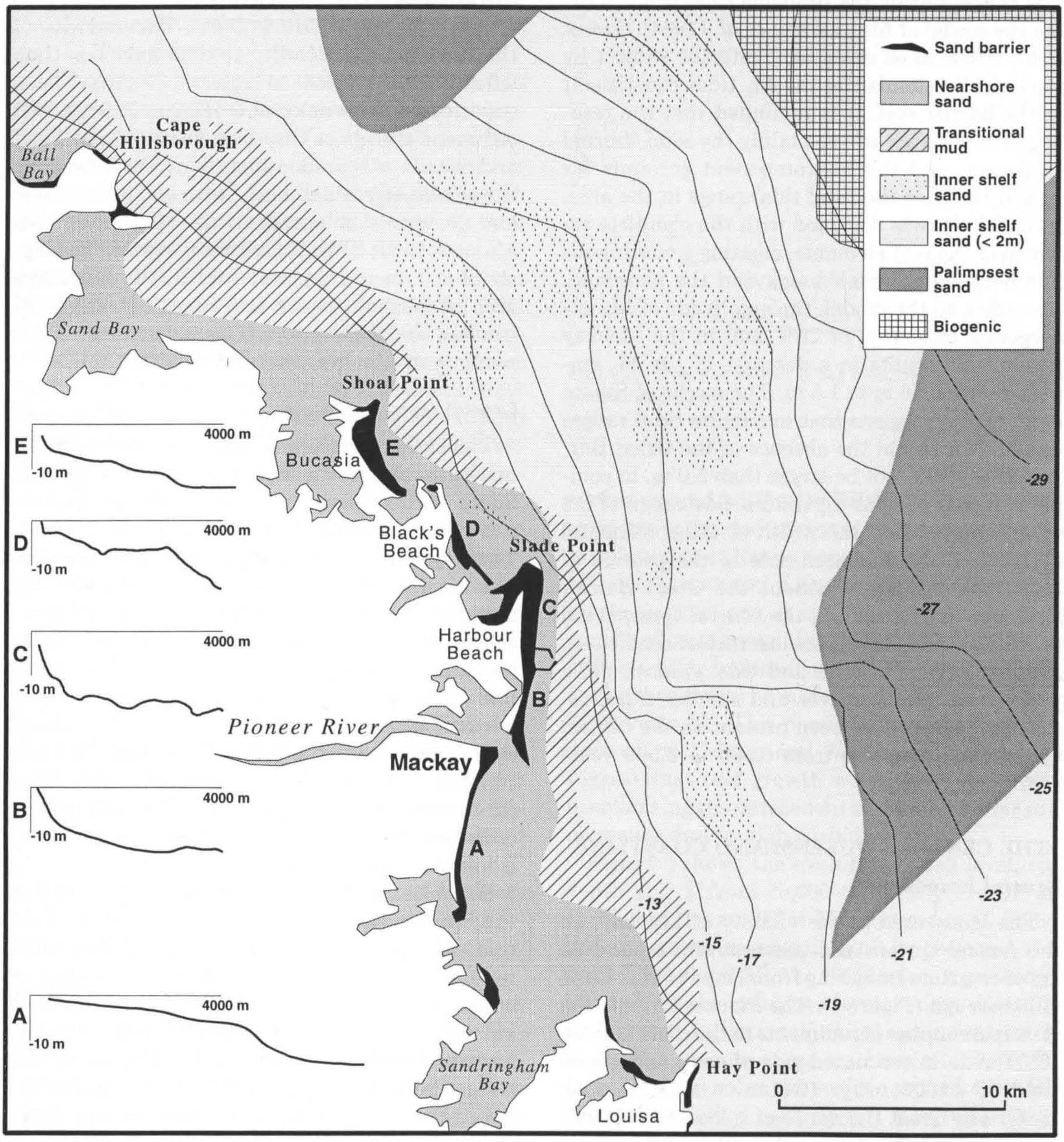

Figure 3. Map of the study area showing the location of barrier systems, Holocene facies, Pleistocene surface and shoreface profiles. The Pleistocene surface is derived from HEGARTY (1983), the shoreface profiles are from JONES (1987) and both these (unpublished) reports have been used to map the Holocene facies.

the beach sediments are relatively coarse and do not contain significant amounts of shelly material. JONES (1987) suggests that the supply of sediment to this part of the coastline from the continental shelf is likely to have been significant in the past and may be continuing at present, but concludes that during the last few thousand years of relatively stable sea level, the Pioneer River was the most important source of sand.

Beyond Slade Point, smaller scale versions of the intertidal/subtidal terrace occur in the sheltered southern parts of the embayments. Beach sediments are fine and contain significant amounts of carbonate material. Seafloor slopes are more gentle than south of Slade Point and a large shallow sand bank extending northwards from Slade Point is the main offshore feature of the Black's Beach embayment (profile D, Figure 3). Due to the presence of offshore rocks and sand banks, simple nearshore concave profiles to about −7 m only occur in the centre of the Bucasia embayment (profile E, Figure 3).

North of the large estuarine system of Sand Bay, several small, low wave energy barrier systems are present on the Cape Hillsborough peninsula. The beach sediments range from very fine to very coarse, reflecting local sand sources, but contain large quantities of carbonate material (> 10%) suggesting that the most important sediment source of these barrier systems is the continental shelf.

Pleistocene Surface and Holocene Stratigraphy

Seismic reflection profiling surveys were conducted in 1979 by the geophysics section of the Geological Survey of Queensland off Mackay in order to determine the shallow seismic stratigraphy of the area. Surface sampling and shallow water drilling assisted in the recognition and interpretation of the seismic sequences and facies. The results reported by HEGARTY (1983) indicate that a strong sub-bottom reflector occurs throughout the Mackay area. The regional reflector is a seismic discontinuity which dips gently to the east and runs roughly parallel to the seafloor. The seismic character is a high amplitude, low frequency cycle and the first cycle generally correlates with the depth where a brown cohesive clay horizon was intercepted during shallow drilling. LESSA and MASSELINK (in preparation) characterize the Pleistocene in the estuaries as an oxidised conglomerate with a mud matrix. A map showing the depth below AHD (mean sea level) to this reflector indicates that a uniform gradient of 1.5 m/km exists over the area (Figure 3). Generally, the gradient of the Pleistocene surface steepens towards the coastline and this is especially apparent in front of Mackay Harbour and Hay Point.

Several Holocene facies have been deposited on top of the Pleistocene reflector and have been extensively sampled by both HEGARTY (1983) and JONES (1987). Figure 3 summarises these studies by showing the distribution of the four relevant postglacial sediment facies identified, namely: nearshore sand, transitional mud, inner shelf sand shoals and a palimpsest layer. An additional inner shelf facies can be recognised where only a thin veneer of Holocene sediments (thickness less than 2 m) is present. The total amount of sediment present on the inner shelf is estimated at 3.1 km^3 (measured from sediment thickness maps provided by HEGARTY, 1983).

The nearshore sand fringes the entire coastline in the study area, extending seawards to a depth below mean sea level of around 10 m with grain sizes in the range of 0.5–2.5 phi, and exhibiting a northward fining trend. The average thickness of this facies is about 1–2 m, and its width varies from approximately 2 km to up to 6 km in the large embayments. JONES (1987) suggests that the sediments are predominantly supplied by the Pioneer River.

A belt of transitional mud lies offshore from the nearshore sand facies and is composed of a mixture of mud, sand and gravel. It has an average width of 2 km, reaching 8 km in Sand Bay and reducing to less than 1 km at Cape Hillsborough. The deposits occur mostly over a depth range of 8 to 12 m and the thickness ranges from 2 m in the south to 6 m in Sand Bay south of Cape Hillsborough. Strong parallel reflections characterise the sediment where it attains a thickness greater than 2 m (HEGARTY, 1983). One of the drillholes reported in HEGARTY (1983) intersected this unit and indicated a 1 m thick, brown/grey cohesive mud, underlain by yellow/brown gravelly sand. According to HEGARTY (1983) and JONES (1987) this transitional mud facies is Holocene in age and results from sedimentation processes which are operative at present sea level. As such, they represent the transition from the nearshore to the inner shelf sands and contain a mixture of mud (suspended load of Pioneer River), fine sand (nearshore), and coarse sand (offshore). A different interpretation of this unit is provided at the end of this paper.

The inner shelf sand consists of medium to coarse-grained sand (−0.5 to 1 phi) with some gravel. The sand has reorganised itself into sand shoals and according to HEGARTY (1983) the shoals are part of an inner shelf sand ridge field. The morphology of the sand waves suggests that the deposits are active. The sand ridges may be reworked sediment from the last glacial low-stands in sea level, or Holocene sediment added to such a deposit, with ridges forming as a result of inner shelf currents. The inner shelf shoal sand generally lies below 9 m water depth but also occurs as offshore sand banks in smaller depths (*e.g.* the large shallow sand bank extending northwards from Slade Point).

A palimpsest layer of sediment lies over much of the Late Pleistocene land surface in the area, thickening towards the east and northeast extremities up to 4–5 m. The unit consists of a medium to fine-grained sand with minor gravel and shell fragments and is interpreted as a combination of relict and modern tidal current deposited sediment (*i.e.* palimpsest).

A large part of the inner shelf zone does not contain more than 2 m of Holocene sediment and only a thin veneer of sediment is found. This facies contains a mixture of nearshore sand, transitional mud, inner shelf shoal sand and palimpsest veneer and it is not unlikely that the Pleistocene surface is exposed over large parts of this area.

BARRIER MORPHOSTRATIGRAPHY

Methodology

Sediment cores were taken using a mobile pneumatic vibrocore unit with a 6 m long, 50 mm diameter stainless steel core barrel. In the field the sediment in the core barrel was extruded into 1 m length plastic sleeves of 65 mm diameter. This procedure destroys the sedimentary structures in sandy deposits and in addition reduces the recovery length because sandy sediment expands in the plastic sleeves. However, since the penetration depth was recorded in the field and the amount of expansion is known (30%), the true elevation and thickness of the sediment units can be obtained. The ground levels of the core sites were determined using a TOPCON total survey station, and these were corrected to Australian Height Datum (AHD = approximate mean sea level) using established benchmarks.

In the laboratory, the cores were cut in 0.5 m lengths, opened and logged. Subsamples were taken from each identified sedimentary unit for textural analysis. All the cores were photographed and several epoxy resin peels were made for future reference. Sediment samples were analysed using a settling tube and sand/mud ratios and carbonate content were determined using standard procedures.

Bucasia Barrier

The Bucasia barrier is oriented in NNW–SSE direction and fronts two estuarine systems (Figure 4). The barrier is approximately 5 km long and bounded by headlands at both sides. The southern end of the barrier is about 800 m wide, but the barrier widens to nearly 2 km towards the north. A smaller estuarine system (Southern estuary) is present at the southern end of the barrier and air photo analysis suggests that in the past it was connected to a larger estuary, towards the north (Northern estuary). Presently, this relict entrance channel is represented by a brackish swamp, and behind this former connection is another small barrier consisting of highly leached medium to coarse white sand of Pleistocene age. An attempt was made to core this barrier but the senior author managed to bog two 4WD vehicles and a tractor crossing the brackish swamp. Pleistocene barriers have been identified by MARTIN (in JONES, 1987) behind the large Holocene barrier system south of Slade Point and behind the Black's Beach barrier.

At the southern end of the barrier, the intertidal beach consists of a steep, reflective upper part and a low gradient dissipative low tide terrace (tidal flat). The latter is part of the ebb tidal delta; and in the low intertidal zone of the tidal flat, sand waves 1 m high and about 20 m long form a succession of ridges indicating transport towards the north. Present day surface beach sediments are medium to coarse sand and carbonate free. In northward direction the overall beach gradient decreases, resulting in a wide, gently sloping dissipative beach with fine to medium sand sized sediments. Several foredune ridges are present at the southern end of the Bucasia barrier, reaching an elevation of 5.5 m AHD. The foredunes are replaced by a transgressive dune field in the north with the height of the dunes increasing up to 10 m AHD as the shoreline becomes more exposed to the southeasterly winds. At the northern end of the barrier, the transgressive dunes overlie bedrock. Further inland, a large number of foredune ridges were identified in the field and on air photos (Figure 4), and their relief increases in a seaward direction. Building practices have largely obliterated the foredune ridges at the southern end of the barrier. Washover deposits are present at the back of the barrier abutting the large Northern estuary, and have largely maintained their lobe-like topography.

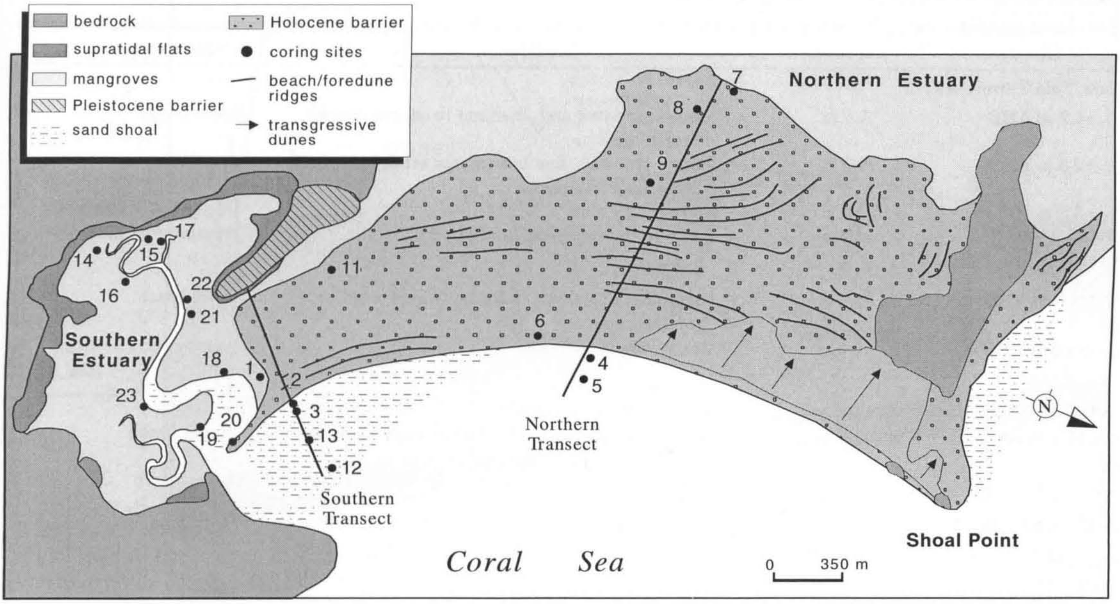

Figure 4. The geomorphology of the Bucasia barrier system with the locations of the core sites.

There is no wave data available for the Bucasia barrier. However, modal wave heights can be estimated taking into consideration the modal significant wave height of Mackay (0.6 m) and the orientation of the Bucasia embayment. The northern end of the barrier is expected to have a wave height slightly less than off Mackay (0.5 m) with a gradual decrease in wave energy level towards the south. The mean spring and neap tidal range for Bucasia are 4.6 and 2.1 m, respectively and the highest astronomical tide level is 3.6 m AHD (AUSTRALIAN NATIONAL TIDE TABLES, 1992).

Two transects were established across the Bucasia barrier, one at the southern end and one further to the north (Figure 4). The coring sites inside the Southern estuary are related to a study of the estuarine evolution (LESSA and MASSELINK, in preparation), but since occasional reference is made to these cores, their locations have also been plotted.

Northern Transect

The northern transect consists of six cores: low tide terrace (#5), runnel between upper beach and low tide terrace (#4), first swale (#6), mid barrier (#10), backbarrier (#8) and the backbarrier mud flat (#7). Generalised core logs are presented in Table 1 and the interpreted stratigraphy of the barrier in this transect is shown in Figure 5.

Seven depositional facies were identified on the basis of textural characteristics, stratigraphic position and heavy mineral occurrence: (1) a fine grained, well sorted, fining downward nearshore sand; (2) a medium to coarse grained, moderately sorted, coarsening downward lower beach sand; (3) a fine to medium grained, well sorted upper beach sand containing heavy minerals and thin coarse sand layers; (4) a fine to medium grained, well sorted eolian sand; (5) a fine grained, poorly sorted, coarsening down washover sand; (6) an organic-rich, cohesive transgressive mud; and (7) an organic-rich estuarine mud. Erosive contacts occur between: (a) the coarse lower beach sediment and the fine nearshore sand; (b) the nearshore sand and the transgressive mud; and (c) the washover sand and the transgressive mud.

Figure 6 shows the average grain size distributions for five of the identified facies (the two mud units have not been included). It is apparent that the lower beach sediment is significantly coarser than the upper beach sand. The transition between these two facies was therefore made on

Table 1. *Generalised core logs and interpretation of the Northern transect of the Bucasia barrier. The elevations in m AHD indicate the upper level of the adjacent unit.*

Elevation	Thickness	Description	Interpretation
Low Tide Terrace #5			
−1.2 m AHD	1.1 m	coarsening down unit, medium to coarse yellow/orange sand	lower beach
−2.3 m AHD	3.0 m	fining down unit, fine brown/grey sand, lower half large shell fragments	nearshore
−5.3 m AHD	0.2 m	brown/black mud	transgressive mud
	0.7 m	light grey muddy sand to sandy mud	transgressive mud
Runnel #4			
+0.5 m AHD	2.5 m	coarsening down unit, medium to coarse yellow/orange sand	lower beach
−2.0 m AHD	3.5 m	fine yellow/brown sand	nearshore
Swale #6			
+2.8 m AHD	0.5 m	fine yellow/orange sand	eolian
+2.3 m AHD	2.6 m	fine yellow/orange sand with low angle lamination and heavy minerals	upper beach
−0.3 m AHD	1.8 m	coarsening down unit, medium to coarse yellow/brown sand	lower beach
−2.1 m AHD	1.1 m	fine yellow/brown sand	nearshore
Mid Barrier #10			
+6.0 m AHD	1.5 m	fine yellow/brown sand	eolian
+4.5 m AHD	3.0 m	fine yellow/brown sand, heavy minerals and patches of coarse sand	upper beach
+1.5 m AHD	1.2 m	medium-coarse yellow/brown sand units	lower beach
+0.3 m AHD	0.3 m	fine yellow/brown sand	nearshore
Backbarrier #8			
+2.9 m AHD	5.5 m	coarsening down unit, fine to medium brown/grey sand, mica	washover
−2.6 m AHD	0.5 m	brown/black mud	transgressive mud
Mud Flat #7			
+2.4 m AHD	2.5 m	coarsening down unit, black/grey mud to brown/grey muddy sand, sand lenses	backbarrier mud flat
−0.1 m AHD	2.5 m	coarsening down unit, fine to medium brown/grey sand, coarse sand at the base	washover
−2.5 m AHD	1.1 m	brown/black mud	transgressive mud

the basis of the texture, since no erosional contact was present between these two units. The grain size distributions of the upper beach sand and the eolian sand are identical and these facies grade into each other. The distinction between upper beach and eolian facies was made on the basis of the occurrence of heavy minerals which tend to accumulate in the upper beach deposits (HAMILTON, 1993) and be preserved underneath the foredunes. In addition, the present day boundary between eolian and upper beach (between #4 and #6) is assumed to be just below the incipient foredune ridge (Figure 5) and above mean high water spring level (MHWS). The washover sediments are slightly coarser and more poorly sorted than the nearshore sand and significantly finer than the upper beach and eolian facies. The fine texture of the washover deposit can be explained by the fact that the washover core (#8) was taken on the distal part of the washover, in the proximity of the backbarrier mud flat.

The transect shown in Figure 5 represents a transgressive and a regressive sequence. The transgressive sequence consists of the transgressive basal mud and washover sediment (and possibly the lower part of the nearshore fine sand unit in #4, #5 and #6). The regressive sequence represents the nearshore, beach, eolian and backbarrier mud facies. Although the transgressive mud layer was only cored in #7, #8 and #5, it is inferred that it underlies the entire barrier and that the barrier transgressed over its own backbarrier

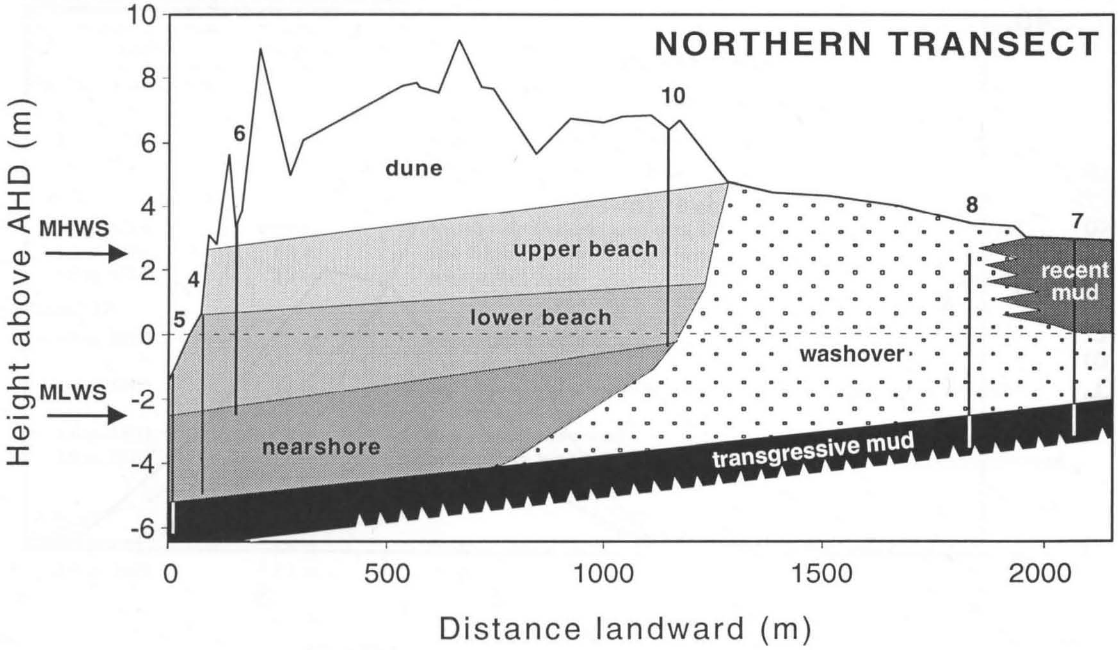

Figure 5. Reconstructed barrier morphostratigraphy of the Northern transect of the Bucasia barrier. MHWS and MLWS refers to mean high water spring and mean low water spring level, respectively.

estuarine deposits (the transgressive mud). Typically, the elevation of the base of the lower beach sand at mid barrier position (#6) is approximately 2 m higher than that of the more recent beach deposits (#5, #4 and #6), suggesting that in the past sea level must have been above the present level. Also, whereas the thickness of the total beach unit for these locations is comparable (4 m) the contribution of the lower beach facies is larger for the recent beach deposits (Table 1).

A large wood fragment was present in core #7 just below the start of the washover facies and embedded in the upper part of the transgressive mud at −2.9 m AHD. This wood fragment was radiocarbon dated at 6,800 ± 140 years BP (SUA#3076).

Southern Transect

The stratigraphy of the Southern transect is quite complex due to the vicinity of the transect to the tidal inlet and the consequent mixing of marine and estuarine deposits. Six cores have been taken (Figure 4); ebb tidal delta (#12), low tide terrace (#13), runnel between upper beach and low tide terrace (#3), first swale (#2), mid barrier (#1) and backbarrier (#11). Table 2 lists the generalised core logs and the stratigraphy of the barrier is shown in Figure 7. It should be mentioned that cores #1 and #11 are actually taken some distance away from the Southern transect; core #1 was taken on the estuarine intertidal area, and core #11 from a drainage ditch.

Six depositional facies were identified on the basis of textural characteristics and stratigraphic position: (1) a fine grained, moderately sorted, fining downward nearshore sand; (2) a medium to coarse grained, moderately to poorly sorted, beach sand containing shell and lithic material; (3) a fine to medium grained, well sorted eolian sand; (4) a composite backbarrier sand representing a mixture of washover, estuarine and tidal inlet sands; (5) a very coarse grained, poorly sorted tidal channel sand, containing abundant pebbles and shell material; and (6) an organic-rich, cohesive transgressive mud. Erosive contacts occur between (a) the tidal channel sediment and other facies and (b) the fine nearshore facies and the transgressive mud. Due to the proximity to

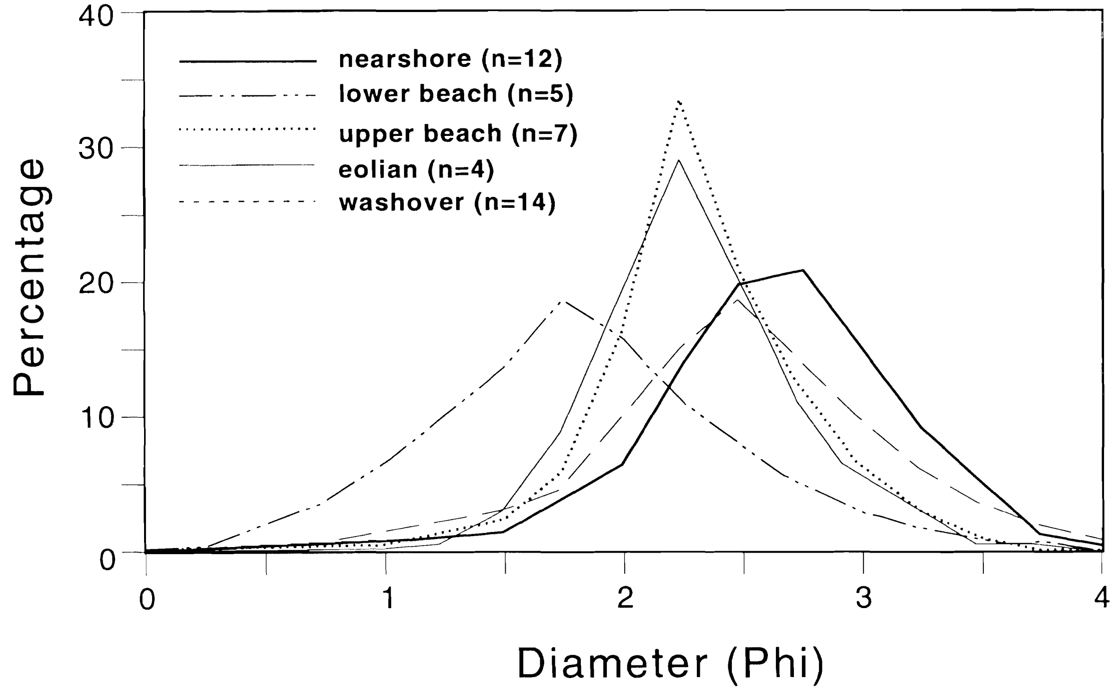

Figure 6. Average sediment distributions of the five identified sandy barrier facies.

the tidal channel, the overall sediment size in the nearshore off the Southern transect is coarser than in the Northern transect and no differentiation could be made between the upper and lower beach facies. In addition, a composite backbarrier sand was identified containing washover, estuarine and tidal inlet sands.

Several uncertainties are associated with the interpretation of the core from the backbarrier (#11) taken very close to the former connection between the Southern and the Northern estuaries (Figure 4). The top half of the core consists of a complex mixture of fine to medium sands and muds and is interpreted as a backbarrier sand. The two mud layers in the lower half of the core are assumed to belong to the transgressive mud facies, due to the similarity in texture, appearance and the correct stratigraphic position. The depositional environments of the coarse sand unit between the two mud layers and the medium sand at the bottom of the core are not known, but their coarseness suggests that they may be tidal channel deposits.

The interpretation of the mid barrier core (#1) also needs a further justification. This coring was performed at the foot of a recent scarp on the barrier. The scarping is caused by erosion of the barrier by a tidal channel and the top 1.5 m of this core, formed by estuarine deposits, is not considered representative of what can be found under the barrier. Field observations suggest that the top of the barrier at this location is formed by eolian deposits.

The stratigraphy of the Southern transect indicates that the entire barrier at this location is underlain by the transgressive mud layer with a thickness of at least 1.5 m (Figure 7). Over a distance of 1.5 km, the elevation of the top of this layer shows a progressive landward decrease in elevation of around 2.5 m. The elevation of the transgressive mud unit at the Southern transect is approximately two meters higher than in the Northern transect which is ascribed to the fact that the transgressive mud layer slopes in the seaward direction in combination with the limited barrier progradation along the Southern transect. In addition, higher wave energy levels at the Northern transect may have eroded part of the

Table 2. *Generalised core logs and interpretation of the Southern transect of the Bucasia barrier. The elevations in m AHD indicate the upper level of the adjacent unit.*

Elevation	Thickness	Description	Interpretation
Ebb Tide Terrace #12			
−1.5 m AHD	0.7 m	coarse yellow/orange sand with lithics/shells	tidal channel
−2.2 m AHD	1.3 m	fine brown/grey sand, some shells	nearshore
−3.5 m AHD	1.6 m	brown/black mud	transgressive mud
Low Tide Terrace #13			
−0.4 m AHD	0.9 m	coarse yellow/orange sand with lithics/shells	tidal channel
−1.3 m AHD	1.3 m	fine brown/grey sand, gravel layer at 0.9 m	nearshore
−2.6 m AHD	2.2 m	brown/black mud	transgressive mud
Runnel #3			
+0.5 m AHD	1.5 m	coarsening down unit, coarse yellow/orange sand, shells and lithics	beach
−1.0 m AHD	0.6 m	very coarse yellow/brown sand, iron stained pebbles	tidal channel
−1.6 m AHD	0.3 m	fine yellow/brown sand	nearshore
−1.9 m AHD	1.4 m	brown/black mud, organic content increases downward	transgressive mud
Swale #2			
+4.5 m AHD	1.5 m	medium yellow/orange sand, mainly quartz	eolian
+3.0 m AHD	3.3 m	alternating layers of medium/coarse yellow/orange sand, shells in coarse sand	beach
−0.3 m AHD	1.1 m	medium/coarse yellow/orange sand, lithic pebbles, very coarse at the base	tidal channel
−1.4 m AHD	0.1 m	sandy mud	transgressive mud
Mid Barrier #1			
+2.8 m AHD	1.3 m	complex unit with fine/medium sand and mud	backbarrier sand
+1.5 m AHD	1.4 m	medium/coarse grey well sorted sand	beach
+0.1 m AHD	1.2 m	complex unit containing mud, fine/coarse sand	backbarrier sand
−1.1 m AHD	0.9 m	brown/black mud	transgressive mud
Backbarrier #11			
+3.1 m AHD	3.4 m	complex unit containing fine/medium sand and sandy muds	backbarrier sand
+0.3 m AHD	0.4 m	brown/black mud	transgressive mud
	0.8 m	coarse sand	? channel
	0.3 m	brown/black mud	transgressive mud
−1.8 m AHD	1.1 m	medium sand	? channel

transgressive mud layer. The northward increase in wave energy level may also explain the less elevated nearshore facies in the Northern transect (−2.5 m AHD in the north versus −2 m AHD in the south).

Apart from the coarser sediments, the estuarine deposits and the smaller progradation in the Southern transect, the stratigraphy of the southern end of the Bucasia barrier is not dissimilar to that of the northern part. In particular, the heights of the washover surface for the two transects are identical (+4 m AHD) and also the elevations of the base of the landward most, and hence oldest, beach deposts are comparable (−0.1 m AHD in core #1 of the Southern transect and +0.3 m AHD of core #10 in the Northern transect). Also, both transects suggest that a higher sea level was present at the end of the PMT since the base of the oldest beach deposits is about 2 m higher than that of the present beach deposit.

Louisa Barrier

The Louisa barrier is located at the southern extremity of the study area. The barrier is about 2 km long, up to 300 m wide and is oriented in NNW–SSE direction (Figure 8). A relatively large, partially infilled estuarine system backs the barrier. An extensive ebb tidal delta in the western side of the barrier is associated with the mouth of the estuary. A very fine sand sheet is present

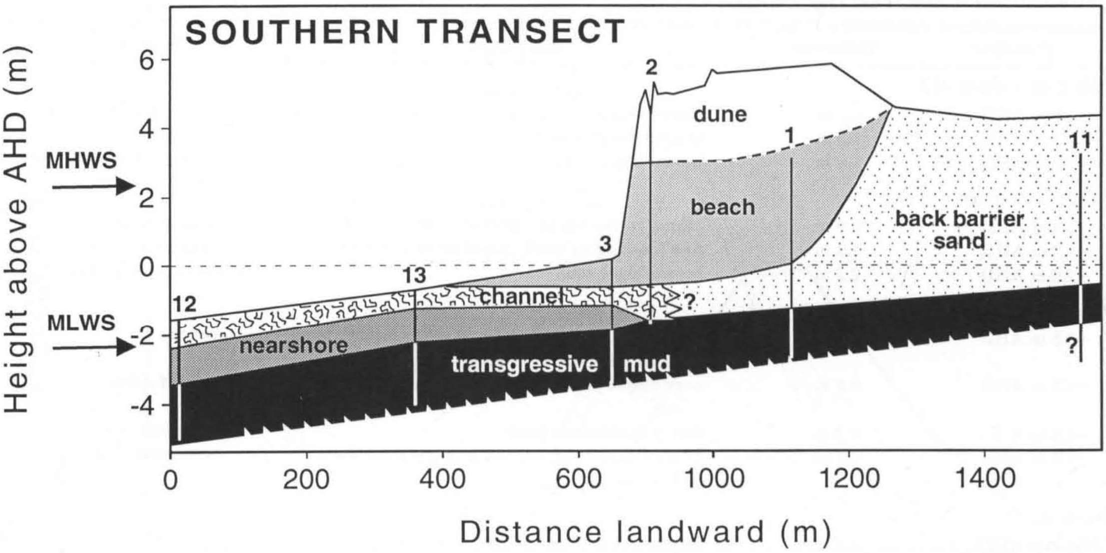

Figure 7. Reconstructed barrier morphostratigraphy of the Southern transect of the Bucasia barrier. MHWS and MLWS refers to mean high water spring and mean low water spring level, respectively.

behind the eastern part of the barrier, and is believed to be of Pleistocene age.

The intertidal zone is characterised by a steep upper part and a low gradient low tide terrace (tidal flat) with a deep runnel separating the two zones. The beach sediments are coarse and contain significant amounts of carbonate. The western part of the barrier is currently being eroded (total erosion > 50 m) as indicated by a 2 m dune scarp, whereas the eastern part of the barrier is characterised by incipient foredune development. An incipient foredune and two relict foredune ridges are present at the front of the uneroded part of the barrier. The relict foredunes are around 1.5 m high and reach an elevation of 7 m AHD. The overall elevation of the barrier increases slightly in a westward direction. Behind the foredunes, a washover deposit abuts the supratidal flats and the intertidal mangrove swamp.

The Louisa barrier is partially sheltered from the predominantly southeasterly waves and modal wave height is estimated at 0.4 m. Wave energy level increases towards the tidal inlet and is probably responsible for the increase in barrier elevation in that direction. The mean spring and neap tidal ranges for Louisa are 4.9 and 2.3 m, respectively, and the highest astronomical tide level is 3.8 m AHD (AUSTRALIAN TIDE TABLES, 1992).

Two transects were established across the Louisa barrier, one in the centre of the barrier and another across the eastern part of the barrier. However, due to limited penetration of the cores in the central barrier transect, caused by an underlaying gravel/boulder layer at 2 m depth, only the eastern transect will be discussed. Four cores were taken from this transect (Figure 8): low tide terrace (#5), beach (#9), first swale (#3) and back-barrier (#4). Generalised core logs are presented in Table 3 and the stratigraphy of the barrier is shown in Figure 9. As in the Bucasia barrier, the estuary behind the barrier was also extensively cored and the locations of these estuarine cores are shown in Figure 8 as well.

Five sedimentary facies were identified: (1) an organic-rich transgressive mud facies (also found at the base of #6, #9b, #24, #20, #19, and #21); (2) a fine grained, nearshore sand; (3) a medium to coarse grained beach sand containing very coarse material at the base of storm layers; (4) a fine grained, very well sorted eolian sand; and (5) a very fine grained, white sand with green/orange mottling facies which is tenatively interpreted as a Pleistocene (terrestrial) dune deposit. The latter

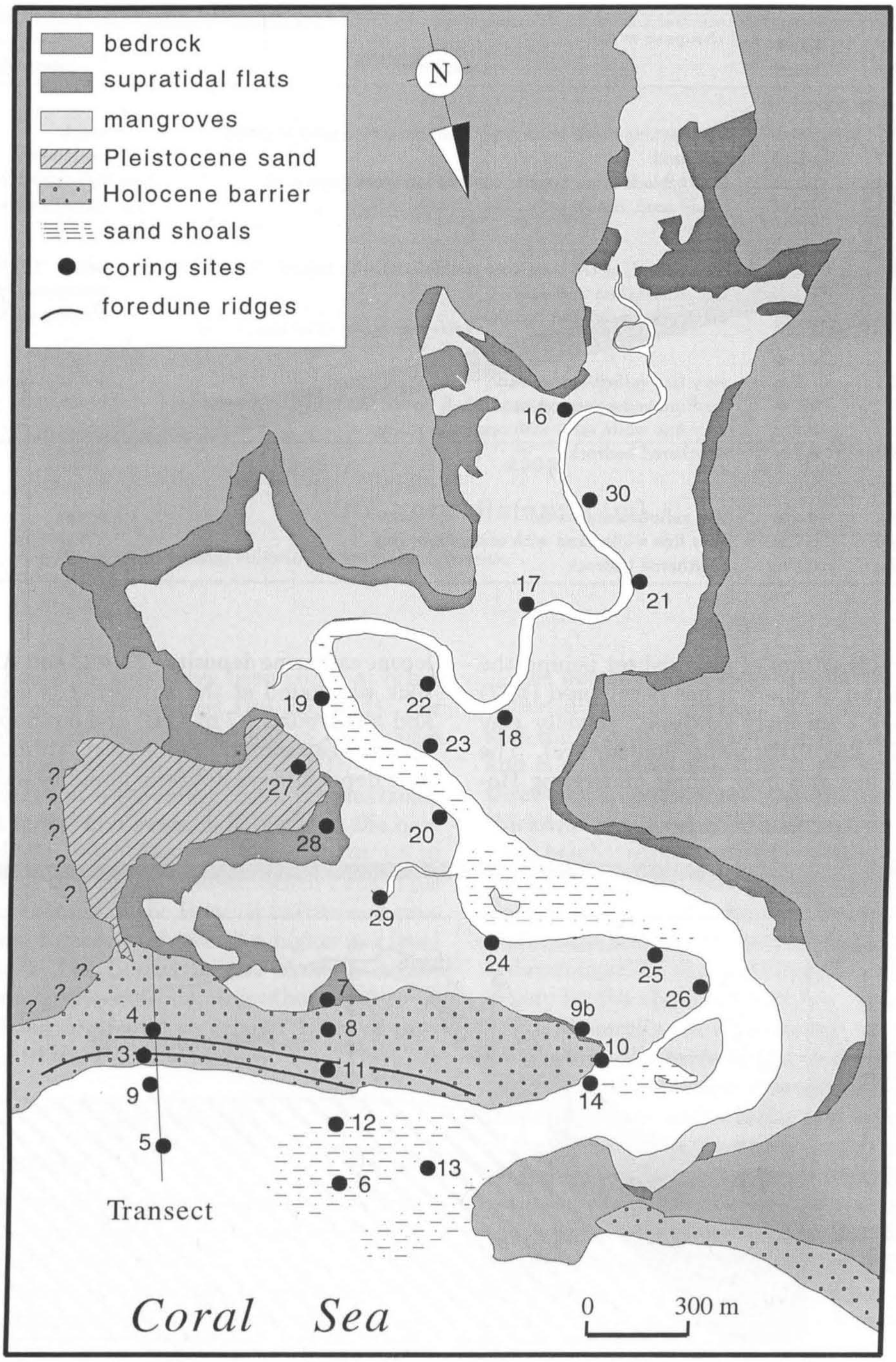

Figure 8. The geomorphology of the Louisa barrier system with the locations of the core sites.

Table 3. *Generalised core logs and interpretation of transect across the Louisa barrier. The elevations in m AHD indicate the upper level of the adjacent unit.*

Elevation	Thick-ness	Description	Interpretation
Low Tide Terrace #5			
−0.8 m AHD	0.6 m	3 coarsening down units from medium/coarse sand to gravel	beach
−1.4 m AHD	1.1 m	fine sand	nearshore
−2.5 m AHD	1.6 m	brown/black mud, organic content increases downward	transgressive mud
−4.1 m AHD	0.1 m	sandy mud, brown/grey	transgressive mud
Beach #9			
+2.5 m AHD	1.5 m	coarse sand, at the base very coarse sand with lithics	beach
+1.0 m AHD	1.1 m	fine sand	nearshore
−0.1 m AHD	0.1 m	very coarse sand and boulders	transgressive layer?
Swale #3			
+4.8 m AHD	1.2 m	very fine yellow/orange sand	eolian
+3.6 m AHD	2.1 m	medium yellow/orange sand, shell layers, lithics in the lower part	beach
+2.5 m AHD	0.8 m	very fine white sand with orange mottling	Pleistocene
+1.7 m AHD	0.3 m	weathered bedrock	bedrock
Backbarrier #4			
+5.8 m AHD	0.4 m	fine yellow/orange sand	eolian
+5.4 m AHD	3.7 m	very fine white sand with orange mottling	Pleistocene
+1.7 m AHD	0.1 m	weathered bedrock	bedrock

is present in the form of a sandsheet behind the barrier (Figure 9) where it has been cored (#27) and overlies a strongly oxidised, gravelly clay (supposedly the Pleistocene landsurface). The white sand has also been found underlying Holocene estuarine deposits (#7, #28 and #29). Bedrock was found at the bottom of two cores (#3 and #4) around 1.7 m AHD and outcrops a small distance eastward of the transect. Although washover deposits have not been found in the transect

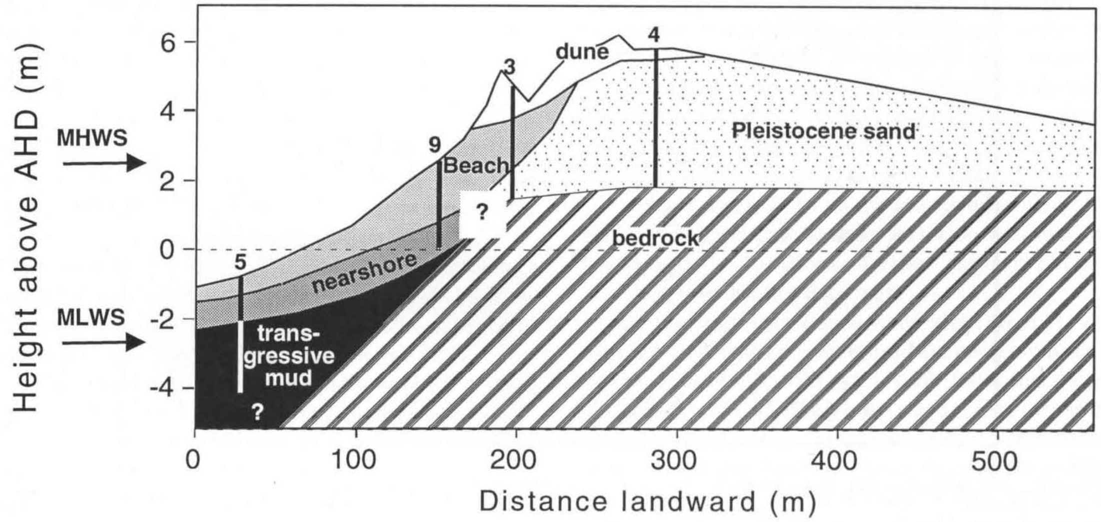

Figure 9. Reconstructed barrier morphostratigraphy of the Louisa barrier. MHWS and MLWS refers to mean high water spring and mean low water spring level, respectively.

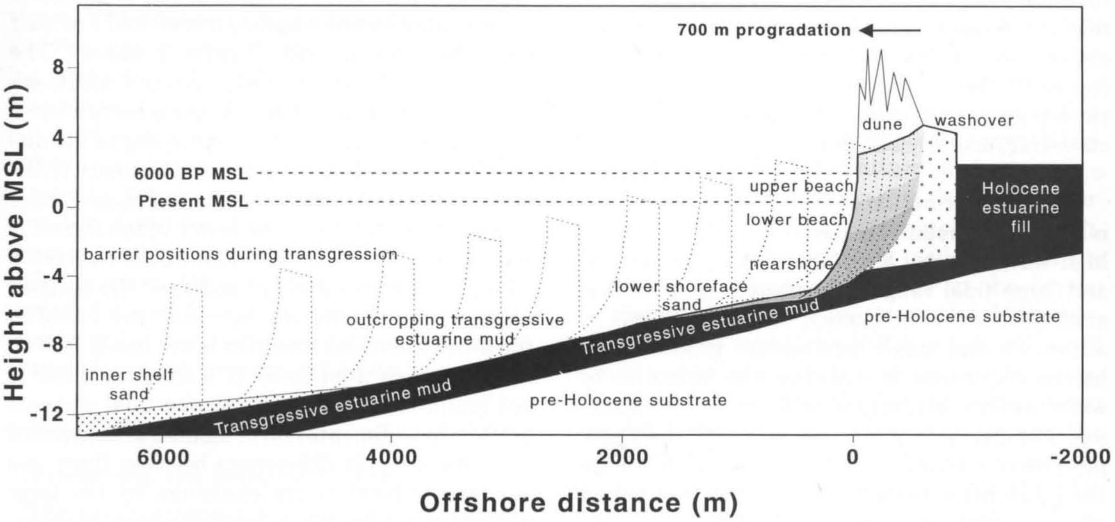

Figure 10. Simulation of the coastal evolution of the Bucasia barrier.

under discussion, they have been cored at other locations further to the west (#8 and #11). The elevation of the washover surface in the Louisa barrier ranges between 4 and 5 m AHD.

A comparison of the coring on the present beach and that in the first swale indicates that the base of the oldest beach deposit (#3) is about 1.5 m higher than that of the present beach (#9). This was also observed in the Bucasia barrier and provides some further evidence of a higher sea level in the past. The Louisa barrier shows a rather limited regressive episode, only about 100 m of progradation compared to around 1 km of progradation of the Bucasia barrier.

DISCUSSION

General Findings

Barrier morphostratigraphy depends on environmental conditions such as sediment availability, continental shelf gradient, wave energy level and tidal range. In general, the thickness of a barrier is a function of (1) the potential thickness of the barrier sequence, which is predominantly a function of oceanographical conditions (waves, wind and tides) and (2) the extent to which the potential thickness is reached, which is determined primarily by the inherited substrate gradient and sediment availability.

The potential thickness of a barrier is the elevation difference between the top of the dunes and the end of the shoreface or the wave base, and is considered the sum of the thicknesses of three facies, eolian, intertidal and nearshore. The thickness of the eolian sequence is dependent on wave height, wind conditions and sediment availability (SHORT and HESP, 1984), the intertidal facies thickness is by definition determined by the tidal range, and the extent of the nearshore facies is determined by the wave height, and to a lesser extent by the sediment size (*e.g.* HALLERMEIER, 1981). Generally, in high energy microtidal environments, the wave height is the most important parameter, because it largely determines the thickness of the eolian facies and especially that of the nearshore facies which constitutes the major part of the barrier sequence (SHORT, 1984). In low energy macrotidal environments, on the other hand, the intertidal facies may be the most significant component of the barrier sequence (see Figures 5, 8 and 10), and the potential barrier thickness will be strongly dependent on the tidal range.

The question as to whether the barrier will reach its potential thickness will depend on the extent of the barrier progradation. In order for the barrier to attain its potential thickness, the entire regressive sequence, from wave base to dune, should be present in a vertical sequence and this requires a significant barrier progradation. The relevant factors herein are the sediment availability and the slope of the substrate. Conditions conducive for the barrier to reach its potential thicknesss are an abundance of sediment and a steep continental shelf gradient (but not too steep; < 1°).

Thus, the formation of thick barrier sequences is favoured by relatively steep continental shelves, high wave energy, large amounts of sediments, and large tidal ranges. Conversely, gentle shelf gradients, low wave energy, limited amounts of sediments and small tidal ranges produce thin barrier sequences. In addition, the formation of wide barrier complexes is promoted by an abundance of sediment, a gentle continental shelf and, preferably, a falling sea level. Narrow barriers, on the other hand, form under conditions of a lack of sediments and a steep continental shelf. Summarising even more, the primary factor controlling the thickness of the barrier is the incident wave height, whereas the continental shelf gradient, sediment supply and sea-level history are the most important factors regarding the width of the barrier complex (for a more extensive discussion on this topic, see for example Roy *et al.*, in press).

The gradient of the continental shelf, the incident wave energy level and the tidal range are causally related. Large tidal ranges are commonly associated with wide continental shelves and low wave energy levels, whereas high wave energy environments are usually associated with narrow, steep continental shelves and microtidal tide ranges. Therefore, it may be suggested that macrotidal barriers are generally characterised by a limited thickness but, given an abundance of sediment supply, can form extensive barrier complexes. Unfortunately, this suggestion cannot be tested since macrotidal barriers have not yet been described in the literature.

In order to investigate the role of tides on the barrier morphostratigraphy, the macrotidal Bucasia barrier is compared to the classic prograded barrier on Galveston Island in the microtidal Gulf of Mexico. Apart from the tidal range, environmental conditions for these two locations (shelf gradient and wave energy level) are comparable. The generalised sequence of the low energy, microtidal prograded barrier system comprising Galveston Island demonstrates a total thickness of around 14 m composed of about 4 m eolian deposits, 1 m beach deposits (swash zone), 1 m upper shoreface deposits (surf zone), 6–7 m lower shoreface sediments (shoaling waves) and 2 m shelf mud (McCubbin, 1982; Figures 2 and 4). The coarsest sediments are found around mean sea level. The sequence of the Bucasia barrier has a smaller thickness, and the partitioning of the different facies to the overall barrier sequence is different. The total thickness of the beach and upper shoreface units (upper and lower beach facies) is around 4 m, which is twice that of the Galveston Island barrier sequence. In addition, the coarsest sediments in the Bucasia sequence are found in the lower intertidal zone (the lower beach facies). Also, the shelf mud facies of Galveston Island is not found in Bucasia since strong tidal currents inhibit the sedimentation of mud sized sediments. The stratigraphic differences between these two low energy barriers are explained by the large difference in tidal range, resulting in more extensive beach and upper shoreface facies in the macrotidal sequence.

Three beach/nearshore facies were identified in the Bucasia barrier as follows: (1) a fine grained, well sorted, fining downward nearshore sand (around 2.7 phi); (2) a medium to coarse grained, moderately sorted, coarsening downward lower beach sand (around 1.7 phi); (3) a fine to medium grained, well sorted upper beach sand containing heavy minerals and thin coarse sand layers (around 2.2 phi). The nearshore sand is located in the subtidal zone, the lower beach sand is found between mean low water spring (MLWS) and MSL, and the upper beach sand is positioned between MSL and mean high water spring (MHWS) swash level. The identification of these three facies corresponds to the results of Jago and Hardisty (1984) who investigated the sedimentology of a macrotidal beach (Pendine Sands, Wales) with morphological and hydrodynamic conditions similar to Bucasia Beach. Jago and Hardisty (1984) divide the foreshore into three zones on morphological and sedimentological grounds: the upper foreshore (from highest spring to neap high tide marks) the middle foreshore (between mean neap high and mean neap low tide mark) and the lower foreshore (below the mean neap low tide mark). Sediments of the upper foreshore are fine (around 2.6 phi), well sorted and contain heavy minerals. Over the mid-foreshore, the surface layer is similar to the upper foreshore sands but the subsurface sands are significantly coarser (around 2.4 phi) and contain large quantities of shell and lithic material. These sands make up almost the entire mid-foreshore section, and according to Jago and Hardisty (1984), the unit represents a storm lag produced under the surf zone at high water. The lower foreshore sands are fine (around 3.2 phi) and poorly sorted. On the basis of the sequence of sedimentary structures Jago and Hardisty (1984) suggest that swash zone processes establish the upper foreshore facies, storm breakers and surf, the mid-foreshore facies and shoaling waves, the lower foreshore facies. Thus, provided coarse material is available, either carbonate or lithic, a coarse lower intertidal beach deposit is typical on macrotidal beaches.

Site-Specific Findings

Stratigraphy and Sediment Supply

The two macrotidal barriers under investigation are characterised by a limited thickness and this is predominantly ascribed to the low wave energy level (wave height < 0.5 m) and the small gradient of the continental shelf (approximately 1/1,000). Underlying both barrier systems is a cohesive, estuarine, organic-rich mud layer and the preservation of this basal mud is indicative of a relatively fast rate of transgression and/or a low gradient of the surface being transgressed (Swift, 1975). A transitional mud facies is present between 8 and 12 m water depth (Figure 3), and it is inferred that this facies is in fact the basal transgressive mud layer exposed on the shoreface. In contrast, Hegarty (1983) and Jones (1987) interpreted this facies as a present-day deposit. This is considered unlikely because: (1) current measurements demonstrate that spring tidal currents in 10–15 m water depth are about 0.6 m/sec (Lessa, 1993; BPA, *unpublished data*) and it is not probable that mud sized sediments are deposited under these conditions and (2) the transitional mud unit was intercepted during shallow drilling as a 1 m thick brown/grey cohesive mud (Hegarty, 1983) and it is not probable that present-day, subaqueous mud deposition forms cohesive sediments.

The nature and the amount of sediment contained in the barriers is different for the two studied systems and reflects different sediment sources. The Louisa barrier contains large amounts of shelly material and is only 300 m wide, whereas the Bucasia barrier has a width of 1 to 2 km and does not contain significant amounts of carbonate material. It is suggested that the Louisa barrier, and also other barriers south of Hay Point (Figure 3), are supplied by the continental shelf, whereas the Bucasia barrier, and neighbouring barriers, are predominantly supplied by the Pioneer River. It is also inferred that the small, carbonate rich barrier systems at Cape Hillsborough are predominantly supplied by the continental shelf.

To pursue the suggestion that the Bucasia barrier and adjacent barriers are supplied by the Pioneer River, a comparison is made between the amount of sediments contained in the barrier complexes and nearshore zones between Sandringham Bay and Cape Hillsborough (Figure 3) and the total amount of sediment discharged by the Pioneer River since the end of the PMT. Four main barrier complexes are present around the mouth of the Pioneer River (Figure 3): the Bucasia barrier, the Black's Beach barrier, the Harbour Beach barrier and the thin barrier system south of the mouth of the Pioneer River. Assuming a mean overall thickness of these barrier systems of 8 m and measuring the subaerial extent of the barriers using aerial photographs, the amount of sediment contained in each of these systems is respectively 5×10^7, 1×10^7, 10×10^7 and 3×10^7 m^3 of sediment. Adding a 50 km long, 3 km wide and 1 m thick nearshore sand facies to the amount of sediment contained in the barrier systems yields a total amount of around 34×10^7 m^3 of sediment. Assuming that this amount of sediment has accumulated over the past 6,000 years, this implies a sediment supply to the coast of 6×10^4 m^3/yr. It should be pointed out that an unknown but probably significant proportion of the sediment in the nearshore area was already present 6,000 years ago in the form of transgressive barriers. The required supply to the coast will therefore be less than 6×10^4 m^3/yr.

Presently, the Pioneer River supplies sand and gravel to the coast at an estimated yield of 4×10^4 m^3/yr (Gourlay and Hacker, 1986). It should be noted that the present location of the mouth of the Pioneer River is very recent (early Holocene; Gourlay and Hacker, 1986), and that the sediment yield was readily supplied to the nearshore area since no paleo valley needed to be filled. In addition, it seems probable that the sediment yield of the Pioneer River during the early stages of its development was larger than present due to incision and erosion associated with the establishment of the new channel. It may be concluded that the nearshore area between Sandringham Bay and Sand Bay was predominantly supplied by the

Pioneer River and sand supply from the inner continental shelf was of secondary importance.

A simulation model was used to illustrate the evolution of the Bucasia barrier (Figure 10). The model is described in COWELL *et al.* (1992) and involves the translation of a predetermined barrier form across a substrate under the influence of falling or rising sea level whilst maintaining its shoreface shape. The pre-Holocene substrate was taken from Figure 3 and the simulated barrier was given a width of 300 m (equivalent to the width of the washover facies) and a thickness of 8.5 m (corresponding to the distance from MHWS level to the toe of the shoreface). Backbarrier estuarine mud is deposited behind the barrier at a rate of 0.8 mm/yr. Figure 10 shows the result of the simulation and demonstrates that during the transgression, the barrier can translate over its own back barrier estuarine mud deposits. The end of the simulated transgression is reached at 6,000 years BP with mean sea level 2 m above present, and the (transgressive) barrier is primarily composed of washover deposits (Figure 10). Under the influence of a gradually falling sea level the barrier subsequently progrades, and in order to accommodate for the approximately 700 m barrier progradation (the average progradation over the embayment), 1 m^3 of sediment per year per unit beach width was added to the system. Due to the lower sea level, the barrier partly erodes its own estuarine mud deposits, reducing the thickness of the transgressive mud deposit. To reproduce what is observed in the field, an inner shelf sand sheet and a veneer of lower shoreface sand was manually added on top of the transgressive estuarine mud, and in addition the barrier was capped with dunes.

The simulation model was run with the purpose to reproduce the observed stratigraphy of the Bucasia barrier and inner shelf (compare Figure 10 with Figure 5) and does not "prove" the proposed evolution. The rate of mud deposition was "tuned" to reproduce the observed 1–2 m thickness, and sediment was added to the prograding barrier to accommodate for the observed width of the present barrier. However, the model demonstrates how the barrier complex can be underlain by its own estuarine mud deposits, and illustrates how this mud can be outcropping at a depth of around 10 m (see Figure 3). In addition, the model highlights the fact that only a small amount of sediment needs to be added to the barrier system to produce the observed progradation. Assuming a width of the Bucasia embayment of 5 km, only 5,000 m^3 of sediment needs to be added per year, and in light of the estimated northward longshore drift in the nearshore area of the barrier south of Bucasia (25,000 m^3/yr; BPA, 1983), the sediment required for the progradation of the Bucasia barrier could easily have been provided by longshore transport.

Sea Level and Morphodynamics

Morphostratigraphies of both investigated barrier systems indicate that higher sea levels were present in the past and that consequently the regressive sequence formed under conditions of a falling sea level. The results from the Bucasia barrier (Figures 5 and 7) suggest that sea level at the maximum of the PMT was approximately 2 m higher than at present, in accordance with the isostatic adjustments proposed by THOM and CHAPPELL (1978), CHAPPELL *et al.* (1982) and HOPLEY (1983). Results from the Louisa barrier (Figure 9) are less conclusive and seem to suggest that sea level was about 1.5 m higher in the past.

A wood fragment present in the transgressive estuarine mud at −2.9 m AHD was dated at 6,800 ± 140 C^{14} years BP (SUA#3067). The mud is interpreted as an intertidal mangrove mud and because the rate of sea-level rise was still relatively fast around this time (> 4 mm/yr; THOM and ROY, 1983), it is suggested that the transgressive mud was deposited in the lower intertidal zone (from MSL to −2 m MSL). It is therefore inferred that the sea level of 6,800 years BP was between −2.9 m AHD and −0.9 m AHD, depending on whether the wood fragment was deposited at MSL or at the lower intertidal limit, respectively. Assuming that sea level rose at a mean rate of 4 mm/yr during the last phase of the postglacial marine transgression, the +2 m AHD sea level at the end of the PMT was reached between 5,600 and 6,100 years BP. This is somewhat later, but within the range, of previous estimates of the end of the postglacial marine transgression of 6,500–6,000 years BP given by CHAPPELL *et al.* (1983), THOM and ROY (1983, 1985) and GAGAN *et al.* (1994).

There are some indications that at the end of the PMT the tidal range may have been smaller and the wave energy level may have been higher than present, supporting the presence of a mid-Holocene wave and tidal window. These indications originate from a comparison of the present and past beach deposits of the Bucasia barrier in the Northern transect. Figure 5 shows that although the total thickness of the beach facies remained approximately constant during barrier progradation (4 m), the contribution of the coarse grained, lower beach sand unit increased, whereas the thickness of the fine to medium grained, upper beach sand unit decreased. The lower beach sand is interpreted as a storm surf zone deposit (JAGO and HARDISTY, 1984) and is located between MLWS and MSL, whereas the upper beach sand is considered a swash zone deposit (JAGO and HARDISTY, 1984) and extends from MSL to MHWS swash level. It is suggested that the thickness of the lower beach unit is related to the tidal range with larger tidal ranges promoting thicker deposits. The thickness of the upper beach sand unit is considered more a function of the wave energy level with higher waves favouring the formation of high berms and thick upper beach deposits. Although the above assumptions are tentative, in conjuction with Figure 5 it may be suggested that at the end of the PMT the tidal range was smaller and the waves were larger. One might argue that, during the early stages of Holocene barrier progradation, the wave energy level could have been less due to the increased amount of headland sheltering and wave energy loss by refraction (see Figure 4). More work is needed to determine whether a wave and tidal window, with respectively larger waves and smaller tidal ranges, has existed during the mid-Holocene.

CONCLUSIONS

The central Queensland coastline represents a macrotidal, fetch-limited and low wave energy environment with a low gradient continental shelf. Offshore sediment and seismic data indicate an inner shelf depleted of Holocene sediments, with a pre-Holocene substratum outcropping in many places. The character of the coastline is controlled by the dominance of either waves or tides and varies along the coast due to the differences in exposure to the waves. The dominance of one over the other is strongly influenced by the alignment of the coast to the predominant southeasterly wind and wave direction. Wide intertidal flats characterise areas sheltered from wave activity whereas sandy barrier systems occur in areas not sheltered by the waves.

The morphostratigraphy of two of these barrier systems was investigated. The studied barriers have many characteristics in common with the more extensively studied microtidal barrier systems and may be classified as prograding and regressive barriers with transgressive and regressive sequences (THOM *et al.*, 1978). The transgressive sequence formed during the Holocene transgression and evolved according to the classic mechanism of "shoreface retreat" as evidenced by the basal mud layer underlying the barrier. The regressive sequence formed under a falling sea level and, for the barrier systems north of the mouth of the Pioneer River, an abundance of fluvial sediment supply. The thickness of the macrotidal barrier sequences is around 10 m, and the limited vertical extent of the barriers is ascribed to the low wave energy level and the gentle gradient of the continental shelf (1/1,000) off the central Queensland coastline.

The morphostratigraphies of the barrier systems indicate that sea level was at 1.5 to 2 m higher than present at the end of the postglacial marine transgression estimated at 5,600–6,100 years BP. This is in general agreement with models of isostatic rebound and field evidence from northern Queensland. The character of the sediments contained in the barriers and the prevailing northward littoral and nearshore drift within the study area suggests that the barriers south of the Pioneer River are predominantly supplied by the continental shelf whereas the barrier systems north of the river mouth are composed of sediments mainly provided by the Pioneer River.

ACKNOWLEDGEMENTS

We would like to thank mastercorer Graham Lloyd and his assistants Robert Brander, Mark McDonald, Andy Short, Ian Turner and Mark F. McDonald for obtaining the vibracores. Mark F. McDonald performed the grain size analysis for the Bucasia barrier. We are grateful to Andy Stephens and Mal Jones (Department of Resources Industries, Queensland) for making the seismic and sedimentological reports for the nearshore area available, and Michael Allen (Beach Protection Authority, Queensland) for providing tidal current data. Prof. J.C. Dionne, Dr. P. Roy, Dr. A.D. Short and an anonymous reviewer gave useful comments to the manuscript and Dr. P. Cowell ran his Shoreface Translation Model for us (Figure 10). We would also like to thank Dr. A.D. Short for providing part of the funds for the field trip (National Greenhouse Advisory Committee) and Dr. P. Roy for financing the radiocarbon date. Additional funds were generously provided by the Department of Geography, University of Sydney. G.M. and G.L. are respectively sponsored by an Overseas Post-graduate Research Scholarship

(OPRS) and a Brazilian Government scholarship (CNPq).

LITERATURE CITED

Australian National Tide Tables, 1992. Papua, New Guinea and Antarctica. *Australian Hydrographic Publication 11.* Canberra: Australian Government Publishing Service.

Beets, D.J.; Van Der Valk, L., and Stive, M.J.F., 1992. Holocene evolution of the coast of Holland. *Marine Geology,* 103, 423–443.

Bernard, H.A.; LeBlanc, R.J., and Mayor, C.F., 1962. Recent and Pleistocene geology of southeast Texas. *In:* Rainwater, E.H. and Zingula, R.P. (eds.), *Geology of the Gulf Coast and Central Texas, Guide Book of Excursions.* Houston Geological Society, 175–224.

Bode, L. and Stark, K.P., 1983. Simulation of tides and currents in the Mackay region. *Proceedings of the 6th Australian Conference on Coastal and Ocean Engineering,* pp. 85–89.

BPA, 1979. *Capricorn Coast Beaches.* Queensland: Beach Protection Authority, 238p.

BPA, 1983. *Sediment Transport Investigation for a Section of the Coast Near Eimeo, North of Mackay.* Queensland: Beach Protection Authority (unpublished).

BPA, 1986. *Wave Data Recording Programme: Mackay Region.* Queensland: Beach Protection Authority.

Chappell, J.; Chivas, A.; Wallensky, E.; Polach, H.A., and Aharon, P., 1983. Holocene paleo-environmental changes, central to north Great Barrier Reef inner zone. *BMR Journal of Australian Geology and Geophysics,* 8, 223–235.

Chappell, J.; Rhodes, E.G.; Thom, B.G., and Wallensky, E., 1982. Hydro-isostasy and the sea level isobase of 5,500 BP in north Queensland, Australia. *Marine Geology,* 49, 81–90.

Cook, P.J. and Mayo, W., 1978. Sedimentology and Holocene history of a tropical estuary, Broad Sound, Queensland. *Bureau of Mineral Resources, Geology and Geophysics, Bulletin no. 170.*

Cook, P.J. and Polach, H.A., 1973. A chenier sequency at Broad Sound, Queensland, and evidence against a Holocene high sea-level. *Marine Geology,* 14, 253–268.

Cowell, P.J.; Roy, P.S., and Jones, R.A., 1992. Shoreface Translation Model; computer simulation of coastal sand body response to sea-level rise. *Mathematics and Computers in Simulation,* 33, 603–608.

Davies, P.J. and Marshall, J.F., 1979. Aspects of Holocene reef growth—Substrate age and accretion rate. *Search,* 10, 276–279.

Davis, R.A. and Hayes, M.O., 1984. What is a wave-dominated coast? *Marine Geology,* 60, 313–329.

Department of Harbours and Marine. 1981. *Mackay Boat Harbour Investigation.* Brisbane: Department of Harbours and Marine (unpublished).

Easton, A., 1970. Tides of the continent of Australia. *Horace Lamb Centre, Research Paper No. 37.* The Flinders University of South Australia.

Flood, P.G., 1983. Holocene sea level data from the southern Great Barrier Reef and southeastern Queensland—A review. *In:* Hopley, D. (ed.), *Australian Sea levels in the Last 15,000 Years.* Department of Geography, James Cook University, Monograph Series, 85–92.

Gagan, M.K.; Johnson, D.P., and Crowley, G.M., 1994. Sea level control of stacked late Quaternary coastal sequences, central Great Barrier Reef. *Sedimentology,* 41, 329–351.

Gourlay, M.R. and Hacker, J.L.F., 1986. *Pioneer River Estuary Sedimentation Studies.* St. Lucia: University of Queensland Printery.

Griffin, D.A.; Middleton, J.H., and Bode, L., 1987. The tidal and longer-period circulation of Capricornia, Southern Great Barrier Reef. *Australian Journal of Marine and Freshwater Research,* 38, 461–474.

Hallermeier, R.J., 1981. A profile zonation for seasonal beaches from wave climate. *Coastal Engineering,* 4, 253–277.

Hamilton, N., 1993. Geomorphological Controls on Coastal Placer Distribution. Unpublished Ph.D. Thesis, Department of Geography, University of Sydney.

Harvey, N., 1986. *The Great Barrier Reef: Shallow Seismic Investigations.* Monograph Series No. 14. Department of Geography, James Cook University.

Hegarty, R. A., 1983. *Results of a Continuous Profiling Survey in the Mackay Area.* Geological Survey of Queensland, Record 1983/19 (unpublished).

Hill, M.C. and Fitzgerald, D.M., 1992. Evolution and Holocene stratigraphy of Plymouth, Kingston, and Duxbury Bays, Massachusetts. *In:* Quaternary Coasts of the United States: Marine and Lacustrine Systems, *SEPM Special Publication No. 48,* 45–56.

Hopley, D., 1982. *The Geomorphology of the Great Barrier Reef: Quaternary Development of Coral Reefs.* Brisbane: Wiley.

Hopley, D., 1983. Evidence of 15,000 years of sea level change in tropical Queensland. *In:* Hopley, D. (ed.), *Australian Sea Levels in the Past 15,000 Years.* Department of Geography, James Cook University, Monograph Series, pp. 93–104.

Hopley, D., 1984. The Holocene high energy window on the central Great Barrier reef. *In:* Thom, B.G. (ed.), *Coastal Geomorphology in Australia.* New York: Academic, pp. 137–150.

Hopley, D. and Kinsey, D.W., 1988. The effects of a rapid short-term sea-level rise on the Great Barrier Reef. *In:* Pearman, G.I. (ed.), *Greenhouse.* Leiden: Brill, pp. 189–201.

Jago, C.F. and Hardisty, J., 1984. Sedimentology and morphodynamics of a macrotidal beach, Pendine Sands, SW Wales. *Marine Geology,* 60, 123–154.

Jones, M.R., 1987. *Nearshore Sediments and Distribution Patterns, Mackay Coast.* Queensland Department of Mines Record Series, 1987/25 (unpublished).

Kraft, J.C., 1971. Sedimentary facies patterns and geologic history of a Holocene marine transgression. *Geological Society of America Bulletin,* 82, 2131–2158.

Lessa, G., 1993. Tidal currents, tidal inlets and sediment dynamics on central Queensland coast; a preliminary analysis. *In:* Thoms, M.C. (ed.), *Catchments and Coasts in Eastern Australia.* Research Monograph No. 5. Sydney: University of Sydney Press, Chapter 5, pp. 50–59.

Lessa, G. and Masselink, G., in preparation. Stratigraphy and Holocene evolution of macrotidal estuaries in central Queensland, Australia—A morphodynamic approach.

MARSHALL, J.F. and DAVIES, P.J., 1984. Facies variation and Holocene reef growth in the southern Great Barrier Reef. *In:* THOM, B.G. (ed.), *Coastal Geomorphology in Australia.* New York: Academic, pp. 125–134.

MARSHALL, J.F. and THOM, B.G., 1976. The sea level in the last interglacial. *Nature,* 263, 120–121.

MCCUBBIN, D.G., 1982. Barrier-island and strand plain facies. *In:* SCHOLLE, P.A. and SPEARING, D. (eds.), *Sandstone Depositional Environments.* Tulsa, Oklahoma: American Society of Petroleum Geologists, pp. 247–280.

MCLEAN, R.F.; STODDART, D.R.; HOPLEY, D., and POLACH, H., 1978. Sea level change in the Holocene on the northern Great Barrier Reef. *Philosophical Transactions Royal Society of London,* 291, 167–186.

MIDDLETON, J.H.; BUCHWALD, V.T., and HUTHNANCE, J.M., 1984. The anomalous tides near Broad Sound. *Continental Shelf Research,* 3, 359–381.

ROY, P.S.; COWELL, P.J.; FERLAND, M., and THOM, B.G., in press. Wave-dominated coasts. *In:* CARTER, R.W.G. and WOODROFFE, C. (eds.), *Coastal Evolution.* Cambridge: Cambridge University Press.

SHORT, A.D., 1984. Beach and nearshore facies: Southeast Australia. *Marine Geology,* 60, 261–282.

SHORT, A.D., and HESP, P.J., 1984. Wave, beach and dune interactions in southeastern Australia. *Marine Geology,* 48, 259–284.

STARK, K.P.; BODE, L., and MASON, L.B., 1984. Simulation of tides and storm surges in the Great Barrier Reef region. *Proceedings 19th International Conference on Coastal Engineering,* pp. 226–242.

SWIFT, D.J.P., 1975. Barrier island genesis; evidence from the central Atlantic shelf, eastern U.S.A. *Sedimentary Geology,* 14, 1–43.

THOM, B.G.; BOWMAN, G.G., and ROY, P.S., 1981. Late Quaternary evolution of coastal sand barriers, Port Stephens-Myall Lakes area, central New South Wales, Australia. *Quaternary Research,* 15, 345–364.

THOM, B.G. and CHAPPELL, J., 1978. Holocene sea-level change: An interpretation. *Philosophical Transactions Royal Society of London,* 291, 187–194.

THOM, B.G.; POLLACH, H.A., and BOWMAN, G.M., 1978. *Holocene Age Structure of Coastal Sand Barriers in New South Wales, Australia.* Duntroon, ACT: Department of Geography, University of New South Wales.

THOM, B.G. and ROY, P.S., 1983. Sea level change in New South Wales over the past 15,000 years. *In:* HOPLEY, D. (ed.), *Australian Sea Levels in the Last 15,000 Years.* Department of Geography, James Cook University, Monograph Series, pp. 64–84.

THOM, B.G. and ROY, P.S., 1985. Relative sea levels and coastal sedimentation in southeast Australia in the Holocene. *Journal of Sedimentary Petrology,* 55, 257–264.

THOM, B.G.; SHEPHERD, M.; LY, C.K.; ROY, P.S.; BOWMAN, G.M., and HESP, P.A., 1992. *Coastal Geomorphology and Quaternary Geology of the Port Stephens-Myall Lakes Area.* Department of Biogeography and Geomorphology, Australian National University. Monography No. 6.