The Development of the Princess Charlotte Bay Chenier Plain

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THE DEVELOPMENT OF THE PRINCESS CHARLOTTE BAY CHENIER PLAIN

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A thesis submitted in fulfilment of the requirements for the degree of Doctor of Philosophy. School of Physical, Environmental and Mathematical Sciences, University College, University of New South Wales.

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Abstract:

Chenier plains record changes in the mode of coastal progradation between periods of mudflat progradation and coarse sediment deposition. There are several environmental forces responsible for these changes. Chenier research worldwide has led to an understanding of these forces in some settings. In Australia, however, the causes of these changes are still not well understood. Several conflicting theories have been proposed to explain the development of a chenier plain in Princess Charlotte Bay on the eastern coast of Cape York, northern Queensland, Australia. These include an internal dynamic, climate fluctuations, and storm activity. Since these were put forward, both our methods and our understanding of Holocene environmental processes have improved. Improvements include new dating methods with luminescence techniques. Improvements to our understanding include a body of research regarding Holocene climate and sea level changes. These have allowed a re-examination of the development of the Princess Charlotte Bay chenier plain.

Data was obtained from a different section of the bay than was examined previously. Optically-stimulated luminescence and radiocarbon ages from eight of the 11 chenier ridges yielded estimates of Holocene chenier ridge building phases. These were: (1) An early phase from around 4000 yr BP – 2000 yr BP (seven ridges built); (2) a phase centred on approximately 1350 yr BP (two ridges built); and (3) a phase since 820 yr BP (one ridge built). Six models for chenier plain formation were tested. It was proposed that the development of the chenier plain has been driven by several environmental forces acting in a morphodynamic hierarchy. From first order to fourth order, these forces are: (1) Holocene climate and sea level changes, (2) tropical cyclone activity, (3) mangrove distribution, and longshore currents, and (4) river channel changes.

A statistical reassessment of Australian chenier ridge ages was also undertaken and revealed periods of non-localised chenier ridge building across multiple northern Australia sites at around 620, 1120, 1850, and 2230 cal yr BP, and around the entire continent between 2400 and 1300 cal yr BP. These results provide further support that sea level oscillations have played a role in chenier plain development in Australia.

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Abstract

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Contents

Originality statement

I hereby declare that this submission is my own work and to the best of my knowledge it contains no materials previously published or written by another person, or substantial proportions of material which have been accepted for the award of any other degree or diploma at UNSW or any other educational institution, except where due acknowledgment is made in the thesis. Any contribution made to the research by others, with whom I have worked at UNSW or elsewhere, is explicitly acknowledged in the thesis. I also declare that the intellectual content of this thesis is the product of my own work, except to the extent that assistance from others in the project's design and conception or in style, presentation and linguistic expression is acknowledged.

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Chapter One - Introduction

1.1 Australian chenier plains and Princess Charlotte Bay

Chenier plains form in settings where mudflat progradation is periodically interrupted by phases of coarse sediment deposition. This results in the building of shore-parallel sandy/shelly ridges, known as chenier ridges. Dating ridges in a chenier plain provides information about palaeoenvironmental changes that have occurred in a particular setting. In Australia, and particularly the northern half of the continent, more chenier plain studies have been conducted than anywhere else in the world. A number of competing models have been put forward to explain their development. Despite this history, knowledge gained from studying chenier plains has yet to be incorporated into general models of coastal behaviour around Australia. For example, in a review of Australian geomorphology to 2004 (Thom and Short, 2006), the word „chenier" appears twice in the context of features that can be found on tropical coasts, and only one review article is cited (Short, 1989). Furthermore, Australian chenier plain studies have decreased since the 1980s. This situation has occurred largely due to a general lack of consensus among researchers as to the most appropriate model to explain the development of Holocene chenier plains in Australia. Two factors that may have contributed to this lack of consensus are (1) appropriateness of dating methods used in previous studies, and (2) the availability of palaeoenvironmental records for comparison to chenier plain data.

The appropriateness of radiocarbon dating of shell material in chenier studies has been questioned by many authors as there are at least two possible sources of error. The first are chemical influences on the shell material due to both the marine environment, and the presence of „old" carbon leached from older chenier ridges being incorporated into new shells (Gillespie and Polach, 1979; Walker, 2005). The second involves temporal differences between when a shell is formed, and when it becomes part of a chenier (e.g. Rhodes *et al.*, 1980; Lees, 1992a; Woodroffe and Grime, 1999).

The second factor that has contributed to lack of consensus among researchers regarding chenier plain development models in Australia has been a lack of palaeoenvironmental records available for comparison to chenier plain data, particularly regarding Holocene climate and sea level changes. For example, although Holocene climate fluctuations have been put forward as a force driving the development of Australian chenier plains, very few studies had reconstructions (e.g. derived from palynology) available for comparison to support this conclusion. Similarly, although past investigations have considered evidence for the effect of sea level changes in chenier plains, practically every Australian study has referred to the sea level curve of Chappell (1983), which suggested a smoothly falling sea level from approximately +1 m to present level over the last 6000 years. Because of this, the possible influence of minor fluctuations/oscillations in sea level has been largely ignored.

Problems with dating methods and lack of records for comparison have become much less important in recent years. Firstly, the use of luminescence techniques to date chenier ridge building events avoids the issues associated with radiocarbon methods. Since luminescence

techniques directly date „last exposure to sunlight" of sand grains, chenier ridge ages from this study should more closely reflect the actual time of chenier ridge building than those obtained in the past. Secondly, and more importantly, a vast body of knowledge now exists with regards to Holocene palaeoclimates, sea level changes and storm activity. These improvements allow more rigorous testing of chenier plain development models than was possible in the past.

In short, Australian chenier plains are once more a fruitful research topic. This study involved a reassessment of the development of one of Australia"s largest chenier plains, in Princess Charlotte Bay on the east coast of northern Queensland, as well as a consideration of broader relationships that may exist across Australian chenier plains. The geomorphic development of the Princess Charlotte Bay chenier plain has been the topic of a number of research papers in the past. The most important of these (Chappell and Grindrod, 1984) concluded that the development of the plain was driven by an internal dynamic and contributed greatly to the advancement of chenier plain studies in Australia. The current study builds on the research of Chappell and Grindrod (1984), and others, in two primary ways. Firstly, a different portion of the bay is examined. Examining a different section of the chenier plain provides a perspective on how processes may differ across the bay, and a more comprehensive consideration of the morphodynamics of the entire chenier plain. Secondly, this study incorporates recent advances in knowledge regarding changes in late Holocene palaeoenvironmental processes. This knowledge was not available to Chappell and Grindrod (1984) and allow reconsideration of the „internal dynamic" model. Broader relationships across Australian chenier plains were also considered by analysing chenier ridge ages from multiple chenier plains in a similar fashion as was undertaken by Lees and

Clements (1987). As with the reassessment of the development of the Princess Charlotte Bay chenier plain, advances in knowledge regarding changes in late Holocene palaeoenvironmental processes allowed a better consideration of possible forcing factors. This exercise provided a broader perspective for the conclusions drawn from the new Princess Charlotte Bay dataset and lessons for chenier plain geomorphology in Australia.

1.2 The value of studying chenier plains

Chenier ridges provide a sensitive record of changes to the depositional mode of shorelines. Because their composition makes them amenable to dating techniques, it becomes possible to reconstruct timings of changes in shoreline processes. This in turn can provide information about palaeoenvironmental changes. The vast majority of chenier plains have formed during the Holocene, with Pleistocene ages having been suggested in just a few cases (e.g. Veen, 1970 in Augustinus, 1989). Examples of palaeo-processes examined in past studies include fluvial avulsions (Russell and Howe, 1935; McBride *et al.*, 2007), climate fluctuations (e.g. Rhodes, 1980; Rhodes *et al.*, 1980; Rhodes, 1982; Lees, 1987; Lees and Clements, 1987; Lees, 1992a; Lees, 1992b), and sea level changes (e.g. Schofield, 1960; Cook and Polach, 1973; Greensmith and Tucker, 1975; Clarke *et al.*, 1979; Chappell *et al.*, 1982; Short *et al.*, 1989). Despite this, it still has not been possible to summarise the information concerning the episodic development of many chenier plains, and the production of even a small number of suitable models has therefore been prevented (Augustinus, 1989).

Given the probability that Australian chenier plains (all of which described so far have been Holocene in origin) have recorded information on palaeoenvironments, and the fact that more chenier plains have been described in Australia than any other country, the lack of consensus mentioned earlier amongst Australian researchers is a significant issue. More certainty and understanding with regard to the development of Australian chenier plains would potentially inform a number of issues in Australian coastal geomorphology. These include the effect of Holocene climate, storm frequency and sea level changes on chenier plain development, and especially their role in nearshore morphodynamics and the depositional balance. Holocene storm frequency is of particular importance in Princess Charlotte Bay as the previously studied transects (i.e. Grindrod, 1983; Chappell and Grindrod, 1984) were assumed by some to yield a record of Holocene storminess (e.g. Chappell *et al.*, 1983; Nott and Hayne, 2001). Examining a different section of the bay where the chenier ridge characteristics are different therefore provides an opportunity to examine this assumption from a different perspective. The debate amongst Australian coastal researchers with regard to smooth or oscillating late Holocene sea levels (Hopley, 1978; Lewis *et al.*, 2008) could also potentially be informed by advances in knowledge regarding the development of chenier plains. Since chenier ridges are built at the shoreline, it is likely that sea level changes (and associated changes in shoreline position) would have played a role or been recorded in some way in plains.

Aside from Holocene processes, the size of the Princess Charlotte Bay chenier plain suggests that Pleistocene events, and particularly interglacial highstands of sea level, may also be recorded there. The Pleistocene record of sea level changes is highly punctuated in Australia, particularly along the northern and northeastern coastlines (Murray-Wallace and

Belperio, 1991; Murray-Wallace, 2002). The discovery of more records in that area could potentially add significantly to our understanding of glacial-cycle sea level changes. This is also beneficial to worldwide knowledge for two reasons. Firstly, more investigations on Pleistocene processes have been carried out in the Northern Hemisphere than in the south, so more southern records are desirable. Secondly, since large ice sheets did not form on the Australian continent during the Pleistocene, the sea level record is dominated by the glacio-eustatic signal (Murray-Wallace, 2002), possibly modulated by the effects of hydro-isostasy (Lambeck and Nakada, 1990; Haworth *et al.*, 2002). This is again a different situation to the northern hemisphere, where large scale glaciation during ice ages has meant sea level changes recorded there are mostly dominated by the effects of glacio-isostatic rebound (Lambeck and Chappell, 2001).

In general it is hoped that the findings of this investigation may enable a better understanding of how chenier plains fit into the wider context of coastal geomorphology. Such knowledge may also enable the prediction of the morphological responses of muddy coastal environments to environmental changes. Given that we expect future changes in climate, sea level and storm activity as a result of global climate change, such knowledge is clearly important.

1.3 Research aim

The primary aim of this research was to investigate the Holocene development of the Princess Charlotte Bay chenier plain. The secondary aim was to consider the broader formation of several chenier plains in Australia and identify possible relationships.

The research was undertaken using new data from a chenier ridge sequence on the western side of the Princess Charlotte Bay plain, data obtained from reassessing results of previous Australian chenier plain studies, and considering these in the context of new palaeoenvironmental knowledge. A detailed explanation of the different aspects of this research is provided in Chapter 3.

1.4 Structure of this thesis

This thesis has been divided into seven chapters, of which this introduction is the first. **Chapter 2** contains a literature review that has been further divided into three sections. Part A is a review of theory relevant to chenier plain geomorphology in general, and a discussion and synthesis of this theory. Part B is a review of palaeoenvironmental knowledge that could be used to test different models of chenier plain development in Princess Charlotte Bay or explain development of the chenier plain through time. Part C considers some issues related to the use of radiocarbon methods in previous studies, and possible alternate methods that may be more suitable for dating chenier ridges. **Chapter 3** describes the research design of this investigation. Specifically this is an explanation of how the research aim was achieved. **Chapter 4** contains a description of the Princess

Charlotte Bay field site, and the methods and procedures followed in the fieldwork, data collection and data analysis phases of the research. **Chapter 5** is a summary of the data collected and results of analyses. It includes aerial photography of the field site and coastline, results of a basic vegetation survey, a topographic profile of the transect, a description of stratigraphic distinctions, and results obtained from dating the Holocene chenier ridges. **Chapter 6** is a discussion and synthesis of the data and evidence collected. This chapter addresses the aim of this investigation and the extent to which it was achieved, and a detailed consideration of the knowledge gathered by the research. **Chapter 7** is a summary of the main findings, conclusions that can be drawn and suggestions for how future research might proceed.

Chapter Two - Literature Review

Because of the scope of this literature review, it is divided into three parts. **Part A** contains a review of theory relevant to chenier plain geomorphology in general. Where possible, more attention has been given to Australian examples. Because much of the theory has developed in other parts of the world, however, this was not always possible. It includes topics such as the development of ideas and definitions, a history of approaches to chenier plain investigations, a comprehensive summary of models and theory regarding chenier ridge/plain development and shoreline processes, and a discussion/synthesis of this theory. **Part B** contains a review of palaeoenvironmental changes relevant to northeastern Queensland that could be used to test different models of chenier plain development in Princess Charlotte Bay, or explain development of the chenier plain through time. It includes consideration of both Pleistocene and Holocene events and change, although it has a Holocene focus. Finally, **Part C** contains a consideration of issues related to the use of radiocarbon methods employed in previous studies and considers alternate methods that may be suitable for dating chenier ridges.

PART A - CHENIER PLAIN GEOMORPHOLOGY

2.1 Introduction: Chenier ridges and chenier plains

2.1.1 Definitions of chenier plains and how they have developed

Chenier ridges were described first by Russell and Howe (1935) along the muddy Gulf of Mexico coastline in southwestern Louisiana. They noted that "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 vegetation cover in which large evergreen oaks form so striking a part that, quite deservedly, the ridges have been called *chenier ridges* by their Creole inhabitants (Russell and Howe, 1935, p449)". The name comes from the word *chène*, French for „oak". Several other studies on the Louisiana chenier plain were conducted in the 1950s (e.g. Howe *et al.*, 1935; Byrne *et al.*, 1959; Gould and McFarlan, 1959) contributing to its status as the world"s archetypal example of a chenier plain coast (Draut *et al.*, 2005). In these „early" years of chenier plain research, another plain along the Amazon-Orinoco coast in South America was also described in detail (e.g. Geyskes, 1952; Brouwer, 1953; Bleackley, 1956).

In what could be called the earliest review article on the topic, Price (1955, p75) introduced the term „chenier plain", comprising "Shallow based, perched, sandy ridges resting on clay along a marshy or swampy, seaward facing, tidal shore, with other…ridges stranded in the marsh behind, forming a belted marsh and ridge plain…". Although it was adapted and refined by later authors (e.g. Curray"s (1969) adaptation added information regarding general morphology, dimensions and composition, while Hoyt (1969) clearly distinguished between barriers and chenier ridges), it was to remain accepted for the next two decades as examples of chenier ridges were identified in more areas around the world and the body of knowledge within the field expanded. Todd (1968) established three conditions associated with the development of chenier ridges; being (1) stability, or recession, of sea level; (2) a variable supply of sediment from rivers that can be redistributed by shoreline processes; and (3) effective longshore currents. More recent studies have identified a further condition necessary: a gentle offshore slope (Zhao, 1989).

Although sufficient to *superficially* describe chenier plains, the early definitions did not imply any subsurface information and thus did not allow for clear distinctions between chenier ridges and other similar coastal features. This confusion was largely laid to rest by Otvos and Price (1979) when they redefined chenier ridges as shallow sand/shell beach ridges resting on silty or clayey deposits and separated from the sea by tidal mud flats. The assertion that chenier ridges are resting *on top of* finer deposits (further explained in Section 2.2) and the emphasis on progradation in Otvos and Price"s (1979) paper implies periods of mudflat progradation punctuated by coarse sediment deposition. This is now the fundamental criterion that distinguishes a chenier plain from other coastal features, and is shown conceptually in Figure 1.

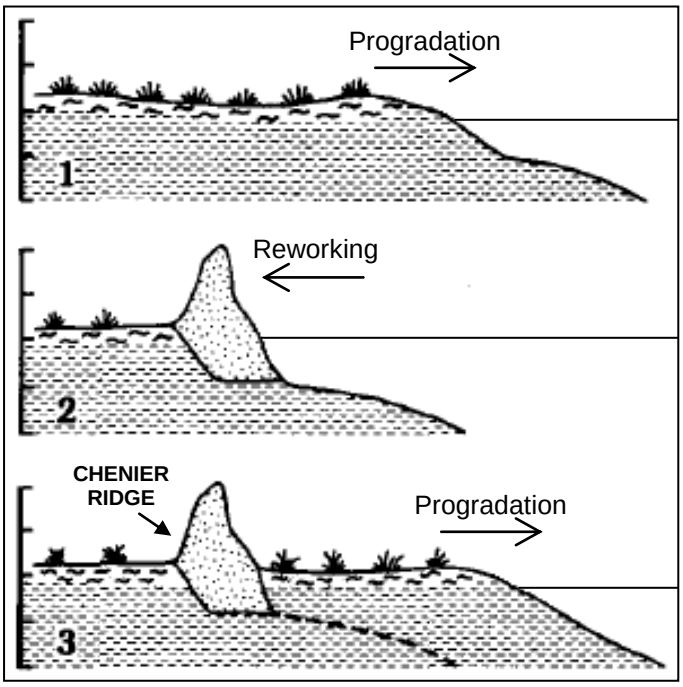

Figure 1. Chenier process model (modified from Hoyt, 1969). Scales have been left off deliberately to make this applicable to chenier ridges with different dimensions that form in different settings.

Otvos and Price (1979) further distinguished between chenier ridges that form on bight coasts and those that form on bay heads or sides. Bight coast chenier plains develop in low to moderate wave energy environments along open ocean shores that receive sufficient sediment from rivers. Dimensions of both individual chenier ridges and the wider coastal plain tend to be larger and sediment is derived from extra-local and regional sources. Bay head/side chenier plains develop in deeply indented bays that offer protection from high energy ocean waves and tend to have large tidal ranges. Individual chenier and plain dimensions tend to be smaller and sediments are locally derived (Otvos and Price, 1979; Augustinus, 1989).

In the most recent study of the Louisiana chenier plain, McBride *et al.* (2007) further refined the definition of Otvos and Price (1979) through the creation of a „geomorphic hierarchy" of features present in that setting. The intention was to enable clear distinctions to be made between „true" chenier ridges (i.e. transgressive ridges formed through winnowing and sorting; Section 2.5.2.1) and features resembling chenier ridges that may have been formed by different processes. The chenier plain is therefore defined as a first order feature (5000 km^2), composed of three second order features (30 to 300 km^2): the chenier complex, beach ridge complex, and spit complex. Individual ridges of each complex type can be further separated into third order features: chenier, beach ridge and spit (McBride *et al.*, 2007). Although extremely useful in this relatively large-scale setting, such a definition may have limited relevance in other, smaller chenier plains where there is less potential for significant variation in longshore processes or where the geomorphic history may have been simpler.

2.1.2 Chenier plains worldwide and Australian study sites

Chenier ridges and chenier plains are azonal features, forming in environments ranging from tropical to subarctic (Augustinus, 1989) and have been identified on every continent except Antarctica. Although geographic distribution and climate are important in determining precipitation and hence flow regimes of rivers, in the context of global chenier plain distribution these factors tend to have more of an influence on the nature of the *underlying* deposit (i.e. mangrove mud/salt marsh/bare mud) than on chenier ridge building *per se* (King, 1972). The worldwide distribution of known chenier plains is shown in Figure 2.

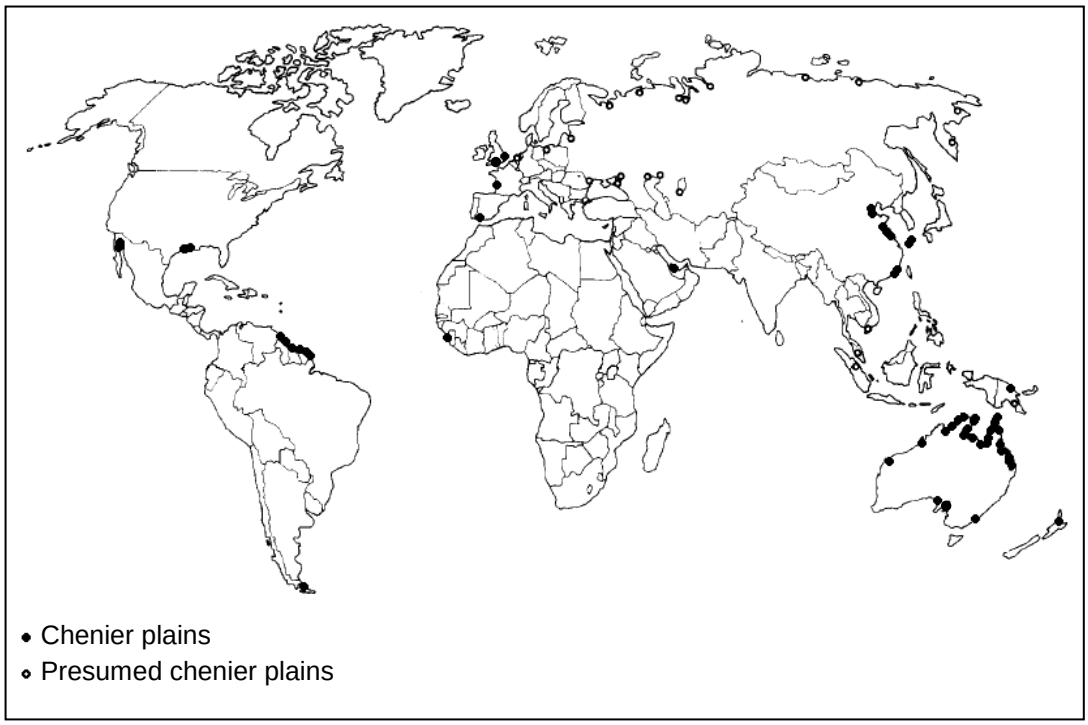

Figure 2. Worldwide locations of chenier plains (modified from Augustinus, 1989).

Since this review is intended to have an Australia focus, more detailed consideration of Australian plains is warranted. The Australian continent is home to more known chenier plains than any other on Earth (Figure 2). The locations where plains have been investigated are shown in Figure 3. It should be noted that although chenier plains are thought to be more numerous than the following list (e.g. Augustinus, 1989; Short, 1989), not all have been described or examined. In Western Australia, chenier ridges have been described in the Fitzroy Estuary, King Sound (Jennings and Coventry, 1973; Semeniuk, 1982) and in Roebuck Bay (O' Connor and Sullivan, 1994). In the Northern Territory, chenier studies have been conducted in The Victoria Delta, Joseph Bonaparte Gulf (Lees, 1985; 1992a); Shoal Bay (Hickey, 1981; Woodroffe and Grime, 1999); Bathurst Island, northwest of Darwin (Fensham, 1993); The Adelaide, Mary and South Alligator Rivers in van Diemen Gulf (Woodroffe *et al.*, 1985a; b; Woodroffe *et al.*, 1986; Woodroffe *et al.*,

1993; Mulrennan and Woodroffe, 1998); Point Stuart in the van Diemen Gulf (Clarke *et al.*, 1979; Lees, 1987) and the Lower Daly River emptying into Anson Bay (Chappell, 1993).

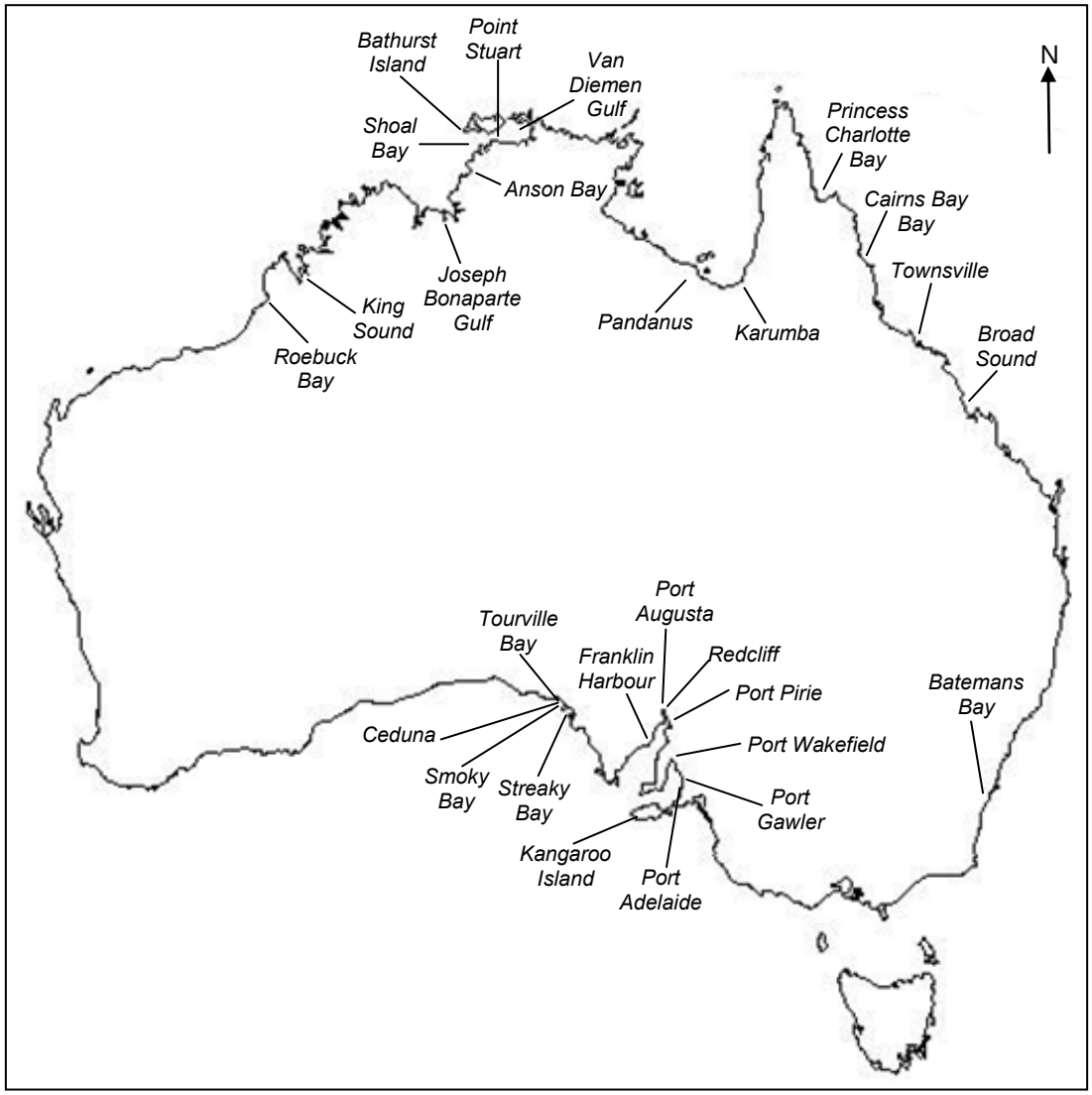

Figure 3. Locations of previous chenier plain investigations in Australia. Some areas contain more than one study site.

In Queensland, chenier studies have taken place near Pandanus and Karumba in the southern Gulf of Carpentaria (Rhodes, 1980; Rhodes *et al.*, 1980; Rhodes, 1982); at Princess Charlotte Bay on the east coast (Grindrod, 1983; Chappell and Grindrod, 1984); in Cairns Bay (Bird, 1969; 1970) and further south at Broad Sound (Cook and Polach, 1973;

Cook and Mayo, 1977). Chenier ridge occurrence has also been suggested along some parts of the Townsville coastal plain (Hopley and Murtha, 1975).

In New South Wales, a small chenier plain found in association with a more extensive beach ridge plain has been briefly described in the Cullendulla Creek embayment, Batemans Bay (Lewis, 1976; Thom *et al.*, 1981; Thom *et al.*, 1986). In South Australia, chenier ridges have been briefly investigated at Streaky, Smoky and Tourville Bays (Short, 1988), with the mouth of Davenport Creek in Tourville Bay receiving further attention in later studies (Belperio *et al.*, 1988), as well as an unconfirmed site at the mouth of the Cygnet River on Kangaroo Island (Short *et al.*, 1989). The occurrence of chenier ridges is suggested at a number of other sites in South Australia including Ceduna, Port Pirie, Port Augusta and Port Adelaide (Belperio *et al.*, 2002).

Although chenier ridges have formed in many different environments, the general requirements of low wave energy and high fluvial input would tend to give a tropical skew to their distribution in Australia, simply because more rainfall in tropical regions means there are more rivers. Tropical areas of even the most developed countries can be extremely difficult to access, and the site for this investigation is one such example. In less developed countries access to such areas may be near impossible. It is, therefore, almost certain that there are many other chenier sites around the world still awaiting discovery and investigation.

2.2 Morphological similarities between chenier ridges and other features

Otvos (2000) defined „beach ridges" as coastal ridges that became isolated from the daily impact of shoreline processes through shore progradation. He considered chenier ridges to be a special type of beach ridge. Because the processes involved in the development of chenier plains differs greatly from those involved in the development of either beach ridge or barrier plains, is it necessary to briefly consider morphological similarities that have in the past lead to confusion between these features.

Beach ridge plains are composed of sandy ridges only, without the intervening (or more importantly, *underlying*) muddy unit that would be present in a chenier plain. The material comprising beach ridges and their bases is usually very similar to that which makes up the adjacent beach (Taylor and Stone, 1996). Morphologically, barrier plains are very similar, in that the ridges generally rest on a base of marine sands that extend well below mean sea level (Hoyt, 1969). Unlike many beach ridge plains, barrier plains often also accumulate fine sediment in the intervening swales. This could give the impression of coarser ridges resting on top of finer sediment, when in actual fact the ridges and bases are composed entirely of sand and the accumulation of fines has occurred only *after* a more seaward ridge has formed. In such settings adequate subsurface investigations must be carried out to reveal the true nature of the ridges in question (Meldahl, 1995).

The confusion described above may have arisen in part, due to the fact that in many settings it is common for chenier ridges to occur in association with other coastal features. This is

particularly so for settings subject to complex morphodynamic forcing regimes, or where chenier ridges are one feature of several that have formed under the same general process trend (e.g. a regressive barrier system). One of the best examples of this is the Louisiana chenier plain, where a large geomorphic system subject to a complex tapestry of littoral drift, relative sea level changes and channel avulsions has resulted in a coastal plain containing chenier ridges, beach ridges, recurved spits, aeolian deposits, storm berms, natural levees and ancient oyster reefs (McBride *et al.*, 2007).

Even in settings where the morphodynamic forcing regimes are relatively simpler than the Louisiana example above, it is common for chenier plains to grade into beach ridge plains due to longshore variations in wave energy or sediment availability. This is the case in Princess Charlotte Bay, where a chenier ridges are found in the central portion of the plain whereas beach ridge plains exist at the western and eastern flanks (Frankel, 1974; Grindrod, 1983; Chappell and Grindrod, 1984). The ability to distinguish between chenier ridges and beach ridges (based on the underlying sediment as described above) is therefore very important.

2.3 Why study chenier plains?

Chenier plains provide a sensitive record of changes to the depositional mode of shorelines. These changes can be driven by a number of environmental processes, for example, changes in sediment supply, river discharge, sea level and storminess, which are described in detail in Section 2.5.1. Furthermore, because their composition makes them amenable to

dating techniques (Clarke *et al.*, 1979; Augustinus, 1989), it becomes possible to reconstruct timings of changes in shoreline processes, which in turn provides information about palaeoenvironmental changes. Examples of how this has happened in the past are given below.

The entire coast of the Gulf of Mexico is undergoing submergence at such a rate that only two rivers (The Mississippi and The Appalachicola) have a sediment supply sufficient to build deltas (Russell and Howe, 1935). This means that a fine balance exists between mudflat progradation in areas receiving fluvial sediment, and subsidence, which causes chenier ridge building in areas where deltas are not actively being built. Accordingly, changes to sediment supply in one region of the plain can tip the balance one way or the other. In this setting, the chenier plains provide an opportunity to study the history of several of the dominant geologic processes affecting the Gulf Coast through the Quaternary (Russell and Howe, 1935; McBride *et al.*, 2007), such as major avulsions of the Mississippi River that have occurred through the Holocene (Section 2.5.1.1).

Chenier ridges may also be particularly sensitive indicators of climate change (Lees and Clements, 1987) and changes between relatively more arid and pluvial conditions have been suggested as responsible from driving chenier plain development in many Australian (e.g. Rhodes, 1980; Rhodes *et al.*, 1980; Rhodes, 1982; Lees, 1992a; b) and overseas (e.g. Anthony, 1989; 2006) studies (Section 2.5.1.3). Some studies have utilised the facies change between sands and muds at chenier ridge bases as indicators of past sea level (e.g. Rhodes *et al.*, 1980; Chappell *et al.*, 1982). Chenier ridges have also been used to estimate the timing of the cessation of the most recent post-glacial marine transgression and

stabilisation of Holocene sea level (e.g. Cook and Polach, 1973; Clarke *et al.*, 1979; Chappell *et al.*, 1982; Short *et al.*, 1989).

Aside from indicating *major* (i.e. glacial cycle) eustatic sea level changes, chenier ridges have also been used for evidence of smaller-scale eustatic fluctuations and/or oscillations (i.e. minor changes that may have occurred since post-glacial sea levels stabilised around 6000 yr BP). For example, Schofield (1960) interpreted different heights of features within the Miranda chenier plain, Firth of Thames, New Zealand, as the result of sea level oscillations that had occurred over the last 4000 years. A similar interpretation was presented for the Essex Chenier Plain in England by Greensmith and Tucker (1975).

Clearly, chenier plain geomorphology has the potential to yield much knowledge about palaeoenvironmental changes along coastlines where they are found. This is particularly so for Australia, where more chenier plains have been described than in any other country. Only with more data and a better understanding of the processes operating in different settings, can this potential to yield knowledge be fulfilled.

2.4 Approaches to the study of chenier plains

There have been four approaches used in studying chenier plains. These are (1) physiographic description; (2) regional stratigraphy; (3) stratigraphic distinctions and (4) coastal process studies and morphodynamics (Chappell and Thom, 1986; Penland and Suter, 1989). In short, this progression represents the development of the field of chenier

plain geomorphology over time, where the older studies tend to be more „descriptive" than recent ones (Augustinus, 1989). The approaches are by no means mutually exclusive, and most studies now use a combination of the following approaches at different stages of investigation.

2.4.1 Physiographic description

The first chenier studies tended to be based on physiographic description and/or classification (e.g. Howe *et al.*, 1935; Russell and Howe, 1935). Authors of modern studies generally utilise physiographic descriptions to introduce and describe their study sites (e.g. Woodroffe *et al.*, 1983; Isla *et al.*, 1991; Lee *et al.*, 1994) or as a preliminary stage in the absence of data (e.g. Anthony, 1989). The descriptions may also be used when detailed stratigraphic data are not required, such as in a description of geomorphological changes following large storms (e.g. Hopley, 1974; Woodroffe and Grime, 1999) or as evidence for different origins of sediment deposits (e.g. Bryant *et al.*, 1992a; Nott, 2004; 2006).

2.4.2 Regional stratigraphy

As more knowledge was gathered regarding the physiography and classification of chenier plains, attention was turned to investigations of regional stratigraphy. This approach has generally been concerned with identification of the properties, composition and faunal

characteristics of sediment facies (e.g. subtidal, intertidal, mangrove muds, and chenier ridges) and is typically a preliminary part of any modern study.

Examples from Louisiana include the detailed facies investigation of Byrne *et al*. (1959) or the geological history compiled by Gould and McFarlan (1959). In Essex a regional stratigraphic approach formed an essential part of an investigation into specific aspects of different sediment facies (Greensmith and Tucker, 1969). This approach also formed a major part of field work for the initial investigation into chenier ridges in northern Sierra-Leone, western Africa (Anthony, 1989).

2.4.3 Stratigraphic distinctions

As chenier plain research progressed it became focused on the stratigraphic distinction of individual ridges. Regional work revealed the internal structure of chenier ridges, and showed that not all have similar vertical sequences (e.g. Penland and Suter, 1989). This enables deeper questions to be asked, for example not just regarding the processes that form chenier plains as a whole, but how those processes vary through time to result in construction of chenier ridges with different internal structure.

A classic paper on the origins of the Louisiana chenier plain used this approach and distinguished between beach ridges and chenier ridges based on internal stratigraphy (Hoyt, 1969). In the Point Stuart chenier plain (Northern Territory) stratigraphic distinctions were made based on carbonate and shell content, grain size and physical dimensions with the aim

of investigating sea level changes (Clarke *et al.*, 1979) and later, coastal processes and climate (Lees, 1987). A modern study focussing on stratigraphic distinctions was conducted in Essex, England and used ground penetrating radar to determine the internal structure of ridges and confirm them as chenier ridges (Neal *et al.*, 2003).

2.4.4 Coastal process studies and morphodynamics

The most modern approach to the investigation of chenier plains is the coastal process or morphodynamics approach (Augustinus, 1989; Penland and Suter, 1989). The morphodynamic approach is the most comprehensive as it considers all aspects (e.g. physical, hydrological, sedimentological and climatic) of a given setting as part of an inter-connected coastal system. In this way it is possible to relate process, time and morphological response to these factors (Chappell and Thom, 1986). This also enables prediction of future changes, a subject that has become a very important aspect of contemporary coastal studies. The other approaches discussed are effectively consolidated in the morphodynamic approach because information on chenier plain physiography and stratigraphy is necessary to consider the effects of coastal processes or morphodynamic responses. Very few modern coastal studies are conducted without some sort of morphodynamic consideration.

In a coastal morphodynamic approach, the central concept is that of four main boundary conditions (i.e. external factors that influence the coastal environment, being sea level, storm activity, sediment inputs and climate). Normal variation about the average value of

any boundary condition may catalyse changes of coastal „state" (Chappell and Thom, 1986). Identification of the „average values" therefore precedes any discussion of how they relate to each other in a given setting. This step can sometimes prove more controversial than the application of the data itself. An example is the establishment of a precise sea-level curve for the late Holocene (e.g. Chappell, 1983; Baker and Haworth, 2000a), which takes into account several interacting factors (e.g. Pirazzoli, 2005). Although not all studies incorporating morphodynamic consideration have the scope to reach the mathematical stage, the generation of equations to model coastal processes is probably the most extreme example, or endpoint, of the morphodynamic approach.

In southeast Asia and northern Australia this approach was used to regionally summarise late Quaternary dynamics of estuarine and deltaic environments (Woodroffe, 2000). Elsewhere in the world the morphodynamic approach has been used to link chenier ridge building and mudflat progradation to changes in fluvial output (e.g. Liu and Walker, 1989; Ying and Xiankun, 1989; Saito *et al.*, 2000), longshore sediment transport variations (e.g. Augustinus, 1980; Augustinus *et al.*, 1989; Augustinus, 2004), climate (e.g. Rhodes, 1980; Rhodes, 1982; Lees and Clements, 1987; Lees, 1992b; Tao *et al.*, 2006), shellfish mortality (Greensmith and Tucker, 1969; Chappell and Grindrod, 1984), storm activity (e.g. Yan *et al.*, 1989; Nott, 2006), and sea level changes (e.g. Schofield, 1960; Greensmith and Tucker, 1975; Clarke *et al.*, 1979; Chen, 1996).

2.5 Theories regarding the development of chenier plains

In any environmental process study it is important to distinguish between the forces that drive the process, and the mechanisms by which the process itself is carried out. Examples of this occur in any field. With regards to the effect of population changes within a species, the driver may be changes in resource availability, but the specific mechanism may be intra-species competition. In fluvial geomorphology, the effect of lateral channel migration may occur in response to a changed flow regime. Here the *driver* of channel migration may be cyclic climate fluctuations causing wet and dry phases, but the actual *process* is carried out by scour of the outer banks and deposition on the inner banks. In chenier plain geomorphology, there are a number of models that reflect different forces driving chenier ridge building, each of which is considered in the following sections. Knowledge regarding the actual physical process by which chenier ridges accrete has developed separately from these models, and is considered separately.

2.5.1 Models of chenier plain development

> "Progradation of chenier plains is not simply a process of smooth accretion because the ridges represent events of coarse sediment deposition separated by periods of mudflat progradation. To explain the alternation of two modes of deposition is the central problem of chenier plain geomorphology" (Chappell and Grindrod, 1984, p199).

Throughout the years there have been nine models proposed to explain the development of Holocene chenier plains (i.e. nine environmental forces that could drive the alternation Chappell and Grindrod (1984) referred to in the above quote). These are river channel switching, mud shoal migration, climate fluctuations, storm activity, tsunami, human impacts, sea level oscillations, variations in shellfish populations, or an internal dynamic. They are considered in the following sections.

2.5.1.1 River channel switching

River channel switching is the oldest model to explain chenier plain development and is well established in the Louisiana literature (Schou, 1967). It was introduced by Russell and Howe (1935) and largely confirmed by the work of Frazier (1967). At a given location, normal mudflat progradation occurs when sediment supply from the Mississippi River Delta is high enough to overcome the normal shoreline distributive agencies of wave action and longshore currents (Hoyt, 1969), and unconsolidated sandy mud is deposited.

Every 1500–2000 years a new major depocentre is initiated by a channel avulsion of the main distributary channel (Frazier, 1967; McBride *et al.*, 2007). When this occurs, sediment supply to the previously prograding area of coastline decreases (Figure 4), and is no longer sufficient to „overcome the normal shoreline distributive processes". Wave action then causes coarse sediment deposition and chenier ridge building (Russell and Howe, 1935; Coleman and Gagliano, 1964; Coleman, 1966; Frazier, 1967; Todd, 1968). McBride *et al.* (2007) recently published an updated model for the development of the Louisiana chenier plain. Although more complex and more fully describing the suite of processes

occurring there, for the sake of describing river channel switching as an individual model, the description thus far is sufficient. McBride *et al.* (2007) is considered later.

More recently the Louisiana model has been shown to be applicable in other settings. The discharges of the Yangtze (Changjiang) and Huanghue (Yellow) Rivers are among the biggest in China, and the world, respectively (Ying and Xiankun, 1989). Chenier ridges have formed in the deltas of both these rivers in response to local sediment deficits, and several authors have pointed to the shifting mouths of the rivers as the cause (e.g. Liu and Walker, 1989; Ying and Xiankun, 1989; Saito *et al.*, 2000).

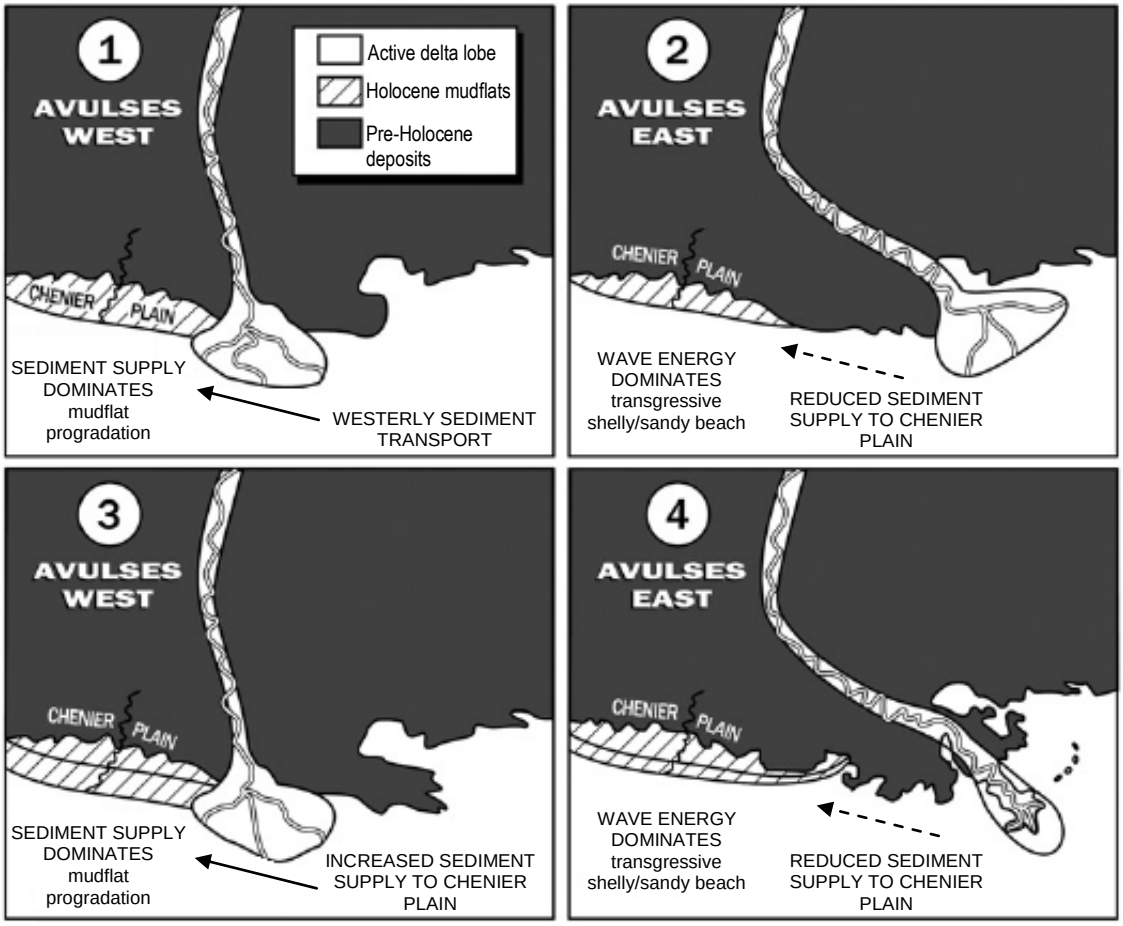

Figure 4. Depositional model explaining chenier plain development through mudflat progradation, wave reworking and ridge development, followed by mudflat progradation, thus creating a chenier ridge (modified from McBride *et al.*, 2007).

2.5.1.2 Mud shoal migration

Mud shoal migration is the process driving chenier plain development along the Amazon-Orinoco coastline in South America. Unlike in Louisiana, the influence of „local" rivers on sedimentation is considered minimal because of the massive sediment load supplied by the Amazon River (Prost, 1989) which amounts to over 1000 million tonnes annually (Eisma *et al.*, 1991). Such a high sediment load results in the formation of shoreface connected mudbanks, 25–30 km in length (Eisma *et al.*, 1991) that regularly migrate westward away from the mouth of the Amazon (Augustinus, 1980; Wells and Coleman, 1981; Froidefond *et al.*, 1988), with a periodicity so regular they have been considered giant mud waves with average wavelengths around 45 km (Daniel, 1988). In this way the fluvial influence of the Amazon River is extended for hundreds of kilometres from its mouth. Figure 5 shows a simplified example of the spacing of the mudbanks along the coastline. The smallest coastal unit in this setting is therefore a mudbank and its eastward adjacent interbank area (Augustinus, 2004).

The movement of these mud waves is driven by the prevailing northeasterly trade winds (Eisma *et al.*, 1991; Allison *et al.*, 2000; Augustinus, 2004) and occurs at around 1.5 km/yr, giving an approximate 30 year duration for the migration of one mudbank unit past a given point (Wells and Coleman, 1981). The movement occurs because mudbanks are more exposed and susceptible to erosion on their eastern sides, and it is thought that sediment from the eastern side of one bank is transported to the western side of the adjacent bank (Augustinus, 1980). In this way an individual bank will tend to be more consolidated and

dense on the eastern side and more fluid on the western side (Wells and Coleman, 1981). When the sediment concentrations on the western side of mudbanks reaches approximately 10 000 mg/L (Wells and Coleman, 1981) a dense suspension called slingmud (also fluid mud or liquid mud) forms. Slingmud is very important in this exposed, open-ocean setting because it tends to change waves from sinusoidal to solitary (Wells and Coleman, 1981), heavily damping their energy in the process.

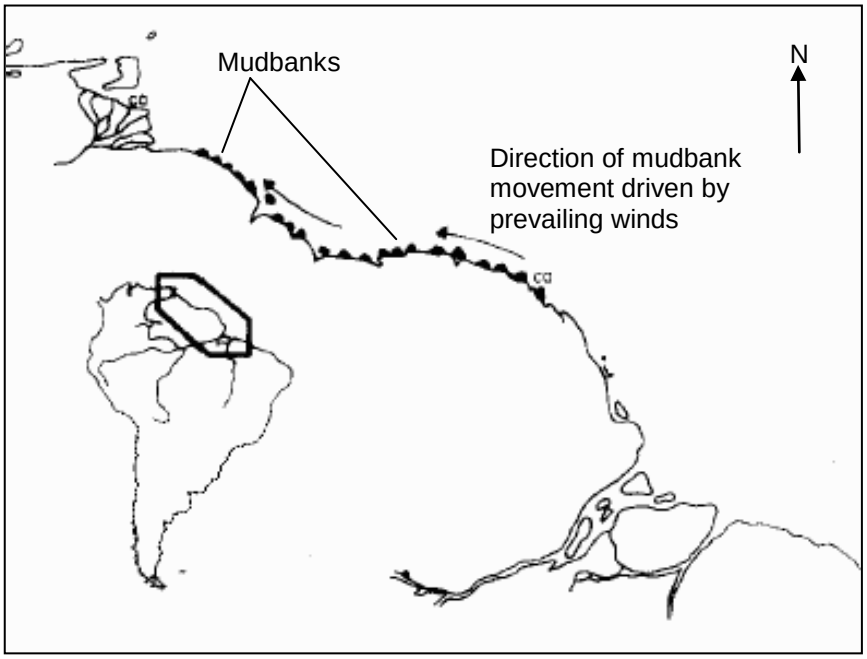

Figure 5. Spacing and movement of mudbanks along the Amazon-Orinoco coastline (modified from Froidefrond *et al.*, 1988).

Cycles of chenier ridge building in this setting are controlled largely by the distribution of the bank and interbank zones: mudflat progradation occurs along coasts fronted by a mudbank and slingmud, while ridge building occurs in the interbank zones exposed to ocean swell (Augustinus, 1980; Froidefond *et al.*, 1988; Augustinus *et al.*, 1989). Chenier ridge building in one section of coastline thus has the potential to repeat itself on a 30 year cycle. A similar process contributes to chenier development in western Africa (Anthony, 1989; 2006).

2.5.1.3 Climate fluctuations

Climate fluctuations have been suggested as a factor driving chenier plain development in a number of studies (e.g. Rhodes, 1980; Rhodes *et al.*, 1980; Rhodes, 1982; Lees and Clements, 1987; Anthony, 1989; Lees, 1992a; Lees, 1992b; Anthony, 2006). Rhodes (1980, 1982) studied a chenier plain in the Gulf of Carpentaria, Australia and concluded that alternating phases of mudflat progradation and coarse sediment deposition were driven by a fluctuating supply of muddy sediment to the coast. When mud supply is high (during more pluvial conditions), coastal progradation and mudflat development occurs. When the supply of mud falls to a critical level (during more arid conditions), wave action causes chenier ridge building (Rhodes, 1982).

Lees and Clements (1987) also emphasized the role of climate fluctuations in the development of Australian chenier plains. They used statistical techniques to compare radiocarbon ages from six chenier plains and were able to identify a significant regional period of chenier construction between 2800 and 1600 yr BP, concluding that this period of chenier construction was related to a climatically driven reduction in fluvial input to the coast.

A problem with the previous application of this model in Australia is that although it is plausible, there has been a lack of palaeoenvironmental records available for comparison. This is not the case now, and rigorously testing the climate model with regards to Australian chenier plains is an avenue of investigation in this thesis (Chapter 3).

2.5.1.4 Storm activity

The role of storms in chenier plain development has been debated since the early Louisiana studies (e.g. Russell and Howe, 1935; Price, 1955; Schou, 1967). The storm interpretation generally hinges on the height of chenier crests above mean sea level, in light of the (relatively small) „normal" wave heights generally found along chenier plain coastlines (e.g. Yan *et al.*, 1989; Nott and Hayne, 2001; Nott, 2004; 2006). Many authors have suggested that chenier ridges are „storm ridges", constructed during either single cyclones or hurricane events, or during phases of increased storminess (i.e. by several closely-spaced storm events). Chenier ridges in San Sebastian Bay, Argentina, for example, are thought to be direct artefacts of storms in the bay (Isla *et al.*, 1991). Yan *et al.* (1989), Lees (1987) and Quaresma *et al.* (2007) similarly attributed chenier ridge building to storm activity. Nott (2004) considered the role of periods of increased storminess, concluding that phases of elevated cyclogenesis through the Holocene were likely to be responsible for chenier ridge construction. Furthermore, he believed it was likely that all chenier ridges have been deposited by storm waves (Nott, 2004).

It is also possible that storms play a more indirect role in chenier plain development. Woodroffe and Grime (1999) suggested that storm events may control the distribution of mangrove populations by reducing the extent of particular stands. On muddy coastlines, the removal of mangrove stands would increase the erosion potential of a particular section of coastline, which may then increase the potential for reworking and chenier ridge building.

Despite the attention given to storms, many authors have explicitly discounted them as constructive agents by highlighting their destructive nature. In an early Louisiana paper, Schou (1967) observed that newly built chenier ridges are normally flattened during hurricanes, with sand and shells being transported by the turbulent water mass. Lees (1992a) similarly commented that it would be difficult to imagine a storm event forming a discrete bank of undamaged shell at high tide level. Chappell and Grindrod (1984) believed the storm explanation to be too simplistic as there exist far fewer chenier ridges in Princess Charlotte Bay, Australia, than there have been storms through the Holocene. Many chenier ridges are also composed entirely of sand and gravel (e.g. Russell and Howe, 1935; Schou, 1967; Bird, 1969; 1970; Cook and Polach, 1973; Jennings and Coventry, 1973; Prost, 1989), and therefore do not conform to Nott"s (2004; 2006) assertion that the presence of abundant shell beds in cheniers is a reason to accept the storm interpretation. Finally, many chenier ridges clearly show a gradual accretion over hundreds of years (e.g. Chappell and Grindrod, 1984; Yan *et al.*, 1989; Ying and Xiankun, 1989; Zhao, 1989). This suggests that although storms may exert an influence in some settings, they are probably not a *primary* driving force in the development of many chenier plains.

With regards to Australian studies at least, Lees and Clements (1987) suggested that it is probably better to consider storms as a „random factor" in chenier plain development. Such a consideration is consistent with considering landward migration (which is driven by storm waves) as ancillary to chenier ridge building (Section 2.5.2.4). Alternatively, it is possible that storms, with their associated surges and temporary elevation in sea level, act as an accretionary force that can be considered part of the „normal" suite of coastal processes a chenier ridge is exposed to. Although these positions are probably better to take

than „chenier ridges are storm ridges", the possibility that they are cannot be fully discounted based on current evidence.

A final point that must be made in relation to considering storms as a „model" is a reassertion of the importance of distinguishing between the forces that drive the process, and the mechanisms by which the process itself is carried out. The storm model involves the consideration of storms as the former (i.e. the factor which drives the alternation of depositional regimes). As has been alluded to above, dismissing this model would not, however, preclude a role for storms as the latter (i.e. as a possible mechanism by which chenier ridges are constructed).

2.5.1.5 Tsunami

Bryant *et al*. (1992a) proposed that ridges at the mouth of Cullendulla Creek in New South Wales were mistakenly interpreted as chenier ridges, and were actually an artefact of Holocene tsunami. This was based on four pieces of evidence; being (1) the position of chenier ridges distal from the shoreline; (2) the chaotic mixing of shell species from several different marine environments (from exposed rock platforms to the continental shelf); (3) the mixing of sand and shell; and (4) the irregular arrangement of radiocarbon and thermoluminescence ages, which do not suggest that the ridges get younger in a seaward direction.

Many problems exist with this model. The reasoning that the ridges are „distal from the shoreline" completely ignores the fact that periods of chenier ridge building alternate with

periods of mudflat progradation. It is likely that when the chenier ridges were formed that they were in fact at the shoreline, and since that time mudflat progradation has occurred. The chaotic mixing of shell species from several different marine environments, from exposed rock platforms to the continental shelf, could be due to reworking and transport of buried shell by storm or other waves after the death of the organism over long periods. Similarly, the „mixing of sand and shell" is likely to simply represent the coarse fraction of unconsolidated mudflat sediment that has been eroded and deposited as a chenier. The degree of shell breakage could be due to wave action and has been noted in chenier ridges in other settings (e.g. Cook and Polach, 1973; Rhodes *et al.*, 1980; Lees, 1987; Woodroffe *et al.*, 1993). Finally, the irregular arrangement of (uncalibrated) radiocarbon and luminescence ages could be explained by the material used for dating and different taphonomic histories, as Bryant *et al*. (1992a) point out. Although the radiocarbon ages do not indicate a seaward decrease in age (as is to be expected from chenier plains), the luminescence ages clearly do. Tsunami experts have also cast doubt on Bryant"s inferred processes (e.g. Felton and Crook, 2003). The occurrence and associated geomorphic impacts of Holocene tsunami notwithstanding, the above considerations suggest the tsunami model for chenier plains is not supported by field evidence.

2.5.1.6 Human impacts

Anthropogenic landscape modification can have an impact on sediment supply to the shoreline, and there are two effects which are relevant to chenier plain development. Large scale land use changes can lead to catchment destabilisation and widespread erosion, leading to increased sedimentation and shoreline progradation. Conversely, the construction

of dams or other fluvial diversion mechanisms can lead to reductions in fluvial input and sedimentation.

Increased sedimentation rates over the last 2000 years have been attributed to Human activity in the Changjiang (Yangtze) Delta, China. It is thought that increasing exploitation of the catchment (i.e. deforestation and farming on the loess plateau to support a rapidly expanding population) led to widespread erosion, which increased the sediment load of the river (Liu *et al.*, 1987; Saito *et al.*, 2001). More recent human impacts in the same catchment have reversed this trend, as sand mining in the middle catchment has increased its capacity as a sediment sink (Chen *et al.*, 2005), reducing coastal sedimentation. It was predicted that the construction of the Three-Gorges dam would have similar consequences (Chen, 1998).

Reductions in sediment output to the coast as a result of catchment modification have been suggested as responsible for the recent development of a third generation of chenier ridges in the northwest Gulf of California. The formation of the two oldest generations has been attributed to diversions of the Colorado River into the Salton Trough over the last 1500 years (i.e. river channel switching). The youngest, having formed in the last 70 years, is attributed directly to the reduction in river flow by the construction of dams and increased irrigation, and associated sediment starvation experienced at the coastline as a result (Kowalewski *et al.*, 1994; Meldahl, 1995). Flessa *et al.* (2004) provided corroborating evidence for these claims, suggesting that the extensive canal system for farms and cities diverts almost all of the Colorado River‟s 16 billion m^3 of flow before it reaches the delta.

2.5.1.7 Sea level oscillations

Minor changes in sea level can affect chenier plain development by inhibiting or encouraging erosion (King, 1972). While there are many studies that have attempted to use aspects of chenier plains to gain information regarding „major" sea level changes, very few have considered the role that more minor (eustatic) sea level changes, fluctuations, or oscillations, may have played in chenier plain development. The main reason for this has been the lack of consensus regarding Holocene sea level curves in many areas, or the assertion that Holocene sea level change has been smooth and linear. This is the case in Australia. Most chenier plain studies conducted in Australia have used Chappell"s (1983) sea level curve, which indicates a smoothly falling sea level from approximately +1–2 m above current sea level over the past 6000 years (Section 2.10.2.3). The effect of eustatic changes in sea level was consequently discounted, or not considered at all, as a driving force behind chenier development in many previous studies (e.g. Chappell and Grindrod, 1984; Lees, 1987; Chappell, 1993; Woodroffe *et al.*, 1993; Mulrennan and Woodroffe, 1998). More recent Australian work (e.g. Baker and Haworth, 2000a) has cast doubt on the linear model and there is a growing body of evidence to support the occurrence of eustatic sea level oscillations during the late Holocene (considered further in Section 2.10.2).

The sea level oscillations model has been applied outside Australia. For example, Greensmith and Tucker (1975) found that chenier ridges on the Essex chenier plain in England were formed during sea level transgressions, while mudflat progradation occurred during regressions. They concluded that the sedimentation patterns in different zones

appeared to reflect minor oscillations superimposed on the major post-glacial marine transgression (Greensmith and Tucker, 1975). Anthony (1989) suggested that sea level oscillations may have affected chenier plain development on the west African coast by modulating the sediment transport capacity of tidal currents. Late Holocene sea level oscillations have also been suggested responsible for chenier formation in some parts of China (Zhao, 1989), and may have played a role in the development of chenier plains in Louisiana (McBride *et al.*, 2007), Donana marshland in Spain (Rodríguez-Ramírez and Yáñez-Camacho, 2008) and in Sierra Leone in west Africa (Anthony, 1989).

2.5.1.8 Fluctuations in shellfish populations or an internal dynamic

As the rivers draining to the Essex chenier plain in England carry very little sediment, Greensmith and Tucker (1969) emphasized the role that offshore shell deposits played by acting as „feeders" for both incipient and established chenier ridges. They noted that chenier growth was most pronounced in the periods immediately following shellfish mass mortality events, which provided shell material to these offshore beds, and then to chenier ridges in turn. Greensmith and Tucker (1969) subsequently replaced this interpretation with the sea level oscillations model (Section 2.5.1.7). Fluctuations in shellfish populations were also invoked by Chappell and Grindrod (1984) to explain chenier formation in Princess Charlotte Bay when they introduced the concept of an „internal dynamic". From this point onwards, only the „internal dynamic" is considered.

The concept of an internal dynamic is centred on a complex interaction between shell availability, muddy sediment input, and width of the mangrove fringe. Particular

combinations of these variables can cause a chenier plain to alternate between a „rapid prograding" mode, and a „cut and recover" mode (Chappell and Grindrod, 1984). The „rapid prograding" mode (low frequency of chenier ridge building) occurs when muddy sediment input is sufficient to suppress shellfish populations and therefore shell (i.e. the primary material comprising chenier ridges in this setting) production. The rapid prograding mode is also associated with a wider mangrove fringe, which focuses sedimentation and acts as a positive feedback mechanism to increase progradation. The „cut and recover" mode (much higher frequency of chenier ridge building) occurs when there has been less muddy sediment input, shellfish populations are relatively higher, and wave action can cause winnowing and sorting. This situation is aided by smaller mangrove populations, which gives less focus to sedimentation and therefore less chance for deposition and progradation. A chenier plain will thus switch between these two modes according to the particular combinations of shell production, mud input and width of the mangrove fringe. The primary factor, however, is shell availability (Chappell and Grindrod, 1984). A version of this model was also used by Woodroffe and Grime (1999) to explain the development of a chenier plain in Shoal Bay, Northern Territory, Australia. The internal dynamic model has been criticised on the grounds that the reliance on „level of muddy sediment input" makes it closely aligned to the climate model (e.g. Lees, 1992a). This is considered further in Chapter 3.

2.5.1.9 Summary and combinations of models

Given that the tsunami model for chenier plain development is not supported by field evidence, and that fluctuations in shellfish populations is closely linked to the internal dynamic, seven plausible models remain. Holocene chenier plain development can be driven by river channel switching, mud shoal migration, climate fluctuations, storm activity, human impacts, sea level oscillations, or an internal dynamic. Although these models have been proposed individually in many settings, combinations are also possible.

For example, on the west African coast, Anthony (2006) proposed that the models of mud shoal migration, climate fluctuations and human impacts interacted to form a complex tapestry of progradation and chenier ridge building. The role of sea level oscillations was also considered in an earlier paper (Anthony, 1989).

Mud shoals along the west African coastline are morphologically similar, but smaller than those described for the Amazon-Orinoco plain (Section 2.5.1.2). They are formed because the numerous small rivers discharging to this coastline all have suspended sediment concentrations sufficient to form shoals of slingmud, which is then transported by the effective longshore currents (Anthony, 1989; 2006). Climate plays a role because the sediment loads of the rivers are dependent on rainfall. Anthony (2006) stated that chenier ridge building is more common during arid periods, during which sediment discharge is decreased and less slingmud can form. Even during these periods, both chenier ridge building and mudflat progradation can occur simultaneously in different locations owing to

differences in local bathymetry and wave penetration. Human impacts in the form of deforestation have differentially modified catchments along this coast. It is also possible that Holocene sea level oscillations played a role by modulating the sediment transport capacity of tidal currents (Anthony, 1989). More recently, McBride *et al.* (2007) proposed that although river channel switching was still the primary force driving development of the Louisiana chenier plain, Holocene sea level oscillations had also played a role in forming the observed palaeoshoreline trends.

2.5.2 Shoreline processes

While early studies in Louisiana were able to identify the cause of alternate phases of mudflat progradation and coarse sediment deposition (i.e. river channel switching), the actual shoreline mechanisms responsible for chenier ridge building were to remain unexamined for some time. Todd (1968, p737) recognised this, stating that "although the geological and oceanographic conditions necessary have been described, [and] individual plains have been described, the actual hydraulic mechanisms involved in the deposition of chenier ridges still await adequate explanation".

The work of Todd (1968) and Augustinus (1980) (among others) showed that there are three main shoreline processes and mechanisms involved in the phases of coarse sediment deposition and chenier ridge building. These are winnowing and sorting, wave deposition and *in situ* accretion, and landward migration. The nature of these processes in different settings can often be revealed by the internal structure of chenier ridges, or the type of

sediment underlying them, and this is also discussed. Despite the array of models described in Section 2.5.1, the theory regarding these specific shoreline processes and mechanisms was accepted and has remained relatively unchanged since it was first introduced by Todd (1968).

2.5.2.1 Winnowing and sorting

Todd (1968) introduced the concepts of winnowing and sorting in his model for the shoreline processes involved in chenier ridge building. In the Louisiana chenier plain, the shifting river mouth leads to changes in sediment supply at locations on the shoreline. At locations which change from being close to the locus of sediment delivery to being remote from it, reduction in the suspended load carried by longshore currents allows for an increase in wave energy. This occurs because waves were previously „damped" to a certain extent by the presence of more suspended sediment. In this way, reduced sediment load leads to an increase in the erosive capacity of marine processes. If this capacity increases beyond a certain threshold, winnowing is induced. This is the removal of the fine (silt and mud) sediment fraction from the unsorted mudflat sediment deposited during a previous episode of progradation. These fines are transported seaward and deposited offshore. Winnowing is followed by sorting, where the coarse (gravel, sand and shell) fraction is concentrated and builds into a chenier ridge (Todd, 1968). These processes have also been referred to as „segregation" (Swan, 1982) and „sediment fractionation", the latter used in relation to the concentration of coarse gravels from a finer sand matrix (Shulmeister and Kirk, 1997). It is important to note that no matter which of the models described in Section 2.5.1 is applied in a particular setting (except the tsunami model), winnowing and sorting is

the shoreline process by which chenier ridges actually accrete. These terms have been applied in a progradational context to the concentration of the coarse sediment fraction from material supplied by longshore currents and beachdrift (e.g. Augustinus, 1980), as well as in reworking/erosional contexts (i.e. Todd, 1968).

2.5.2.2 Wave deposition, in situ accretion and underlying deposits

Winnowing and sorting yields a supply of coarse sediment. This is then incorporated into an actual chenier ridge by accretion processes at the shoreline. While it is generally accepted that wave action is involved in some way, the role of storms, and particularly periods of temporarily elevated sea level brought about during storm surges or spring tides, has been debated. Russel and Howe (1935) originally asserted that „higher than normal" wave energy was responsible for casting coarse gravel and sand grains onshore. Price (1955) instead suggested that chenier ridges are formed during periods where sea level is temporarily raised, as in storms or spring tides, when the zone of wave action was elevated. This results in sand being reworked into underwater bars, which are the chenier ridges once sea level recedes. Todd"s (1968) theory incorporated both of these possibilities. He suggested that coarse sediment bodies (i.e. incipient chenier ridges) could accumulate under normal conditions in shallow foreshore water (i.e. in the sub- and intertidal zones), and that this might be supplemented by continued accretion during „constructive" events such as storm or spring tides.

Studies on the Amazon-Orinoco chenier plain in South America have shown that the zone of accretion may depend on the relationship between grain size of the coarse sediment and

shoreline conditions (i.e. average wave heights and strength of currents). In eastern Surinam the sandy fraction of the unconsolidated mudflat sediment is coarser than in the west. This is because sands in the east are primarily supplied by local rivers, while those in the west also contain a fraction of very fine Amazon-borne pelite (Augustinus, 1980). The coarser chenier ridge sands in the east tend to be deposited by wave swash action as strandlines, with chenier ridges forming around high tide level. In the west, however, longshore currents are responsible for distributing the finer sands. Chenier ridges there tend to form emergent longshore bars in shallow nearshore water (Augustinus, 1980; 1989; Augustinus *et al.*, 1989). These differences might be explained by the tendency of finer sand to stay in suspension longer, and also result in differences in their internal structure (Section 2.5.2.3). If it remains so through any swash-backwash cycle it will only be deposited in a zone not subject to wave action. This will be at the low water mark (i.e. in the shallow subtidal, or nearshore zone) because the zone between the low and high water marks is subject to wave action through a tidal cycle. Conversely, the coarse sands tend to form chenier ridges as strandlines at the high water level because they will be more likely to fall out of suspension at the upper limit of swash, and will therefore be gradually moved to the upper limit of wave action through a tidal cycle. In this way, the wave action may impose a sorting effect that leads to different average grain sizes in chenier ridges built at different ends of the zone affected by swash processes (i.e. shallow subtidal and supratidal).

The actual zone of accretion, and hence the possible processes that may be involved in ridge building, differs greatly between individual chenier plains. This is shown by considering the variety of facies that chenier ridges have been shown to overlie. These include subtidal muds (Byrne *et al.*, 1959; Gould and McFarlan, 1959); low tide muds

(Rhodes, 1980; Rhodes *et al.*, 1980; Rhodes, 1982); a range of intertidal facies (Bird, 1969; 1970; Anthony, 1989; Lee *et al.*, 1994; Vilas *et al.*, 1999) including mangrove muds (Cook and Polach, 1973; Chappell and Grindrod, 1984; Short, 1988); and high tide and supratidal muds (Greensmith and Tucker, 1969; Liu *et al.*, 1987; Neal *et al.*, 2002). Chenier ridge building can also be supplemented by aeolian activity in the same way that isolated beach ridges can continue to grow into sand dunes (King, 1972; Taylor and Stone, 1996). While little aeolian activity has been documented in the Amazon-Orinoco examples, some Louisiana chenier ridges are topped by clay deposits more than a metre in depth (Price, 1955). Aeolian deposits are also associated with some chenier ridges in Princess Charlotte Bay, Australia (Chappell and Grindrod, 1984).

2.5.2.3 Internal structure

The internal structure of established chenier ridges can be used to evaluate the processes responsible for their construction and the nature of their longer-term development (Neal *et al.*, 2002). They can also be used to help distinguish chenier ridges from other types of deposits, such as tsunami deposits (Bryant *et al.*, 1992a) or middens (Sullivan and O' Connor, 1993; Stone, 1995). Internal structure often consists of a number of bedded strata. The thickness/depth, direction, and angle of dip of these strata reflect different processes operating while the ridges were being built. For example, seaward dipping structures (beachface deposits) form by swash/backwash of waves too small to overtop the chenier crest and generally indicate periods of beach face sedimentation or progradation (Augustinus, 1980; Neal *et al.*, 2002). Swash/backwash of normal wave action is mainly responsible for developing distinct subparallel to low-angle, seaward dipping, planar

laminations (Reinson, 1984). Horizontally bedded structures are far less common but have been described as representing deposition of suspended sediment under extreme tides that completely inundate the chenier ridge (Yan *et al.*, 1989). Landward dipping structures are formed by washover of waves large enough to overtop the chenier crest and are common in chenier ridges that have undergone landward migration (Augustinus, 1980; Neal *et al.*, 2002). The angle of such structures can also yield information about the deposition conditions. For example, Shinn (1973) described chenier ridges that display gently angled landward dipping beds (<5°) that suddenly become far steeper (~30°) a few centimeters from the chenier ridge base. Similar structures have also been described by Augustinus (1980) who stated that their formation is dependent on the level of water behind the ridge in question. When water levels are low the washover structures fan out evenly and form parallel bedding because there is no abrupt change to flow velocity. If still water is present, however, the sudden slowing down of running water moving into stagnant water results in sediment settling at angles such that maximum internal friction is achieved (Augustinus, 1980).

2.5.2.4 Landward migration

In some locations it is common for chenier ridges to be driven landward by washover processes once accretion of the ridge has occurred (Augustinus, 1980; 1989; Augustinus *et al.*, 1989). Although this occurs in many chenier plains (e.g. Schou, 1967; Greensmith and Tucker, 1969; Augustinus, 1989; Lee *et al.*, 1994; Park *et al.*, 1996; Woodroffe and Grime, 1999; Neal *et al.*, 2002; Rodríguez-Ramírez and Yáñez-Camacho, 2008) because it is not observed in *all* settings it is only ancillary to chenier plain development (Augustinus,

1989). The nature of the underlying deposits can reveal whether landward migration is a significant process in a particular setting. For example, chenier ridges overlying marsh deposits (Schou, 1967; Greensmith and Tucker, 1969; Augustinus, 1989; Neal *et al.*, 2002) could not have formed *in situ* (since the marsh zone would not be subject to shoreline processes) and must have been driven landwards on to the marsh only after it had formed behind the prograding shoreline. In some cases chenier ridges have been shown to overlie several different facies, indicating differential movement along the chenier length (e.g. Coleman, 1966; Shinn, 1973).

Other evidence for landward migration can be found in the morphology and internal structure of chenier ridges. Landward migration causes a change in the morphology of chenier ridges (Augustinus, 1989) because of the dissipation of wave energy from the seaward side to the landward side. When waves are sufficiently large to overtop a chenier ridge, sediment is scoured by wave run-up on the seaward side, transported over the crest to the landward side and deposited. This process leads to a steeper profile on the seaward side and a gentle landward dipping backshore (e.g. Augustinus, 1989; Lee *et al.*, 1994). It also tends to cause a degree of sorting such that the chenier sediment is finer on the landward side than to seaward (e.g. Byrne *et al.*, 1959).

2.6 Discussion: A fundamental condition for chenier plain development

In considering the forces which drive chenier plain development (as opposed to the actual shoreline processes; Section 2.5), many authors have emphasised the need for a fluctuating sediment supply (e.g. Curray, 1969; Hoyt, 1969; Cook and Polach, 1973; Rhodes, 1982; Augustinus, 1989; Short, 1989). While this is indeed a contributing factor in many settings, overemphasising it may hide more fundamental relationships that are conceptually useful in examining chenier plains. Instead of considering any specific conditions necessary for chenier plain development (e.g. Todd (1968); Section 2.1.1) let us simply focus on the assertion that chenier ridges represent a change between periods of mudflat progradation and coarse sediment deposition. The fact that chenier plains generally contain more than one ridge suggests that the factors driving development of a particular chenier plain are capable of intermittently causing switching between the dominance of either force under certain conditions. If they are capable of intermittently switching, then the two states must exist in somewhat of an (albeit constantly tipping) balance. In this context, the models described in Section 2.5.1 simply represent different forcing factors which act to change this balance and, therefore, the depositional regime. Mudflat progradation will be encouraged by environmental changes that encourage the overpowering of marine distributary processes, such as pluvial climate phases which increase in the fluvial discharge and sediment load, or forces which act to decrease wave energy (let us call such environmental changes „progradational forces"). Conversely, coarse sediment deposition will be encouraged by environmental changes that encourage winnowing and sorting by marine distributary processes, such as such as storms, sea level transgressions or other forces that increase the magnitude, or relative importance of, wave energy, or arid climate

phases that reduce precipitation, fluvial discharge, and the availability of fine sediment for muddy progradation at the shoreline (let us call these environmental changes „accretion forces"). That switches occur between the dominance of either force is the most important criterion for allowing the development of chenier ridges, and hence a chenier plain.

Lees and Clements (1987) called these forces „fluvial" and „marine" forces respectively. The names have been changed here (to „progradation" and „accretion" forces) to more aptly reflect the intent of the concept, as the use of the words „fluvial" and „marine" could be misleading (i.e. it is not necessarily fluvial forces which act to encourage muddy progradation, nor is it strictly marine forces that act to encourage coarse sediment deposition). It is important to note that while this model is conceptually useful for highlighting a broad process relationship that chenier plains have in common, and possibly explaining the existence of chenier plains at certain locations, it cannot on its own account for the complex processes involved in the development of every chenier plain. This is because while the relationship may hold, the actual processes driving the balance will be different, and interact in different ways, in different settings. The thresholds for such a change will therefore vary between different chenier plains depending on the morphodynamic context. This concept can explain the general distribution of chenier plains around the world and previous chenier plain classifications, provide a conceptual development model that applies to all chenier ridges, as well as explain some changes observed in the progradational nature of coastal plains more generally.

On a world scale, for example, the sediment load (which could be considered one way of quantifying „progradation force") of the Normanby River draining into Princess Charlotte

Bay is relatively small (Chapter 4). A large chenier plain has developed there (Grindrod, 1983; Chappell and Grindrod, 1984), however, because normal wave energy (i.e. the „accretion force") of this north-facing bay in the lee of the Great Barrier Reef is low enough to have allowed progradation to occur. This low energy setting can be compared to Amazon-Orinoco Chenier Plain on the eastern coast of South America (e.g. Augustinus *et al.*, 1989). The existence of a muddy progradational plain on what could be considered an unlimited-fetch coastline, and subject to higher energy ocean waves than most chenier settings, may seem strange were it not for the massive sediment load of the Amazon River, which is over 1000 Mt/yr^{-1} (Czaya, 1983). The load is so high that it suppresses wave energy through the formation of slingmud (Section 2.5.1.2). Furthermore, the combination of this extreme sediment load and strong longshore currents means the fluvial influence of the river, and hence the nearshore zone in which this progradation force is sufficient to balance the marine force (thereby allowing switches between them on a smaller scale) is extended hundreds of kilometres from its mouth. On a smaller scale at any one location along the shoreline of the chenier plain, the dominance of either the progradation or accretion force, and hence the depositional mode, is modulated by the passage of mud banks (Section 2.5.1.2). At any one location, the passage of the mud banks acts to cause relative changes in wave energy, and therefore switches in the relative dominance of either force.

The above considerations regarding wave energy versus sediment load can also be applied to the classifications of Otvos and Price (1979) of bay head and bight coast chenier plains (Section 2.1.1). In this new context, these are probably not separate „classes" of chenier plains so much as plains that simply experience different degrees of fluvial and marine

forces. For example, the bight coast chenier plains on the Amazon-Orinoco and Louisiana coastlines have more exposure and therefore higher marine energies, and also happen to occur on coastlines that receive some of the largest fluvial discharges in the world from The Amazon and Mississippi Rivers, respectively.

As well as simply defining the conditions necessary for chenier plains development, this approach can also provide simple explanations for some changes observed in the progradational nature of coastal plains more generally. For example, the plain at the mouth of Cullendulla Creek in Batemans Bay, Australia shown in Figure 6 contains two chenier ridges at the most landward locations then becomes a beach ridge plain as one moves towards the sea (Lewis, 1976; Thom *et al.*, 1981). Figure 6 shows how these two chenier ridges are oriented approximately 45° differently from the beach ridges such that they point in a more easterly direction than the southeast facing beach ridges.

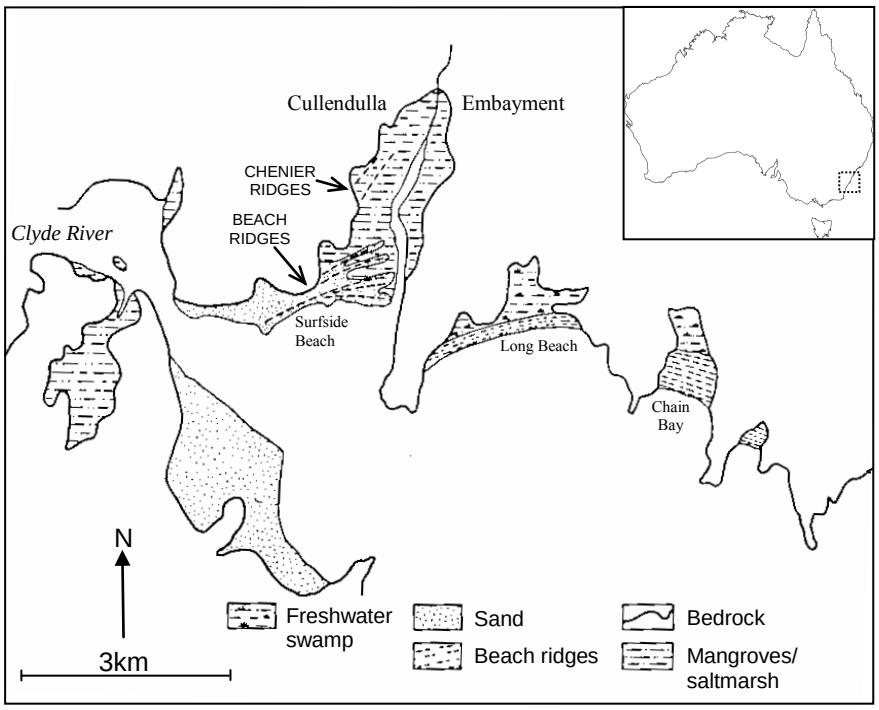

Figure 6. Configuration of chenier and beach ridges in Batemans Bay, New South Wales (modified from Thom *et al.*, 1986).

This orientation difference would have likely resulted from the progressive infilling of the narrow northerly section of the embayment. As the shoreline moved seaward due to progradation it became exposed to dominant southeast swells. This would have resulted in a gradual increase in marine energy at the shoreline (i.e. an increase in the accretion force relative to the progradation force). Once the plain prograded to a certain point, the accretion force became dominant and chenier ridges could no longer form because mud would have constantly been exported offshore as opposed to being deposited in a mudflat. At this point, the chenier plain progradation ceased and beach ridges began to form. A similar explanation can be invoked to explain the lack of net progradation in the Mary River Plain in Northern Australia over the last 2000–3000 years (Mulrennan and Woodroffe, 1998). In this setting, the chenier plain would represent the fine (albeit periodically *tipping*) balance that existed between progradation and accretion forces for 3000–4000 years after the post-glacial marine transgression. Once the plain prograded to a certain point the progradation force was insufficient to continue progradation, but effectively balanced the accretion force, and the plain has existed since. The present static nature of this plain indicates that the forces are in equilibrium.

In summary, the most important condition for the development of chenier plains is a fine (albeit intermittently switching) balance between progradation and accretion forces. Dominance of the accretion forces leads to coarse sediment deposition (forming chenier ridges), while dominance of the progradation forces causes muddy progradation. The specific morphodynamic processes influencing this balance will be different in different settings.

2.7 Summary

Part A of this literature review has summarised current knowledge regarding chenier plain geomorphology (with an Australian focus where possible). Chenier ridges are defined as shore parallel, sandy or shelly shallow based ridges, resting on silty or clayey deposits and separated from the sea by tidal mud flats (Otvos and Price, 1979). They are azonal and can occur in association with other coastal features, in which case it may be necessary to define a geomorphic hierarchy (e.g. McBride *et al.*, 2007). Although chenier plains have been described on every continent but Antarctica, far more have been studied in Australia than in anywhere else. Approaches to the study of chenier plains have varied from simple physiographic description to descriptions of regional stratigraphy and stratigraphic distinctions. More recent studies use a morphodynamic approach that has allowed links between coastal processes and their geomorphic impacts to be better understood. Seven plausible models exist to explain development of Holocene chenier plains: river channel switching, mud shoal migration, climate fluctuations, storm activity, human impacts, sea level oscillations, or an internal dynamic. Combinations of these are also possible, as Anthony (1989; 2006) and McBride *et al.* (2007) showed. There is no evidence that chenier ridges are emplaced by tsunami. The models for chenier plain development can be examined independent of the shoreline processes involved in chenier ridge building. The main processes (introduced by Todd, 1968) are winnowing (removal of fine material) and sorting (concentration of coarse material to form a chenier ridge). The level of chenier ridge accretion depends on the relative nature of the coarse sediment and shoreline energy, and can occur anywhere from the shallow subtidal (nearshore) zone to the high-tidal and

supratidal zones. The internal structure of chenier ridges can offer clues into the processes involved in their accretion. Landward migration is an ancillary feature of chenier ridges and is not observed in all settings. The nature of the underlying deposits can often indicate if it has occurred. It was concluded that the most important condition for the development of chenier plains is a fine (albeit intermittently switching) balance between progradation and accretion forces. Dominance of the accretion forces leads to coarse sediment deposition (forming chenier ridges), while dominance of the progradation forces causes muddy progradation. The specific morphodynamic processes influencing this balance will be different in different settings, but are represented by the seven plausible chenier plain models. The concept of a balance between progradation and accretion forces can replace some previous definitions and is a far simpler way of defining chenier ridges in the context of other coastal geomorphological features.

PART B - PALAEOENVIRONMENTAL CHANGES IN NORTHEASTERN QUEENSLAND AND PRINCESS CHARLOTTE BAY

2.8 Introduction

The existence of an internal dynamic was the original model proposed to explain the development of the Princess Charlotte Bay chenier plain (Section 2.5.1.8). Since this was put forward, many palaeoenvironmental studies have been undertaken which now enable other chenier plain models to be more critically assessed in Princess Charlotte Bay. Part B of this review contains a review of palaeoenvironmental knowledge that could be used to test the applicability of the models described in Part A (Section 2.5.1) in Princess Charlotte Bay. This includes reconstructions of palaeoclimates (including sea surface temperature and storm activity) and sea level change through the Pleistocene and Holocene. Although Holocene chenier plain development is the primary focus of this investigation, the size of the greater Princess Charlotte Bay plain makes it likely that it has formed over a considerably longer time than this, and consideration of some Pleistocene events is therefore warranted.

2.9 Palaeoclimates and change

Climate reconstructions have been published based on data from many locations close enough to be relevant to a study of Princess Charlotte Bay palaeoenvironments (Figure 7). Late Pleistocene climates (since around 40 000 yr BP) are summarised first, with a particular focus on aeolian episodes and climate reversals during the most recent deglaciation period. A more detailed consideration of Holocene climate reconstructions follows, including consideration of sea surface temperature changes, the El Nino Southern Oscillation, the Medieval Warm Period and the Little Ice Age.

It must be noted that it is not the intention of this chapter to specifically identify the „true" pattern of climate change for the Princess Charlotte Bay region. What is intended is to summarise all knowledge that could be used in the interpretation of ages obtained from a chenier plain. Because of this, any study that has discussed past climates will be considered, regardless of whether the findings presented conflict with others. This should provide the best grounding for testing models that could apply to Princess Charlotte Bay. The terrestrial sites in Queensland referred to in both this and the previous sections are shown in Figure 7.

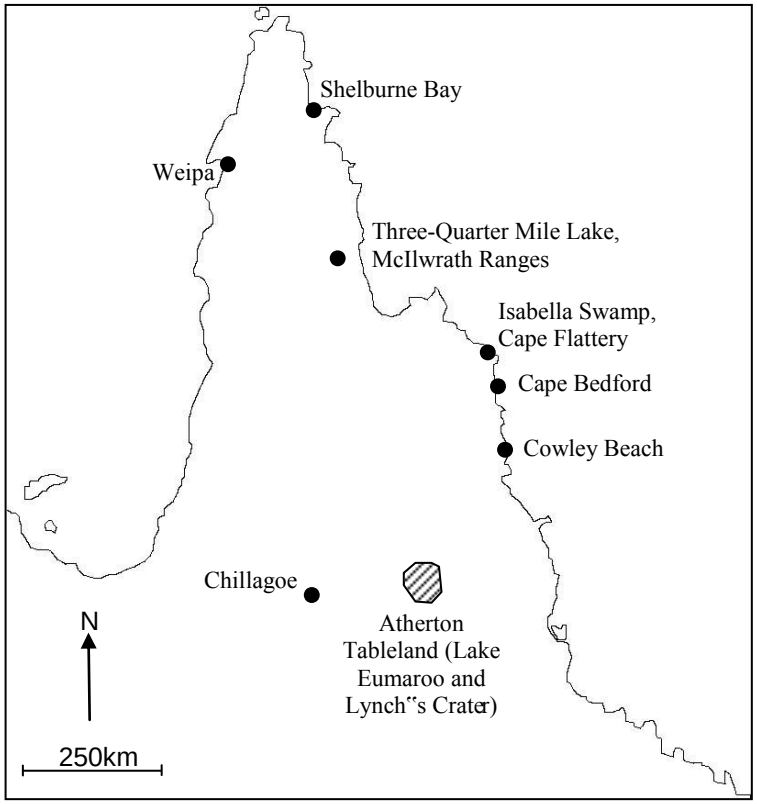

Figure 7. Location of Queensland sites referred to in this and previous sections.

2.9.1 Late Pleistocene climates in northeastern Queensland

2.9.1.1 The Last Glacial Maximum and aeolian activity

Based on pollen evidence, conditions in Australia grew progressively cooler and drier from around 75 000 to between 30 000 and 20 000 yr BP due to a combination of glacial-cycle climatic factors, and increased distance of sites to the coastline during lower sea levels (Chappell, 1987; Williams *et al.*, 2006). Vegetation changes suggest conditions attributable to the Last Glacial Maximum occurred between 23 000 and 18 000 yr BP (Turney *et al.*, 2006a; b; c) although geomorphological evidence suggests windier and drier conditions

could have persisted until as late as 14 000 yr BP (Kershaw and Nanson, 1993; Bryant *et al.*, 1994; Thom *et al.*, 1994). Evidence from a variety of sources suggests conditions on the Atherton Tableland were much drier (up to 50% less precipitation) and cooler (2–5°C) than present around the time of the Last Glacial Maximum (Haberle, 2005).

Of particular interest in the context of a coastal geomorphological study of the Queensland coastline, is that although the dominant wind direction during the last glacial period was still from the southeast (Figure 8), it was stronger as a result of intense anticyclonic conditions over much of the continent and enhanced subtropical westerly winds (Shulmeister *et al.*, 2004; Turney *et al.*, 2006a). This resulted in the activation and emplacement of a number of aeolian dune units along the northeastern Queensland coast. For example, a dune building event occurring from 24 000 to 17 000 yr BP has been identified along this stretch of coastline based on examination of dunes at Cape Bedford, Cape Flattery and Shelburne Bay (Lees *et al.*, 1990). A unit dating to around 30 000 yr BP has also been identified at Shelburne Bay.

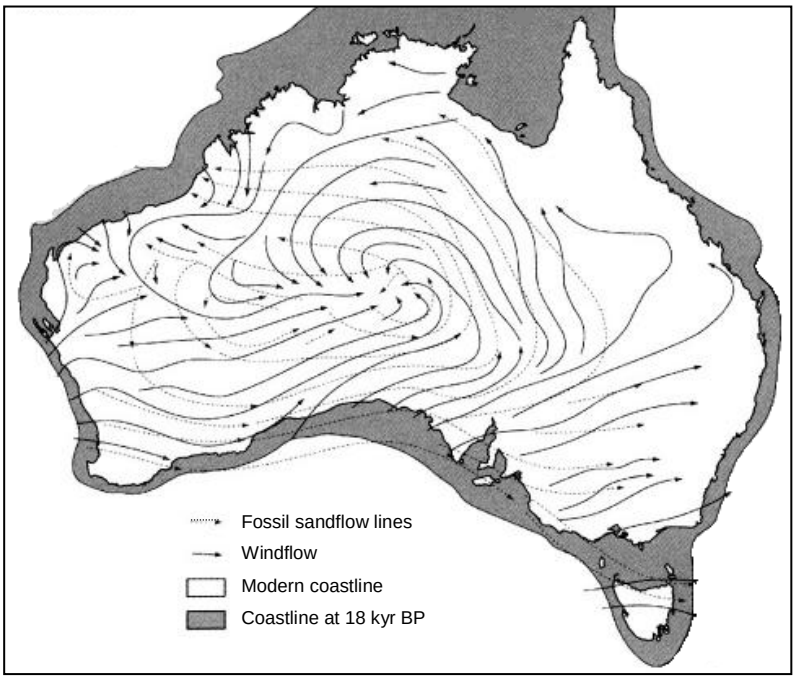

Figure 8. Palaeowind directions after Thiede (1979) and Wasson (1987).

Bryant *et al.* (1994) combined ages of dune units from 90 sites in eastern Australia to determine the most frequent age for aeolian activity. The results are shown in Figure 9, and suggest that the most frequent age for aeolian activity in the Princess Charlotte Bay region is around 30 000 yr BP.

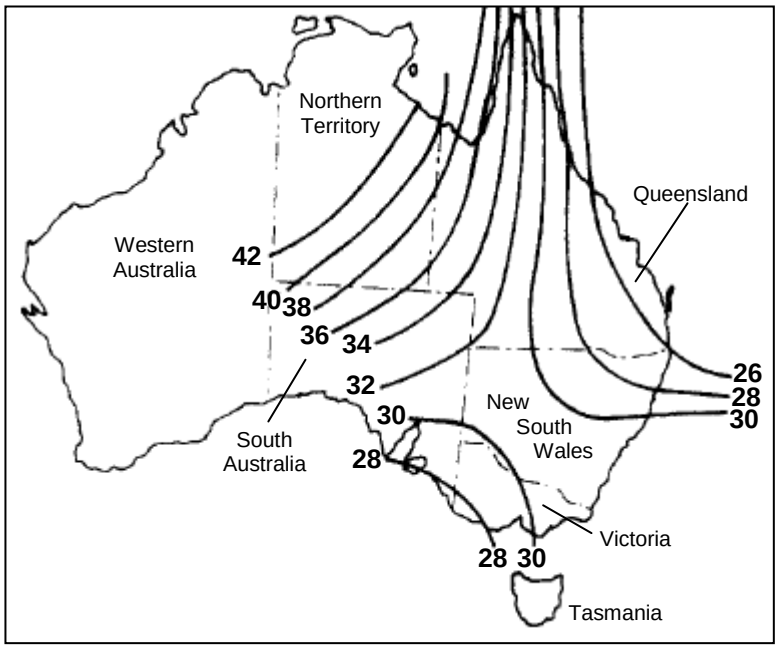

Figure 9. Most frequent ages for aeolian activity during the last glacial period across eastern Australia (modified from Bryant *et al.*, 1994). Ages shown are kyr BP.

2.9.1.2 Post-glacial warming and climate change to the early Holocene

Australian records suggest that the end of the onset of post-glacial warming occurred between 17 000 and 14 000 yr BP (Haberle, 2005; Turney *et al.*, 2006a; b). Based primarily on Northern Hemisphere records, the post-glacial warming trend is generally thought to have been episodic and punctuated by at least one cooling trend. The Younger Dryas cold reversal is an example. This climatic fluctuation has received much attention and represents an abrupt return to near-glacial conditions between 12 900 and 11 500 yr BP (Peteet, 1995; Andres *et al.*, 2003). Controversy exists, however, as to the climatic expression of the Younger Dryas in the Southern Hemisphere (Peteet, 1995), and furthermore whether or not it was a global event (e.g. Lambeck and Chappell, 2001; Andres *et al.*, 2003; Hormes *et al.*, 2003).

Records compiled from the Southern Hemisphere do suggest that cooling events have punctuated post-glacial warming. Deglacial records compiled from east Antarctic ice cores have recorded two significant cooling events punctuating the general warming trend, the most well marked of which is the Antarctic Cold Reversal, lasting from around 14 000 to 12 500 yr BP (Jouzel *et al.*, 2001). Using a marine sedimentary sequence obtained from the Great Australian Bight, Andres *et al.* (2003) suggested that two cold reversals recorded between 13 100 and 11 100 yr BP were likely to be associated with drops in sea surface temperatures, large scale shifts in the strength and position of the westerly winds, and could be associated with the Younger Dryas. Records from northeastern Australia and Queensland also show evidence of climatic fluctuations punctuating post-glacial warming. For example, a dry reversal is recorded between 14 000 and 10 000 yr BP at Chillagoe (Figure 7) (Turney *et al.*, 2006a). Pollen and diatom records from Lake Eumaroo and Lynch‟s Crater, both on the Atherton Tableland, also indicate a phase of lower effective precipitation between 13 800 and 11 500 yr BP (Haberle, 2005; Turney *et al.*, 2006c; Tibby and Haberle, 2007). The occurrence of climatic fluctuations around these times is supported by the activation of aeolian units at a number of Australian sites. For example, Lees *et al.* (1993) attributed the stabilisation of dunes near Weipa (Figure 7) at 11 200 yr BP to a minor sea level regression around the time of the Younger Dryas, while Woodroffe *et al.* (1992) found reworking of sand sheets had occurred at 10 700 yr BP on the Cobourg Peninsula.

Generally, most studies have found that warming and increased precipitation occurred beyond these events into the early Holocene (e.g. Shulmeister, 1999; Prebble *et al.*, 2005). For example, diatom records reveal high lake levels between 9500 and 4500 yr BP in Lake

Eumaroo on the Atherton Tableland (Tibby and Haberle, 2007) while a gradual increase in rainforest pollen at the same site until between 7300 and 6300 yr BP suggests warming (Haberle, 2005). Sites very close to Princess Charlotte Bay also record more reliable rainfall into the early Holocene (e.g. Stephens and Head, 1995; Luly *et al.*, 2006).

2.9.2 Holocene climates and change in northeastern Queensland

2.9.2.1 Late Holocene climates

The late Holocene is generally considered a period of increased climatic variability, and a number of climate reconstructions for this period have been developed for the Queensland region based on analysis of pollen and diatom records (e.g. Overpeck, 1996; Shulmeister, 1999; Haberle *et al.*, 2001; Prebble *et al.*, 2005). The locations of the sites referred to are shown in Figure 7.

Shulmeister (1999) summarised the findings of several climatic investigations from across Australasia and presented a conceptual model of Holocene palaeoclimates for the Pacific Basin based on effective precipitation. This model can be summarised by: (1) a gradual increase in effective precipitation (presumably in association with rising post-glacial sea levels and increasing coastal influence) up until around 5000 yr BP; (2) a period of maximum effective precipitation from around 5000 to 3700 yr BP; (3) an apparent increase in climatic variability since 3700 yr BP; and (4) a possible increase in effective precipitation within the last 3000 to 2000 years (Prebble *et al.*, 2005). The findings of other studies are in general agreement with this. For example, pollen and diatom records from

Lake Eumaroo on the Atherton Tableland indicate a period of climatic stability and higher rainfall through the early to mid Holocene, with increased variability around 3900 yr BP and a possible increase in rainfall in the last 1000 years (Haberle, 2005; Tibby and Haberle, 2007).

Stephens and Head (1995) analysed pollen assemblages in a core taken from Isabella Swamp, near Cape Flattery and also found evidence to support an increase in effective precipitation since 2700 yr BP. Pollen and diatom records from Three-Quarter Mile Lake on the flanks of the McIlwrath Range (the closest site to Princess Charlotte Bay from which records have been obtained), indicate a pattern of Holocene lake and vegetation change that is consistent with those described above, although the writers concede that the location of the study site may make it less sensitive to minor changes in rainfall (Luly *et al.*, 2006). Taking into account the different sensitivity of many sites spread across eastern Australia, in terms of precipitation, the following summary of Holocene change emerges: from the early to mid Holocene rainfall was gradually rising. A mid Holocene climate optimum (maximum relative effective precipitation) was reached from as early as 8000 yr BP and lasted until around 5000 yr BP, when conditions began to become more variable. A further increase in variability occurred after 3000 yr BP (Donders *et al.*, 2007). Although this is generally recorded as lower levels of precipitation, some sites on the northeastern Queensland coastline appear to have recorded higher levels of rainfall since then (e.g. Stephens and Head, 1995).

„Increased climatic variability" is a problematic statement, however, as the actual climatic manifestation of increased variability can differ greatly between sites. Generally speaking,

early Holocene climate records are overwhelmed by the effects of post glacial sea level rise (Shulmeister, 1999; Andres *et al.*, 2003; Prebble *et al.*, 2005), adding to the agreement in these records from sites across a large region. It could therefore be expected that regional and local forcing mechanisms would have become more important in the late Holocene, adding variability into phases determined for broad regions such as those described by Shulmeister (1999). This variability would be further enhanced by the possible „noise" effects introduced into records from increased Aboriginal occupation and land use in the late Holocene (Kershaw, 1983; Haberle and David, 2004; Tibby and Haberle, 2007). Despite these issues, the general result of increased variability appears to have been a net reduction in rainfall.

Using a global approach, Mayewski *et al.* (2004) examined around 50 globally distributed records arguing that they revealed evidence of as many as six periods of significant climate change through the mid to late Holocene. These are generally characterised by polar cooling, tropical aridity, and major atmospheric circulation changes. Although no Australian records were included in the analysis, these events were said to have occurred during the time periods 9000–8000, 6000–5000, 4200–3800, 3500–2500, 1200–1000 and 600–150 yr BP. It is obvious from comparison to the reconstructions described above that even if these events actually occurred, not all of them are present in Australian pollen and diatom records. It would still be interesting to compare the timings of these events to any ages obtained in the current study of Princess Charlotte Bay.

2.9.2.2 The El Nino Southern Oscillation

A major factor driving late Holocene climatic variability is changes to the strength and periodicity of the El Nino Southern Oscillation (ENSO). ENSO circulation patterns are characterised by a trans-Pacific modulation of pressure in the tropics. These are in turn related to changes in the intensity of trade winds and hence tropical convection (McGlone *et al.*, 1992). ENSO events are manifested as sea surface temperature anomalies. During an „El Nino‟ phase, sea surface temperatures in the western Pacific are lower than usual which leads to reduced evaporation and drought conditions for Indonesia, Papua New Guinea and northeastern Australia. Conversely, „La Nina‟ phases result in higher sea surface temperatures and generally more pluvial conditions in the western Pacific (McGlone *et al.*, 1992; Cane, 2005).

Changes to the nature of El Nino events through the Holocene have likely played a large role in determining Holocene climatic changes and variability, and studies have found that a stepwise increase in the frequency and magnitude of El Nino events occurred at 5000 yr BP and again at 3000 yr BP (Turney *et al.*, 2006c; Donders *et al.*, 2007). This finding is consistent with the general pattern of Holocene climate changes described in Section 2.9.2.1 (e.g. Shulmeister, 1999). Aside from precipitation regimes, changes to sea surface temperatures modulated by ENSO can also have an effect on storm frequency and magnitude, which can modulate the timing of periods of coastal erosion and reworking (e.g. Dingler and Reiss, 2002). Along the Australian east coast, storm frequency could be expected to be higher during La Nina events in which sea surface temperatures are

relatively higher and hence there is more potential energy available for storm formation. Records of Holocene storm frequency are considered further in Section 2.9.3.

2.9.2.3 The Medieval Warm Period and the Little Ice Age

The Medieval Warm Period and the subsequent Little Ice Age are two more recent climatic events that have received much attention in the literature. The Medieval Warm Period is commonly described as a period lasting from approximately 1200 to 600 yr BP where Northern Hemisphere temperatures were anomalously warm. In Europe, it was generally characterised by warm, dry summers and mild winters (Bradley *et al.*, 2003). The Little Ice Age is the period following this, approximately 700 to 200 yr BP, in which summer temperatures in particular were significantly lower across much of the Northern Hemisphere (Matthews and Briffa, 2005). Although historical records from the Northern Hemisphere, and particularly Europe indicate that these events certainly occurred and were manifested as temperature anomalies, there has been considerable debate about their expression and timing in other parts of the world (e.g. Bradley *et al.*, 2003; Matthews and Briffa, 2005). Part of the uncertainty undoubtedly springs from the simple fact that far fewer records exist for the Southern Hemisphere (Cook *et al.*, 2002).

Despite this, anomalous climate events have been identified and linked to the Medieval Warm Period or Little Ice Age in a number of Southern Hemisphere studies. The possibility that these events may have affected Southern Hemisphere (or possibly Princess Charlotte Bay) precipitation regimes necessitates their consideration in this review.

The Medieval Warm Period has been identified in a number of Southern Hemisphere studies as a generally warmer and/or drier period. For example, Mauquoy *et al.* (2004) examined a bog core taken from the southern tip of Argentina and noted drier conditions between 1020 and 980 yr BP, while tree ring records from Patagonia suggest a warm period occurred between 920 and 750 yr BP (Villalba, 1994). Verschuren *et al.* (2000) examined lake sediments in equatorial east Africa and found that significantly drier conditions existed from 950 to 680 yr BP. A variety of records from New Zealand also indicate that a prolonged warm period occurred from around 900 to 600 yr BP (e.g. Wilson *et al.*, 1979; Cook *et al.*, 2002; Williams *et al.*, 2004).

The Little Ice Age has similarly been identified in a number of Southern Hemisphere studies as a generally colder and/or wetter period. For example, Patagonian tree ring records indicate a prolonged cold period from 630 to 340 yr BP, while moraine studies from the same region confirm that a glacial advance culminated between 400 and 200 yr BP (Villalba, 1994). Lake sediments in equatorial east Africa also indicate a relatively wetter climate between 680 and 100 yr BP (Verschuren *et al.*, 2000).

Even fewer records specifically from Australia exist. An extended warm period manifested by elevated summer temperatures between 1100 and 500 yr BP was identified from analysis of Tasmanian tree rings (Cook *et al.,* 2000). This broadly correlates to quoted timings of the Medieval Warm Period. This finding is consistent with the suggestion of Shulmeister *et al.* (2004) that Australia experienced reduced westerly flow during the period preceding the Little Ice Age, and enhanced westerly flow during the Little Ice Age itself. Coral cores from the Great Barrier Reef were used by Hendy *et al.* (2002) to

reconstruct 420 years of sea surface temperature changes in the tropical western Pacific. Particularly relevant to Princess Charlotte Bay was the suggestion of Hendy *et al.* (2002) that during the Little Ice Age, the tropical western Pacific may have experienced warming and increased evaporation under the influence of strengthened trade winds during this period.

2.9.2.4 Late Holocene sea surface temperatures

Sea surface temperature reconstructions should be considered along with the more general climate reconstructions described in the previous sections because of the link between them and oceanic evaporation, and hence precipitation (i.e. warmer temperatures mean more evaporation, hence more precipitation in coastal catchments, with relatively less rainfall when temperatures are cooler). The following reconstructions are based on coral palaeothermometry and oxygen isotopes of fixed biological indicators.

Gagan *et al.* (1998) studied skeletal Sr/Ca and oxygen isotope ratios in a coral sample from the Great Barrier Reef dated at having formed 5350 yr BP, and found that ocean temperatures at that time were 1°C warmer than at present. Baker *et al.* (2001b) found changes in sea surface temperatures (based on oxygen isotope ratios of fixed biological indicators in Port Hacking, New South Wales) after this time tended to occur in sync with sea level oscillations (Section 2.10.2.4), such that cooling occurred in association with a fall, and warming with a rise. Relatively colder conditions were present from 5200 to 4500 yr BP From 4500 to 3800 yr BP there was a rise in sea surface temperature, followed by

cooling again after 3800 BP. Although it is possible some warming occurred after 2000 yr BP, temperatures are thought to still have been colder than present (Baker *et al.*, 2001b).

Reconstructions of more recent sea surface temperature changes have added to the debate regarding the different spatial expression of climate changes during the Medieval Warm Period and Little Ice Ages (Section 2.9.2.3). Hendy *et al.* (2002) examined coral records from the Great Barrier Reef and found cooler (-0.2 to -0.3°C relative to present) temperatures persisted until around 300 yr BP. Following this, a century scale warming of around 0.4°C occurred centred on 200 BP, after which a cooling trend was again apparent until 100 yr BP. Shorter records from Raratonga in the South Pacific gyre identified a similar trend (Linsley *et al.*, 2000).

2.9.3 Holocene tropical cyclone activity in the Princess Charlotte Bay region

The role of storms in chenier plain development has received much attention in the literature (Section 2.5.1.4). To briefly summarise, in the past it has been suggested that chenier ridges may be either storm ridges (i.e. constructed during a single storm), or accumulations resulting from „phases of storminess". In the Princess Charlotte Bay region, the concept of a „storms" can be refined to „tropical cyclone", as these will be the major storm events capable of effecting geomorphic change. Reconstructions of Holocene tropical cyclone activity have been undertaken for the Princess Charlotte Bay/northern Queensland region and will be crucial for testing the applicability of either of the storm models in that setting.

It has been assumed by some authors that chenier ridges represent a proxy record of storminess, or variations in storminess, through the Holocene (e.g. Nott and Hayne, 2001). It is important to note that the sequence primarily used for such analyses is that studied by Chappell and Grindrod (1984) (i.e. the ridge sequence on the northeastern side of Princess Charlotte Bay where the chenier plain grades into a beach ridge plain). Given that an aspect of this investigation is to test competing models of chenier plain development in Princess Charlotte Bay, the possibility for circular logic is very real. More generally, this possibility also applies to other geomorphic storm indicators used (e.g. shingle, detrital coral and shell, chenier and beach ridges), all of which are marine features as opposed to independent proxies such as records derived from palynology. With these problems in mind, the investigations have generally concluded that there has been no variation in tropical cyclone frequency in the late Holocene (i.e. the last 3000 to 5000 years), and furthermore that the return interval of individual major events (i.e. tropical cyclones rated between category 2 and 5 on the Saffir-Simpson scale; Nott and Hayne, 2001) is between 80 and 300 years (Chappell *et al.*, 1983; Chivas *et al.*, 1986; Hayne and Chappell, 2001; Nott and Hayne, 2001). Super cyclones are hypothesised to have occurred every 200–300 years along all parts of the Great Barrier Reef (Nott and Hayne, 2001).

The nature of the records used to reconstruct Holocene tropical cyclone activity, however, means that such findings are far from certain. Forsyth (2010) has argued that reconstructed records of tropical cyclone activity may, to a certain extent, be an artefact of the coastal feature examined, and that it is possible that different types of beach ridge systems reveal different aspects of the long term cyclone climatology. This is because the nature or sensitivity of a system to record the occurrence of events varies. For example, Forsyth

(2010) developed a reconstruction of tropical cyclone activity using a record from a beach ridge plain at Cowley Beach (Figure 7). In contrast to other records from shell and coral rubble ridge plain systems (which only recorded the most intense storms), the finer pure sand beach ridges recorded a larger range of storm intensity and indicated that there may have been considerable variations in the intensity of tropical cyclones through the late Holocene. Specifically, Forsyth (2010) identified a phase of high intensity events (with ridges being deposited by category 4–5 tropical cyclones) between 5010 and 3380 yr BP, a phase of lower intensity events (category 1–4) between 3380 and 1550 yr BP, and a return to higher intensity events (category 4–5) between 1550 and 130 yr BP. While it is therefore possible that the *frequency* may have been relatively constant (as suggested by earlier authors), there may have been considerable variations in *intensity*.

2.10 Sea level changes

The following sections contain a review of knowledge regarding sea level changes on both glacial and late Holocene timescales. Glacial-cycle sea level changes and preserved Australian coastal deposits from earlier interglacial periods are considered first. Following this is a review of Holocene sea level changes (specifically those occurring from the end of the most recent post-glacial marine transgression), including curves that have been developed for the Australian east coast. Finally, the effects of hydro-isostasy and tectonic movements on apparent sea level along the northeastern Australian coast are briefly considered.

2.10.1 Glacial cycle sea level changes and interglacial highstands

Although the periodicity has varied through time, the alternation of glacial-interglacial cycles has been a feature of Earth‟s climate since about 3 Myr BP. During this time, change in continental ice volume of 50–60 million km^3 have resulted in global sea level changes with amplitudes of 120–140 m (Lambeck and Chappell, 2001; Lambeck *et al.*, 2002; Caputo, 2007). Various sea level curves spanning the last three to four glacial cycles, a period extending to around 350 000–450 000 yr BP (Caputo, 2007), are shown in Figure 10. Clearly, there are a very limited number of times during the last 450 000 years that sea levels have been as high as they are at present.

Although Figure 10 provides good information on the timings of interglacial highstands, because no Australian sites or data have been included in the analysis it provides very little information on the actual level of these highstands in Princess Charlotte Bay. A commonly cited height is between 0 and 5 m above present level, a record obtained from the Huon Peninsula in Papua New Guinea (Esat *et al.*, 1999). Previous interglacial shoreline deposits, however, may help to further constrain this level around the Australian coastline. Relatively few examples of coastal deposits from previous interglacial periods are preserved in Australia, particularly for the north and northeastern Australian coastlines (Murray-Wallace and Belperio, 1991; Woodroffe *et al.*, 1992). Of those that have been used to estimate previous interglacial sea levels, the most interesting to consider in terms of a chenier study include several beach ridge deposits (e.g. Murray-Wallace and Belperio, 1991; Woodroffe

et al., 1992; Ward, 2006). The physical characteristics of these and other last interglacial coastal deposits will be considered in Chapter 3.

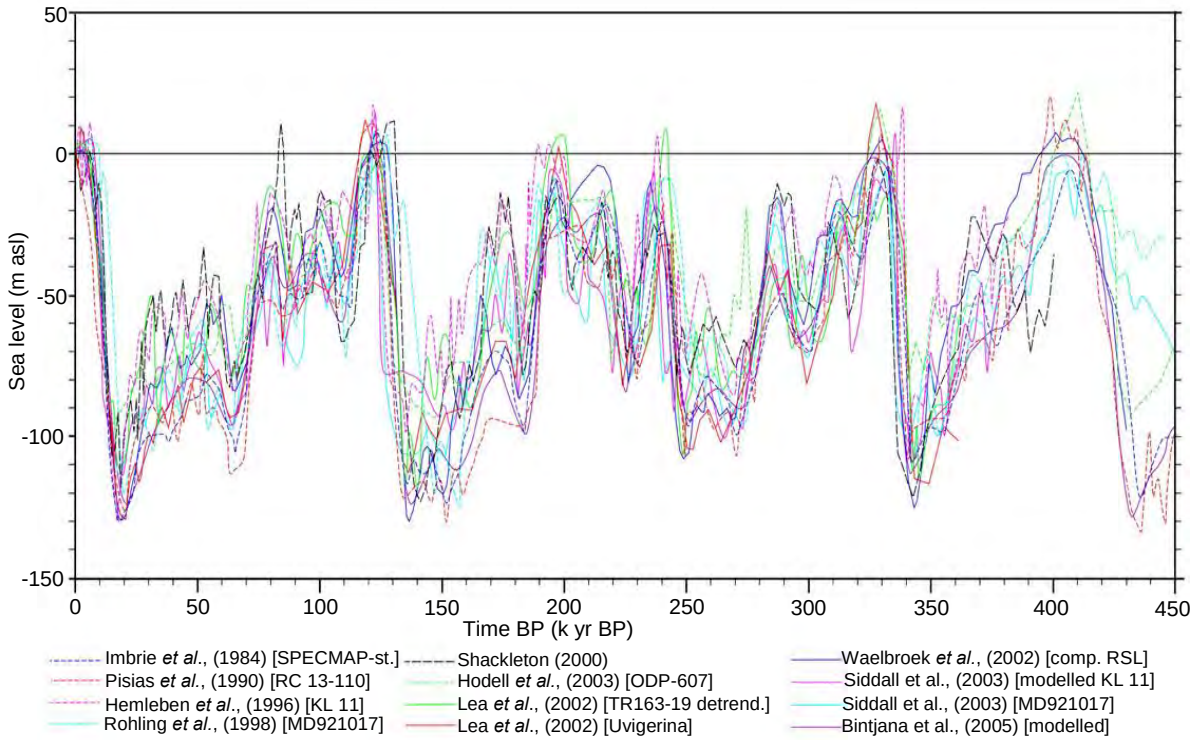

Figure 10. Various records and models of sea level changes over the last 450 000 years (modified from Caputo, 2007).

Woodroffe *et al*. (1992) examined last-interglacial shore platforms and preserved beach ridge deposits on the Cobourg Peninsula in the Northern Territory. Their findings supported the theory that at some stage during the last interglacial, sea levels were within 1–2 m of their present position for a substantial period of time. The features they investigated are also interesting in demonstrating the potential for reoccupation of coastal landforms formed in previous interglacial periods. This inheritance is an important feature in explaining the shoreline that is seen today (Woodroffe *et al.*, 1992). More recently, Ward (2006) examined coastal dunes and strand plains near Noosa (southeast Queensland) and found evidence of four preserved interglacial shorelines (with ages of around 125 000, 245 000, 335 000 and

410 000 yr BP). These deposits implied relative sea levels very close to present levels at 1, 0, 1 and 3 m above present levels, respectively (Ward, 2006).

The existence of last interglacial shorelines around the Australian continent was reviewed by Murray-Wallace and Belperio (1991). They obtained ages from a number of preserved coastal deposits (mostly from South Australia) and used these to provide estimates of sea level around 125 000 yr BP. They concluded that in contrast to the commonly cited height of +6 m, the Australian sites consistently indicated a height of +2 m, although this ranged from -2 m around the Great Barrier Reef to +32 m in Tasmania due to regional differences in neotectonic deformation. Evidence from older deglaciation periods was generally harder to delineate due to poorer preservation. Murray-Wallace and Belperio (1991) suggested that the general lack of preservation across northern and northeastern Australia may be in part due to shelf subsidence combined with burial of last interglacial sediments by actively prograding Holocene strata. A slight downwarping of around 2 m occurring along the north/northeastern coastlines of Australia was also identified by Bryant (1992) using trend surface analysis of last interglacial sea-level elevations.

Previous interglacial periods aside, the most recent post glacial rise in global sea level began around 20 000 yr BP (Figure 10). The ~130 m rise after this time was marked by stillstands and possibly minor regressions. An example of one such stillstand/regression is associated with the Younger Dryas cold reversal around 12 000 yr BP (e.g. Lambeck *et al.*, 2002), the climatic effects of which were discussed in Section 2.9.1.2. This is of little relevance to this review, however, as sea levels were between 50 and 60 m below present levels when they occurred, meaning present-day coastal sites were still far removed from

any shoreline processes. The global post-glacial marine transgression ceased in the mid to late Holocene, and it is from that point until the present in which most chenier plains worldwide have formed (Augustinus, 1989; Stanley and Warne, 1994).

2.10.2 Holocene sea level changes

Generally speaking, coastal progradation will not occur under a transgressive sea level as erosion and reworking dominate (Section 2.5.1.7). Establishing the timing of the cessation of the Holocene post-glacial marine transgression is therefore crucial as this will mark the point in time from which progradation of coastal plains could begin. This is considered first, followed by a discussion of various late Holocene sea level curves that have been developed for the eastern Australian coastline.

2.10.2.1 Cessation of the post-glacial transgression

Older studies tended to date the cessation of the post-glacial marine transgression at some point between 6000 and 4500 yr BP (e.g. Hopley, 1978; Chappell *et al.*, 1982; Chappell, 1983) and although some older ages were obtained from features presumed to be Holocene (e.g. 6980 ± 130 yr BP for a notch in a Pleistocene dune calcarenite), such ages were viewed as enigmatic (Hopley, 1975 in Hopley, 1978). It is doubtful if this age would be considered in the same light given more recent findings, which suggest it could have been much earlier. Stanley and Warne (1994) compiled many ages indicating time of delta

initiation around the world. They found that that globally, the post-glacial marine transgression ceased between 8500 and 6500 yr BP. The best estimate for the timing of this event around the Australian coastline was 6800 yr BP (Stanley and Warne, 1994). More recent studies from a range of sites around the Australian coastline have similarly constrained the timing to the period between 8000 and 6200 yr BP (Sloss *et al.*, 2005; 2007; Collins *et al.*, 2006; Horton *et al.*, 2007; Lewis *et al.*, 2008; Woodroffe, 2009). 6800 yr BP was maintained as the estimate of the timing of the post-glacial marine transgression for this review. It is now well accepted that sea levels at this time were between +1 and +2.8 m relative to present, and that they have since fallen (Chappell, 1983; 1987; Baker and Haworth, 1997; Haworth *et al.*, 2002; Sloss *et al.*, 2005; 2007; Horton *et al.*, 2007; Lewis *et al.*, 2008; Woodroffe, 2009).

2.10.2.2 Smooth or oscillating late Holocene sea level curve?

Until relatively recently, the most commonly referred to sea level curve for the northeastern Australian coastline has been the linear model of Chappell (1983), which indicates a smoothly falling sea level from around 6000 yr BP until present. As mentioned in Section 2.5.1.7, this has resulted in the possible effects of sea level oscillations on the development of Australian chenier plains being largely ignored.

More recent evidence has been interpreted as indicating that sea level change for the east coast of Australia since 6800 yr BP was not smooth, and instead may have been characterised by a number of minor fluctuations or oscillations (the „oscillating" model; e.g.

Baker *et al.*, 2001a) superimposed on the general falling trend. Given that this concept was explicitly rejected in previous Australian chenier plain studies (Section 2.5.1.7), it is of great importance to review possible evidence so that its role can be thoroughly investigated in the current study. The linear model is summarised first, followed by a review of evidence that supports the oscillating model for the Australian east coast.

2.10.2.3 Smooth sea level curve

In the 1960s and through to the 1980s the smooth, non oscillating or „Shepard" Curve was the preferred way to represent late Holocene sea level changes (Kearney, 2001). Pirazzoli (1991) summarised Holocene sea level change in northeastern Australia by synthesising data from ten publications between 1979 and 1989. All used a linear line of best fit for their data, indicating a smooth sea level fall from +1–2 m above present since the cessation of the post-glacial marine transgression, which is now known to have occurred around 6800 yr BP. The most commonly used in Australian chenier plain studies has been that of Chappell (1983), who examined microatoll development in north Queensland and concluded a straight line was the simplest that fitted the results obtained (Figure 11).

Chappell (1983) did, however, state that the smoothly falling sea level may not be the only interpretation. Chappell (1987) later conceded that evidence for sea level oscillations may be difficult to find as the erosional effects of transgressive/regressive cycles means it is likely to be destroyed. With respect to his 1983 data, he believed that although oscillations were unlikely to have occurred in the period 6000–4000 yr BP, they could not be ruled out from 4000 yr BP to the present owing to a lower density of datapoints (Chappell, 1987).

Furthermore, Pirazzoli (1991) warned that conclusions such as „a straight line is the simplest that fits the data" can be dangerous as they introduce a subjective aspect to the interpretation of the dataset and of the errors in it.

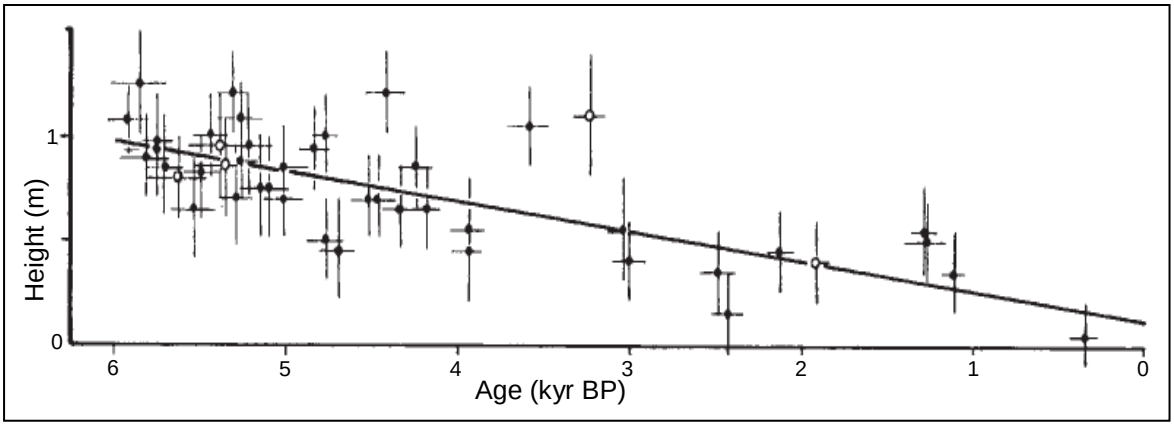

Figure 11. Age-height plot of north Queensland microatoll sea-level data and smooth model of sea level change (modified from Chappell, 1983). Vertical bars span maximum and minimum estimates of past mean low water spring tide level (MLWS) relative to present MLWS, for each station. Horizontal bars span ±1σ for radiocarbon age error.

A problem with the smooth sea level model is uncertainty arising from the range of geomorphic, biological and systematic variance in sea level indicators used (Pirazzoli, 1991; Baker and Haworth, 2000a). The shortfalls in such methods have been covered by maximising standard error margins in sea level height estimations by enveloping curves with variance in the order of 1m or greater, and using the argument that the „real" sea level curve lies somewhere within the envelope (e.g. Chappell, 1983; Chappell *et al.*, 1983). As mentioned, the consequence of this was that the linear model was identified as the simplest that fitted the data. This approach may, however, have disguised more subtle (i.e. <1 m) changes, or oscillations, in sea level during the late Holocene. The use of different indicators with smaller error margins has overcome these problems (Baker and Haworth, 2000a) and has provided evidence of such oscillations.

2.10.2.4 Oscillating sea level curve

The idea that late Holocene sea level change occurred in a series of oscillations, as opposed to a linear fall, was introduced largely by the work of Fairbridge (1960). Interestingly, it is the period from 6000–4000 yr BP in which Chappell (1987) believed no oscillations occurred that more recent studies suggest most oscillations may have taken place. The supporting evidence comes from a variety of indicators including fossil molluscs in transgressive deposits, various stratigraphic assemblages, mangrove roots, coral microatolls, and fixed biological indicators from locations including Magnetic Island, the northeastern Queensland coast, various sites along the NSW coast, as well as further afield sites in the Pacific Islands and Brazil (i.e. Baker and Haworth, 2000a; b; Baker *et al.*, 2001a; Sloss *et al.*, 2007; Lewis *et al.*, 2008).

Initially, statistical regressions on a variety of data, including that used by Chappell (1983) to develop the linear model, demonstrated that neither the smooth or oscillating models could claim statistical exclusivity, and furthermore that the simple linear model contained too many shortcomings for use as an interpretative tool for late Holocene sea levels (Baker and Haworth, 2000a). With regards to the eastern Australian data, the best fit was found for the quintic polynomial regression model. The oscillating model, for which various „estimates" were provided by the regression analysis, was then tested against biostratigraphic assemblages of relic intertidal formations at a variety of sites in eastern Australia and beyond. Evidence from each location suggested a broad similarity in the timing of sea level changes. These were summarised as including a rapid sea level rise after

4500 yr BP to a maximum of around +1.7 m, with at least one marked fall after 3900 yr BP and a possible small positive oscillation around 1900 yr BP (e.g. Baker and Haworth, 2000a; b; Baker *et al.*, 2001a).

This generalised pattern has been supported and refined by the findings of more recent studies. Sloss *et al.* (2007) reviewed an extensive range of sea level indicator data from eastern Australia, concluding that the post-glacial highstand of +1.7 m attained by 7000 yr BP lasted until around 2000 yr BP, and that during this time a series of minor oscillations probably occurred. The precise nature of these oscillations was said to be difficult to quantify because of problems associated with accurately determining a host of palaeo-conditions that may have changed since the mid Holocene, such as wave climate and shoreline morphology (Sloss *et al.*, 2007). Further problems arise from the fact that there are systematic differences between individual sea level indicators, mostly as a result of variances in their growth range within the tidal zone (Lewis *et al.*, 2008). These problems aside, the most recent review by Lewis *et al.* (2008) has presented a similar generalised pattern of late Holocene sea level change. They suggest that sea level reached its maximum by 7000 BP, followed by two 0.3–1.0 m oscillations (4800–4500 and 3000–2700 yr BP) and a fall to present levels after 2000 yr BP, as shown in Figure 12. Although finding no evidence for negative oscillations, Woodroffe, S. (2009) presented a similar general pattern of relative sea level change based on foraminifera from Cleveland Bay, northern Queensland. She suggested sea level have been relatively stable above +1.5 m from 6200 cal yr BP until at least 2300 cal yr BP, however, noted that oscillations within the highstand could not be ruled out entirely.

Sea level curves developed for data from further afield sites in Western Australia, South Australia, the Pacific Islands and Brazil (Figure 12) have largely agreed with the generalised patterns described above. Although the degree of predicted vertical change differs, the timings of the changes (particularly the hypothesised regression followed by a transgression between approximately 5000 and 2000 yr BP) appear very similar. The curves from French Polynesia also suggest that a recent stillstand may have occurred meaning that sea level was higher until as recently as 500 yr BP (Figure 12).

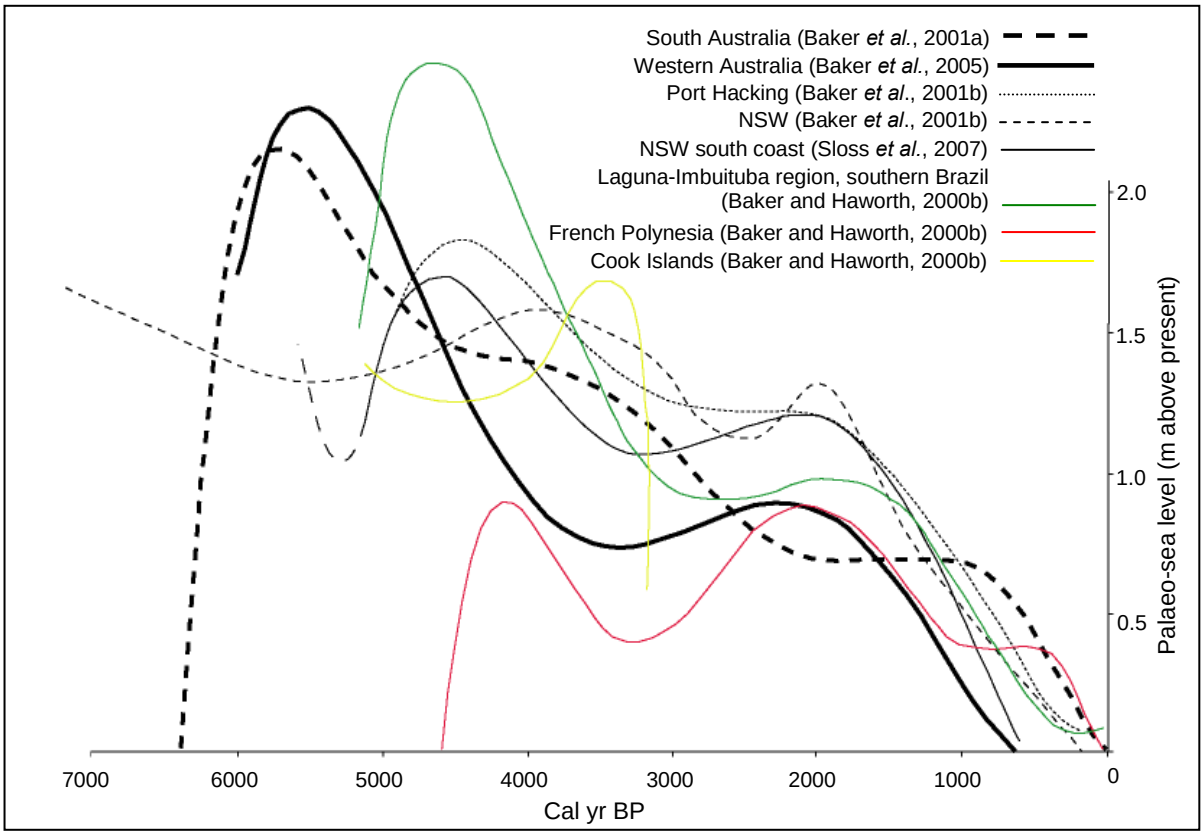

Figure 12. Examples of possible oscillating sea level curves developed for Australia, the Pacific Islands and Brazil.

The body of evidence indicating sea level variability through the late Holocene has grown to a point where some authors believe the debate regarding the „smooth versus oscillating" models may be outdated (e.g. Kearney, 2001). The occurrence of eustatic oscillations is

further supported by studies that have identified possible driving mechanisms, for example changes in Antarctic ice volume (Goodwin, 1998). Lewis *et al.* (2008) contains an excellent summary of factors that may drive small scale sea level changes in the late Holocene.

2.10.2.5 Possible geomorphic expressions of sea level oscillations

Several authors have directly linked observed geomorphic changes to late Holocene sea level oscillations. These links have helped to move the debate over a smooth versus oscillating sea level curve beyond the sea level curves described in the previous section by providing physical evidence of geomorphic changes that appear to have been driven from a marine source. The timings of these changes provide evidence for at least some of the oscillations hypothesised by the curves, and boost the argument against a smooth curve. The examples considered below include regional investigations of dune destabilisation and localised investigations on beach ridge development. It should be noted that none of the features referred to in this section have been used specifically to reconstruct sea level curves (e.g. transgressive deposits, stratigraphic assemblages, mangrove roots, coral microatolls and fixed biological indicators; Section 2.10.2.4), rather they are geomorphic features that have been dated and investigated independently of palaeoenvironmental records.

Lees (2006) investigated the timing of coastal dune development across northern and eastern Australia and found that there were synchronous dune mobilisation events in sites between Cape St. Lambert in the northwest and Kurnell in the southeast. These occurred

between 3400 and 2900 yr BP, possibly again between 2500 and 2000 yr BP and again in the last 1000 years, and the coastal dune fields appeared to have a marine trigger (Lees, 2006). Geomorphic evidence for either coastal erosion or higher sea levels during the mid to late Holocene has been reported from a number of investigations into beach ridge development. Bryant *et al.* (1992b) explained the occurrence of a Holocene beach deposit near Sandon Point, New South Wales by inferring sea levels over 1 m higher than present had persisted until 1520 yr BP. This is in contrast to the smoothly falling sea level curve (Chappell, 1983) which suggests sea level would be considerably lower at that time (Figure 11; Section 2.10.2.3). Nott (1996) put forward a similar interpretation for a beach ridge sequence near Darwin, which was dated as having formed between 3390 and 2020 radiocarbon yr BP (3713–2108 cal yr BP). He concluded that the evidence suggested that sea levels may not have fallen progressively from the mid Holocene, and specifically, did not fall to its present position until after 2000 yr BP. More recently, Brooke *et al.* (2009) used OSL to examine a beach ridge sequence within Moreton Bay (southeast Queensland) and identified a period of shoreline erosion that occurred between 2600 and 1700 yr BP, followed by progradation until the present. A possible link to sea level was noted, although other mechanisms were not dismissed. Switzer *et al.* (2009) also found a beach ridge sequence in Batemans Bay had developed under the influence of sea levels around 1–1.5 m higher than present, which persisted from 5000 yr BP until between 2500 and 2000 yr BP. The fact that these features are beach ridges, which form under a different depositional regime to chenier ridges, is irrelevant. What is important is that they constitute evidence for geomorphic changes driven by sea level regimes that would appear to be at odds with the smooth sea level hypothesis, and which are consistent with some of the oscillating curves put forward (Figure 12; previous section).

2.10.2.5 Hydro-isostasy and tectonic effects

The effect of regional hydro-isostatic adjustments resulting from post-glacial ocean water loading on the continental shelf has received much attention in past Australian Holocene sea level and chenier plain investigations (e.g. Chappell *et al.*, 1982; Rhodes, 1982; Chappell, 1983; Chappell and Grindrod, 1984). Since the degree of adjustment is largely a function of coastline geometry, it has been used to explain regional variations in sea level records, particularly differences in the level of the mid to late Holocene highstand (e.g. Chappell *et al.*, 1982; Lambeck and Nakada, 1990).

The degree of hydro-isostatic adjustment in the Princess Charlotte Bay region has been estimated as approximately 1.0 m since 5500 yr BP, equating to ~0.2 mm/yr (Chappell *et al.*, 1982; Chappell, 1987). In the southeastern portion of the Gulf of Carpentaria a similar adjustment is estimated at closer to 0.5 mm/yr (Chappell *et al.*, 1982; Chappell, 1987). If the hydro-isostatic deformation model is accepted, then Pleistocene shorelines may have undergone a similar amount of warping during the glacial low sea levels when oceanic load was removed from the shelf. Only with the reapplication of this load during the Holocene would they be resuming a near-horizontal disposition (Hopley, 1978). Haworth *et al.* (2002) believe that the degree of hydro-isostatic adjustment may have been overemphasised in the past. In any case, in terms of factors that may have affected shoreline processes, a gradual (smooth) uplift of 1 m since 6000 yr BP would be minimal compared to the more rapid oscillations discussed in Section 2.10.2.4.

Tectonic movements can also play a role in observed sea level changes. Eastern Queensland is not recognised as a zone of high seismicity, however, and there is no need to invoke tectonic movements to explain Holocene sea level changes in this region (Lambeck and Nakada, 1990).

2.11 Human occupation and fire in the Princess Charlotte Bay region

In Princess Charlotte Bay chenier development due to the advent of European environmental impacts is inconceivable due to the lack of development and anthropogenic modification of the Normanby-North Kennedy Catchments. The existence of shell middens in the area, however, point to a history of pre-European Aboriginal occupation (Chappell and Grindrod, 1984). Some Australian palaeoenvironmental studies have found increases in sedimentation rates associated with the onset of widespread Aboriginal land use, particularly burning, in catchments (e.g. Haberle and David, 2004) and there is evidence to suggest that the vegetation in some coastal areas of northeast Queensland has been manipulated by fire through the mid to late Holocene.

Hill and Baird (2003) investigated the wet tropics region of northeastern Australia and found evidence that traditional environmental management techniques had been used prior to European occupation, in particular to maintain open patches within the rainforest matrix and the position of the rainforest/sclerophyll boundary. There is evidence to suggest that the intensity of firing in these areas was most pronounced between 3700 and 2000 yr BP, and that activity has dropped since then (Haberle and David, 2004). Hopkins *et al.* (1993; 1996)

drew similar conclusions and suggested that the maximum sclerophyll expansion into rainforest zones occurred from 4000–3500 yr BP in a time when climate would have been pushing vegetation change in the opposite direction. By 1000 yr BP this had stopped and rainforest was again expanding.

2.12 Summary

Part B of this review summarised palaeoenvironmental knowledge that could be used to test the applicability of different models of chenier plain development in Princess Charlotte Bay. This included reconstructions of palaeoclimates (including sea surface temperature and storm activity) and sea level change through the Pleistocene and Holocene.

Climatic conditions attributable to the last glacial period may have persisted until as late as 14 000 yr BP in the Queensland area. The warming trend into the Holocene may have been punctuated by minor cooling events, which appear to have been recorded in a number of dune activation events along the Queensland coastline. Post glacial warming in the Princess Charlotte Bay area continued after this with Holocene optimum (wet) conditions lasting from as early as 8000 yr BP to as late as 5000 yr BP, when conditions began to become more variable. A further increase in variability occurred after 3000 yr BP. Although this is generally recorded as lower levels of precipitation, some sites on the northeastern Queensland coastline appear to have recorded higher levels of rainfall since then.

Changes in the El Nino Southern Oscillation are in agreement with this general pattern. Other events that may have affected late Holocene climates include the Medieval Warm Period and Little Ice Age climate shifts. Although sea surface temperature changes during the Holocene have been identified from sites further south on the New South Wales coast, evidence suggests Holocene storm frequency in the Princess Charlotte Bay region has been relatively constant, with an 80 to 300 year return interval of major storms.

With regards to sea levels, the last interglacial high sea level occurred around 125 000 yr BP. Coastal deposits from previous interglacial periods are generally poorly preserved, particularly on the northern and northeastern Australian coastlines. With regards to Holocene sea levels, it is generally agreed that sea level reached a peak around 6800 yr BP. Although past Australian chenier plain studies have referred to the smooth curve of Chappell (1983), more recent evidence suggests that there may have been a series of minor oscillations superimposed on the general falling trend. Hydro-isostatic and tectonic influences on sea level are minor in the Princess Charlotte Bay area.

PART C – DATING TECHNIQUES

2.13 Introduction

All previous Australian chenier plain studies have utilised radiocarbon dating techniques to date the emplacement of chenier ridges, primarily using whole shells and shell fragments (Section 2.13). Although a modest body of work has been built around these methods, issues exist that introduce a degree of uncertainty into the ages obtained for chenier ridges (and indeed other types of beach ridges). These issues can be avoided by the use of luminescence dating techniques that, as yet, have not been widely applied in other chenier plain studies.

It must be stated at the outset that is certainly not the author‟s intention to dismiss the value of radiocarbon ages or present the method as unreliable. Indeed, radiocarbon ages were obtained in this study to compliment luminescence ages that appeared unreliable, and also to allow other comparisons (i.e. between the dating methods, and between the results of this study and those of Chappell and Grindrod, 1984) to be made. Furthermore, radiocarbon ages from other Australian chenier plain studies have been used in a statistical analysis which forms a crucial piece of work for this thesis. The intention of the following section is simply to acknowledge issues that have been noted at length in previous studies, and justify the attempt at using an alternate dating method.

The following sections summarise two of the issues associated with the use of radiocarbon dates in chenier plain studies, and outline the principles behind luminescence techniques.

2.14 Issues associated with using radiocarbon dating in chenier studies

Two issues associated with using radiocarbon dating of shell material in chenier ridges to date the time of ridge emplacement are (1) chemical influences on the shell material due to the marine environment (i.e. marine reservoir effects), and (2) the temporal difference between when a shell is made, and when it becomes incorporated into a chenier ridge (i.e. associating sample with event). Although other sources of error and complications exist (e.g. Hopley, 1978; Walker, 2005), these two appear to be the most commonly cited in the context of chenier plain studies.

2.14.1 Marine reservoir effects

Radiocarbon dating methods obtain ages for material by comparing the amount of ^{14}C isotopes found in a sample (which decay over time) to that which would be expected in a modern, fresh sample (i.e. one in which apparently no ^{14}C decay has taken place since it is living and constantly being replenished) (Walker, 2005). At the surface of the ocean, the incorporation of ^{14}C isotopes into seawater occurs constantly during exchange reactions between the atmosphere and the ocean. When this surface water sinks, it is no longer „replenished" with ^{14}C, decay occurs, and the seawater will have an apparent age (Gillespie and Polach, 1979). Since marine molluscs use the carbonate in sea water to construct their

shells, radiocarbon ages obtained from some shells can be falsely old, particularly in areas where the age effect is particularly pronounced. For example, in the equatorial eastern Pacific the apparent age difference is approximately 600 years (Shackleton *et al.*, 1988). This is known as the „marine reservoir effect" and is rectified by subtracting a correction factor that varies spatially according to ocean water movements such as upwelling, surface currents, thermohaline circulation and other factors (Walker, 2005). In Australia the correction factor is thought to be 450 ± 35 years (Gillespie and Polach, 1979) and, where applied, this value has been utilised by every Australian study to date.

Problems do exist, particularly for estuarine environments where fresh water mixing occurs and some authors have debated whether or not to apply the correction factor in certain settings (e.g. Woodroffe *et al.*, 1985a; Chappell, 1993; Stone, 1995). It is also conceivable that the factors involved in ocean water movements (e.g. upwelling, surface currents, thermohaline circulation etc), and therefore the most appropriate correction factor, would vary considerably around the Australian coastline. Furthermore, recent evidence suggests that as well as spatial variations, the correction factor may also vary temporally (Walker, 2005).

2.14.2 Associating sample with event

The second issue arising from the use of radiocarbon methods to date marine shell in chenier ridges is the uncertainty regarding the temporal difference between when a shell was created (i.e. when the organism which created the shell was alive) and when its shell

became incorporated into a chenier ridge. This issue has been noted in several past Australian chenier plain studies (e.g. Rhodes *et al.*, 1980; Lees, 1987; Lees and Clements, 1987; Kowalewski *et al.*, 1994; Woodroffe and Grime, 1999). Because many chenier ridges can be formed by winnowing and sorting from existing mudflats (Section 2.5), this time period could be considerable, for example hundreds or thousands of years. This could lead to falsely old ages for chenier ridge emplacement. For example, Lees (1992a) obtained ages from what appeared to be „fresh" shell material from the most modern foredune ridge in the Victoria Delta in northern Australia. The ages suggested the shell material was between 4000 and 6000 years old. Clearly, therefore, these ages could not be used to date emplacement of a modern ridge. In an attempt to deal with this problem, Lees (1987) had previously suggested using the youngest age from a chenier ridge as the most representative of its „true" age. This would make it less likely that falsely old ages would be obtained from ridges formed at a later date from older material that had been reworked.

2.14.3 A possible alternative to radiocarbon methods

Kowalewski *et al.* (1994) considered the two issues discussed above and formed the belief that it is impossible to use dated shells to determine exactly when a given generation of chenier ridges started to form. Put simply, the age of shell or shell fragments does not necessarily reflect the age of the sediments in which they are found (Belperio *et al.*, 1984). Furthermore, even if these issues could be confidently remedied, the fact that some chenier ridges are composed entirely of sand or gravel (e.g. Russell and Howe, 1935; Schou, 1967; Bird, 1970; Cook and Polach, 1973; Jennings and Coventry, 1973), and therefore have no

shell material to provide ages in the first place, makes radiocarbon techniques immediately inappropriate in some settings.

The above issues notwithstanding, in many settings radiocarbon methods appear to have yielded reasonable depositional chronologies, and have allowed the development of a vast body of coastal geomorphological knowledge. Ideally, however, the preferred method would be the one which was free of these issues. For chenier ridges, this may mean a method which can estimate the time of chenier ridge emplacement by directly dating the age of the sediments forming the ridge, as opposed to dating material found in association with those sediments such as marine shell or mangrove fragments. Luminescence techniques are capable of this, and appear to provide an alternative to radiocarbon methods. Luminescence techniques are discussed in the following sections.

2.15 Luminescence dating techniques

Luminescence dating was developed in the early 1960s as a technique to date archaeological samples such as heated rocks and pottery (Forman, 1989; Duller, 1996). Wintle and Huntley (1980) later found that by making some small modifications to the method, it could be applied to sediments that had been exposed to daylight at time of deposition. Sediment dating by luminescence was subsequently developed through the 1980s. Only in recent years, however, has it come into wider use by geoscientists. This was due to a combination of lack of willingness to change away from traditional (trusted) radiocarbon techniques, and methodological problems while luminescence techniques were

being developed (e.g. Gibbons, 1997). Luminescence dating has since been applied extensively to various types of sediments including volcanic products, cave speleotherm carbonate, loess, glacial lake deposits, deep ocean sediments, and coastal sediments containing quartz and feldspar (Aitken, 1985; Forman, 1989; Wintle, 1997; Walker, 2005).

Luminescence techniques may offer a promising avenue for the advancement of chenier plain studies in Australia. Although as yet no studies have utilised them to date chenier ridges, because these techniques are capable of dating quartz grains directly they completely avoid problems related to using shell material (Section 2.15). Furthermore, because the event they are dating is „last exposure to sunlight" of mineral grains (Huntley *et al.*, 1985; Prescott and Robertson, 1997; Stokes, 1999), the ages obtained may reflect time of chenier ridge emplacement more accurately. Although this position seems reasonable, for ages obtained using luminescence techniques to be reliable, chenier ridge sediment must be sufficiently exposed to sunlight at the time of deposition. Relative to radiocarbon techniques this caveat appears to be the only major drawback of luminescence, and influenced the methods used and results obtained in this study. This is discussed further in Chapter 5.

2.15.1 Reliability of luminescence dating techniques

In order for luminescence techniques to be employed over radiocarbon methods, their reliability must be established. To do this, several investigations have compared luminescence ages with those obtained from radiocarbon (and other methods) in dating

samples that are undoubtedly chronologically matched (e.g. Prescott and Robertson, 1997). They include comparisons with Uranium-Thorium dating (e.g. Bryant *et al.*, 1990; Nanson *et al.*, 1991), radiocarbon (e.g. Belperio *et al.*, 1984; Gardner *et al.*, 1987; Readhead, 1988; Nott and Price, 1991; Prescott and Robertson, 1997; Hormes *et al.*, 2003; Argyilan *et al.*, 2005) or both of these techniques (e.g. Bryant *et al.*, 1992b).

The similarities have been very encouraging. Particularly in comparison to radiocarbon, the results have shown „reasonable agreement" (Belperio *et al.*, 1984), „no significant differences or contradictions" (Gardner *et al.*, 1987), „concordance" (Readhead, 1988), „good agreement" (Nott and Price, 1991), „mutual supportiveness" (Bryant *et al.*, 1992b) „agreement to an acceptable degree" (Prescott and Robertson, 1997) and „consistency" (Hormes *et al.*, 2003). Argyilan *et al.* (2005) also found agreement between the methods and concluded that luminescence provides a credible alternative to radiocarbon for dating littoral and aeolian landforms. A final advantage of luminescence over radiocarbon is that as well as obtaining results in the form of ages and associated error ranges, post-hoc analysis of the data can be used to assess the reliability of luminescence results (Aitken, 1985). These are not available for radiocarbon, where the result is in the form of an age range only.

The positive results of these comparisons and other advantages have helped to establish the potential of luminescence dating as not only a reliable geochronological tool in chenier plain studies, but also one that may be a more appropriate alternative to traditional radiocarbon methods. The following section briefly considers some of the main concepts behind two commonly utilised luminescence techniques. More comprehensive reviews of

luminescence theory and practise can be found in Aitken (1985), Forman (1989), Duller (1996), Stokes (1999), Duller (2004) and Walker (2005).

2.15.2 A brief overview of luminescence concepts

Luminescence dating methods work on the principle that materials lying in close proximity to naturally occurring radioactive isotopes such as uranium, thorium or potassium-40 (such as in buried contexts) will be subject to low levels of radiation through environmental radioactive decay and cosmic radiation (Aitken, 1985; Walker, 2005; Lian and Roberts, 2006). In mineral crystals such as quartz and feldspar the radiation causes ionisation of atoms and electrons freed during this process may become trapped at defects or „holes" in the crystal lattice. This leads to a build up of energy at these „electron traps" (or „luminescence centres") (Huntley *et al.*, 1985; Forman, 1989; Bøtter-Jensen, 1997; Duller, 2004). The energy stored in these traps increases through time as the crystal is subjected to a constant rate of environmental radiation (this is referred to as the „dose rate" and generally has units of yr^{-1}). Since electron traps can be (re)emptied by controlled exposure to heat or light in the laboratory, the amount of energy stored in them can be used to quantify the time that has elapsed since the traps were last emptied due to exposure to sunlight or heat (Walker, 2005; Lian and Roberts, 2006). Archaeological dating of pottery, for example, quantifies time elapsed since the pottery was fired. The release of energy in the laboratory occurs in the form of an emission of light and is referred to as the „luminescence signal". It can be quantified using a photomultiplier tube, a light-sensitive instrument that converts photons released from luminescence centres into electrical signals (Aitken, 1985; Bøtter-

Jensen, 1997; Stokes, 1999). The strength of luminescence signal is related to the amount of environmental radiation that the quartz has been subjected to since it was last exposed to sunlight. This amount is known as the equivalent dose (D_e). The actual way in which the equivalent dose is determined varies between methods. This is demonstrated and explained in Section 4.6.1.

Once equivalent dose and dose rate are known, the luminescence age can be determined using the equation below:

$$\text{Age} = \frac{D_e\,(\text{Gy})}{\text{Dose rate (Gy/yr)}}$$

There are two main ways that luminescence from trapped electron populations can be released in the laboratory, and these form the bases of different luminescence techniques.

2.15.3 Thermoluminescence and Optically Stimulated Luminescence

Two common techniques in luminescence dating are known as thermoluminescence (TL) and optically-stimulated luminescence (OSL). TL, as the name suggests, uses heat to empty the electron traps and generate a luminescence signal. OSL uses particular wavelengths of light. The main implication of this difference is that different types of electron traps within the crystal matrix are targeted. TL measures luminescence signals from a wide range of traps, some of which are sensitive to light and some not. OSL measures only those sensitive to light.

There are two consequences of this difference. Firstly, the TL signal will never reach absolute zero, even after prolonged exposure to light. This adds an element of uncertainty regarding the „residual" signal. Secondly, sediments dated using OSL need much less exposure time (in the order of seconds) to reset the luminescence clock than those dated using TL, which take hours to have their signal reset (Duller, 1996; Prescott and Robertson, 1997; Stokes, 1999; Duller, 2004; Walker, 2005; Lian and Roberts, 2006). This has important implications for which technique is chosen for any particular investigation. TL has generally been applied to aeolian sediments such as sand dunes (e.g. Gardner *et al.*, 1987; Readhead, 1988; Nott and Price, 1991; Bryant *et al.*, 1992b; Huntley *et al.*, 1993; Lees *et al.*, 1995) as complete resetting of the luminescence signal is rarely, if ever a problem (Prescott and Robertson, 1997; Wintle, 1997; Duller, 2004). OSL is more appropriate for sediment that may not have experienced complete exposure to the sun, for example in fluvial settings where factors such as turbidity or shading from vegetation may influence exposure (Forman, 1989). Because OSL can avoid problems of incomplete exposure to a greater extent than TL, it has become the preferred method for dating sediment in geomorphic investigations (Walker, 2005).

2.16 Summary

Part C of this review has described two of the issues associated with using radiocarbon dating for chenier studies, and outlined the principles behind luminescence techniques which may be used as an alternate dating method.

The two most commonly cited issues involved in using radiocarbon methods in chenier plain studies are marine reservoir effects, and temporal differences between sample age and time of chenier ridge building. While these issues certainly do not make radiocarbon methods unreliable, they nevertheless remain as issues that may introduce uncertainty into ages for chenier ridge emplacement. Since luminescence techniques are capable of dating the time that quartz grains in chenier ridges were last exposed to sunlight, they avoid these problems and as such, appear to constitute a possible alternative for dating chenier ridges. Confidence in the reliability of ages obtained using luminescence dating comes from the fact that luminescence and radiocarbon techniques have been shown to yield similar results in locations where they are used to date samples that are undoubtedly chronologically matched. There are two main types of luminescence dating, thermoluminescence and optically-stimulated luminescence, and the latter has become the preferred method for geomorphic investigations. Both will be employed in this investigation.

Chapter Three - Research Design

3.1 Introduction

As stated in Chapter 1, the primary aim of this research was to investigate the Holocene development of the Princess Charlotte Bay chenier plain, and the secondary aim was to consider the broader formation of several chenier plains in Australia and identify possible relationships. The research was undertaken using new data from a chenier ridge sequence on the western side of the Princess Charlotte Bay plain, data obtained from reassessing results of previous Australian chenier plain studies, and considering these in the context of new palaeoenvironmental knowledge.

The current chapter describes the analysis framework through which the research was conducted. Firstly, the new data obtained from Princess Charlotte Bay was used to consider the applicability of various models for chenier plain development by testing hypotheses based on those models. Second, the method used to reassess the results of previous Australian chenier plain studies is briefly introduced.

3.2 Chenier plain development models in Princess Charlotte Bay

In Chapter 2 it was concluded that there are seven different individual models which have been used to explain the worldwide development of Holocene chenier plains. These were river channel switching, mud shoal migration, climate fluctuations, storm activity, human impacts, sea level oscillations or an internal dynamic (Section 2.5.1).

The last of these models (the internal dynamic) was previously put forward to explain the development of the Princess Charlotte Bay chenier plain. Although the concept of an internal dynamic provides valuable insights, and the study it was derived from has greatly advanced Australian chenier plain geomorphology, in light of the large amount of new palaeoenvironmental knowledge that now exists it is appropriate to firstly test the applicability of other six models before rethinking and reworking the concept of an internal dynamic. This systematic approach was taken to increase the likelihood of identifying factors that were not incorporated into Chappell and Grindrod‟s (1984) work, particularly since the data necessary to confirm or discount many of the other models was not available to Chappell and Grindrod (1984). Considering the models individually may therefore reveal influencing factors that were not incorporated into Chappell and Grindrod‟s (1984) work.

It may be obvious at the outset that some of these models have an extremely low likelihood of playing a significant role in the development of the Princess Charlotte Bay chenier plain. For the sake of prudence, however, all models were considered. The applicability of the other models of chenier plain development was considered by testing hypotheses derived from them. These hypotheses are stated below:

Hypothesis (1): The Holocene development of the Princess Charlotte Bay chenier plain has been driven by **river channel switching**.

Hypothesis (2): The Holocene development of the Princess Charlotte Bay chenier plain has been driven by **mud shoal migration**.

Hypothesis (3): The Holocene development of the Princess Charlotte Bay chenier plain has been driven by **climate fluctuations**.

Hypothesis (4): The Holocene development of the Princess Charlotte Bay chenier plain has been driven by **tropical cyclone (i.e. storm) activity**.

Hypothesis (5): The Holocene development of the Princess Charlotte Bay chenier plain has been driven by **human activity**.

Hypothesis (6): The Holocene development of the Princess Charlotte Bay chenier plain has been driven by **sea level oscillations**.

These hypotheses were tested by setting up scenarios of what would be expected from the topographic, stratigraphic, temporal, and other data that were collected from the Princess Charlotte Bay chenier plain, if the hypotheses were to be accepted. The scenarios are described in the following sections.

3.2.1 Hypothesis (1): River channel switching

In settings where this model is applicable, chenier ridges form on sections of the coastline that have experienced a dramatic decrease in sediment supply due to a major shift in the location of a river mouth (Chapter 2). The existence of abandoned palaeochannels, or occurrence of distinct groups of chenier ridges would be suggestive of this process having occurred in Princess Charlotte Bay.

The existence of abandoned palaeochannels of any of the rivers draining into Princess Charlotte Bay would suggest that the locations of river mouths have shifted. This would in turn imply that sediment distribution across the bay has varied through time and may have influenced the depositional mode at certain locations. The presence of spatially distinct groups of chenier ridges is another characteristic of settings in which the river channel switching model has been accepted. Given the size of the rivers and associated deltas in settings where this model has been applied, channel diversions often mean major shifts in the output location (i.e. the location where progradation occurs). This also leads to major shifts in the location of sections of coastline where chenier ridges are built, hence the distinct clusters. For example, groups of chenier ridges in the Louisiana chenier plain are up to 200 km apart (Penland and Suter, 1989). In the Changjiang (Yangtze) delta there are two groups to the north and south of the main channel separated by over 300 km (Yan *et al.*, 1989) (Chapter 2). With a width of 50 km, such separation is not possible in Princess Charlotte Bay. This does not mean, however, that a „scaled down" version of the model is not theoretically possible. The instantaneous nature of major channel diversions (in the

context of Holocene development) means that in settings where this model is applicable, one would also be unlikely to find long, continuous chenier ridges that span a significant portion of the delta.

It is likely that the existence of abandoned palaeochannels or groups of chenier ridges would be apparent on aerial photos of Princess Charlotte Bay. It was assumed that analysis of these would provide a satisfactory answer as to whether river channel switching has occurred or played a significant role in development of the plain.

3.2.2 Hypothesis (2): Mud shoal migration

The mud shoal migration model has been primarily applied along the Amazon-Orinoco chenier plain in South America. The model involves the modulation of coastal processes by large, shore face-connected mudbanks that migrate slowly along the shoreline. This results in changing wave properties such that waves are more energetic and erosive in inter-bank zones, and are damped in areas where mud concentration is higher (i.e. the sections of the shoreline protected by the mudbanks, where progradation occurs) (Section 2.5.1.2).

The mudbanks themselves are large enough to be represented in bathymetric charts (e.g. Augustinus, 1980), and are therefore likely to be apparent in aerial photography of the shoreline. Furthermore, the effect they have on the wave action means that longshore variations in wave activity would also be apparent in aerial photography. If this model was to be applied in Princess Charlotte Bay, one would expect similar variations in both

longshore sediment concentrations and wave activity to be apparent in aerial photography of the shoreline.

3.2.3 Hypothesis (3): Climate fluctuations

The reason climate fluctuations are a plausible driver of chenier plain development is that climate, and specifically rainfall, plays such a large role in determining the balance of fluvial and marine forces (Section 2.5.1.3). Simply, more rainfall means higher fluvial discharge, which in turn encourages mudflat progradation. The key requirement for linking chenier plain development to climate fluctuations is therefore to find that chenier ridges tend to be formed during arid phases, and that progradation occurs during pluvial phases (e.g. Lees and Clements, 1987). Several reconstructions of Holocene climate trends were described in Section 2.9.2, including general Holocene climates, the El Nino Southern Oscillation, the Medieval Warm Period and Little Ice Ages, and sea surface temperatures. These provide many avenues by which to test the climate model in Princess Charlotte Bay.

3.2.4 Hypothesis (4): Tropical cyclone activity

The hypothesis that Holocene development of the chenier plain has been driven by tropical cyclone activity can be broken down into two separate sub-hypotheses. The first is that „chenier ridges are storm ridges" (i.e. are created during one event). The second is that

„chenier ridges develop during *phases* of cyclones" (i.e. increased storminess leads to a period of increased coastal reworking) (Section 2.5.1.4).

With the problems regarding circular logic derived from the use of marine deposits and chenier ridges as records in mind , both of these hypotheses can be tested by comparing the frequency and timing of chenier ridge building to the reconstructed frequency and intensity of Holocene tropical cyclone activity., Regarding frequency, it has been suggested that the return interval of individual major tropical cyclones (i.e. between 2 and 5 on the Saffir-Simpson scale; Section 2.9.3) has been between 80 and 300 years (Chappell *et al.*, 1983; Chivas *et al.*, 1986; Hayne and Chappell, 2001; Nott and Hayne, 2001). Nott and Hayne (2001) hypothesised that super cyclones occurred every 200–300 years along all parts of the Great Barrier Reef (Nott and Hayne, 2001). Regarding intensity, Forsyth (2010) identified a phase of high intensity events (with ridges being deposited by category 4–5 tropical cyclones) between 5010 and 3380 yr BP, a phase of lower intensity events (category 1–4) between 3380 and 1550 yr BP, and a return to higher intensity events (category 4–5) between 1550 and 130 yr BP. For the tropical cyclone hypothesis to be true we would therefore expect to see similar frequency of ridge building events, or ridge building events being more pronounced during periods of elevated intensity.

It should be highlighted that the tropical cyclone hypothesis model states that cyclone activity has been primarily responsible for driving development of the chenier plain (i.e. for causing the switches between mudflat progradation and coarse sediment deposition). Should this hypothesis be dismissed, it would not in any way preclude the possibility that tropical cyclones may have played a lesser role in assisting the accretion of chenier ridges.

Therefore, while the depositional chronology of the chenier ridge series will allow the hypothesis to be tested, physical parameters such as grain size, the degree of sorting of the coarse chenier ridge sediment, or the height of ridges, may allow the other possible roles of tropical cyclones (e.g. as an accretionary force, which lies outside the strict scope of the hypothesis) to be elucidated.

3.2.5 Hypothesis (5): Human activity

Although it seems unlikely that Aboriginal land use patterns could have caused a reduction in sediment supply on a scale necessary to allow chenier ridge building at the shoreline, it is possible that catchment destabilisation through intensified burning could have led to increased sedimentation rates (e.g. during periods of most intense burning; Section 2.11). Although this would not directly lead to chenier ridge building, it would be interesting to consider in the context of other forces that may have affected sedimentation rates.

Evidence for Aboriginal impacts may therefore constitute increased sedimentation rates until 4000–3500 yr BP (and possibly peaking then) with a drop in the following 2000 years as burning levels dropped (2.11). To confidently attribute these changes to Aboriginal impacts alone would further require that any changes can be conclusively shown to have occurred out of sync with other environmental changes.

3.2.6 Hypothesis (6): Sea level oscillations

The sea level oscillations model states that chenier ridges would be formed during transgressions, while progradation would occur during regressions (Section 2.5.1.7). Oscillating sea level for a number of locations around Australia and the Pacific Ocean have been developed in recent studies (Section 2.10.2.4), and these can be compared to data collected from Princess Charlotte Bay to assess the possible role of sea level oscillations in that setting.

Although the picture of sea level change is subtly different depending on the location the record is obtained from and the proxy method used, generally speaking the consensus among those who have developed these records is that the changes are eustatic (2.10.2.4). Because of this, the similarity in the actual sea level changes experienced at many locations, and particularly around the Australian coastline where sea level changes due to isostatic factors are thought to be relatively minor, may be more similar than the curves might suggest. This is due to the reality that local factors would act to influence the way in which the (more common) eustatic signal would be recorded at different sites. For this reason, sea level curves from a number of locations around Australia and the Pacific Ocean were used to test the sea level hypothesis.

Two particular criteria would ideally have to be fulfilled to support this model in Princess Charlotte Bay. Firstly, the timings of chenier ridge building must agree with transgressive phases reconstructed in oscillating curves. For example, if this model was correct then on

the basis of the curves shown in Figure 12 (Section 2.10.2.4) we would expect chenier ridge building to possibly occur sometime between 4800 and 2000 yr BP as these were when transgressions would have been occurring. Secondly, since the base levels of chenier ridges can be used as an indicator of past sea levels (Section 2.3), the base levels of Princess Charlotte Bay chenier ridges must be consistent with the sea level heights reached at the peaks of reconstructed transgressive phases. For example, it would be expected that the base levels of any chenier ridges formed around 6800 yr BP would be +1–2 m above that of modern ridges. This interpretation would be conditional on finding no evidence of landward migration in the Princess Charlotte Bay chenier ridges, and that all ridges can be shown to overlie the same muddy unit. This would suggest that the Princess Charlotte Bay chenier ridges could be used to indicate sea levels in the same way as has been done in previous studies (e.g. Rhodes *et al.*, 1980; Chappell *et al.*, 1982).

3.2.7 A possible combination of other models

Although several past studies have put forward a single model to explain the development of chenier plains in a particular setting, others have shown that this is not always the case. For example, Anthony (2006) and McBride *et al.* (2007) have both shown that the models of chenier plain development are not necessarily mutually exclusive and put forward explanations based on the interaction of factors such as sea level oscillations, climate change, and river channel switching (Section 2.5.1.9). Given that the models all represent environmental forces that can all be expected to have at least some impact on shoreline processes, it seems entirely possible that any new model developed will incorporate some

consideration of multiple environmental forces, and will therefore incorporate multiple (individual) models. Indeed, the concept of an internal dynamic is, in itself, a model based on a combination of multiple factors.

Testing the applicability of the other models using the hypotheses described above is not, therefore, intended to indicate an assumption that only one model will be applicable. It is being done in the interests of identifying factors that may have played a role in the development of the Princess Charlotte Bay chenier plain, and to inform a critical assessment of the plausibility of earlier theories regarding chenier ridge formation.

3.3 Reassessing the results of previous Australian chenier plain studies

Compiling and reassessing the results of previous Australian chenier plain studies was undertaken to reveal broader scale relationships or processes that may exist for chenier plains across a large region, and also to give insights as to how the processes that have operated at Princess Charlotte Bay compare and relate to other Australian chenier plains.

The reassessment was essentially an updated version of the statistical analysis of Australian chenier ridge ages across northern Australia performed by Lees and Clements (1987). This involved applying a randomisation procedure to determine „clumps" in chenier ridge ages and therefore periods of non-random chenier ridge building across multiple sites. The analysis undertaken in this study, however, includes far more chenier ridge ages (i.e. *all*

available radiocarbon ages from previous Australian chenier ridge investigations) as well as some improvements to the statistical method. It is described in detail in Section 6.4.

Aside from providing an important source of data for considering possible relationships between Australian chenier plains, the results of this exercise also allowed a critical consideration of the conclusions put forward by Lees and Clements (1987), and more broadly, the conclusions of other Australian chenier plain investigations. For example, the most common model put forward in past Australian chenier plain studies was the climate model (i.e. that chenier ridge building is more common during arid phases where sediment supply to the shoreline is reduced; Section 2.5.1.3). As was explained in Chapter 1, a problem has been a lack of palaeoenvironmental knowledge available for comparison to verify this. The availability of newer palaeoenvironmental records (Part B, Chapter 2), however, means that the suitability of the climate model can now be considered more critically.

3.4 Summary

This chapter described the analysis framework for the primary and secondary aim of this research (i.e. investigating the Holocene development of the Princess Charlotte Bay chenier plain, and considering the formation of chenier plains in Australia and identifying possible relationships). This involved two research aspects.

Firstly, the new data obtained from Princess Charlotte Bay was used to consider the applicability of various models for chenier plain development. This was achieved through testing a number of hypotheses representing different chenier plain development models. Second, the method used to reassess the results of previous Australian chenier plain studies was introduced. This involved a statistical analysis of chenier ridge ages around the Australian coastline.

Chapter Four - Methods

4.1 Introduction

The new data obtained for this investigation were collected in field work conducted across two seasons at Princess Charlotte Bay between August and September of 2007 and 2008. One transect in the southwestern portion of the bay was investigated and is described in Section 4.2. This chapter describes the methods used in acquiring and processing these new data. It has been divided into four sections. The first is a site description of Princess Charlotte Bay and the study site, including details of site location, geology, field seasons, climate, and marine and fluvial characteristics. The second describes the processes of air photo analysis and surveying of the study site transect. The third describes the methods involved in the stratigraphic analysis of the chenier plain, and specifically how distinctions were made between stratigraphic units. Finally, the various dating techniques employed in this study are described.

4.2 Site description and regional setting

4.2.1 Location, geology and regional setting

Princess Charlotte Bay is located on the eastern coast of Cape York Peninsula, Queensland, approximately 220 km NNW of Cairns at 14° S, 144° E (Figures 13 and 14). Opening to the north, it is approximately 50 km wide, making it one of the largest bays on the Australian coastline. To the east of Princess Charlotte Bay, between Bathurst Head and Cape Melville, are the highlands of the Bathurst and Melville Range. These highlands reach to over 1400 m in height and comprise a dissected plateau of horizontally bedded Mesozoic sandstone that rests on a granite basement. The upper Normanby River catchment, which drains to the bay, is dominated by these sandstones and conglomerates (Bryce *et al.*, 1998). Outliers of this plateau can be found on the Flinders Group of islands to the north of Bathurst Head. These islands rise from steep sandstone cliffs on all sides and vegetation is sparse due to poor soil development (Frankel, 1974; Grindrod, 1985). To the west are the central uplands which lie roughly in the centre of the Cape York Peninsula. This feature represents the northernmost extent of the Great Dividing Range and is composed mainly of weathered Proterozoic metamorphic materials and intrusive granites (Grindrod, 1983).

The area to the south and southwest of Princess Charlotte Bay is the Laura Plain. It consists of a gently rising plain which occupies a structural basin between the eastern ranges and the central uplands to the west. It contains a mixture of weathered terrestrial deposits, the most extensive of which comprise early to late Pleistocene alluvium and

colluvium (Galloway, 1970). This plain extends to the main Cape York Dividing Range and is generally less than 100 m above sea level (Grindrod, 1983). It contains open forest, swampland and isolated riverine jungle and tropical rain forest (Frankel, 1974), and is dissected by a complex drainage system which feeds northward to Princess Charlotte Bay. Closer to the shoreline are the coastal plains comprising Quaternary alluvium, marine and dune deposits. The topography of the Princess Charlotte Bay region and some of the features mentioned in this section are shown in Figure 13.

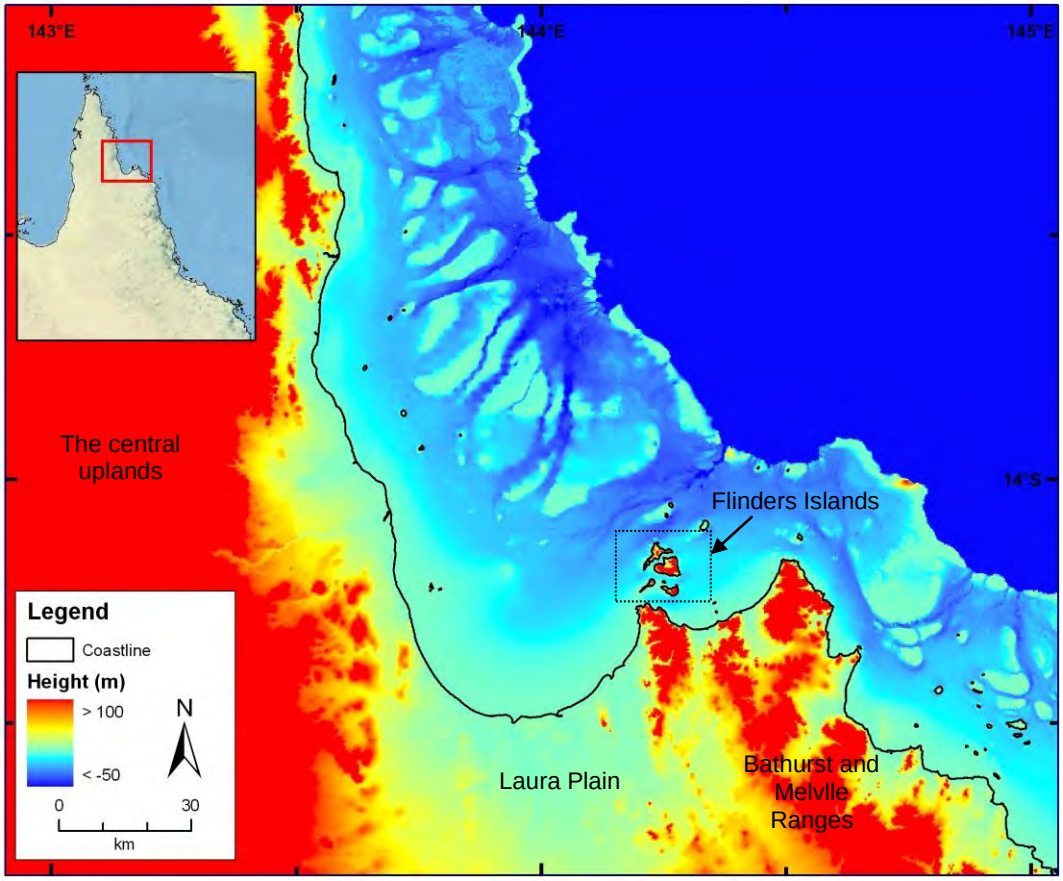

Figure 13. Topography and bathymetry of the Princess Charlotte Bay region (Geoscience Australia, 2009).

4.2.2 Climate

The climate of the Princess Charlotte Bay region can be considered subhumid to humid tropical and has a very marked seasonal regime. The major atmospheric systems that influence regional climate are the subtropical high pressure belt and the intertropical convergence zone, and it is the seasonal latitudinal migration of these components that largely dictate weather patterns (Grindrod, 1983).

During the May to October dry season the anticyclonic weather systems associated with the subtropical high pressure belt provide for a stable climate with relatively cool temperatures and predominant southeasterly winds. Mean daily temperatures range between 16 and 27° during this time (Gunn, 1970). Precipitation is generally confined to light localised occurrences east of the eastern highlands. The average number of rainy days in the dry season is seven and rainfall in this period generally does not exceed 50 mm.

The southward displacement of the intertropical convergence zone during the November to April wet season brings much warmer and wetter conditions due to the predominance of unstable equatorial and tropical maritime air and northeasterly winds. Of the 1000–1200 mm annual rainfall, the bulk (~95%) falls in the wet season. Tropical cyclones are also common during the wet season and are another (albeit erratic) source of precipitation. Tropical cyclones in the region generally originate in the Coral Sea and move in a westerly or southwesterly direction to deliver torrential rain and high seas to coastal northern Australia (Gunn, 1970). Local convectional thunderstorms and tropical cyclones originating

in the Coral Sea often occur at the beginning and end of the wet season respectively. Mean daily temperatures are 22 to 35° C in the wet season (Gunn, 1970).

The primary local climatic control is the eastern highlands (i.er. Bathurst and Melville ranges). These intercept the prevailing, moisture-laden southeasterly winds from the Coral Sea such that precipitation is relatively much higher in the eastern coastal areas and drops considerably to the west.

4.2.3 Fluvial characteristics

The rivers that drain into Princess Charlotte Bay originate in the eastern highlands to the east, the uplands to the west and at the southern extent of the vast plain to the south. They form a catchment that extends more than 150 km inland and contain numerous channels which coalesce into the four that empty into the bay itself. From west to east these are the North Kennedy, Bizant, Normanby and Marrett Rivers (Figure 14). All are generally very sinuous. Two of these (the Bizant and Marrett Rivers) do not extend more than 10–15 km inland and past authors have only considered the fluvial output of the Normanby or Normanby-North Kennedy Catchments as being significant (e.g. Frankel, 1974; Bryce *et al.*, 1998; Ridd *et al.*, 1998).

All rivers in the area display a marked seasonality in their flow regime and generally cease to be active by the end of the dry season (which lasts from approximately May to October), although a number of perennial waterholes strung along the major channels are a feature of

the area (Galloway, 1970). Even though dry season flow is relatively minor, the rivers still display a relatively high discharge compared to other Australian rivers. For example, the Normanby River, with a catchment is approximately 13 600 km^2 in area, has an annual fluid discharge of approximately $6x10^6$ ML. The annual sediment discharge of the Normanby River is approximately 7530 m^3 (Galloway, 1970; Sahl and Marsden, 1987; Bryce *et al.*, 1998).

4.2.4 Marine characteristics

Water depths in the bay are regular with a gradual offshore slope, and irregularities are only found in the vicinity of offshore reefs (Frankel, 1974). The gradient of the lower intertidal and shallow subtidal muds in the eastern region of the bay is approximately 1:1000 (Chappell and Grindrod, 1984) and the offshore gradient approximately the same (Frankel, 1974). The offshore gradient in the western region of the bay is approximately 1:1500 (Frankel, 1974). The water depth decreases gradually and uniformly to approximately 11 m around 40 km north of the mouth of the North Kennedy River and 25–30 m in the lee of offshore reefs (Frankel, 1974). Figure 13 (Section 4.2.1) shows the gradual offshore slope and lack of bathymetric features from the shoreline to the offshore reefs. The possible exception is what appears to be a broad (10–15 km across) lobe-like feature which protrudes in a north east direction, the apex of which is located approximately 40 km north of the mouth of the North Kennedy River.

As the bay is in the lee of the Great Barrier Reef, a number of shelf and fringing reefs and islands exist offshore, the closest of which is approximately 50 km northward from the mouth of the Normanby River (Figure 14). The northward orientation of Princess Charlotte Bay and the Great Barrier Reef provide much protection from the dominant southeasterly waves, and under normal conditions waves are generally very small. Wave data is shown below in Table 1.

Table 1. Mean and maximum wave height and periods at the mouths of the four rivers draining into Princess Charlotte Bay under normal, non-tropical cyclone conditions. Source: NLWRA, 1998.

River (Figure 14)	Mean wave height (m)	Mean wave period (s)	Max. wave height (m)	Max. Wave period (s)
Marrett	0.152	3.58	1.0	8.4
Normanby	0.152	3.58	1.0	8.4
Bizant	0.381	3.336	1.4	7.9
North Kennedy	0.411	3.312	1.5	7.4

During storm conditions waves are often much larger and more powerful, but will vary greatly depending on the location and strength of the storm. Typical storm surges on the Queensland coast are between 2 and 5 m in height (Queensland Government, 2005).

The Princess Charlotte Bay tidal regime is mesotidal and semidiurnal, with an average spring range of 2.2 m. There is a strong diurnal inequality in tidal elevation, which is most apparent during spring tides (Bryce *et al.*, 1998). Mean maximum spring tides for the area range from 0 to 3.0 m above datum, with mean sea level close to 1.5 m above datum (Grindrod, 1985). Longshore currents in Princess Charlotte Bay are only present on the western side on the bay and travel eastwards and northwards (Figure 14). Frankel (1974) noted that the flood tide surface currents tended to approach the shoreline from a

northeasterly direction while the ebb tide surface currents tend to move in a north westerly direction. Based on this, and the extremely small wave heights, the longshore currents appear to be tide-driven.

4.2.5 The chenier plain and study site

The Princess Charlotte Bay plain has developed due to the Quaternary deposition of fluvially-transported terrigenous sediment, and contains occasional chenier ridges composed of shell and sand, some of which are located a significant distance inland. A mangrove fringe is also found at some locations on the shoreline. The chenier plain is widely flooded in the wet season. This may be further aggravated by artificially high sea levels associated with high spring tides or storm surges which result in extensive backing up of fresh water in the fluvial systems (Grindrod, 1983). Only the highest spring tides inundate the plain in the dry season. During this time dessicated algal mats often form on the exposed supratidal mudflats and deflation during strong winds is common.

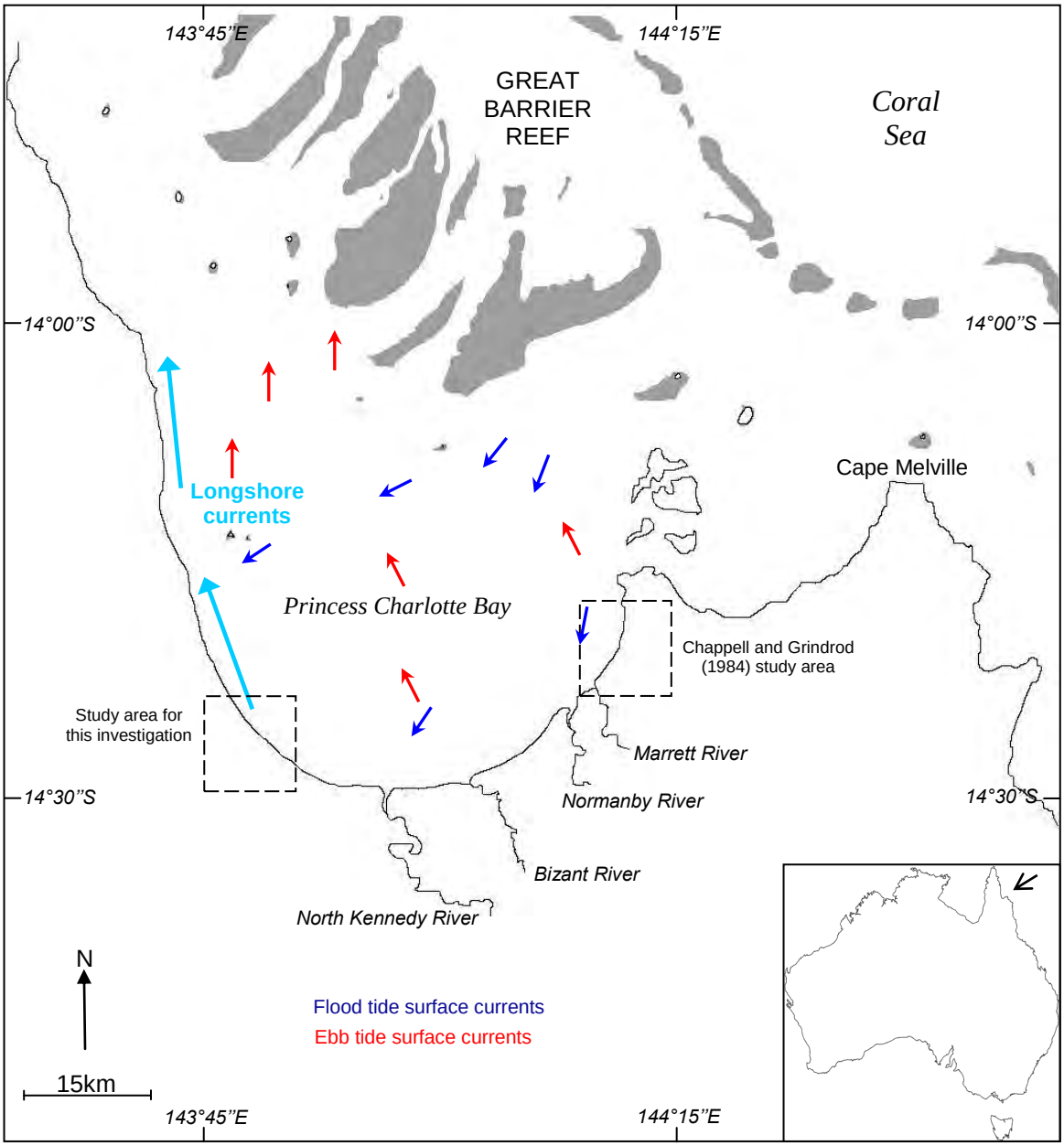

Figure 14. Location of Princess Charlotte Bay within Australia and the region. The locations of the current study site, and that of Grindrod (1983) and Chappell and Grindrod (1984). The location and direction of longshore currents and tidal surface currents are taken from Frankel (1974).

The general position of the study site within Princess Charlotte Bay was the major difference between this study and that conducted by Chappell and Grindrod (1984). The study site for this investigation was located in the southwestern corner of the Bay, to the west of the mouth of the Normanby River (Figure 14). As was briefly mentioned in Chapter

1, actually finding an appropriate and accessible transect in this area proved exceedingly difficult. Thick vegetation and lack of roads meant that there were very few routes that went all the way to the coastline, and the many crocodiles in the area made crossing rivers dangerous. A safe route was eventually found to a location in which one 2.6 km transect running perpendicular to the modern coastline and crossing a number of chenier ridges could be examined (Figure 14). The sequence in this area appeared to be most representative of the chenier plain sequence in this portion of the bay. Its location is shown in detail in Chapter 5.

4.3 Air photo and imagery analysis

Air photo and imagery analysis was carried out to identify locations of geomorphic features in the Princess Charlotte Bay region and to check for the presence of abandoned river channels or shallow, shoreface-attached mud shoals. The photos were obtained from the Queensland Department of Natural Resources and Water. They were captured in 1991 and are approximately 1:50 000 in scale. Stereo pairs were analysed using a mirror stereoscope.

CNES:SPOT satellite imagery was also analysed to compliment the analysis of aerial photography. The imagery was sourced from Google Earth (Google, 2009, 2011). Using Google Earth provided an excellent opportunity to examine geomorphic features at different spatial scales and identify relationships that might not be obvious from the fixed scale and poorer resolution afforded by the aerial photos. A number of images were saved (Chapter 5) and used to illustrate aspects of the discussion (Chapter 6).

4.4 Topographic surveys

4.4.1 Surveying methods used in previous studies

Previous studies have surveyed chenier plains using optical surface levelling (e.g. Woodroffe *et al.*, 1993; Mulrennan and Woodroffe, 1998; Woodroffe and Grime, 1999), for example using dumpy levels and Carr staffs (e.g. Fensham, 1993). These techniques have enabled accuracies to within 5 cm to be achieved (e.g. Belperio *et al.*, 2002). Although relatively accurate, the main problem with these manual techniques is that they are very time and labour consuming in areas where vegetation density prevents a clear line of sight, such as in Princess Charlotte Bay. This necessitates more relocations of the survey equipment and greatly increases the potential for operational errors, particularly since at least two people are required for data collection. A more modern surveying method which avoids these problems is to use a differential global positioning system (DGPS). This was the preferred method for most of the topographic surveys conducted in this investigation. A brief summary of the DGPS technique and background is given in the following section.

4.4.2 DGPS and RTK DGPS

The basic concept behind GPS positioning is that if the distance from a receiver to three or more GPS satellites is known, along with the satellite locations, then the location of the point can be determined by resection (El-Rabbany, 2002). DGPS is based on the use of two (or more) receivers, where one (stationary) reference or base receiver is located at a known point and the reference of the remote receiver is to be determined. The known position of the reference receiver is used to calculate corrections to the GPS derived position. These corrections are then relayed to the roving receiver and allow the computation of its position (Langley, 1998). The precision of differential techniques arises because the measurements of two (or more) receivers simultaneously tracking a particular satellite generally contain the same errors and biases, and this similarity is increased the closer together the receivers are located. By taking the difference between the measurements of the two receivers (hence „differential" positioning), the errors can be largely removed (Langley, 1993; El-Rabbany, 2002).

There are several types of differential positioning techniques (e.g. El-Rabbany, 2002). One of the most advanced currently available is called „real time kinematic" (RTK) GPS and is capable of determining coordinates with millimetre to centimetre level accuracy (both vertically and horizontally) in real time (i.e. without post-processing of data) (Langley, 1998), depending on the degree of signal obstruction. RTK GPS is capable of surveying a large number of unknown points within 10–15 km vicinity (El-Rabbany, 2002).

4.4.3 Surveying the Princess Charlotte Bay transect

The ease of roaming surveying able to be conducted by one person, and real time acquisition of data with millimetre-level accuracy made the use of RTK GPS appropriate for the topographic surveys conducted in Princess Charlotte Bay and it was the method used in this investigation.

The unit used to survey the Princess Charlotte Bay transect was a Thales Z-MAX RTK GPS unit. In static/rapid static mode it was capable of achieving accuracies to 1 cm horizontally and 2 cm vertically with recording intervals as little as a few seconds. Data points were logged at regular intervals along flat sections of the transect, and at every noticeable change in surface gradient. The one location on the transect where the DGPS signal was lost was on the modern beach below the level of the base station. Optical survey methods were employed in this area only. Further details of the surveys are provided in Chapter 5.

4.5 Stratigraphic analysis and descriptions

Manual augering was carried out at various intervals along the entire transect to identify the locations of chenier ridges, and enable their depth and sedimentary characteristics to be recorded. As the ridges were the primary focus, auger holes were widely separated in locations where it was obvious that no ridges existed (i.e. in inter-ridge zones) and more

concentrated on the ridges themselves. Chenier basal depth was defined as the depth at which muddy sediment appeared in the auger head.

This process also afforded the opportunity to examine and note differences in the physical properties of coarse chenier ridge sediment. Differences in colour, degree of weathering, soil profile development, shell content, grain size, sorting, compaction, and presence/absence of a carbonate basement were noted in the field to establish avenues for more detailed sediment examination in the laboratory. Dating pits dug into a number of chenier ridges (Section 4.6.1.1) also allowed for closer examination of the profiles suggested by the augering results and sampling of chenier sediment at various depths.

4.6 Dating techniques

Both luminescence and radiocarbon techniques were used to determine a chronology for the Holocene chenier ridge sequence in Princess Charlotte Bay. There were two reasons for this. Firstly, obtaining some ages using radiocarbon techniques allowed comparisons (i.e. between the dating methods, and between the results of this study and those of Chappell and Grindrod, 1984) to be made. Secondly, as mentioned in Section 2.15.2, ages obtained using luminescence techniques are unreliable if chenier ridge sediment is not sufficiently exposed to sunlight at time of chenier building to reset the luminescence signal, and this appeared to have been the case for some ages. This is discussed further in Chapter 5.

The following sections describe the sample collection and laboratory methods for each of the dating techniques employed in this investigation.

4.6.1 Luminescence dating

Thermoluminescence (TL) techniques were initially used to attempt to date the emplacement of chenier ridges. The results of these (discussed in Chapters 5 and 6) suggested that optically-stimulated luminescence (OSL) may be a more appropriate method and it was subsequently also used. The methods used in sample collection and laboratory analysis for both techniques are described in the following sections.

4.6.1.1 Sample collection in the field

Samples for TL and OSL dating were collected by digging pits up to 1.6 m deep into the seven most seaward ridges (i.e. the youngest seven), and the most landward ridge (i.e. the oldest) in the sequence. Samples from each chenier ridge were removed from at least 30 cm above the muddy layer underneath to avoid variations in dose rate due to differences in sediment properties (i.e. so that the dose rate calculated would be representative of that found in the homogeneous chenier ridge sands, not that which is in close proximity to a major stratigraphic break). Steel tubes 25 cm long with a diameter of 6 cm were filled with sediment by hammering them into walls of the pit at this depth. Upon removal they were sealed and stored away from light to preserve the environmental moisture content and,

importantly, avoid any exposure to light. In addition to this, grab samples of chenier ridge sediment were removed from within a 30 cm radius of where the dating sample was taken for the purpose of dose rate calculations in the thermoluminescence procedure. These were placed in plastic bags and sealed.

4.6.1.2 Thermoluminescence laboratory procedure

Thermoluminescence dating was carried out by Mr David Price at the University of Wollongong. Samples were dated using a method that combines the additive and regenerative technique, and provides a means to check for a change in TL sensitivity due to the laboratory analytical technique (Pers. comm., Price, 2008). Depending on the grain size of the sediment comprising the sample, the regenerative method only was also used on some samples. The procedure is described in further detail in Nanson *et al.* (1991).

Sample preparation and equivalent dose estimation

Samples were opened under low intensity yellow-orange light to prevent signal resetting. Material used for estimation of D_e was removed from the centre of the metal tubes to avoid sediments that may have been exposed to light during field sampling. Samples were first sieved to separate the 90–150 or 90–180 μm grain size fraction, chemically cleansed in dilute hydrochloric acid, etched in 40% w/w hydrofluoric acid and subjected to heavy liquid separation. The quartz in each sample was then divided into two parts, one of which was bleached under a Philips MLU 300 W laboratory UV sunlamp for 24 hours. This

effectively removed all of the previously acquired TL leaving only the residual signal. Aliquots of the bleached quartz were then deposited onto a series of aluminium discs and a number of these incrementally irradiated using a calibrated 90Sr plaque source to replicate the TL naturally accumulated in the original sample (the regenerative technique). Aliquots of the natural (unbleached) quartz were also deposited on discs and all samples heated to 500° C at a controlled rate in a high purity nitrogen atmosphere. The light emitted (TL) was recorded using an EMI 9635QB photomultiplier tube fitted with a Chance Pilkington heat filter and a Corning 7-57 blue transmitting filter. The TL output was quantified in the form of glowcurves, which relate it to temperature. The overall TL output from the bleached and irradiated discs enabled a regenerative growth curve to be constructed, which related it to radiation dose. The average TL from the natural (unbleached) samples was then used to estimate D_e (e.g. Figure 15).

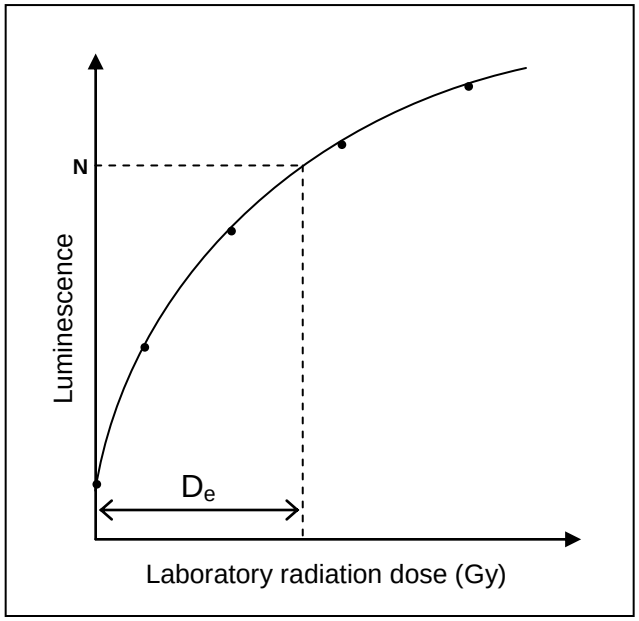

Figure 15. The regenerative method for estimating De. The natural TL of the sample (N) is compared to the TL signal from aliquots that have been bleached then progressively irradiated (points in curve) (Walker, 2005).

Given sufficient quartz grains of the desired size, a further 6 aliquots of natural quartz were deposited on discs and incrementally irradiated (additive method) to provide a check for change in sensitivity. If there had been no sensitivity change these TL values would be expected to lie on the regenerated growth curve constructed from the bleached and irradiated aliquots.

Dose rate determination

Annual dose rates were measured using specific activity values for uranium and thorium using thick source alpha counting and assuming secular equilibrium. Samples taken from material collected in grab samples was first crushed to an extremely fine grain size and placed in immediate contact with a 42 mm scintillation screen. This screen was sealed in an alpha counting cell, which in turn was positioned on a photomultiplier tube assembly. Because certain daughters within the uranium and thorium decay chains are gaseous, a three week waiting period was necessary before introducing the cell into the counter to allow the decay chains to be re-established. The amount of potassium present in the sample was determined by means of atomic emission spectroscopy. Sample rubidium levels were assumed to be 50 ppm in each case and the cosmic ray contribution 150 μGy/yr. Compared to the radiation flux derived from the major contributors (i.e. uranium, thorium and potassium) these only played a minor contributing role (Pers. comm., Price, 2007).

4.6.1.3 Optically-Stimulated Luminescence laboratory procedure

Optically-stimulated luminescence dating was carried out by Dr Kathryn Fitzsimmons at the Australian National University in Canberra (samples K2050 to K2053; „ANU samples" below) and Dr Matt Cupper at the University of Melbourne (samples MELA0311A to MELA0311I; „UM samples" below).

Sample preparation and equivalent dose (D_e) estimation

Samples were opened and processed under low intensity red light or yellow-orange sodium vapour lamps to prevent signal resetting. Material used for estimation of D_e was removed from the centre of the metal tubes to avoid sediment that may have been exposed to light during field sampling. Samples underwent slightly different treatments depending on the laboratory; both techniques isolated pure quartz from the natural sediment. ANU samples were first treated with hydrochloric acid to remove carbonates and hydrogen peroxide to remove organic material. The 180–212 μm size fraction was separated by sieving. Minerals more dense than quartz were then removed by density separation using sodium polytungstate at 2.68 g/cm^3. The remaining material was then etched in 40% w/w hydrofluoric acid for 100 minutes (ANU samples) or 40 minutes (UM samples) to remove feldspars and the outer rinds of quartz which had received alpha radiation.

Equivalent dose measurements were undertaken using automated Risø TL-DA-12 and TL-DA-15 readers with photomultiplier tubes fitted with Hoya U-340 filters. Irradiation was

undertaken using calibrated $^{90}Sr/^{90}Y$ beta-sources. Light stimulation was provided by clusters of blue light-emitting diodes (ANU samples) or green light from a solid-state laser (UM samples) (Bøtter-Jensen *et al.*, 2000).

D_e was estimated using the single-aliquot regenerative-dose (SAR) protocol (e.g. Murray and Roberts, 1998; Murray and Wintle, 2000; 2003) for all samples. The SAR protocol measured the natural OSL signal and the signals arising from laboratory-administered radiation doses, and included sample preheating to remove thermally unstable signals, and built-in checks for thermal transfer of charge and signal recycling. It also corrected for sensitivity changes in the quartz. Samples were exposed to IR stimulation prior to OSL stimulation and measurement in order to check for feldspars (ANU samples). D_e was calculated using the central age model of Galbraith *et al.* (1999).

Dose rate determination

The beta and gamma components of the radiation dose rate were estimated from measured concentrations of uranium, potassium and thorium concentration. These were analysed using inductively coupled plasma mass spectrometry (U, Th) and inductively coupled plasma optical emission spectrometry (K) at Genalysis Laboratories, Perth (ANU samples) and instrumental neutron activation analysis (U, Th and K) at Becquerel Laboratories, Mississauga, Ontario, Canada (UM samples). Dose rates were determined using the conversion factors of Adamiec and Aitken (1998). Beta dose rate attenuation by moisture was accounted for using the present-day water content. The cosmic ray component of the dose rate was determined using the formulae of Prescott and Hutton (1994).

4.6.2 Radiocarbon dating

The high shell content of the most seaward chenier ridge (ridge 11) in Princess Charlotte Bay meant that luminescence techniques were less appropriate for dating this ridge (discussed in Chapters 5 and 6) and radiocarbon methods were instead used. Two ages were obtained from this ridge using samples of marine mollusc shell taken approximately 1.0 m and 1.3 m depth. One further radiocarbon age was obtained from a depth of 1.0 m in ridge 8 for the sake of providing more data for comparisons between the dating methods and with results of Chappell and Grindrod (1984).

Radiocarbon dating was carried out at the Radiocarbon Dating Laboratory, University of Waikato, New Zealand. Physical pretreatment involved cleaning surfaces and sample washing in an ultrasonic bath. Samples were then tested for aragonite recrystallisation. Chemical pretreatment involved acid washing samples using 2M w/w hydrochloric acid for 100 seconds, rinsing and drying. Conventional ages were determined using standard radiometric dating based on a Libby half-life of 5568 years with correction for isotopic fractionation applied. Calibrated ages were obtained using OxCal version 3.10 using the calibration curve of Hughen *et al.* (2004), which includes a correction for local ocean surface 14C variability (i.e. marine reservoir effect) (Ulm, 2002).

Chapter Five - Results

5.1 Introduction

This Chapter presents the new findings and data obtained from air photo analysis, field investigations and laboratory dating. Aerial photography and imagery of the study site and wider region is presented first. Characteristics of the deposits found in the western region of Princess Charlotte Bay are then described in order of oldest to youngest. This includes consideration of common vegetation assemblages, stratigraphic relationships and, where applicable, chronological data.

5.2 Aerial photography, satellite imagery, and regional observations

5.2.1 Aerial photography and images

Aerial photography and satellite imagery of the southwestern portion of Princess Charlotte Bay are shown in Figures 16 to 23. The aerial photography was acquired on 9 October 1991. The two air photos overlapped (as shown in Figure 18) such that Figure 16 showed the northwestern portion of the area, while Figure 17 showed the area to the southeast.

Figure 16. Aerial photography of Princess Charlotte Bay (1 – northwestern portion). The imagery was captured on 9 October 1991 by the Beach Protection Authority, Queensland. © The State of Queensland (Department of Environment and Resource Management), 2009.

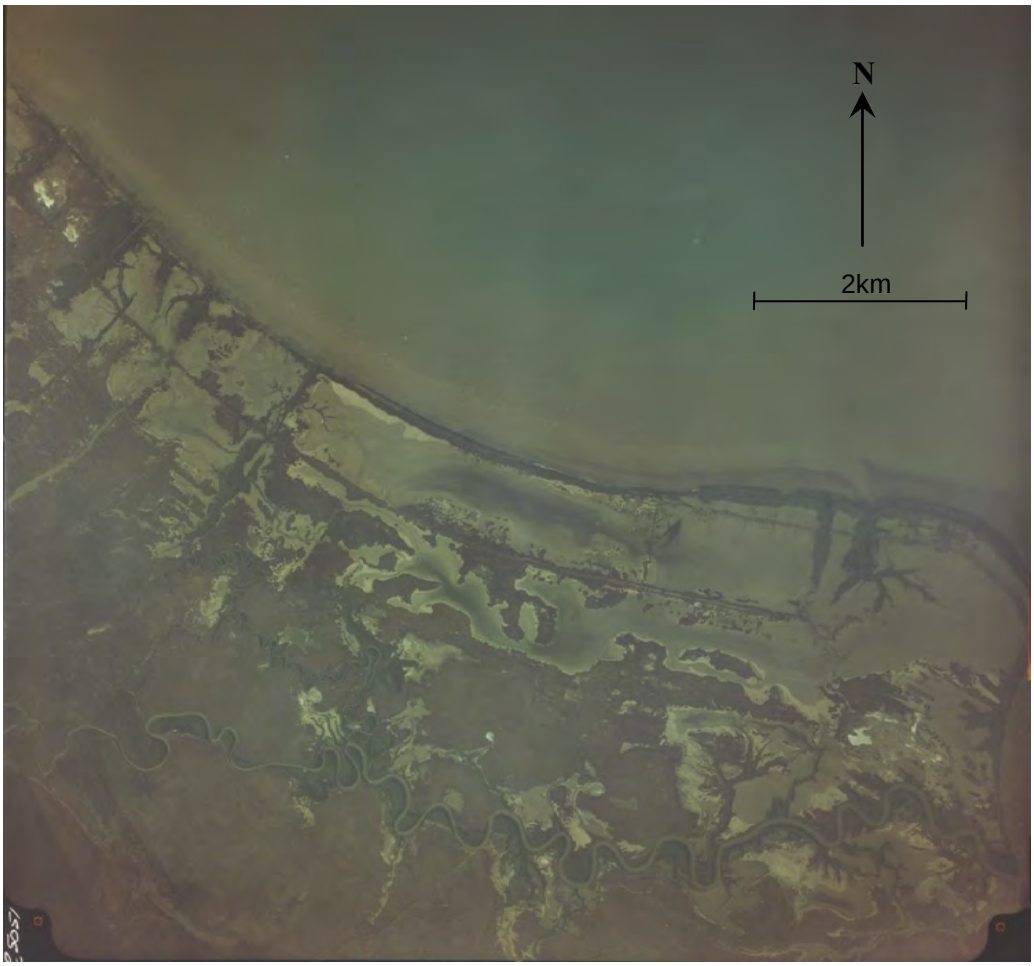

Figure 17. Aerial photography of Princess Charlotte Bay (2 – southeastern portion). The imagery was captured on 9 October 1991 by the Beach Protection Authority, Queensland. © The State of Queensland (Department of Environment and Resource Management), 2009.

CNES:SPOT satellite imagery (Figures 18 – 23) also provided an excellent view of the Princess Charlotte bay study site and showed all the depositional features that are discussed in this investigation. Images were captured at several different scales and therefore show a number of different features. Figure 18 shows the relative locations of the (larger scale) Figures 19 – 23. Figure 19 shows the western portion of the bay and the current study site. Figure 20 shows the central portion of the bay to the west of the North Kennedy River. Figure 21 shows recurved spits that exist to the east of the North Kennedy River. Figure 22 shows similar features further east, near the mouth of the Bizant River. Figure 23 shows the

furthest east region of the bay and includes the area studied by Grindrod (1983) and Chappell and Grindrod (1984). Observations based on the images are discussed in the following section.

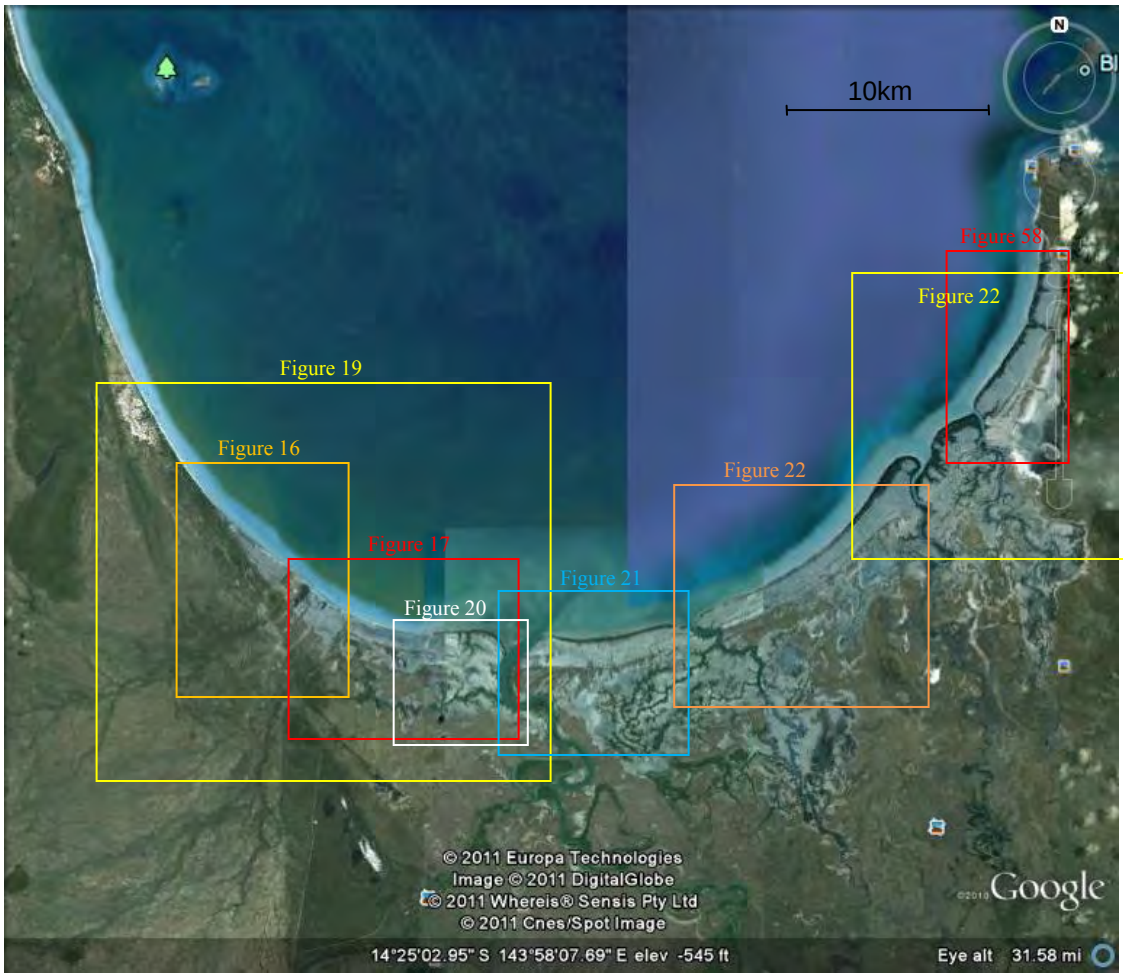

Figure 18. CNES:SPOT satellite imagery of the entire Princess Charlotte Bay shoreline (Google, 2011). The image is a composite of separate images obtained between early 2004 and late 2005. The approximate areas encompassed by Figures 16, 17 and 19 – 23 are indicated.

Figure 19. CNES:SPOT satellite imagery of the southwestern portion of Princess Charlotte Bay (Google, 2009). The imagery was acquired on April 23, 2004. The coastal transect examined has been marked as a small black line as indicated.

Figure 20. CNES:SPOT satellite imagery of the central portion of Princess Charlotte Bay (Google, 2011). The imagery was acquired on April 23, 2004. Examples of discontinuous ridge fragments referred to in text have been marked with arrows.

Figure 21. CNES:SPOT satellite imagery of the area immediately east of the mouth of the North Kennedy River (Google, 2011). The imagery was acquired on June 12, 2005. Note the recurved spits that coalesce into a common ridge front moving eastwards from the river mouth.

Figure 22. CNES:SPOT satellite imagery of the area immediately to the west and further east of the mouth of the Bizant River (Google, 2011). The imagery was acquired on December 7, 2005. Note the recurved spits that appear to coalesce into a common ridge front moving eastwards from the river mouth.

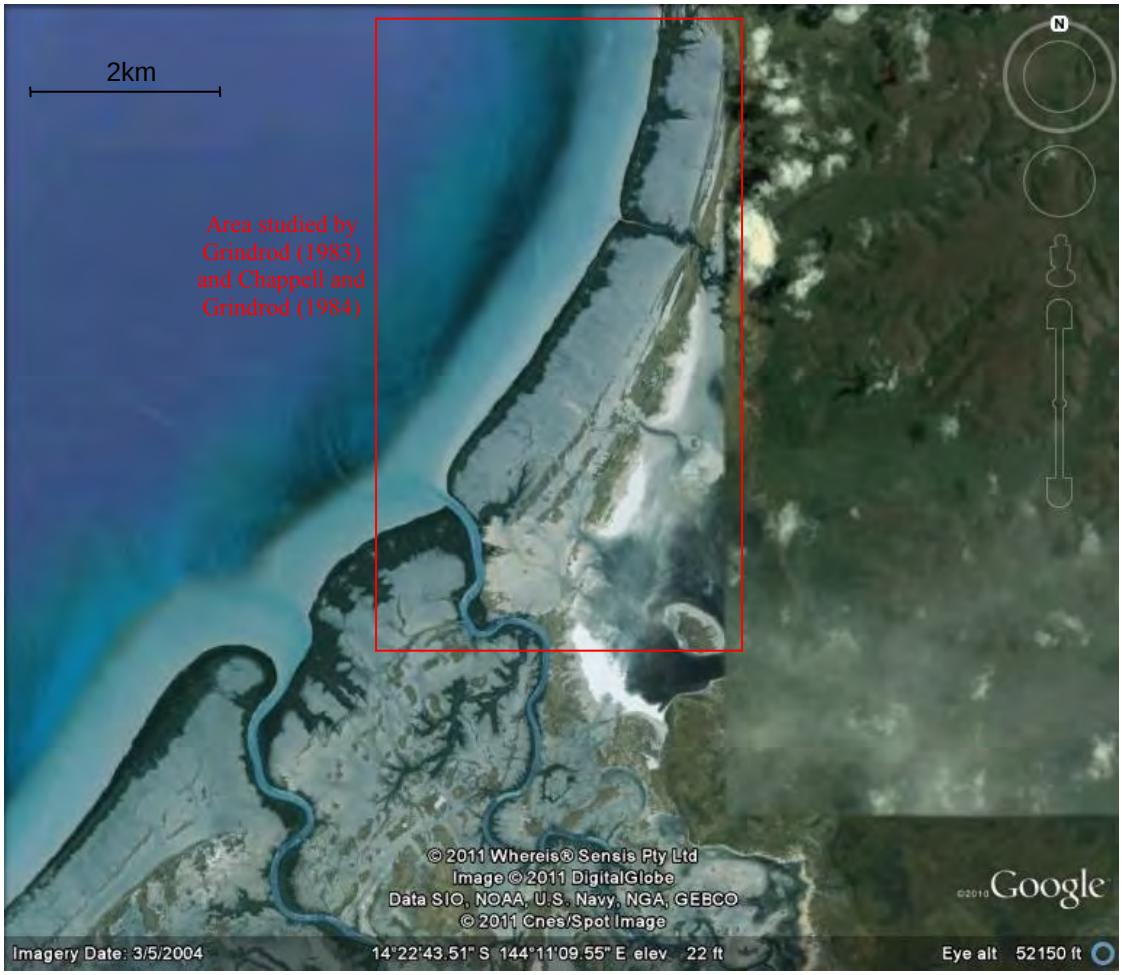

Figure 23. CNES:SPOT satellite imagery of the area immediately to the west and further east of the mouth of the Bizant River (Google, 2011). The western portion of the image was acquired on July 12, 2005 and the eastern portion on 5 March 2004. The region studied by Grindrod (1983) and Chappell and Grindrod (1984) has been indicated. A more detailed representation of their transects and the ridges that occurred there is given in Figure 58, Section 6.2.6.

5.2.2 Regional observations on ridges in Princess Charlotte Bay

A number of observations can be made from analysis of the aerial photographs and images shown above. (Note: In the paragraphs below, the divide between the „western" and „eastern" areas of the plain is arbitrarily defined as an imaginary line which runs directly south from the large mouth of the North Kennedy River in the centre of the Princess Charlotte Bay shoreline).

Chenier ridges exist in both the eastern and western regions of Princess Charlotte Bay and show widening of the spacing between individual ridges as one moves towards the centre of the bay, presumably simply due to the availability of finer sediment. Broadly speaking, ridge preservation appears to be better on the eastern and western flanks of the plain (i.e. the locations in which Grindrod (1983)/Chappell and Grindrod (1984), and the current investigation, respectively, occurred), than in the central area of the plain. The presence of what appear to be discontinuous ridge fragments through the central areas (e.g. Figure 19), and that these ridge fragments appear to share similar longitudinal axes to the better preserved ridges to the east or west, suggests that ridges are likely to have occurred there in the past but have been reworked.

Comparing the eastern and western flanks where obvious ridge morphology exists, the ridges to the west appeared to be better preserved (i.e. are more continuous). Furthermore, the ridge sequence to the west appears to extend further inland. The most landward ridges on the western side of the plain were distinguishable by distinct changes in vegetation occurring 5 to 10 km inland from the shoreline (Figures 16, 17 and 19). These darker areas

of vegetation occurred in discontinuous lines, and appeared to be oriented parallel to the shoreline. Features like these are not obviously identifiable on the eastern side of the plain. Two possible explanations for these observations are (1) that the presence of three rivers on the eastern side compared to no rivers on the western side means that the potential for reworking has been significantly greater on the eastern side; or (2) simply, that the landward extent of the progradational plain is further inland in the western region than in the east (i.e. progradation has been occurring for a longer time in the west than in the east, meaning that older geomorphic features are preserved there). Considering the location of the rivers, and the landward extent of the coastal plain shown in Figure 13, both would seem possible.

One of the most continuous ridge features on the western side of the plain is the first discernable ridge landward of the current shoreline. Apart from two stretches of approximately 2 km and 1.5 km, (Figure 20), this ridge can be traced from the north western corner of the progradational plain to the mouth of the North Kennedy River, a distance of more than 20 km (e.g. Figures 16, 17 and 19 – 23). In the context of the lack of ridge preservation through the central part of the plain, the fact that a relatively seaward (i.e. young) ridge should be so well preserved is interesting. It may indicate either a relatively significant ridge building event which affected more of the shoreline than other ridge building events in the past, or it may simply be that, as a relatively young ridge, rivers have not had as much time to rework it yet. The continuity of this ridge from the mouth of the North Kennedy River eastwards is more difficult to determine. It is clear, however, that a number of discontinuous ridge fragments do exist there, and these may share a common longitudinal axis to the continuous ridge in the west.

Finally, while for the most part the ridges in Princess Charlotte Bay appear to be generally parallel to both the shore and each other (apart from the gradual ridge coalescing that is expected as one moves away from the centre of the plain), there are some exceptions that are worth noting. Two areas where such exceptions are particularly obvious in the imagery include the area adjacent to the east of the mouth of the North Kennedy River (e.g. Figure 21), and an area approximately 3 km east south east of the Bizant River (e.g. Figure 22). Two geomorphic features exist in these areas that are not readily identifiable in other areas of the bay. Firstly, they both contain recurved spits. Such features are commonly found on other chenier plains at river discharge points (e.g. McBride *et al.*, 2007). Secondly, at both locations (although more apparent at the first) the recurved spits appear to coalesce into a common ridge front that is very distinct, is continuous for between 3 and 6 km, and is separated from the shoreline by an area of mudflat/samphire flat between 1 and 2 km in width. The distance from the shoreline of this common ridge front appears similar to that of the obvious continuous ridge present on the western side (described earlier in this section).

5.3 Regional morphostratigraphy landward of the transect

The Princess Charlotte Bay progradational plain has formed from the Quaternary deposition of fluvially-transported terrigenous sediment (Section 4.2.1). Rivers and their associated tributaries have cut channels several metres deep into the sedimentary sequence and reveal that it consists of leached and weathered muds up to 3 m in depth overlying an older surface of highly lithified sandstone. On this sandstone surface were nodules between 2 and 5 cm in diameter (Figure 24). This surface has the appearance of an indurated

palaeosol and appears to extend over the entire area as it was encountered at a number of river crossing locations *en route* to the modern coastline.

Figure 24. Example of basal lithified sandstone surface exposed in river channels in the Princess Charlotte Bay area. The nodules can be seen particularly well in the bottom right hand corner of the photograph. Running horizontally across the centre of the photograph is the overlying leached and weathered mud layer. The location of this photograph is shown in Figure 25.

The muddy sedimentary sequence described above appears to extend over most of the region. At some locations, however, isolated and distinct surficial deposits were also identified based on vegetation changes observed in aerial photographs and satellite imagery (Section 5.2). The locations of these features are shown diagrammatically in Figure 25. One was examined at the location shown in Figure 25.

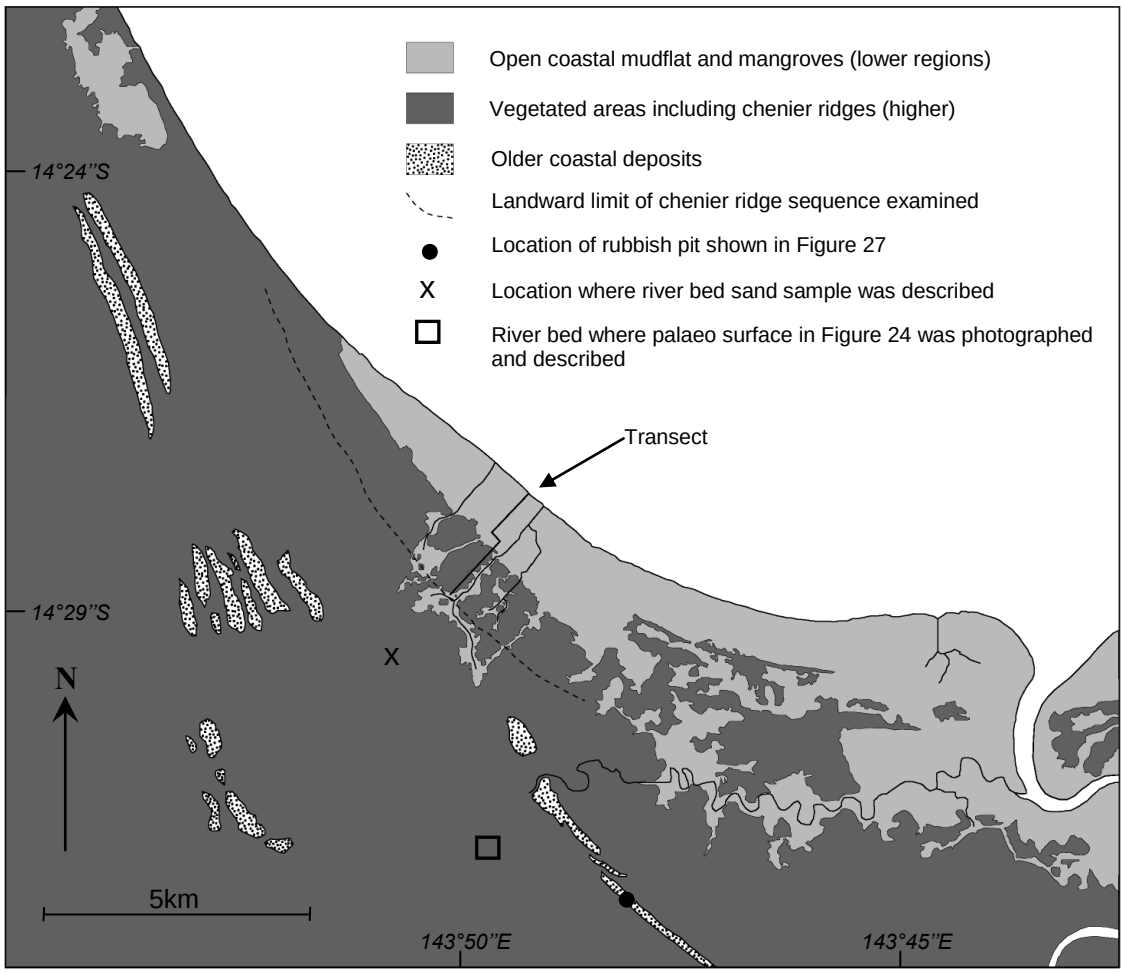

Figure 25. Representation of the Princess Charlotte Bay area showing location of the older coastal deposits and other features referred to in these sections. This figure was traced from the CNES:SPOT imagery shown in Figure 18. Specific coordinates of features are given in Appendix A.

As the imagery had suggested, the vegetation change between that found on the surrounding plain and that found on this deposit was very abrupt. Various grasses and ground cover characteristic of low lying areas were found on the plain, while the deposit itself supported a very dense, mature vine forest where large *Eucalyptus* and *Melaleuca* spp. were common (Figure 26).

Figure 26. Example of distinct vegetation changes which helped to identify the location of very old, shore parallel sandy deposits in Princess Charlotte Bay. The trees growing on the sandy deposit (background) are in stark contrast with the grasses on the surrounding mudflats (foreground).

A rubbish pit dug at this location also enabled the soil profile to be briefly examined. The deposit consisted of strongly weathered, compacted and podsolised quartz sand and gravel with a high proportion of quartz silt. The depth of the deposit was not determined, but it was clear from the exposed section that it was at least 2 m, and contained a very well developed podsol sequence. The most notable feature of this sequence was a mottled deep red (10 R 3/6) B horizon approximately 1 m below the surface. Figure 27 shows part of the soil profile for this ridge.

Figure 27. The stratigraphy of a very old deposit in Princess Charlotte Bay was revealed where a rubbish pit had been dug. The location of this pit is shown in Figure 22.

Although this deposit occupied a much greater surface area than the younger chenier ridges found along the transect, it appeared to be perched on top of the surrounding plain in a similar way. Furthermore, although it is clearly discontinuous, its linear nature suggested that it may be a remnant palaeo-chenier ridge. Based on the vegetation changes observed in the imagery, a number of such deposits are likely to exist in the more landward areas of the progradational plain.

5.4 Vegetation, stratigraphy and chronology of landscape units

Figure 28 shows the location of the coastal transect. Figure 29 shows a stratigraphic cross section of the transect, including the ages obtained from individual chenier ridges. Elevations are given relative to the point at which the base of the modern beach intersects the intertidal mudflat. This point approximates a level 10–20 cm above mean low water spring tide level (MLWS). Justification for this is given below in Section 5.4.5. The „dogleg" shape of the transect was necessary to avoid having to survey through a large mangrove forest found closer to the shoreline (Figures 16, 17 and 19). This was a localised feature occurring along a small tidal inlet at that location and did not affect interpretations regarding Holocene development.

The thick vegetation in the area necessitated the use of two base stations for the DGPS survey (Section 4.4.3). To ensure there were no discrepancies between the two survey legs a significant area of overlap was maintained in which data points were collected relative to both receiver positions. The results of each leg yielded the same topographic changes along the transect, adding confidence to the accuracy of the topographic profile shown in Figure 29.

Moving from the rear of the transect seaward, approximately 11 closely spaced chenier ridge crests were crossed before moving into an open coastal samphire flat containing one more chenier ridge, and finally the coastline and modern beach. The results of vegetation assessments, augering, and profile inspection while digging the dating pits, enabled the chenier ridges along this transect to be divided into two distinct groups based on sediment

characteristics and degree of podsolisation. These were a landward group in which the sediment generally appeared weathered and dull (chenier ridges 1 to 5) and a seaward group in which the sediment generally appeared unweathered (chenier ridges 6 to 11). The possible exception to this general distinction is ridge 10. This is discussed below in Section 5.4.1. Ages were obtained from chenier ridge 1 (i.e. the most landward of the chenier ridges along the transect), 4 – 9, and 11 (Figure 29).

Compared to the other landscape units identified along the transect (i.e. the samphire flat/saltmarsh, and modern beach and intertidal mudflat), the two groups of chenier ridges are relatively similar. That similarity notwithstanding however, the degree of difference in the sediment and vegetation means that they have been considered as separate landscape units in the following sections. The characteristics of these three landscape units and the data collected for each is described in the following sections.

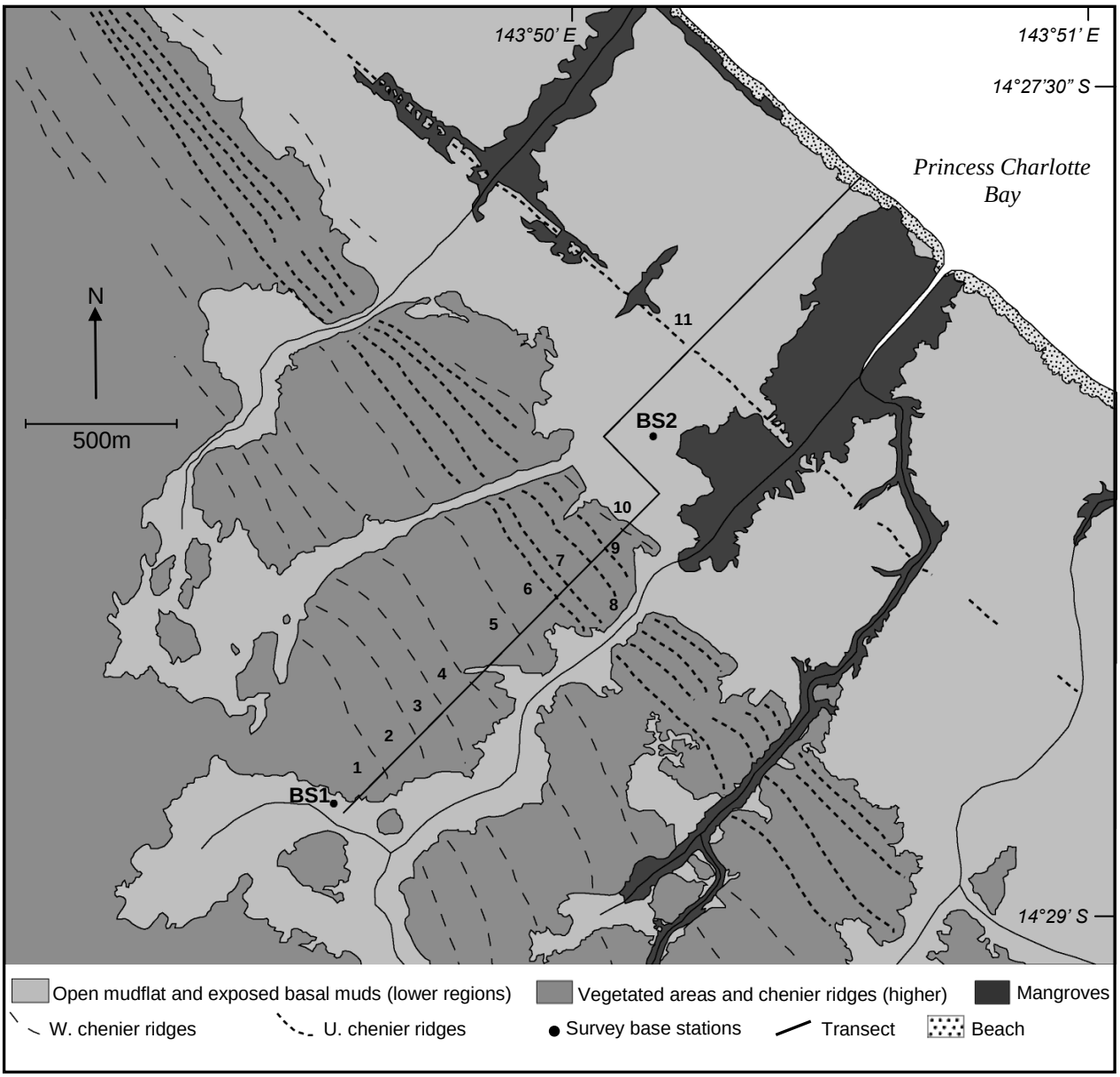

Figure 28. Plan view of the Princess Charlotte Bay study site with chenier ridges labelled. Specific coordinates of the extents of the transect, pivot points and DGPS base stations are given in Appendix A. „W. chenier ridges" refers to the cheniers that appeared to consist of weathered and dull sediment. „U. chenier ridges" refers to the cheniers that appeared to consist of unweathered sediment.

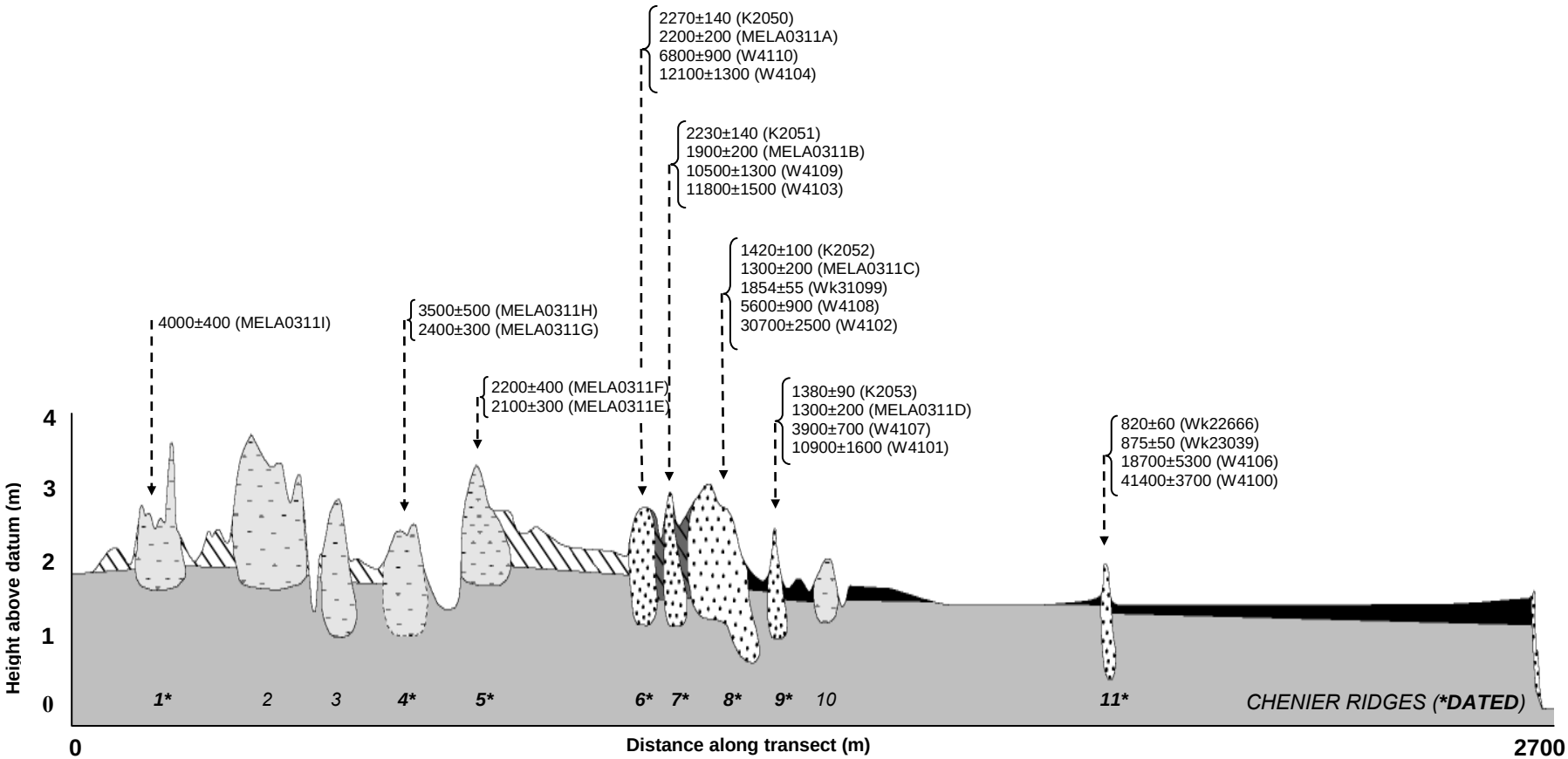

Figure 29. Stratigraphic cross section of the transect. All dating samples were removed from locations along the transect line. Specific coordinates of the locations of dating sample pits are given in Appendix A. Error terms represent 1σ.

5.4.1 Vegetation and general descriptions of chenier ridges 1 to 5 and 10

Chenier ridges 1 to 5 and 10 (Figures 28 and 29) were found to be between 0.5 and 2.5 m in depth. Their crests reached up to 3.7 m above datum while base levels were between 1.0 and 1.7 m above datum. They were between 30 and 150 m wide and were slightly steeper on their landward sides. The ridges nearer the middle of the series (i.e. ridges 2 to 5) appeared to be better preserved and had not been subject to significant reworking. This landward series of ridges differed from the seaward series (ridges 6 to 11) by having less defined vegetation zones, and more notably, very different stratigraphic profiles. There was evidence of considerable bioturbation on all chenier ridges along the transect including scrub-fowl mounds, and disturbance and pits dug by macropods and wild pigs, which were both abundant in the region.

Compared to the vegetation observed on ridges 6 to 11, the vegetation found through the portion of the transect containing ridges 1 to 5 was mostly mixed. Distinct assemblages could only be made out in the topographically highest and lowest areas, and even there no distinct break could be identified, rather a change in the proportions of particular species was noticed (e.g. Figure 30). Ground species included *Typha* spp., *Imperata cylindrica* and various grasses. Where there was more light penetration, grasses such as *Themeda*, *Rottboellia* and *Sorghum* spp., and *Sporobolus virginicus* were more common. Shrubs included *Cleistanthus*, *Breynia* and *Terminalia* spp. between 1 and 4 m in height. The overstorey consisted of mature *Melaleuca leucadendra*, *Acacia holosericea*, *Cochlospermum gillivraei*, *Terminalia* and *Eucalyptus* spp. between 5 and 20 m in height.

Figure 30. Example of mixed vegetation assemblages found on ridges 1 to 5. Here *Melaleuca* (typically distinct at lower elevations on ridges 6 to 11) was mixed with tropical coastal vine forest.

Although most overstorey trees along the transect appeared similar in size with trunk diameters between 5 and 40 cm (depending on the species), the largest and most established trees were all found growing on ridges 1 to 5. This was particularly noticeable towards the rear of the sequence, where *Melaleuca* and *Eucalypt* species with trunk diameters between 40 and 90 cm were common (e.g. Figure 31). Several species of vine were also present in the area including *Maclura cochinchensis*. In terms of vegetation, ridge 10 was different to all other ridges in the „weathered" group (and indeed the unweathered group also) in that the only vegetation on its surface was scattered grasses (Figure 33).

Figure 31. Large *Eucalyptus* growing on ridge 1. Trees of this size were only found growing on ridges 1 to 5.

Stratigraphically, ridges 1 to 5 consisted of deeply weathered and compacted quartz sands and gravels with a high proportion of quartz silt (Figure 32) and well developed podsol sequences. The stratigraphic profile was very similar for each ridge examined. They contained a darker A1 horizon between 0.3 and 0.5 m in depth, which overlay an A2 horizon of bleached pale yellow (2.5 Y 8/4) sands and quartz silt. This horizon became paler down the profile to a depth of 1.0 to 1.5 m, at which point a B horizon of mottled and piped orange sands (7.5 YR 7/8) was found. At the base of each of the chenier ridges an indurated and weakly lithified deep red (10 R 3/6) basement was found. Below this (i.e. underlying each chenier ridge), grey intertidal (basal) muds were found. This indicated that the ridges were likely to have formed *in situ* and had not been subject to significant landward migration (see Section 2.5.2.4).

Figure 32. Weathered sands in ridges 1 to 5 and 10. This photograph shows a pit dug into ridge 5. The two small holes in either side of the facing sides are where samples were removed for luminescence dating

Ridge 10 differed from ridges 1 to 5 as it was only 0.5 m in depth, and more notably, was found seaward of chenier ridges containing unweathered sand (i.e. 6 to 9, and 11). It consisted of weathered and weakly cemented grey quartz sands and gravels, which quickly graded to the deep red basement described above for the other landward ridges. It appeared to be a remnant chenier ridge that had the bulk of its (previously overlying) body reworked, leaving behind only the indurated, weakly lithified base (Figure 33).

In topographically low areas between chenier ridges 1 to 5, finer and darker grey cemented muds and soil were found. These appeared to be an indurated version of the more recent swale infill or supratidal clays found between chenier ridges 6 to 11. This indurated infill was undergoing active erosion between ridges 2 and 3, and also between 4 and 5 (e.g. Figure 34). In these swale areas wet grey basal muds were exposed.

Figure 33. This photograph was taken looking west off the transect on ridge 10. Note how the ridge did not rise very far above the surrounding mudflat, was discontinuous and relatively narrow.

Figure 34. This photograph was taken in the swale between ridges 5 and 6 looking northwest. Note how the indurated mudflat in the background was being eroded and reworked.

Although all chenier ridges in the sequence examined along the transect (i.e. in the wider region of southwestern Princess Charlotte Bay) were subject to minor erosion and reworking as local rivers and tidal channels cut paths through them, the southeastern flanks of ridges 4 and 5a appeared to be eroding at a faster rate. This was presumably occurring due to tidal and/or storm surge penetration from an adjacent inlet that ran parallel to the transect approximately 300 m to the southeast (Figure 28).

5.4.2 Vegetation and general descriptions of chenier ridges 6 to 11

(Note: In the following section, „ridges 6 to 11" refers to ridges 6 to 9, and 11. Ridge 10 is excluded from this discussion, as it was considered in the previous section). Chenier ridges 6 to 11 (Figures 28 and 29) were between 1.3 and 1.8 m in depth. Their crests reached up to 2.9 m above datum while base levels were between 0.4 and 1.4 m above datum. They were between 27 and 122 m in width and were slightly steeper on their landward sides. The vegetation found along the chenier ridges 6 to 11 was similar to that on ridges 1 to 5a, but occurred in more distinct locational assemblages (i.e. compared to the ridges 1 to 5a, the vegetation on the ridge crests and sides of ridges 6 to 11 showed more variation to that found in the swales). As with ridges 1 to 5, evidence of bioturbation was found on all of the seaward chenier ridges.

Tropical coastal vine forest occurred on the chenier ridges themselves (Figure 35). Common shrubs included *Cleistanthus*, *Breynia* and *Terminalia* spp. between 1 and 4 m in height. The overstorey was composed of trees including *Acacia holosericea*,

Cochlospermum gillivraei, *Terminalia* spp. *Eucalyptus* and *Pandanus* spp. between 5 and 15 m in height with trunk diameters between 5 and 40 cm. Several species of vine were also present in this area including *Maclura cochinchensis*. In some areas on the chenier ridges where the vine forest assemblage thinned out and light penetration was greater, grasses such as *Themeda*, *Rottboellia* and *Sorghum* spp., and *Sporobolus virginicus* were common. There were also several large loose sandy areas that were unvegetated. Such areas were not seen on any of ridges 1 to 5. Chenier ridge 11 (i.e. the most seaward ridge) displayed a much less developed version of this assemblage compared to the ridges to landward, with *Osbornia octodonta* immediately adjacent on both the seaward and landward sides and primarily the grasses covering the ridge itself. Some scattered small trees existed at other locations along this ridge, but in far less abundance than the other ridges.

Figure 35. This photograph shows some of the vegetation assemblage on ridge 8. Note the grasses in the foreground and tropical coastal vine forest to the rear.

Figure 36. This photograph was taken at the seaward side of ridge 8 looking southeast (i.e. the swale between ridges 8 and 9). Note the *Avicennia* and *Melaleuca*, common in the topographically lower locations. The bare patch in the background was the sand exposed on the ridge crest.

In the swales between ridges 6 and 7, and 7 and 8, ground cover was sparse but included *Typha* spp. and *Imperata cylindrica* along with occasional tropical coastal vine forest grasses. The overstorey was primarily mature *Melaleuca leucadendra* and occasional *Pandanus*. *Avicennia marina* occurred at lower swale elevations (e.g. in between ridges 8 and 9, and 9 and 10; Figure 36).

Stratigraphically, chenier ridges 6 to 11 were primarily composed of undifferentiated, poorly sorted angular to sub angular quartz sands and gravels (Figure 37). Their profiles contained a shallow O horizon which became thinner in a seaward direction (from 0.9 m in ridge 6 to 0.1 m in ridge 11). Roundness of the quartz grains tended to decrease in a seaward direction, and the grain size through each stratigraphic profile tended to become finer from bottom to top. The grains were mostly „fresh" looking (i.e. appeared to have not been subject to significant weathering) and were very similar to that found on the modern beach (described below). As with ridges 1 to 5 and 10, ridges 6 to 11 all overlay the same grey intertidal (basal) muds. This indicated that they had formed *in situ* and had not been subject to significant landward migration.

Figure 37. Unweathered sands in ridges 6 to 11. This photograph shows the pit dug into ridge 8.

5.4.3 Individual ridge characteristics and dating results

Aside from the dimensions, and where obtained, ages, there was very little difference between chenier ridges 1 to 5 (see description in Section 5.4.1). All ridges tended to be composed of the same deeply weathered and compacted quartz sands and gravels with a high proportion of quartz silt.

Ridge 1 [4000 ± 400 yr BP (MELA0311I)] was 75 m wide, 1.1 m in depth and had a base level around 1.8 m above datum (Figure 29). Ridge 2 [not dated] was 137 m wide, 1.9 m in depth and had a base level around 1.8 m above datum (Figure 29). Ridge 3 [not dated] was 49 m wide, 1.8 m in depth and had a base level around 1.3 m above datum (Figure 29). Ridge 4 [3500 ± 500 yr BP (MELA0311H); 2400 ± 300 yr BP (MELA0311G)] was 79 m wide, 1.6 m in depth and had a base level around 1.3 m above datum (Figure 29). Ridge 5 [2200 ± 400 yr BP (MELA0311F); 2100 ± 300 yr BP (MELA0311E)] was 48 m wide, 1.6 m in depth and had a base level around 1.8 m above datum (Figure 29).

Between the crests of ridges 6 to 8, swale infill was found that contained a mixture of unweathered quartz sands and gravels reworked from nearby crests and a large organic component. In the swales between ridges 8 and 9, and 9 and 10, black supratidal muds were exposed (described below). Ridge 10 [not dated] was 24 m wide, 0.5 m in depth and had a base level around 1.3 m above datum (Figure 29).

Between ridges 5 and 6 there was an abrupt change in the nature of the sediment composing the ridges. In short, from ridge 6 seawards (not including ridge 10) the sand and gravel

composing the ridges appeared much less weathered and did not display the same degree of profile development.

Ridge 6 [2270 ± 140 yr BP (K2050); 2200 ± 200 yr BP (MELA0311A); 6800 ± 900 yr BP (W4110); 12 100 ± 1300 yr BP (W4104)] was 63 m wide, 1.5 m in depth with a base level around 1.3 m above datum. Ridge 7 [2230 ± 140 yr BP (K2051); 1900 ± 200 yr BP (MELA0311B); 10 500 ± 1300 yr BP (W4109); 11 800 ± 1500 yr BP (W4103)] was 30 m wide, 1.7 m in depth with a base level around 1.3 m above datum (Figure 29).

The profiles of ridges 6 and 7 contained an O horizon 0.9 and 0.5 m in depth, respectively, overlying unweathered quartz sands and gravels. Towards the base of the ridges 6 and 7, very leached and weathered quartz (similar to what was found in ridges 1 to 5a) were found interbedded with the unweathered sediment. In ridges 6 and 7 a cemented carbonate basement was also encountered just above the underlying basal muds. This layer was commonly 0.1 m in thickness and was very difficult to dig or auger through. Interestingly, ridges 6 and 7 were the only ridges from the seaward group (6 to 11) to display this characteristic and were also the only ridges from this group that did not contain beds of marine shell at their bases, making them completely devoid of shell material.

Ridge 8 [1420 ± 100 yr BP (K2052); 1300 ± 200 yr BP (MELA0311C); 5600 ± 900 yr BP (W4108); 30 700 ± 2.5 yr BP (W4102); 1854 ± 55 cal yr BP (Wk 31099)] was 122 m wide, 1.8 m in depth and had a base level around 1.4 m above datum (Figure 29). Its profile contained an O horizon 0.5 m in depth overlying unweathered quartz sands and gravels. The ridge body also displayed a downward dipping section extending seaward beneath the

adjacent samphire flat. Although no shell material was found in the main body of this ridge a distinct bed of marine shell was found at the ridge base. This consisted of whole and broken shell fragments of various intertidal molluscs.

Between the crests of ridges 6 to 8, swale infill was found that contained a mixture of unweathered quartz sands and gravels reworked from nearby crests and a large organic component. In the swales between ridges 8 and 9, and 9 and 11, black supratidal muds were exposed (described below).

Ridge 9 [1380 ± 90 yr BP (K2053); 1300 ± 200 yr BP (MELA0311D); 3900 ± 700 yr BP (W4107); 10 900 ± 1600 yr BP (W4101)] was 35 m wide, 1.3 m in depth and had a base level around 1.2 m above datum (Figure 29). Its profile was very similar to that of 8, containing an O horizon 0.4 m in depth overlying unweathered quartz sands and gravels. A distinct bed of marine shell was also found at the ridge base.

Ridge 11 [820 ± 60 cal yr BP (Wk 22666); 875 ± 50 cal yr BP (Wk 23039); 18 700 ± 5300 yr BP (W4106); 41 400 ± 3700 yr BP (W4100)] was 27 m wide, 1.4 m in depth and had a base level around 0.6 m above datum (Figure 29). Its profile contained an O horizon 0.1 m in depth overlying unweathered quartz sands and gravels with a significant marine shell component, and was the only ridge to contain shell through the entire profile. This ridge also displayed a downward dipping section extending seaward beneath the adjacent samphire flat similar to ridge 8 and a bed consisting almost entirely of marine shell at its base.

5.4.4 Samphire flat and saltmarsh

The samphire flat/saltmarsh stretched from the seaward side of ridge 10 to the rear of the modern beach, a distance of approximately 1.4 km (Figure 28). Although it contained one isolated chenier ridge (ridge 11), this was part of the ridge sequence and as such was considered in the previous section. The samphire flat/saltmarsh area contained a distinctly different vegetation assemblage and stratigraphy to other landscape units in the area.

Immediately seaward of ridge 10 were occasional *Melaleuca leucadendra* saplings and *Imperata cylindrica.* Moving further seaward, in stratigraphically higher areas patches of samphire communities occurred. Aside from this, and the small area of chenier ridge vegetation on ridge 11, this zone was composed almost entirely of *Typha* spp. (Figure 38). *Avicennia marina* is occasionally found, particularly near watercourses. Approximately 200 m east of the transect this zone changes abruptly to a *Rhizophora/Ceripos* closed forest around a watercourse and lower lying section of the coastal salt marsh.

Figure 38. Samphire flat/saltmarsh. This photograph was taken looking landward along the transect from just behind the modern beach. The line of trees in the background marked the position of chenier ridge 11.

Very dark (almost black) organically rich supratidal clays occurred in the upper 10–40 cm of the samphire flat/saltmarsh, and became progressively thicker in a seaward direction. Underlying these, grey intertidal clays and muds were found. They are described in the next section.

5.4.5 Modern beach and intertidal mudflats

The final landscape unit identified along the Princess Charlotte Bay transect was the coastline itself, consisting of a beach and foredune fronted by intertidal mudflats. Vegetation found on the top and landward side of the beach included trees of mature

Avicennia marina, *Lumnitzera racemosa*, *Ceriops decandra* and *Thespesia polulnea* (Figure 39).

Figure 39. Vegetation on top of the modern beach. This photograph was taken looking north at the point where the transect reached the shoreline. The intertidal mudflat can be seen in the right background.

The beach sediments were composed of poorly sorted, very angular to angular quartz sands and gravels, and a large component of whole or broken shells of various intertidal molluscs. The quartz component of these sands was almost identical to that sampled from a dry minor stream draining to Princess Charlotte Bay at the location shown in Figure 25. Augering revealed that the beach sands were a coarse unit resting on top of grey intertidal (i.e. basal muds). It is therefore appropriate to consider the modern beach as an incipient chenier ridge.

Seaward of the beach were intertidal muds which, depending on the tidal level, were exposed for a distance before the waterline. During the August/September field work periods when this landscape unit was observed (i.e. during the dry season) the distance was estimated as being between 100 to 200 m. The exact distance could not be determined as the topographic survey had to be ceased at the base of the beach. This was because of the danger involved in venturing out on to the intertidal muds (i.e. the tendency to sink and become stuck, and first hand sightings of crocodiles beyond the waterline). The only vegetation in this zone was occasional juvenile *Avicennia marina* (Figure 40). The top of the intertidal muds (i.e. the interface between these muds and the base of the beach, and the transect datum level) was between 10 and 20 cm above the level of MLWS. Justification for this is provided below.

Firstly, the moist surface indicated that it was clearly subject to marine inundation during the dry season. Secondly, although there was no significant mangrove fringe where the transect met the shoreline at this location, the unit was situated at a lower level than that of mangrove populations nearby (located approximately 100 m to the southeast; Figure 28). Based on these two observations it appeared analogous to the „lower intertidal muds" described by Chappell and Grindrod (1984), which were situated between MLWS and the level of the mangroves. The waterline observed during the dry season field work periods can be assumed to represent MLWS. Based on a gradient of 1:1000 for the intertidal muds (Chappell and Grindrod, 1984), and an estimated maximum distance of 100–200 m to the waterline, the elevation of the interface between the beach and the intertidal muds would be approximately 20 cm above MLWS. The author concedes that this is an approximation only and introduces some uncertainty into comparisons between the elevations quoted for this

transect, and others that are more closely tied to the Australian Height Datum or similar. It was, however, unavoidable.

Figure 40. The intertidal mudflat. This photograph was taken looking northwest from the point at which the transect reached the shoreline.

The intertidal mudflat consisted of wet grey muds (GLEY 6/10B) with primarily clay sized particles (<8 Φ or <39 µm), although quartz sand and gravel, and whole and broken shells of various small intertidal molluscs, were common. These muds appeared to represent „basal muds" for the entire region, as a unit identical in colour and composition, but less saturated, was exposed at the landward extent of the transect in a large open mudflat. Compressed grey basal muds were also encountered at the base of each chenier ridge along the transect, and were exposed in locations where parts of ridge bodies had been removed by tidal or storm surge reworking, such as between ridges 4 and 5 (Figure 28).

5.5 Dating results

5.5.1 Thermoluminescence results

Thermoluminescence dating was undertaken on two samples each from ridges 6, 7, 8, 9, and 11, producing a total of 10 ages. The results of this are shown in Table 2. Because of the coarseness of the chenier ridge sediment, in all cases it was necessary to increase the grain size range used from the normal 90–125 µm to 90–180 µm to obtain enough quartz for analysis. Due to a limited amount of quartz in the grain size required, the combined additive/regenerative technique (which allowed testing for a change in quartz sensitivity due to laboratory methods) was only possible in two samples (W4107 and W4110), the rest being dated using the regenerative technique only.

No change in sensitivity was detected in samples W4107 and W4110. Combined with the fact that a change in sensitivity is rare in Australian quartz (Pers. comm., Price, 2008), this provided sufficient evidence to assume sensitivity change due to laboratory methods were not a source of error in any samples used in this analysis. Little confidence was placed in the TL results from ridge 11 (W4100 and W4106) as the large spread of data (σ = 3700 and 5300 years, respectively) indicated sample heterogeneity (a possible reason for this is given in Chapter 6). All other TL samples displayed lengthy temperature plateaux between 300–500° and TL growth curves having r-square coefficients approaching unity. The TL characteristics of sample W4103 are shown in Figure 41.

Table 2. Thermoluminescence measurements and results. Assumed values for Rb and cosmic ray contribution are given in Section 4.6.2.2 (Pers. comm., Price, 2008).

Chenier ridge	**Sample depth (m)**	**Sample lab code**	**Specific activity (Bq/kg U+Th)**	**K (%)**	**Moisture content (% by weight)**	**Dose rate (µGy/yr)**	**D_e (Gy)**	**TL age (yr BP ± 1σ)**
6	1.1	W4104	23.2 ± 0.6	0.705 ± 0.005	0.7±3	1387 ± 27	16.8 ± 1.7	12 100 ± 1300
		W4110	22.0 ± 0.4	0.760 ± 0.005	0.7±3	1424 ± 27	9.7 ± 1.3	6800 ± 900
7	1.1	W4103	26.9 ± 0.9	0.690 ± 0.005	0.5±3	1448 ± 29	17.1 ± 2.1	11 800 ± 1500
		W4109	22.9 ± 0.4	0.710 ± 0.005	0.3±3	1388 ± 26	14.6 ± 1.8	10 500 ± 1300
8	0.85	W4102	28.8 ± 0.9	0.600 ± 0.005	0.7±3	1386 ± 29	42.5 ± 3.5	30 700 ± 2500
		W4108	27.4 ± 0.5	0.645 ± 0.005	0.6±3	1415 ± 27	7.9 ± 1.2	5600 ± 900
9	0.75	W4101	25.7 ± 0.7	0.640 ± 0.005	1.1±3	1358 ± 28	14.9 ± 2.2	10 900 ± 1600
		W4107	25.7 ± 0.8	0.615 ± 0.005	0.9±3	1342 ± 28	5.3 ± 0.9	3900 ± 700
11	0.7	W4100	28.3 ± 0.9	0.365 ± 0.005	5.2±3	1067 ± 27	44.2 ± 3.8	41 400 ± 3700
		W4106	23.8 ± 0.8	0.450 ± 0.005	6.7 ±3	1050 ± 26	19.6 ± 5.5	18 700 ± 5300

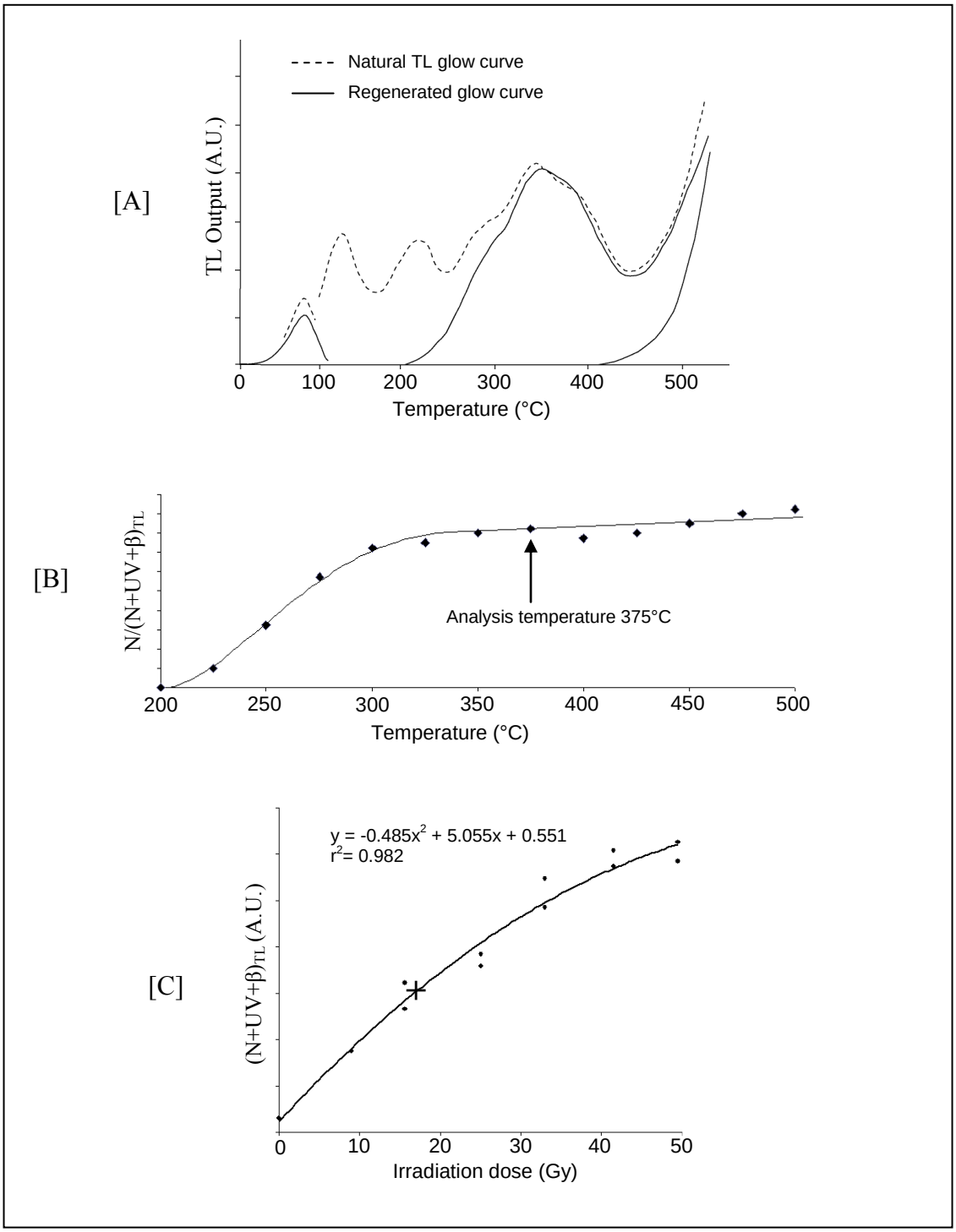

Figure 41. TL characteristics for sample W4103. [A] Comparison of natural and laboratory induced (regenerated) TL glow curve following 24 hour bleaching period beneath an ultraviolet lamp. Laboratory dose was approximately 23.6 Gy; [B] Plotted comparison of natural TL with laboratory induced TL following UV bleach showing plateau region extending between 300 and 500°C; [C] TL growth curve at 375°C (sample W103). The data points are those TL outputs induced by laboratory irradiation following UV bleaching and used to construct the growth curve. The data value (+) represents the mean natural TL value used to determine D_e (Pers. comm., Price, 2008).

The TL characteristic gave confidence that the ages obtained represent actual depositional events in which sufficient exposure occurred to effectively reset any previously acquired TL signal. No corrections were made to account for a residual TL signal, there being no modern surface analogue available for this purpose. This could potentially mean the TL ages represent an overestimation of the timing of the most recent deposition.

Of considerable concern was the fact that there was a difference between the TL ages obtained for samples taken from the same pit, which remains at this time unexplainable and without precedent (Pers. comm., Price, 2009). Ignoring the ages from ridge 11 (41 400 ± 3700, W4100; and 18 700 ± 5300, W4106), which can be dismissed for other reasons, this concern particularly applies to the paired ages from samples W4104 (12 100 ± 1300 yr BP) and W4110 (6800 ± 900 yr BP) taken from ridge 6, W4102 (30 700 ± 2500 yr BP) and W4108 (5600 ± 900 yr BP) obtained from ridge 8, as well as W4101 (10 900 ± 1600 yr BP) and W4107 (3900 ± 700 yr BP) obtained from ridge 9. The ages yielded by these samples all lie well outside the acceptable age uncertainty limits. As was briefly mentioned in Section 4.6.1, erratic results such as these cast great doubt on the reliability of TL to provide age estimates for the chenier ridges, and prompted the use of OSL techniques. Further justification for considering the TL ages as unreliable is provided in Chapter 6.

5.5.2 Optically-Stimulated Luminescence results

Optically-stimulated luminescence dating was undertaken on one sample from chenier ridge 1 and two samples each from ridges 4, 5, 6 to 9. In total this yielded 13 OSL ages. The results of this are shown in Table 3. Sample behaviour was generally reliable and reproducible, yielding bright, rapidly decaying OSL signals and good recycling in response to the duplicate regenerative dose within the single aliquot regenerative dose (SAR) protocol. Thermal transfer of charge appeared to be minimal. No feldspar was detected in any of the samples.

Table 3. Optically-stimulated luminescence measurements and results (Pers. comm., Fitzsimmons and Cupper, 2009, 2010).

Chenier ridge	Sample depth (m)	Sample lab code	U (ppm)	Th (ppm)	K (%)	Cosmic contribution (Gy/kyr)	Dose rate (Gy/kyr)	D_e (Gy)	OSL age (yr BP ± 1σ)
1	1.1	MELA 0311I	0.51 ± 0.04	2.99 ± 0.10	0.39 ± 0.01	0.17 ± 0.02	0.86 ± 0.08	3.4 ± 0.2	4000 ± 400
4	0.6	MELA 0311H	0.90 ± 0.05	3.80 ± 0.12	0.58 ± 0.02	0.19 ± 0.02	1.19 ± 0.11	4.2 ± 0.4	3500 ± 500
	0.6	MELA 0311G	1.02 ± 0.05	4.22 ± 0.13	0.71± 0.02	0.19 ± 0.02	1.36 ± 0.12	3.3 ± 0.3	2400 ± 300
5	0.4	MELA 0311F	1.24 ± 0.07	4.81 ± 0.15	0.81 ± 0.03	0.20 ± 0.02	1.55 ± 0.14	3.4 ± 0.5	2200 ± 400
	0.4	MELA 0311E	0.96 ± 0.04	4.45 ± 0.14	0.73 ± 0.02	0.20 ± 0.02	1.39 ± 0.13	2.9 ± 0.2	2100 ± 300
6	1.1	K2050	0.66 ± 0.03	3.06 ± 0.15	0.66 ± 0.03	0.17 ± 0.02	1.15 ± 0.06	2.6 ± 0.1	2270 ± 140
	1.0	MELA 0311A	0.51 ± 0.04	3.18 ± 0.11	0.66 ± 0.02	0.18 ± 0.02	1.12 ± 0.10	2.5 ± 0.3	2200 ± 200
7	1.1	K2051	0.72 ± 0.04	3.47 ± 0.17	0.60 ± 0.03	0.17 ± 0.02	1.13 ± 0.06	2.5 ± 0.1	2230 ± 140
	1.0	MELA 0311B	0.60 ± 0.04	3.53 ± 0.12	0.63 ± 0.02	0.18 ± 0.02	1.14 ± 0.10	2.1 ± 0.2	1900 ± 200
8	0.9	K2052	0.79 ± 0.04	4.01 ± 0.20	0.69 ± 0.03	0.18 ± 0.03	1.27 ± 0.06	1.8 ± 0.1	1420 ± 100
	0.95	MELA 0311C	0.86 ± 0.05	4.23 ± 0.14	0.86 ± 0.03	0.18 ± 0.02	1.45 ± 0.13	1.9 ± 0.2	1300 ± 200
9	0.8	K2053	0.55 ± 0.03	2.86 ± 0.14	0.59 ± 0.09	0.18 ± 0.03	1.06 ± 0.06	1.5 ± 0.1	1380 ± 90
	0.85	MELA 0311D	0.60 ± 0.04	3.36 ± 0.11	0.64 ± 0.02	0.18 ± 0.02	1.14 ± 0.10	1.5 ± 0.1	1300 ± 200

D_e was calculated based on the dose distributions as illustrated on the radial plots shown in Figures 42 to 54 (Galbraith *et al.*, 1999). The arcuate Y-axis on the plots corresponds to D_e and the left hand Y-axis the number of standard deviations. The X-axis corresponds to relative error and precision. The shaded area represents 2σ from the calculated D_e. All samples yielded Gaussian distributions with no substantial evidence of incomplete bleaching or mixing, which, for example, could be identified by multiple populations or skewed distributions. Confidence in the reliability of the OSL ages was also gained from the fact that where two ages were obtained from a single chenier ridge, the ages overlapped at 1σ for all ridges but one. The ridge where the two ages did not overlap at 1σ was ridge 4 (3500 ± 500 yr BP, MELA 0311H and 2400 ± 300 yr BP, MELA 0311G). In this case the ages still overlapped at 2σ, and the difference between the 1σ limits of the age ranges was only 300 years. In the context of the ages obtained for ridges formed before and after (i.e. 1 and 5), this did not considerably affect interpretations.

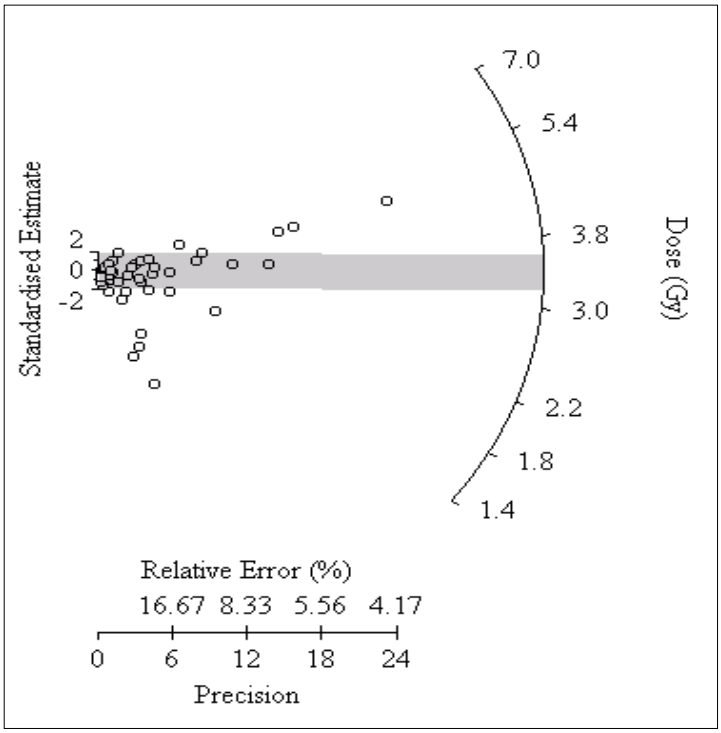

Figure 42. Radial plot of D_e dose distributions for sample MELA 0311I (chenier ridge 1) (Pers. comm., Cupper, 2010).

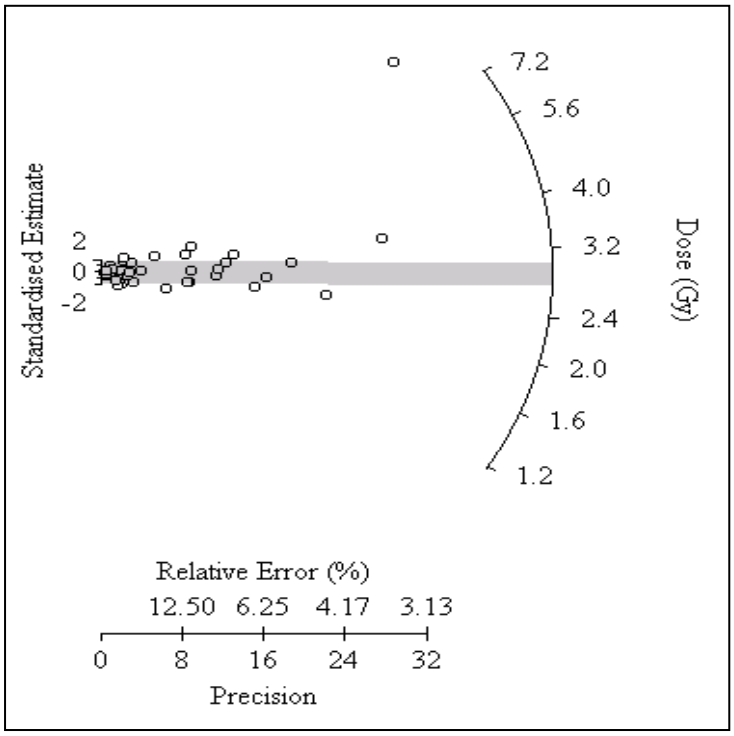

Figure 43. Radial plot of D_e dose distributions for sample MELA 0311H (chenier ridge 4) (Pers. comm., Cupper, 2010).

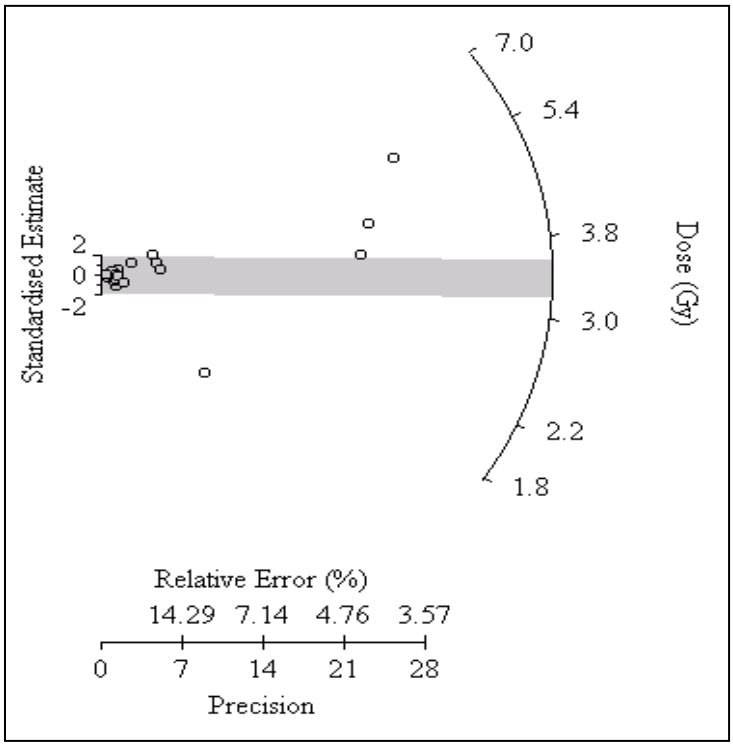

Figure 44. Radial plot of D_e dose distributions for sample MELA 0311G (chenier ridge 4) (Pers. comm., Cupper, 2010).

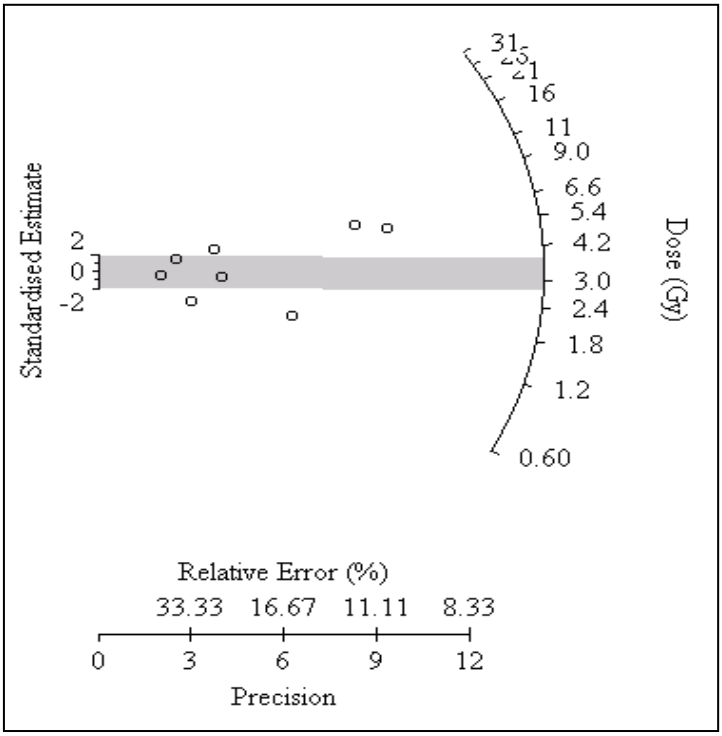

Figure 45. Radial plot of D_e dose distributions for sample MELA 0311F (chenier ridge 5) (Pers. comm., Cupper, 2010).

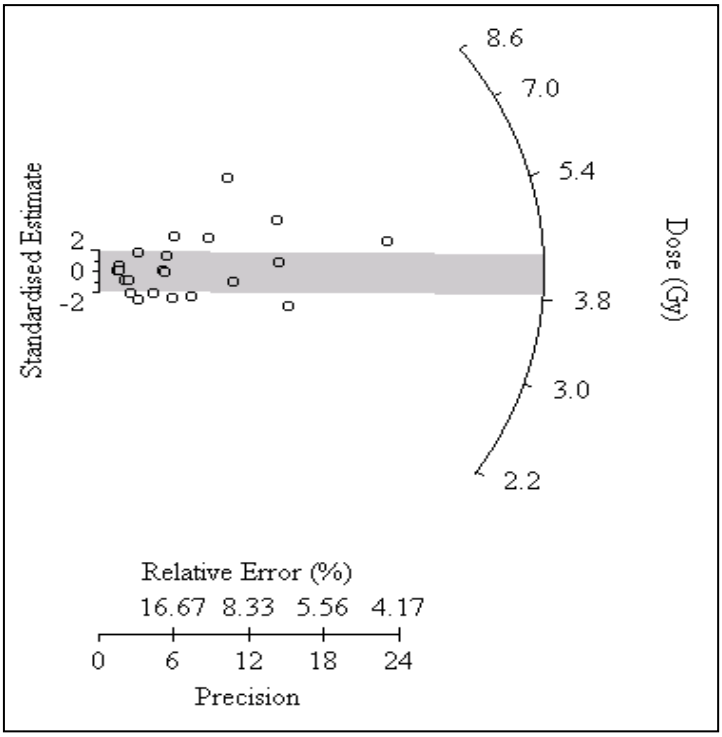

Figure 46. Radial plot of D_e dose distributions for sample MELA 0311E (chenier ridge 5) (Pers. comm., Cupper, 2010).

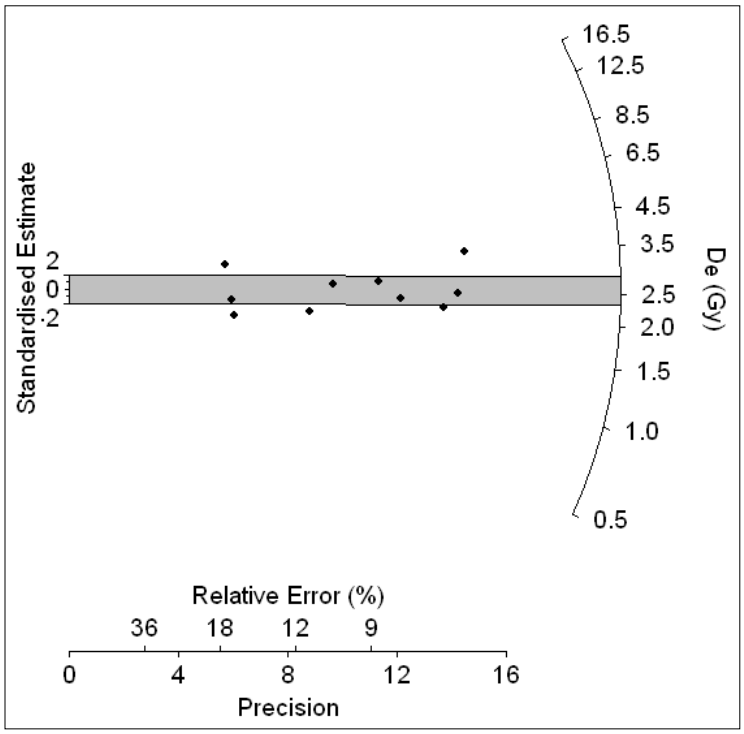

Figure 47. Radial plot of D_e dose distributions for sample K2050 (chenier ridge 6) (Pers. comm., Fitzsimmons, 2009).

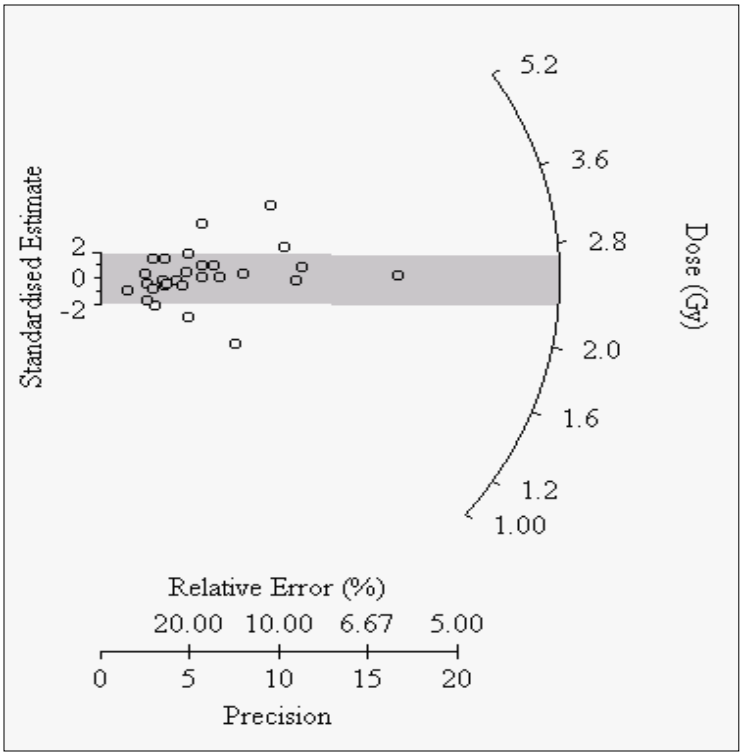

Figure 48. Radial plot of D_e dose distributions for sample MELA 0311A (chenier ridge 6) (Pers. comm., Cupper, 2009).

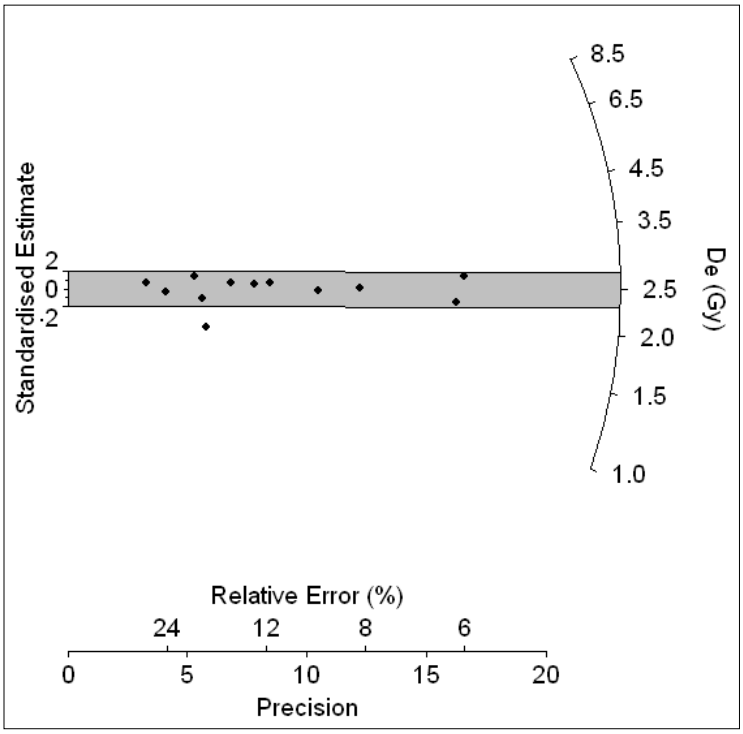

Figure 49. Radial plot D_e dose distributions for sample K2051 (chenier ridge 7) (Pers. comm., Fitzsimmons, 2009).

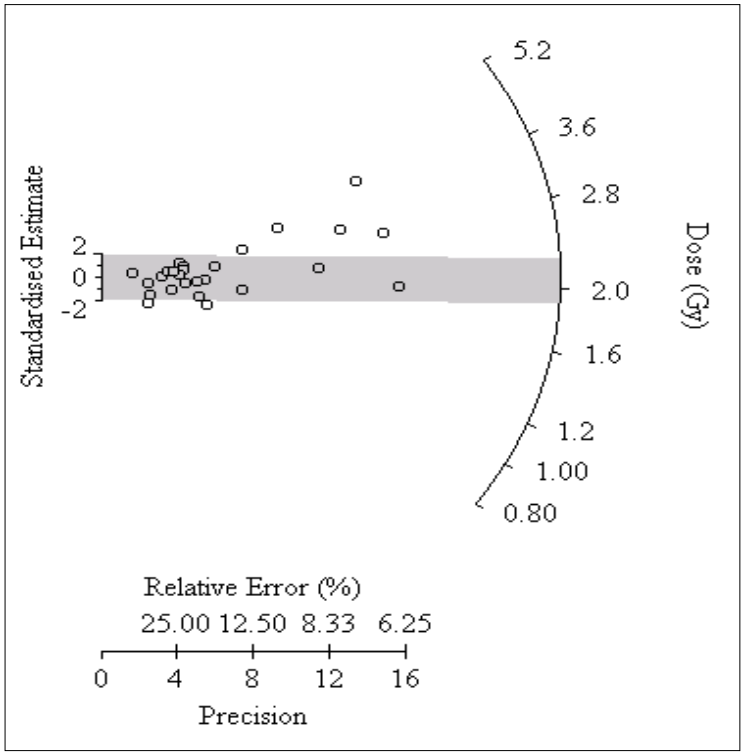

Figure 50. Radial plot of D_e dose distributions for sample MELA 0311B (chenier ridge 7) (Pers. comm., Cupper, 2009).

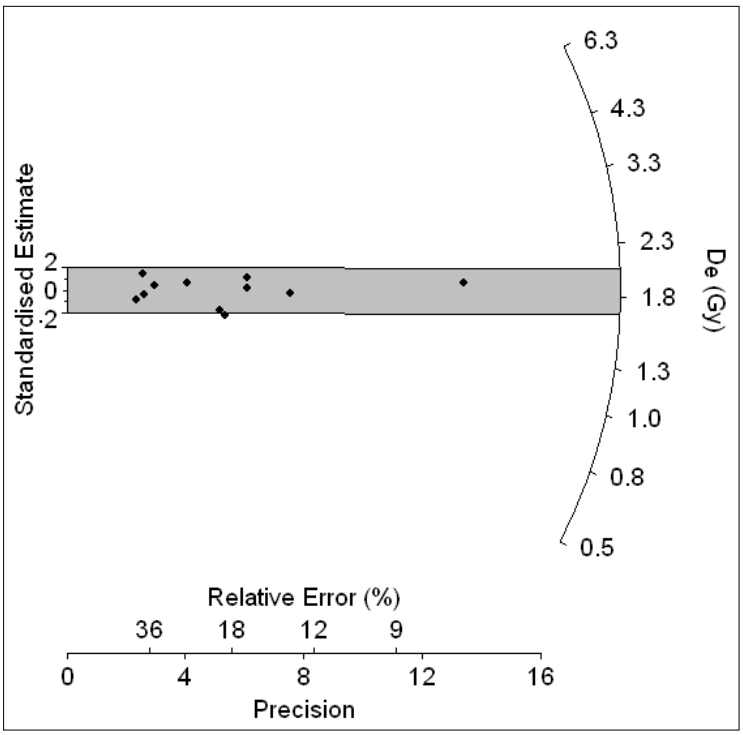

Figure 51. Radial plot D_e dose distributions for sample K2052 (chenier ridge 8) (Pers. comm., Fitzsimmons, 2009).

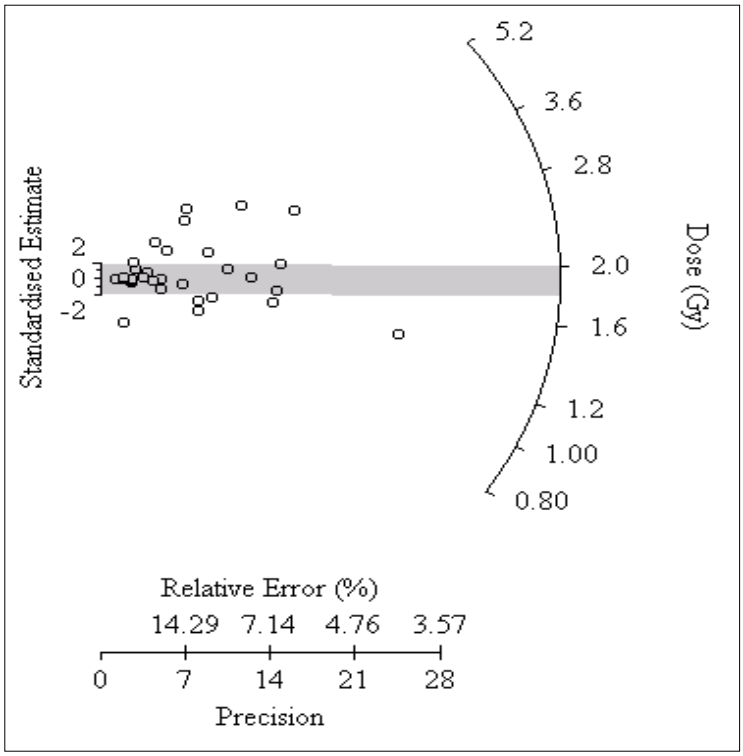

Figure 52. Radial plot D_e dose distributions for sample MELA 0311C (chenier ridge 8) (Pers. comm., Cupper, 2009).

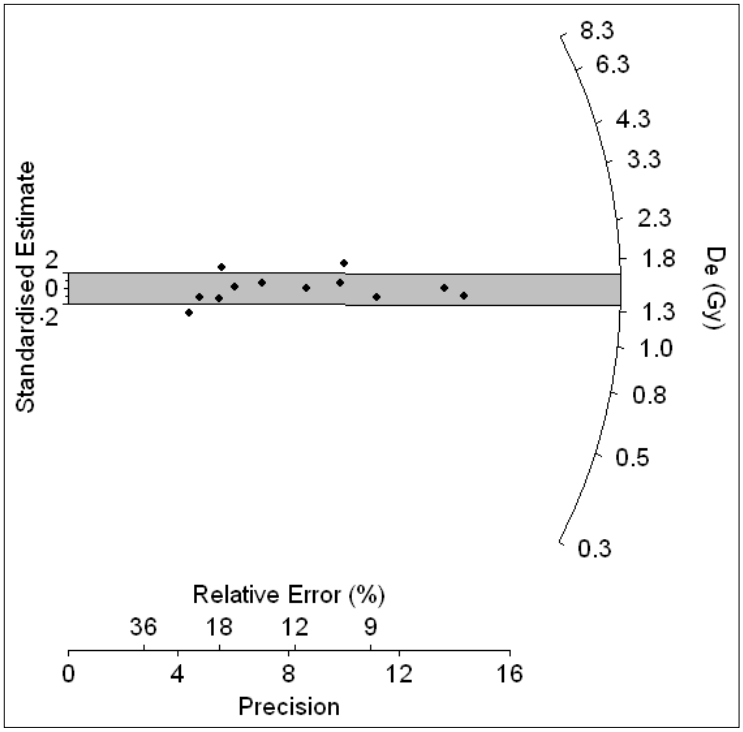

Figure 53. Radial plot D_e dose distributions for sample K2053 (chenier ridge 9) (Pers. comm., Fitzsimmons, 2009).

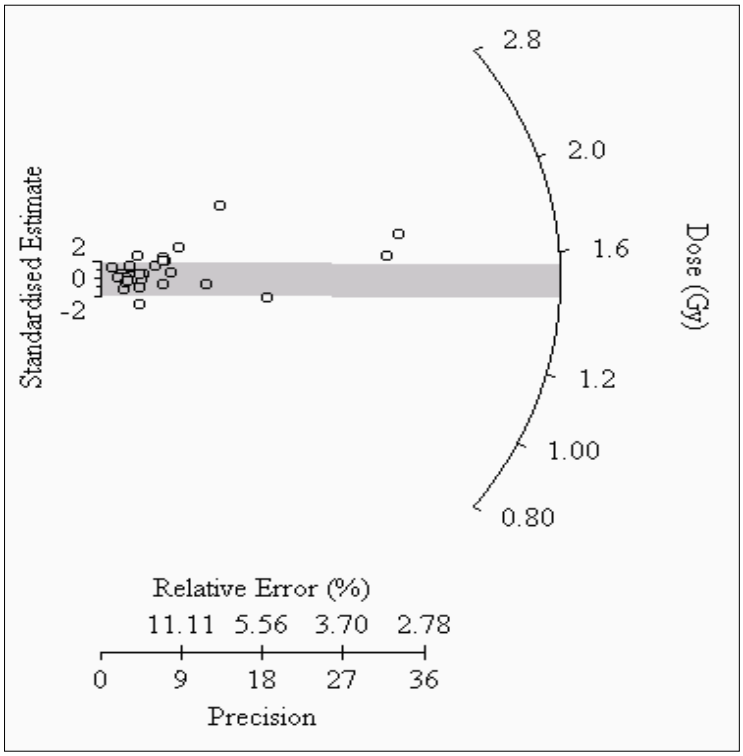

Figure 54. Radial plot D_e dose distributions for sample MELA 0311D (chenier ridge 9) (Pers. comm., Cupper, 2009).

5.5.3 Fire and bioturbation as sources of error in luminescence results

The error introduced by partial resetting of the luminescence signal by fire on the surface of deposits was considered to be negligible in this environment for several reasons. Firstly, since chenier ridges form at the shoreline, it is only after they have become established (i.e. finish accreting, and are isolated from the shoreline by a progradation phase) that they develop vegetation assemblages that might support fire. This means the sediment composing the chenier ridge body would not be exposed to fire while the ridge is being built. Second, the vegetation communities observed on the chenier ridges (Section 5.4) seems unlikely to develop fuel loads that would support large fires. Third, even if fire did occur in the chenier ridge communities, the fact that samples were removed from a minimum depth of 0.4 m (and generally around 1 m) below the surface would avoid the zone that would have been exposed to significant heat.

Bioturbation in the form of pig and kangaroo diggings, and scrub fowl mounds, were noticeable on ridges 6 to 11. The depth from which the samples were removed, however, would theoretically avoid any effects on the reliability of the luminescence ages as such diggings were only noticeable on ridges that supported vegetation (particularly grasses), and vegetation would only grow after the ridges had finished accreting. Furthermore, the most significant source of bioturbation (from wild pigs) would only have occurred recently since European settlement of the region, and again removing samples from a depth of at least 0.4 m (and generally around 1 m) depth should have avoided the zone that had been subject to bioturbation.

5.4.4 Radiocarbon results

Radiocarbon dating was carried out on three samples of marine shell, one obtained from the base of ridge 8 and two from the base of ridge 11. The results of these are shown in Table 4. The radiocarbon statistics and calibration curves for each sample (Wk 31099, Wk 22666, and Wk 23039) are shown in Figures 55, 56 and 57, respectively. Conventional ages were determined using standard radiometric dating based on a Libby half-life of 5568 years with correction for isotopic fractionation applied. Calibrated ages were obtained using OxCal version 3.10 using the calibration curve of Hughen *et al.* (2004), which includes a correction for local ocean surface 14C variability (i.e. marine reservoir effect) (Ulm, 2002).

Table 4. Conventional and calibrated radiocarbon dates (Pers. comm., Hogg, 2008, 2011).

Chenier ridge	Lab code	Sample depth (m)	Conventional age (yr BP ± 1σ)	68.2% (1σ) min age (cal yr BP)	68.2% (1σ) max age (cal yr BP)	95.4% (2σ) min age (cal yr BP)	95.4% (2σ) max age (cal yr BP)	Median age (cal yr BP ± 1σ)
8	Wk 31099	1.0	2263 ± 39	1807	1916	1737	1971	1854 ± 55
11	Wk 22666	1.0	1293 ± 39	770	890	720	920	820 ± 60
11	Wk 23039	1.3	1349 ± 36	830	930	780	970	875 ± 50

For all three ages there was a minor difference (up to 10 years) between the median ages of the 68.2% (1σ) range and the 95.4% (2σ) age ranges. For Wk 31099 the median ages for these ranges were 1862 cal yr BP and 1854 cal yr BP respectively. For Wk 22666 the median ages for these ranges were 830 cal yr BP and 820 cal yr BP respectively. For Wk 23039 the median ages were 880 cal yr BP and 875 cal yr BP respectively.

In all instances the ages accepted for interpretation (and quoted from this point forward) were the youngest median age ± 1σ. The youngest range was accepted to minimise the

possibility for error introduced by falsely old ages (following the logic described in Section 2.14.2), but in all cases the difference was so small that it did not affect interpretation.

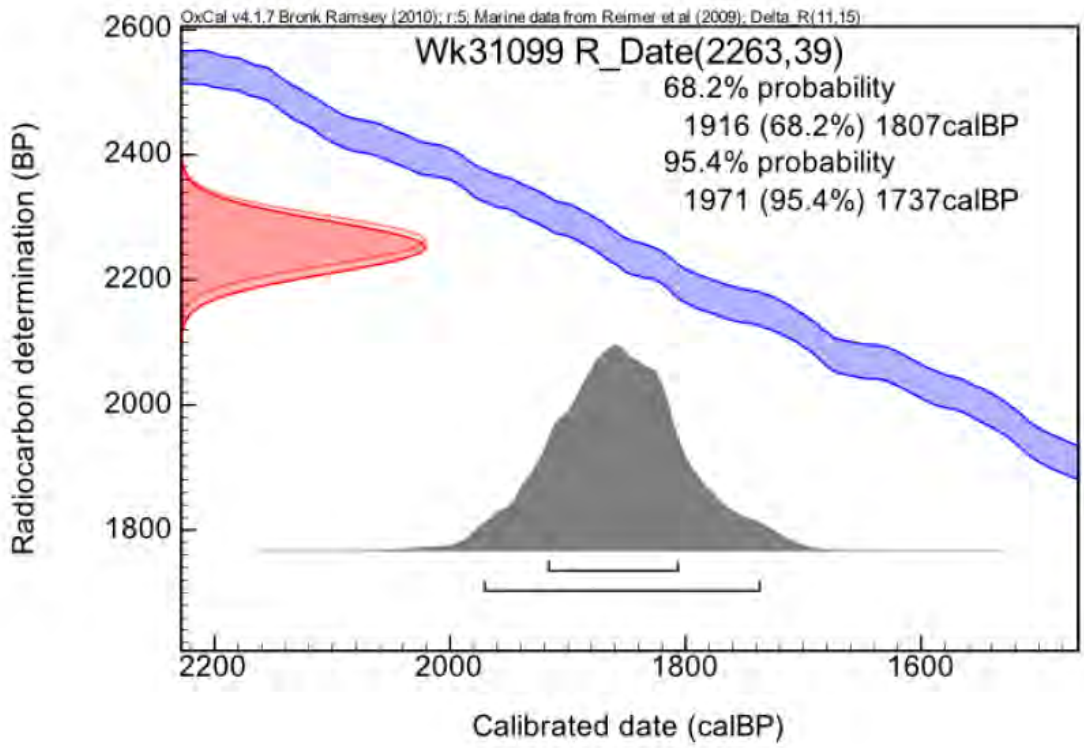

Figure 55. Radiocarbon statistics and calibration curve for sample Wk 31099 (Pers. comm., Hogg, 2011).

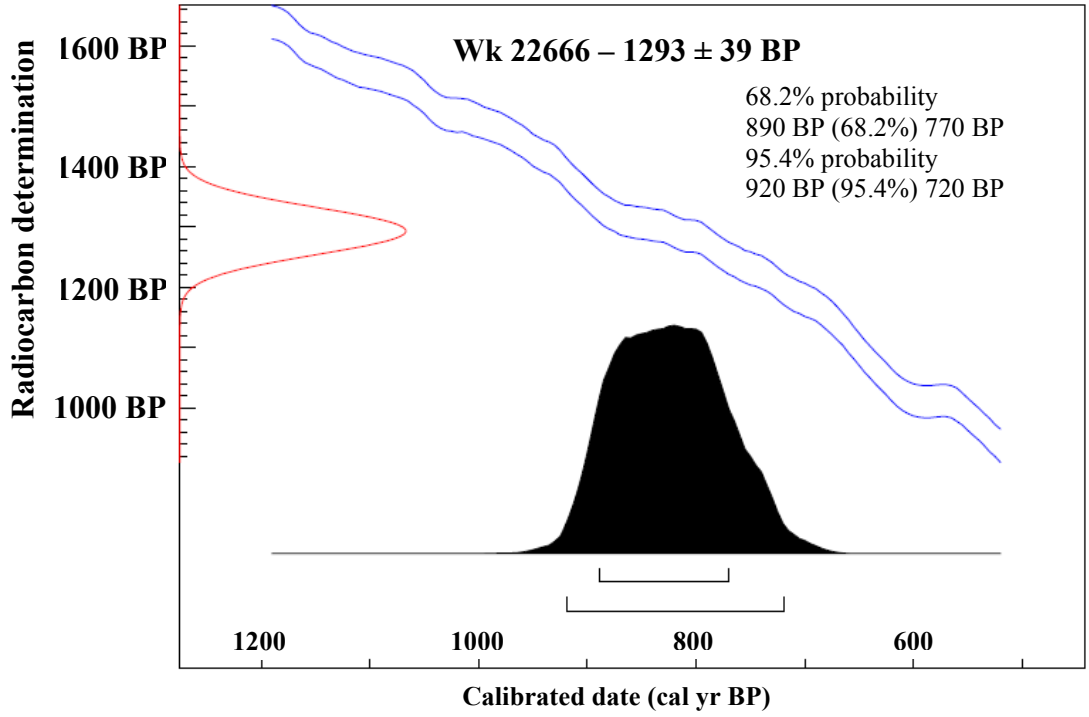

Figure 56. Radiocarbon statistics and calibration curve for sample Wk 22666 (Pers. comm., Hogg, 2008).

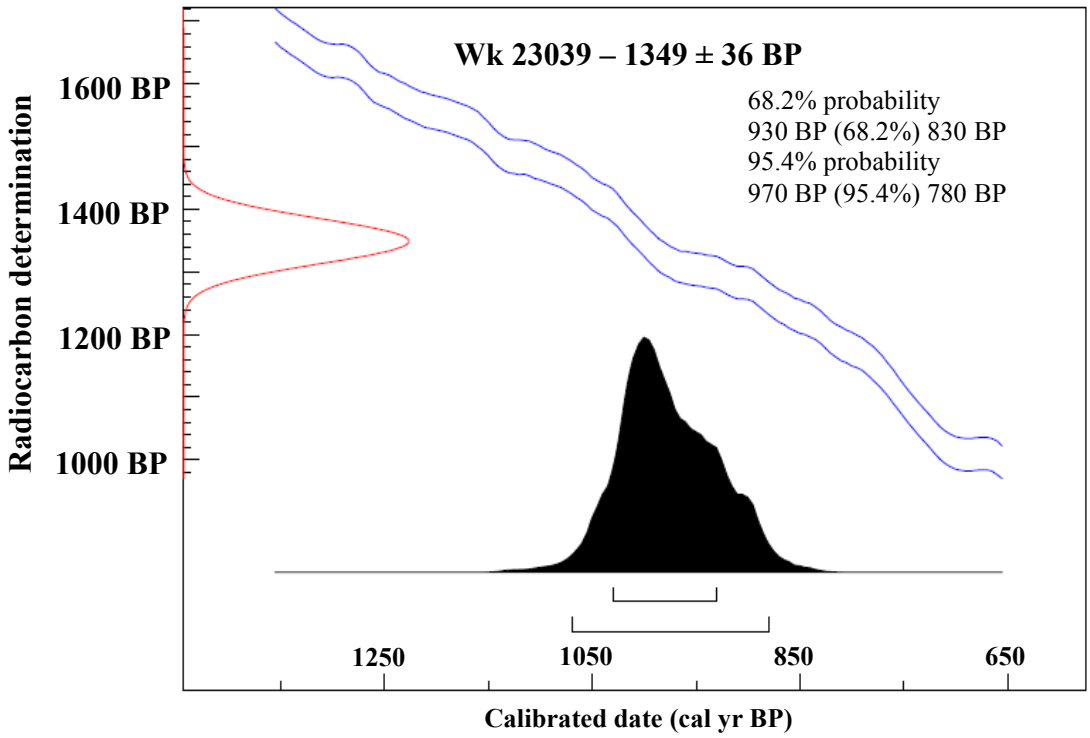

Figure 57. Radiocarbon statistics and calibration curve for sample Wk 23039 (Pers. comm., Hogg, 2008).

Chapter Six - Discussion

6.1 Introduction

This chapter contains a discussion of the results of this research (as described in Chapter 5) in the context of the research design described in Chapter 3. This chapter has been structured in the following way: Section 6.2 considers the radiocarbon, TL and OSL dating results obtained, the best chronology to accept for the chenier ridge sequence and some comparisons between different dating methods used as well as the results of this study and Chappell and Grindrod (1984). Section 6.3 considers the evidence for each of the six hypotheses for the Holocene development of the chenier ridge sequence in Princess Charlotte Bay using the data presented in Chapter 5. Section 6.4 contains a reassessment of the results of previous Australian chenier plain studies. Section 6.5 contains a synthesis of findings, including a new morphodynamic model for the development of the Princess Charlotte Bay chenier plain and observations regarding chenier plain development in Australia.

6.2 Dating results and depositional chronology

Several ages from the chenier ridge sequence were obtained using the three different dating methods (Section 5.5). As may be obvious already, it is likely that not all of these ages are reliable, and that not all are suitable for use in a discussion of the Holocene chronology. The following sections discuss some of the issues involved in interpreting the chenier ridge age dataset. These issues are: the age of ridge 4, the difference between the TL and OSL ages for ridges 6 to 9, and the difference between TL and radiocarbon ages for ridge 11. Considering these issues enabled the formation of a depositional chronology and identification of phases of chenier ridge building. The dating results also enabled a brief discussion of the results obtained by OSL and radiocarbon methods. Finally, comparisons were made between this new chronology from western Princess Charlotte Bay, and that presented by Chappell and Grindrod (1984) from the eastern side.

6.2.1 OSL results for chenier ridge 4

The OSL ages obtained from chenier ridge 4 (3500 ± 500 yr BP, MELA 0311H and 2400 ± 300 yr BP, MELA 0311G) were the only paired OSL ages that did not overlap at 1σ. They did, however, still overlap at 2σ, meaning that they can both be used as a reasonable estimate (albeit slightly less confident than that for the other ridges for which OSL ages were accepted) of the age of ridge 4. A possible reason for this difference is that the ridge accreted over a longer period of time than the other ridges. Although plausible, in the absence of more ages this suggestion cannot be confirmed and must remain tentative.

In situations where differing radiocarbon ages are obtained from a single chenier ridge, Lees (1987) suggested using the youngest age as the most representative of the „true" age. This would make it less likely that falsely old ages would be obtained. Although this suggestion was made in the context of radiocarbon ages, his conclusions could possibly be applied to this situation, particularly since there is nothing in the characteristics of the ages to suggest that either of them do not represent actual depositional events. Put another way – it seems far more likely that ages would be falsely old than falsely young. This is because it seems unlikely that a situation would arise in which the luminescence signal would be reset without a corresponding geomorphic event. Fire or bioturbation would be possible candidates for such an error, but these appear unlikely to have affected the chenier ridges in Princess Charlotte Bay (Section 5.5.3). With this reasoning in mind, one could have concluded that ridge 4 was formed somewhere around 2400 ± 300 yr BP, MELA 0311G, and possibly slightly earlier than this.

As the ages overlap at 2σ, however, the actual age of the ridge would be likely to lie somewhere within the age range defined by the median points of the two ages. That is, somewhere between 3500 yr BP and 2400 yr BP. For the sake of establishing an age which can confidently be used in the discussions that follow, the central point of these (i.e. 2950 yr BP) was accepted as the best estimate of the actual age of the ridge, and the error term defined by the length of time from 2950 yr BP to either 3500 yr BP or 2400 yr BP (i.e. 550 years). The best estimate of the age of ridge 4 was therefore 2950 ± 550 yr BP. The 550 year difference between this and the other possible age to accept for the ridge (i.e. 2400 ± 300 yr BP) is significant, but in the context of the ~6000 year sequence of Holocene coastal

events that are discussed, it does not introduce an amount of uncertainty that would prevent conclusions being drawn.

6.2.2 TL versus OSL results for chenier ridges 6 to 9

Apart from the TL ages from chenier ridge 11 (W4100 and W4106; Table 3; discussed in the following section), there was nothing in the *individual* characteristics of the other luminescence ages and associated plots/curves (Section 5.5) to suggest that any of them did not represent actual depositional events (i.e. events in which sufficient exposure occurred to effectively reset and remove any previously acquired luminescence signal). The clear differences between the TL and OSL ages obtained from ridges 6 to 9, however, necessitate a brief discussion on the most appropriate ages to accept and use in interpretation. The following discussion relates only to the ages obtained from those ridges that were subject to both TL and OSL techniques (i.e. the ages for chenier ridges 6 to 9, shown in Figure 29), providing an explanation as to why some of the ages obtained from these ridges were not used in later discussions. It is centred around the concept that just because the characteristics of a luminescence age suggest that it represents an actual depositional event, does not mean that this depositional event is the most recent one, or the one in which the chenier ridge was built. An explanation for this would be that prior to the most recent depositional event (i.e. the one in which the chenier ridge was built) the luminescence signal may not have been properly reset by sunlight.

Firstly, the five Pleistocene ages obtained from ridges 6 to 9 (10 900 ± 1600, W4101; 30 700 ± 2500, W4102; 11 800 ± 1500, W4103; 12 100 ± 1300, W4104; 10 500 ± 1300, W4109; Table 2) clearly could not provide any information about late Holocene events by mere virtue of the fact that they are Pleistocene. There are two possible explanations for the depositional events they represent. Their associated error terms make it possible that they represent deposition during the early Holocene (e.g. W4101, W4103, W4104 and W4109 could represent ages as young as ~9000 yr BP). If so, then this would have occurred when sea level was 10–30 m lower than present (Section 2.10.1), and could therefore be associated with large storm events. Although such storm events could have yielded a source of sediment for chenier ridge building in the late Holocene, it is inconceivable that they would have played any role in the Holocene development of the chenier plain. Alternatively, the Pleistocene ages may not be the result of „coastal" events at all, but could simply represent some other reworking event in the late Pleistocene or early Holocene. Whatever reworking event these dates represent, as stated previously they cannot provide information about Holocene events because they are not Holocene ages. Accepting them as reliable (i.e. accepting that chenier ridges 6 to 9 formed between 10 000 and 30 000 years ago) would not only go against the other (far more plausible) ages obtained, but also a vast body of well established Australian coastal geomorphological knowledge. Putting these dates to one side in a discussion of the formation of the Holocene sequence therefore seems prudent.

There were 11 Holocene luminescence ages for ridges 6 to 9 (Figure 29). The three derived from TL indicate deposition occurred between 6800 and 3900 yr BP (W4110, W4108 and W4107; Table 2), while the eight derived from OSL indicate deposition occurred between

2500 and 1700 yr BP (K2050-1, MELA0311A-B), and again between 1520 and 1100 yr BP (K2052-3, MELA0311B-C; Table 3). If the TL and OSL ages were obtained from significantly different stratigraphic levels then there may be some scope for accepting both. This is not the case (Tables 2 and 3), and one set must be therefore accepted as the most representative of the timings of chenier ridge building. There are three reasons to preference the OSL ages over the TL ones. Firstly, although all ages decreased in a seaward direction, only OSL ages overlapped at 1σ. The three Holocene TL ages were all partnered with Pleistocene ages obtained from the same pits at the same level (Section 5.5.1) making them appear erratic. Second, as was mentioned in the previous section, using the youngest age is likely to yield the best estimate of the „true" age, and it seems far more likely that the TL ages would be falsely old than the OSL ages be falsely young. Third, and this reason is closely related to the assertion made in the previous sentence, there are differences between the way in which TL and OSL signals are reset. These differences have implications for the ability of each technique to reveal a particular depositional event Compared to OSL, TL needs a far longer exposure time to reset the luminescence signal (Section 2.15.3). This means that where all other variables are the same (e.g. sediment type, process, setting etc.), an event that causes resetting of the TL signal will generally have to be of higher energy or duration than that which causes resetting of the OSL signal. Put another way, it is conceivable that the same deposition event could cause resetting of the OSL signal, but not of the TL signal. In this case, that deposition event could only be revealed using OSL. Attempting to use TL would instead reveal a previous event which was sufficient to reset the TL signal. This explanation would appear to offer a plausible reason for the differences between the TL and OSL ages. As was stated earlier, the fact that all TL and OSL samples were obtained from similar depths in each chenier ridge (Tables 2 and 3), and using

identical sampling methods, also suggested that the differences between the TL and OSL ages were derived from the methods themselves rather than indicating different patterns or timings of deposition. Aside from leading to the conclusion that the OSL ages should be accepted over the TL ones, this also suggests that TL is likely to be an inappropriate method for dating chenier ridges.

With the above reasoning in mind the OSL ages were tentatively accepted as providing better estimates of the timings of the most recent deposition events (i.e. the times at which chenier ridges 6 to 9 were built) than the TL ages.

6.2.3 TL versus radiocarbon results for chenier ridge 11

The TL characteristics of the ages obtained from chenier ridge 11 were the only ones that indicated sample heterogeneity, and suggested that the ages did not represent actual deposition events (Section 5.5.1). A possible explanation for this was the large amount of shell material through the chenier profile (Section 5.4.3), which would have contained sand that had not been exposed to sunlight at deposition. This problem appeared to be unique to ridge 11, as the other ridges have not only yielded reliable OSL ages but TL ones too. While this explanation for obtaining unreliable TL ages from ridge 11 is only one possibility (albeit a plausible one, given the prevalence of shell throughout the entire profile which is unique to ridge 11), the risk that it would also prevent reliable OSL ages being obtained from that ridge was used as justification for the use of radiocarbon methods on that ridge only. Although radiocarbon ages can only be considered maximum ages for a

chenier ridge (Section 2.14), the fact that the ages obtained for ridge 11 (820 ± 60 cal. yr BP, Wk22666; 875 ± 50 cal. yr BP, Wk23039) were still younger than those from ridges landward (i.e. there is no reversal in ages along the ridge sequence) made them a reliable upper limit for the age of ridge 11.

6.2.4 Comparison between OSL and radiocarbon ages for ridge 8

The results obtained from ridge 8 allow a comparison and brief discussion of the results obtained from radiocarbon and OSL methods as they apply to chenier ridge 8. It is regrettable that only one radiocarbon age was obtained from this ridge, and that further radiocarbon ages were not obtained from other ridges subject to OSL dating techniques. The reason for this is that at the time when field work was conducted, radiocarbon methods were never intended to be used at all, so shell samples were not obtained from every ridge. Despite this, a preliminary discussion based on the single age that was obtained may provide directions for the future use of OSL techniques on chenier ridges, as well as provide some indication of the comparability of the results of this study to other chenier ridge studies that have employed radiocarbon methods.

The main issue involved in interpreting radiocarbon ages involves a temporal difference between when a shell is formed, and when it is incorporated into a ridge (Section 2.14.2). Even if this difference were as small as possible, one would still expect some lag between the time of shell formation and the time at which it is finally deposited. This is because it is generally accepted that chenier ridges are built from coarse material that is winnowed and sorted from a muddy matrix (Section 2.5.2.1), and a shell will only become part of this

muddy matrix after the mollusc dies and the shell is deposited. Assuming the OSL ages are an accurate estimate of the time of ridge building, the fact that the radiocarbon age (1854 ± 55 cal yr BP) is approximately 430–550 years older than the OSL ages could be simply explained by this temporal lag. Given that the OSL ages both overlap at 1σ, and their characteristics showed no substantial evidence of incomplete bleaching or mixing (Section 5.5.2), this explanation would seem plausible. Put simply, the difference between the ages obtained from the different methods can be considered reasonable, and neither result (in terms of OSL versus radiocarbon) casts doubt on the reliability of the other.

Although these results are by no means substantial, this preliminary conclusion does suggest that OSL and radiocarbon results may be capable of yielding similar depositional chronologies in certain settings.

6.2.5 Holocene ridge building chronology in western Princess Charlotte Bay

With the reasoning of the previous sections in mind, and the fact that only OSL ages were obtained for chenier ridges 1, 4, and 5 (and further, that these ages generally seem reasonable; Section 5.5.2) the OSL and radiocarbon ages were accepted as providing a tentative chronology for the Holocene chenier ridge sequence. This chronology is described below.

The chenier sequence began forming with the establishment of the most landward ridge (ridge 1) at around 4000 yr BP. In the ~1500 years following this, two more chenier ridges

were built (ridges 2 and 3). Ridge 4 was built at around 2950 ± 5500 yr BP. Ridge 5 was formed between 2100 ± 300 and 2200 ± 400 yr BP. Ridge 6 was built between 2270 ± 140 and 2200 ± 200 yr BP. Ridge 7 was built between 2230 ± 140 and 1900 ± 200 yr BP. Ridge 8 was built between 1420 ± 100 and 1300 ± 200 yr BP. Ridge 9 was built between 1380 ± 90 and 1300 ± 200 yr BP. The calibrated and corrected radiocarbon ages provided an upper limit of the age of ridge 11 at 820 ± 60 cal yr BP.

The ages described above allow three general phases of chenier ridge building to be identified. These phases are: (1) An early phase lasting from when the ridge sequence began to form (around 4000 yr BP) until around 2000 yr BP in which at least seven chenier ridges (1 to 7) were built; (2) a phase centred on approximately 1350 yr BP in which at least two ridges were built (8 and 9); and (3) a final phase since approximately 820 yr BP in which one final ridge (11) was built. The fact that an incipient chenier (i.e. beach) currently exists at the shoreline likely means that the system is presently in a fourth phase of ridge building.

6.2.6 Comparison with the results from eastern Princess Charlotte Bay

Clearly, there are differences in the number and configuration of ridges at different locations in the bay (Section 5.2.2). Such differences were even evident between the two eastern transects examined by Chappell and Grindrod (1984) (e.g. Figure 58), in which notable differences existed between chenier ridge height, spacing, dimensions and age.

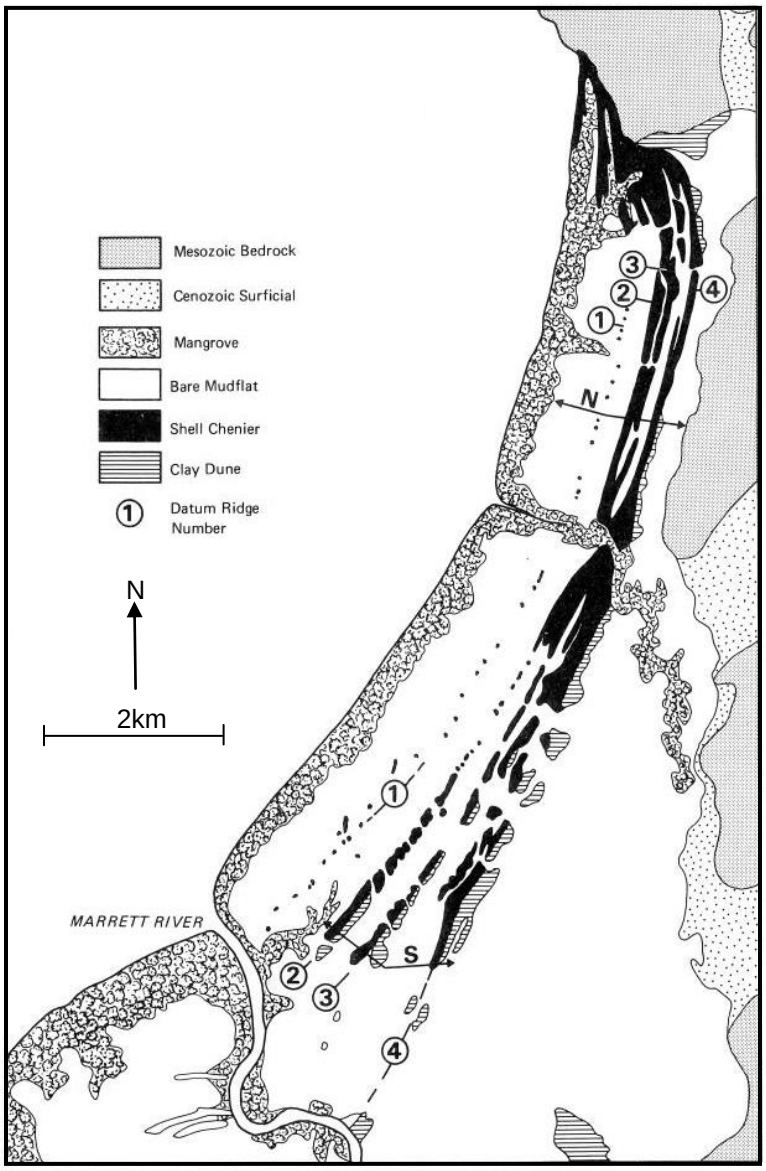

Figure 58. Diagram representing the chenier ridge sequence in eastern Princess Charlotte Bay. Modified from Chappell and Grindrod (1984). The „N‟ and „S‟ denote the northern and southern transects respectively. The location of this area within the bay is shown in Figure 18 (Section 5.2.1).

Furthermore, although some ridges appear to be better preserved and exist for longer distances than others, no single ridge is continuous for the entire distance from east to west (Section 5.2.2). This means that it is impossible to directly compare the ages of individual ridges from different areas of the bay with any certainty. This problem is compounded by

the fact that this new chronology in the west has been obtained using mostly OSL methods, while the western chronology was developed using uncalibrated radiocarbon ages. Despite these difficulties, a discussion which broadly compares the depositional chronologies, the nature of ridges, and the nature of the underlying sediments from the eastern and western areas of Princess Charlotte Bay will still be useful and may allow broad relationships to be identified. These relationships can then provide insights in to localised variations in morphodynamic processes across the bay.

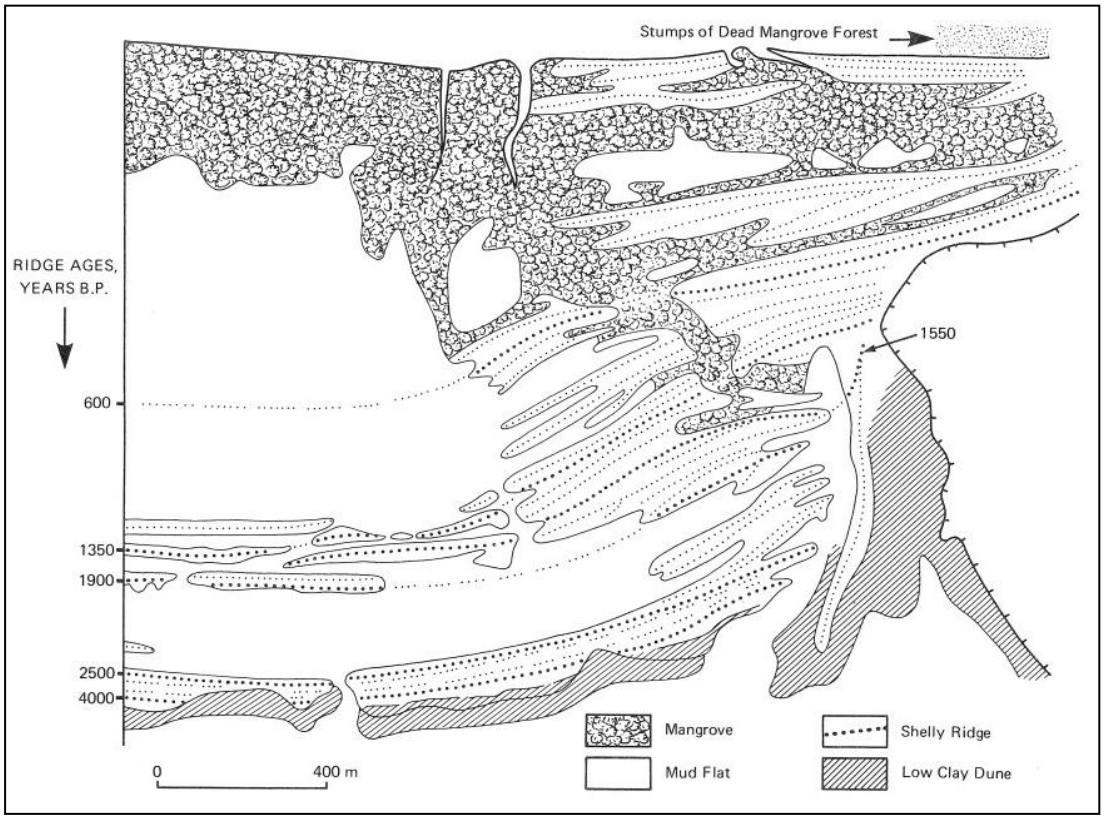

Figure 59. Datum ridges and ages in eastern Princess Charlotte Bay (northern part of study area) as presented in Chappell and Grindrod (1984).

The depositional chronology from eastern Princess Charlotte Bay was first presented in Grindrod (1983) and discussed further in Chappell and Grindrod (1984). It was based on

two transects, the locations of which are shown in Figure 58. Although there were differences in the ridges observed at both transects, five general chenier ridge crests were identified, these being referred to as „datum ridges" for convenience (Figures 58 and 59). The (uncalibrated) ages of ridge building were approximated for these datum ridges, based on several radiocarbon ages. The most landward datum ridge was given an age of 4000 yr BP. Moving seaward, the ages of the other datum ridges are approximately 2500 yr BP, 1900 yr BP, 1350 yr BP and 600 yr BP (Figure 59). To allow comparison with the new, western chronology, these ages were calibrated. The translation from these uncalibrated datum ridge ages to calibrated approximate individual ages is shown in Table 5.

Table 5. Translation from uncalibrated datum ridge ages to approximate calibrated ages for the eastern ridge sequence. The first column contains datum ridge ages given in Chappell and Grindrod (1984) (i.e. the ages in Figure 59). The second column shows the individual ages obtained from chenier ridges along the eastern ridge sequence (Grindrod, 1983; Chappell and Grindrod, 1984). These ages were used in the reanalyses done by Lees and Clements (1987), and in Section 6.4 of this thesis. The third column shows the calibrated individual ridge ages (after Klein *et. al.*, 1982). The fourth column shows the ages used in the discussion comparing ridge ages in eastern Princess Charlotte Bay to those in western Princess Charlotte Bay. They are the averages of the individual calibrated ages rounded to the nearest 10.

Datum ridge age	**Uncalibrated individual ridge age (yr BP)**	**Calibrated individual ridge age**	**Age used for comparison (cal yr BP)**
600	710 ± 70	665 ± 55	610
	510 ± 80	550 ± 105	
1350	1670 ± 70	1560 ± 165	1540
	1540 ± 115	1515 ± 190	
1900	1900 ± 70	1845 ± 125	1970
	1960 ± 70	1920 ± 185	
	2170 ± 80	2150 ± 210	
2500	2460 ± 135	2470 ± 320	2470
4000	4030 ± 90	4552 ± 289	4550

A number of observations can be made from comparing these results to the depositional summary provided in the previous section. The most obvious is that ridge building in the

west has occurred at a higher rate than ridge building in east (i.e. at least 10 ridges since 4000 yr BP compared to five ridges since 4550 yr BP, respectively). Interestingly, despite this major difference, and taking into account the possibility that radiocarbon may slightly overestimate actual ages compared to OSL, the timings of ridge building appear to be generally similar, and the ages shown for the eastern areas agree very closely with the phases of ridge building identified in the west. For example, the approximate eastern ages of 4550 cal yr BP, 2470 cal yr BP and 1970 cal yr BP occur during the early phase identified in the west that lasted from around 4000 yr BP until around 2000 yr BP. As well as this, the later eastern ages of 1540 cal yr BP and 610 cal yr BP agree with the later western phases centred on approximately 1350 yr BP and since approximately 820 yr BP respectively.

As in the east, the crest heights of ridges in the west also tend to decrease as they get younger, although the height change is not consistent for every ridge, particularly those towards the rear of the transect (i.e. ridges 1 to 5). Chappell and Grindrod (1984) stated that the (more gradual) height change they observed could be attributed to the accumulation of aeolian and occupation overburden on older ridges, rather than actual differences between crest heights. There were no aeolian deposits identified on the western ridge sequence, and this may explain the observed differences in height changes. Chappell and Grindrod (1984) stated that the only chenier ridge with a distinctly lower crest was the youngest (~600 yr BP) ridge. This is the same as the current transect, although the crest of ridge 9 is also relatively low compared to landward ridges 1 to 8.

Aside from observations related to ridge configurations or the physical dimensions of ridges, there are also differences in the composition of ridges such that ridges in the east tend to be composed entirely of shell, while the ones in the west tend to be composed of coarse sand and gravel.

Data regarding the ridges aside, one of the most interesting differences between the results obtained by Grindrod (1983) from eastern Princess Charlotte Bay, and the current results from the west, is the nature of the underlying sediments. Grindrod (1983) provided detailed descriptions of the sediment underlying the ridge sequence in the east. In short, the chenier sequence there overlies an organic intertidal (mangrove) facies. This is completely different to the sediment underlying ridges along the western transect, which in all locations was observed to be grey intertidal muds. This is an important observation, and provides an important insight into how chenier formation processes have differed across the bay. It is discussed later.

6.3 Testing hypotheses for individual chenier plain models

The concept of an internal dynamic was previously put forward by Chappell and Grindrod (1984) to explain the development of the Princess Charlotte Bay chenier plain. Since this occurred, a vast body of knowledge has been developed regarding Holocene palaeoenvironmental changes. Before attempting to rethink this concept, it was therefore appropriate to test the applicability of other six models. This systematic approach was taken

to increase the likelihood of identifying factors that were not incorporated into Chappell and Grindrod‟s (1984) work (Section 3.2).

The following sections consider the evidence in support of each of the six hypotheses described in Section 3.2. The evidence used includes the new data and observations presented in Chapter 5. Where possible, evidence from Chappell and Grindrod‟s (1984) investigation is also considered, as well as observations made from comparing the two datasets.

6.3.1 Hypotheses (1): River channel switching

Previous studies have linked river channel switching to chenier plain development in settings where channel avulsions occur (i.e. there is a significant sudden shift in the position of the river mouth (Section 2.5.1.1). While evidence for abandoned palaeochannels (e.g. cutoff bends) and minor shifts in the position of river mouths is apparent in aerial photography and satellite imagery (e.g. Figures 16 – 23; discussed in Section 5.2.2), there is no evidence for *major* avulsions and significant changes in the position of the mouths of any of the four rivers draining into Princess Charlotte Bay.

The observations that can be made from aerial photography and satellite imagery (Section 5.2.2) suggest that the role of rivers in the development of the chenier plain has been simply to simply to rework ridge sequences over time, rather than control where the ridges have formed. The only areas where the position of the river mouth appears to have an influence

on ridge morphology are those areas adjacent to the river mouths themselves. In these areas (e.g. Figures 21 and 22; Section 5.2.1), recurved spits appear to be present. The existence of such features in these locations, compared to the absence of them elsewhere, suggests the position of the river mouth has been relatively stable as the plain has prograded over time. In the eastern and western areas of the progradational plain where ridge sequences are preserved, the ridges also tend to be oriented in relatively continuous and generally parallel lines (albeit gradually coalescing at the eastern and western extents of the plain) over long distances (e.g. Figures 16, 17, 19, 23 and 58), and have also formed at generally similar times (Section 6.2.6). This is not consistent with the ridge configuration expected where delta channel switching is the major controlling factor (i.e. clusters of discontinuous chenier ridges in distinct age groups; Section 3.2.1).

No evidence therefore exists that river channel switching has been a primary factor in the Holocene development of the chenier plain at Princess Charlotte Bay, and Hypothesis (1) is rejected.

6.3.2 Hypotheses (2): Mud shoal migration

The mud shoal migration model seems unlikely to be applicable in Princess Charlotte Bay because of the small suspended sediment concentration in the nearshore zone and lack of longshore currents compared to locations where it *is* applicable (e.g. the Amazon-Orinoco coastline). For example, the annual sediment discharge of the Normanby River (7530 m^3; Section 4.2.3) is negligible compared to the more than 1000 million t/yr supplied by the

Amazon River (Section 2.5.1.2). This relatively small discharge means suspended sediment concentration in the near shore zone is likely to be insufficient to form slingmud, a critical requirement for this model (Section 2.5.1.2).

Another requirement is the existence of a suitable longshore current. The majority of the Princess Charlotte Bay shoreline where chenier ridges are present (including both the eastern area examined by Chappell and Grindrod (1984), and the western section examined now) is protected from the persistent seasonal winds and the longshore currents generated there are either non-existent or minimal (Figure 14; Section 4.2.5). Furthermore, examination of aerial photography and imagery (Figures 16 – 23; Section 5.2), a bathymetric chart of Princess Charlotte Bay produced in an earlier sedimentation study (Frankel, 1974), and a modern bathymetric chart (Figure 13; Section 4.2.1) revealed no regular features that could be interpreted as mud shoals. Hypothesis (2) is therefore rejected.

6.3.3 Hypotheses (3): Climate fluctuations

The key requirement of the climate model is that chenier ridge building occurred during arid phases, while progradation of the poorly sorted mud flat occurred during pluvial phases. If climate fluctuations played a major role in the development of the Princess Charlotte Bay chenier plain, the general climate (Section 2.9.2.1) and ENSO (Section 2.9.2.2) reconstructions that have been published for the northern Queensland region, and the limited evidence for changes in sea surface temperature (Section 2.9.2.4), imply that

progradation would have been more prevalent from the time of sea level stabilisation around 6800 yr BP, until around 3700 yr BP when conditions became more variable. The decrease in rainfall and drop in fluvial flow associated with this would have increased the likelihood of chenier ridge formation after this time. A possible increase in effective precipitation in the last 1000–2000 years would have subsequently encouraged progradation, with a possible punctuation associated with the Medieval Warm Period and/or Little Ice Age events between 1200 and 200 yr BP (Section 2.9.2.3).

Although there is no data to allow speculation of events that occurred prior to 4000 yr BP, the agreement between the age obtained for the earliest Holocene ridge (4000 ± 400 yr BP; MELA0311I) would appear to coincide with the change to more arid and variable conditions at around 3700 yr BP (Section 2.9.2.1). Furthermore, the early phase of ridge building which starts at this time persists until around 2000 yr BP, a time at which some authors have hypothesised an increase in rainfall (e.g. Stephens and Head, 1995). As this early phase is the one in which the most chenier ridges were built, the prospect of a role for climatic forcing seems reasonable. Linking chenier ridge building to climate fluctuations after 2000 yr BP is more difficult as fewer definitive climatic fluctuations have been identified. The general consensus that climate is likely to have been „more variable" (Section 2.9.2.1) does, however, allow scope for a possible link to be retained.

The global Holocene arid climate phases identified by Mayewski *et al.* (2004) (i.e. 9000–8000, 6000–5000, 4200–3800, 3500–2500, 1200–1000 and 600–150 yr BP) appear to offer a broadly similar picture to the one described above (i.e. they may explain chenier ridge building occurring between 4000 and 2000 yr BP), with the added possibility that the most

recent arid phase beginning at around 600 yr BP may also provide an explanation for the development of a chenier ridge in the last 1000 years.

The data certainly do not allow the acceptance of the climate model outright, because there is no evidence for multiple fluctuations that would have led to multiple ridges. The evidence does, however, point to a possible modulating role for climate changes, because chenier ridges have only formed in the time since Holocene climate in the north Queensland region is thought to have become more arid and variable. Based on this reasoning, the possibility that the climate model could be incorporated into a new morphodynamic model for chenier plain development in Princess Charlotte Bay should be retained.

6.3.4 Hypotheses (4): Tropical cyclone activity

The tropical cyclone activity model states that the development of chenier plains has been driven primarily by tropical cyclone activity (i.e. that chenier ridges are effectively storm ridges built by the passage of tropical cyclones, or phases of tropical cyclone activity). If true, it would mean that there should be agreement between either: the frequency of ridge forming, and the frequency of tropical cyclone events; or between the periods of ridge forming and periods of elevated tropical cyclone intensity (Section 3.2.4).

Holocene tropical cyclone frequency is generally thought to have been uniform along the north Queensland coast since 5000 yr BP. The return interval of individual major events is

hypothesised to have been between 80 and 300 years (Chappell *et al.*, 1983; Chivas *et al.*, 1986; Hayne and Chappell, 2001; Nott and Hayne, 2001). In particular, super cyclones are hypothesised to have occurred every 200–300 years along all parts of the Great Barrier Reef (Nott and Hayne, 2001; Section 2.9.3). This hypothesised uniform frequency, and the fact that tropical cyclones occur at a range of intensities, effectively means that the concepts of „(individual) tropical cyclone events", and „phases of storminess" can be considered the same (i.e. each phase of storminess consists of one supercyclone and a suite of smaller cyclone events). If Nott and Hayne"s (2001) hypothesised frequency is accurate, this would mean that there have been more than 16 supercyclones, or 16 phases marked by individual supercyclones, during the time that the Holocene chenier ridge sequence has been developing. Given that 11 ridge crests were identified along the western transect, the frequency of ridge formation compared to supercyclone frequency could be considered generally similar (i.e. ridge building has occurred in the same order as major tropical cyclones).

Although the *number* of ridges built since 5000 yr BP may be generally similar to the number of tropical supercyclones/phases of cyclone activity, the fact that it was possible to identify three phases of ridge building (Section 6.2.5) shows that the development of ridges has been constrained to particular periods. This means that the *frequency* of ridge formation has not been uniform, which contrasts with the uniform tropical cyclone frequency hypothesised by Nott and Hayne (2001), and others described in Section 2.9.3. If the frequency of individual ridge formation *within* phases of ridge building is considered, however, then more agreement may be found. For example, during ridge building phase 1 (4000 yr BP to 2000 yr BP; Section 6.2.5), seven chenier ridges are built. Based on this

data, a role for a force which operates on a cycle in the order of 300 years is therefore possible.

With regards to tropical cyclone intensity, Forsyth (2010) identified a phase of high intensity events (with ridges being deposited by category 4–5 tropical cyclones) between 5010 and 3380 yr BP, a phase of lower intensity events (category 1–4) between 3380 and 1550 yr BP, and a return to higher intensity events (category 4–5) between 1550 and 130 yr BP. Taken alone, these changes in intensity do not appear to be reflected in the periods of chenier ridge building, as ridges have been built during both high and low intensity phases.

The reasoning above would suggest that while tropical cyclone activity does not appear to have been a primary factor controlling development of the chenier plain, it seems possible that it has played some role. This is based primarily on two observations. Firstly, the height of ridges (i.e. ridge crests) is far beyond the height reached by „normal" wave activity and larger waves must have played an accretional role. Secondly, the number of ridges is approximately similar to the number of supercyclones hypothesised to have occurred, and the frequency of ridge formation during phases of ridge building may be similar to the frequency of supercyclones or supercyclone phases.

Although the „height of ridge crests above normal wave height" aspect of the preceding discussion would seem to have few weaknesses, a possible problem with the „tropical cyclone frequency" aspect of it is that the tropical cyclone records are all sourced from coastal features. Furthermore, one of the studies (Nott and Hayne, 2001) based their conclusions on a record sourced from Princess Charlotte Bay. The possibility for circular

logic may therefore exist. There are three observations, however, which argue against the existence of a degree of circular logic significant enough to make the above conclusions unreasonable. Firstly, although the records used were indeed coastal, the vast majority of them were coral shingle ridges that consisted of clearly identifiable individual units, each of which was deposited in a single tropical cyclone event, and which accreted under the influence of tropical cyclones alone (e.g. Chappell *et al.*, 1983; Chivas *et al.*, 1986; Hayne and Chappell, 2001; Nott and Hayne, 2001). This is not the case for chenier ridges sequences, which tend to be composed of much finer sediment, and the development of which is driven by more factors than tropical cyclone activity alone (Section 2.5). Second, although Nott and Hayne (2001) did indeed use data from Princess Charlotte Bay, this was only one component of their overall dataset, their conclusions being based on records derived from eight sites in total. Third, the fact that the sequence from Princess Charlotte Bay was referenced as being from a paper that was „*in the press*", yet was never subsequently published (See Hayne and Chappell (reference number 11) in Nott and Hayne, 2001), means questions exist regarding the validity of the data used.

The discussion of circular logic aside, it should be reinforced that the preceding discussion concluded that tropical cyclone activity did *not* appear to have been a primary factor controlling development of the chenier plain. Despite the (based on the above considerations, very minimal) possibility of a certain degree of circular logic, therefore, it is still reasonable to suggest that it seems possible that tropical cyclones have played some role controlling the development of the chenier plain.

Based on this reasoning, therefore, the possibility that the tropical cyclone model could be incorporated into a new morphodynamic model for chenier plain development in Princess Charlotte Bay should be retained.

6.3.5 Hypotheses (5): Human activity

More intense Aboriginal burning would theoretically have caused higher sedimentation rates (thereby encouraging progradation) by increasing catchment instability. Aboriginal burning in the region of the Princess Charlotte Bay catchment is thought to have been highest from around 4000 yr BP until as late as 2000 yr BP (Section 2.11). Given that this is the period in which most chenier ridges were built (i.e. chenier ridge building phase 1; Section 6.2.5), it seems highly unlikely that this burning could have had any influence on the development of the chenier plain. There is therefore little evidence to support the human activity model, and Hypothesis (5) is rejected.

6.3.6 Hypotheses (6): Sea level oscillations

In its simplest form, the sea level oscillation model states that chenier ridge building occurs during transgressions, while mudflat progradation occurs during regressions (Section 2.5.1.7). The interpretation generally put forward for a late Holocene oscillating sea level in Australia, as well as nearby sites in the Pacific and Brazil, is firstly a rise to a maximum of up to 2 m above present levels by 6800 yr BP, followed by two 0.3–1.0 m oscillations

(4800–4500 and 3000–2700 yr BP) and a fall to present levels after 2000 yr BP (Section 2.10.2.4). Changes subsequent to this are difficult to detect due to a paucity of data. Sea level curves from a number of locations around Australia and the Pacific Ocean, with the new chenier ridge ages superimposed, are shown in Figure 60. The rationale for using more curves than just those developed for the Australian east coast was explained in Section 3.2.6.

Some interesting observations can be made from this data. Firstly, generally speaking chenier ridge building appears to have occurred more often during periods when sea level was oscillating (i.e. from as early as 6000 yr BP until around 2000 yr BP; Figure 60), than when it has been generally regressing (i.e. after 2000 yr BP; Figure 60). For example, during times when the curves show an oscillating sea level with still stands and/or transgressive phases, at least seven chenier ridges were built (ridges 1 to 7, denoted as „ridge building phase 1" in Section 6.2.5; see Figure 60), compared to four during times of regression (ridges 8 to 11, ridge building phases 2 and 3; see Figure 60). If geomorphic evidence of erosion episodes is accepted, as the authors suggest, as indicating a later transgression or later persistence of higher sea levels (Section 2.10.2.5) then perhaps all ridges but one (ridge 11, ridge building phase 3; see Figure 60) may have formed during transgressive or stable phases. Although fewer ridges were formed in the eastern region of Princess Charlotte Bay, the times at which they were built allow a similar observation to be made. The formation of chenier ridge 11 after 820 ± 60 yr BP (Section 6.2.5) is difficult to tie to sea level oscillations due to a lack of data for this period (Section 2.10.2.4). One curve, however, does suggest a minor stillstand around this time (French Polynesia; Baker and Haworth, 2000b).

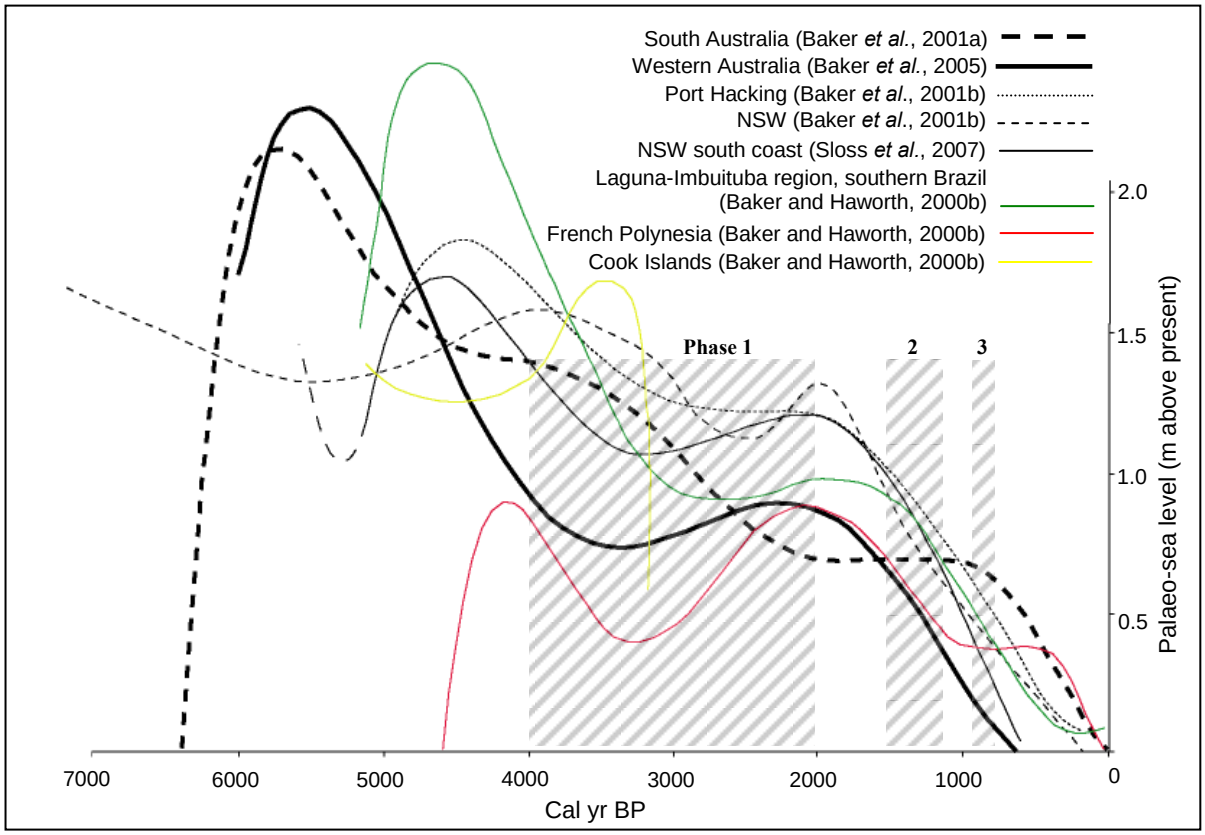

Figure 60. Examples of possible oscillating sea level curves developed for Australia, the Pacific Islands and Brazil with chenier ridge building phases (Section 6.2.5) superimposed.

Support for the influence of sea level oscillations on the development of the chenier plain is also found in the accelerated progradation rate in Princess Charlotte Bay after 1300 yr BP. This rate is around 1 m/yr from 1300 yr BP to present compared to 0.35 m/yr between 4000 yr BP and 1300 yr BP). Although there is less sea level data for this period, the curves are in agreement that it was generally a regressive period, and in muddy coastal environments this would have led to accelerated progradation rates.

The base levels of chenier ridges, which in the past have been used as a sea level indicator, also appear to generally match with the hypothesised changes in sea level. At 4000 yr BP when ridge 1 was built, sea levels would have been approximately 1.5 m higher than

present (Section 2.10.2.4), a figure that closely matches the height of the base of that ridge above datum (Figure 29). Moving seaward, the base levels of chenier ridges 2–4 (which correspond to the period between 4000 yr BP and approximately 2100 yr BP) then appear to reduce and elevate by around 0.5 m (Figure 60), a trend which is broadly consistent with the negative sea level oscillation hypothesised for that period. Moving further seaward, at around 1.1–1.3 m above datum, the base levels of chenier ridges 5–11 also broadly follow the hypothesised sea level trend from 2000 yr BP to the present, decrease in line with the hypothesised regression for that period (Figure 60). Although in themselves such agreement is not strictly evidence for the sea level oscillations model, it is still consistent with a sea level role.

The data certainly do not allow the acceptance of the sea level oscillations hypothesis outright, because there is no evidence for multiple oscillations that would have led to multiple ridges. Other issues that complicate the matter include the lack of ridge building prior to 4000 yr BP (given that the peak of the post glacial transgression was thought to be around 6000 yr BP) and the formation of the most seaward ridge after 820 cal yr BP during a time when sea level is generally thought to have been regressing. The evidence does, however, point to a possible modulating role of sea level changes. Based on this reasoning, the possibility that Holocene sea level oscillations could be incorporated into a new morphodynamic model for chenier plain development in Princess Charlotte Bay should be retained.

6.4 Reassessing the results of previous Australian chenier plain studies

Compiling and reassessing the results of previous Australian chenier plain studies was undertaken to reveal broader scale relationships or processes that may exist for chenier plains across a large region, and also to give insights as to how the processes that have operated at Princess Charlotte Bay compare and relate to other Australian chenier plains.

The method of reassessing the results of previous chenier plain studies was originally developed by Lees and Clements (1987). They compiled chenier ridge ages from a number of past studies and applied a randomisation procedure to determine the probability of „clumps" in chenier ridge ages. This was then used to determine the probability of periods of non-localised chenier ridge building across multiple north Australian sites. It was an innovative exercise which allowed them to both reassess previous data, as well as reveal broader scale relationships that might not have been obvious from looking at results individually.

Since they completed this, many new chenier ridge ages have been collected, and this has created the opportunity for a new, more thorough analysis to be carried out. The analysis undertaken in this study, however, includes far more chenier ridge ages (i.e. *all* available radiocarbon ages from previous Australian chenier ridge investigations) as well as some improvements to the statistical method. The following sections describe the method used in the Lees and Clements (1987) analysis, the new clump test and how it differs, the data used in analysis, the results obtained, and an evaluation of the clumping hypotheses.

6.4.1 Previous reassessment exercise: Lees and Clements (1987)

Lees and Clements (1987) used radiocarbon ages from six dated chenier ridge sequences in their analysis. Combining the individual chenier ridge ages from each individual sequence allowed them to identify age ranges in which chenier ridge building was occurring (e.g. combining individual chenier ridge ages of 2000 ± 200 yr BP and 2100 ± 100 yr BP would yield an interval of 2200–1800 yr BP). The age ranges from the six chenier ridge sequences used in Lees and Clements (1987) is shown in Figure 61. Patterns of chenier ridge building were then tested using a chi-square statistic and „Robinson" clump test statistic, defined as the maxima of sums of consecutive occurrence of chenier ridge building at three (or more) out of a possible six sites in any 200 year time interval (i.e. at least half the sites had a chenier formed during the time interval). The null hypothesis was that chenier ridges form by random, local events. They divided the continuous variable (time) into 200 year intervals, reasoning that this maximized the observable differences and, hence minimized the observed significant level of the chi-square statistic. The distribution of chi-square and „Robinson" clump test statistics were generated by randomising the starting points for age occurrence groups (i.e. the age ranges shown in Figure 61) with respect to time so that measures of statistical significance could be applied. They used a minimum age of 0 and maximum of 6600 yr BP as bounds and completed 1000 full randomisations. From each full randomisation, the occurrence totals in 200 year time intervals across the six chenier plains, the chi-squared, and the Robinson clump statistic were calculated (Lees and Clements, 1987). By completing this exercise, they identified 2800–1600 yr BP as a

significant period of chenier ridge construction and linked this to climate fluctuations (Section 2.5.1.3).

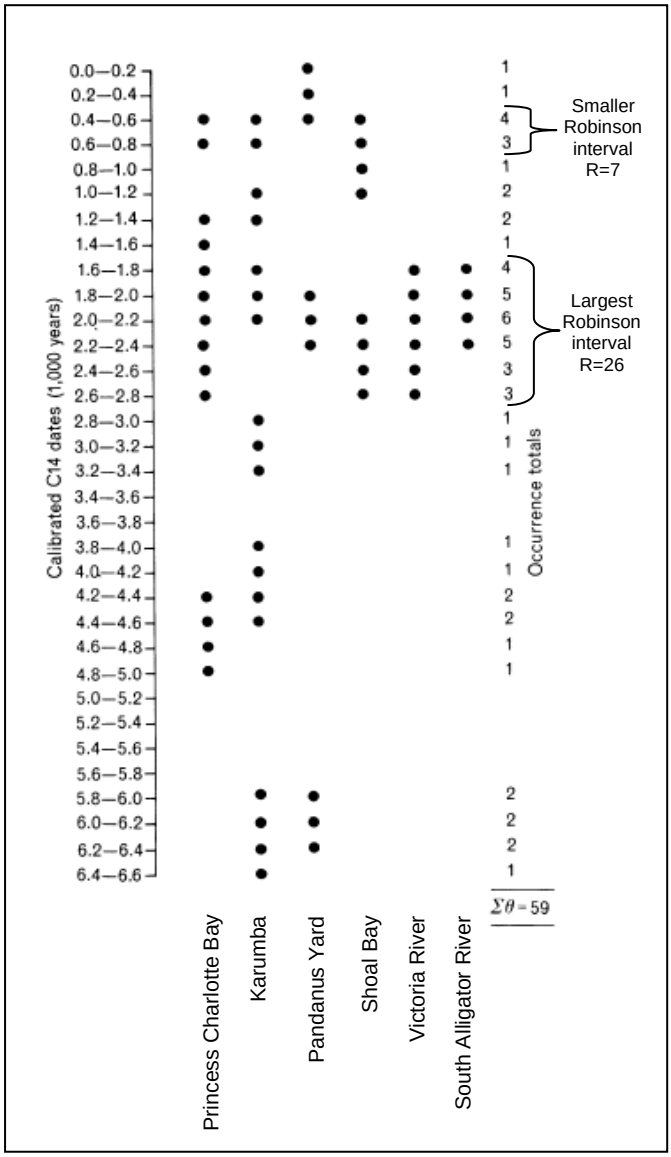

Figure 61. Occurrence patterns of chenier ridge formation used in the Lees and Clements (1987) analysis.

6.4.2 Clumping and the test statistic

The analysis procedure used in this investigation is described in detail in O"Neill (2009). It is similar to the Robinson clumping test used by Lees and Clements (1987), but the advent of greater computational power since that time has allowed some key improvements to be made. As in Lees and Clements (1987), occurrence groups for environmentally-corrected and calibrated chenier ridge age ranges were determined by the ranges of chenier ages from particular sites. The ranges used were the mean age ± 1σ. Although it is possible that the true age of chenier ridges lay somewhere outside this range, because the procedure requires data in the form of age ranges, it was thought that using ± 1σ ranges would produce more meaningful results than ± 2σ or greater, as larger ranges would increase the likelihood that the age ranges from different sites would overlap. In particular, unlike the Robinson test, the test used in O"Neill (2009) allows the age brackets to be determined by the data itself, using the start and end points of the age ranges (as shown in Figure 62). For this reason it was called the „exact method". The test also does not require any selection of a minimum level of clumping over the sites to allow measurement. Two advantages of this are that data accuracy is not lost by grouping the data into externally determined age brackets, and the results of analysis are not affected by externally-determined values. Determining the age brackets in this manner has implications for the strength of the test and interpretation of results, and these are discussed in a later section.

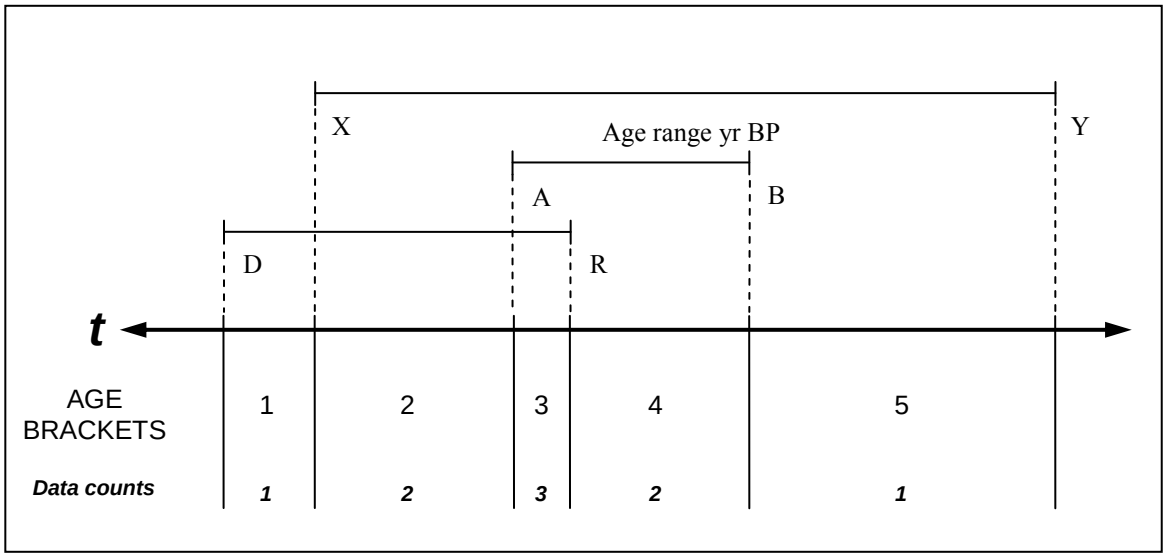

Figure 62. Conceptual diagram showing how the windows for analysis were determined by the data ranges, rather than setting an arbitrary grouping value. In this example, the ranges of three chenier ridge ages (X-Y yr BP, A-B yr BP and D-R yr BP) result in the age brackets (or windows) used in analysis being D-X yr BP (1), X-A yr BP (2), A-R yr BP (3), R-B yr BP (4) and B-Y yr BP (4). „Data counts" are the number of chenier age intervals which fall over an age bracket.

For any arbitrary set of values in the time interval, the „length" of the set and the „mass" of the set were measured with respect to the observed data. The length was measured by measuring the proportion of the time interval occurring within the set. The mass was measured as the proportion of the observed data intervals that fell over the set of interest. The ratio of the mass of the set to the length of the set then gave a measure of the intensity of the set: an intensity of one meant that the amount of observed data falling over this set was what would be expected if the data were exactly uniform over the entire range; a higher or lower value indicated more or less data (respectively) falling over this set than would be expected if the data were uniform over the entire range. In a set of data which was exactly uniform over the entire range, all sets in the range would have had an intensity of one. In a more clumped set of data, however, there would be some sets with high intensity values (greater than 1) and some sets with low intensity values (less than 1).

The intensity value enables a measure of the overall degree of „clumping" over the entire range of the data to be determined, clumping meaning chenier ages occurring concurrently across multiple sites. It was obtained by looking at the highest intensity value that occurred for any connected set with a stipulated minimum length, and then integrating the inverse of this value, which will be a value between zero and one, over all possible values of this minimum length (O"Neill, 2009). This lead to a clumping statistic between zero and one: a clumping value of zero occurred if the data were exactly uniform over the entire range, with higher clumping values indicating a greater level of clumping of the observed sets.

6.4.3 Hypothesis testing

The null hypothesis tested was that chenier ridges were formed by localised events that applied to only one site and did not affect chenier ridge building at other sites (i.e. that chenier ridge building across multiple sites was effectively a random process). Because localised events do not affect other sites, they would be expected to lead to a chenier ridge pattern with no probabilistic difference in the amount of chenier ridge building across multiple sites during different time periods. If this is the case then this would result in the length and expected mass (i.e., expected value of the mass) of sets being equal. This produces the following null and alternative hypotheses:

$\boldsymbol{H_0}$**:** For all sets: Length = Expected Mass

$\boldsymbol{H_A}$**:** For some set: Length ≠ Expected Mass

The null hypothesis in this case corresponded to the hypothesis that the chenier ridges were formed only by localised events. The alternate hypothesis is that at least some of the chenier ridges were formed by non-localised events. This could include processes that operated on much larger (e.g. continental) scales, thereby affecting multiple sites at the same time.

Regardless of which hypothesis is true, some areas in the data would be expected to have a higher or lower mass (actual mass, not expected mass) than the length. However, if the alternate hypothesis is correct, then there will be some periods with higher expected mass relative to their length. In this case it would be expected that some time periods did contain significantly more (or less) chenier ridge formation than others, compared to the null hypothesis. This led to a situation in which higher values of the clumping statistic used in O'Neill (2009) gave evidence in favour of the alternative hypothesis. By determining the distribution of this test statistic under the null hypothesis, the hypotheses could be tested through the observed *p*-value.

The null hypothesis was tested by using the observed data and randomising the starting points for chenier age occurrence groups with respect to time, thereby replicating the null hypothesis. Assuming sufficient randomisations were completed, the distribution of the clump statistic values obtained should have accurately approximated the true distribution of the clump statistic under the null hypothesis (i.e. the distribution of the clump statistic if chenier ridge building was a random process). Upon completion, this was assessed by visual inspection of the histograms and considering expected properties such as smoothness and unimodality. A minimum age of 0 and maximum of 6600 yr BP were used as bounds at

which chenier ridge age ranges were cut off so as not to bias clumping towards the centre of the timeline. The procedure was repeated 50 000 times to approximate the distribution of the test statistic under the null hypothesis using the statistical computing program R 2.4.0.

6.4.4 Data used in analysis

The Australian sites from which chenier ridge ages have been obtained are shown in Figure 63. The specific ages and calibrated ranges used in the analysis are shown in Tables 6 (northern Australia) and 7 (southern Australia). A representation of the combined age ranges from overlapping chenier ages at each site are shown in Figure 64. The randomisation procedure was applied to two subsets of the northern Australian data. One included all chenier ridge age ranges from northern Australia. The other included only the original data examined by Lees and Clements (1987) and was performed for comparative purposes.

With regard to the analysis on the Lees and Clements (1987) data, although they originally stated that all ages used had been environmentally corrected, this was not the case for the Woodroffe *et al.* (1985a) and Lees (1984) ages from the South Alligator River and Victoria Delta, respectively. Woodroffe *et al.* (1985a) justified using uncorrected ages by noting that the shells were estuarine in origin therefore the standard correction factor of 450 years may be inappropriate. In the light of past confusion regarding whether or not to apply this correction (Section 2.14.1), it has been applied to these ages for the present analysis, which includes the „new" northern Australian data, but left as the original (uncorrected, but

calibrated) ages for the re-analysis of the Lees and Clements (1987) data. Another difference in the data used was that Lees and Clements (1987) only used whole dated sequences (e.g. chenier ridge ages from Point Stuart were excluded in their original work as only alternate ridges were dated). In the new analysis, all available chenier ridge ages were included. All ages in both analyses were calibrated (Klein *et al.*, 1982).

Figure 63. Locations of areas from which the chenier ridge ages in Tables 4 and 5 have been obtained. Some areas contain more than one study site.

Table 6. Chenier ridge ages from northern Australia. Sites are presented in order as they are found from west to east. Ages used in Lees and Clements (1987) analysis are indicated with *.

Publication	Location	Chenier ridge ages (yr BP)	Lab codes	Corrected and calibrated ages (cal yr BP)	Chenier age range (cal yr BP)
O'Connor and Sullivan (1994)	Roebuck Bay	1140 ± 90*	WK2742	643 ± 98	740 – 545
		2810 ± 110*	WK2734	2428 ± 318	2745 – 2110
Lees (1992a)	Victoria Delta, Joseph Bonaparte Gulf	1880 ± 95	BETA5048	1860 ± 240	2100 – 1620
		1210 ± 135	ANU4886	1130 ± 230	1360 – 900
		1590 ± 140	ANU4887	1598 ± 283	1880 – 1315
		2760 ± 95	ANU4881	2955 ± 225	3180 – 2730
		2020 ± 290	ANU4884	2048 ± 663	2710 – 1385
Lees (1984)*	Victoria Delta, Joseph Bonaparte Gulf	2065 ± 345*	BETA5049	1590 ± 495	2085 – 1095
		2415 ± 315*	BETA5048	1955 ± 590	2545 – 1365
Chappell (1993)	Daly River, Anson Bay	3990 ± 90	ANU4514	3873 ± 248	4120 – 3625
		2830 ± 80	ANU4512	2438 ± 293	2730 – 2145
		2780 ± 80	ANU4511	2420 ± 295	2715 – 2125
		2180 ± 100	ANU4510	1657 ± 292	1949 – 1365
		2270 ± 90	ANU4509	1748 ± 213	1960 – 1535
Hickey (1981)*	Shoal Bay	670 ± 110	BETA1170	635 ± 95*	730 – 540
		850 ± 100	BETA1167	825 ± 190*	1015 – 635
		1060 ± 110	BETA1171	963 ± 213*	1175 – 750
		2350 ± 120	BETA1168	2425 ± 295*	2720 – 2130
		2340 ± 120	BETA1169	2425 ± 295*	2720 – 2130
Mulrennan and Woodroffe (1998)	Mary River, near Point Stuart	3690 ± 190	ANU6657	3480 ± 420	3900 – 3060
		3280 ± 340	ANU6658	2958 ± 578	3535 – 2380
		2800 ± 360	ANU7142	2355 ± 625	2980 – 1730
		2220 ± 70	BETA55848	1718 ± 213	1930 – 1505
		1030 ± 60	BETA55849	605 ± 90	695 – 515
Lees (1987)*	Point Stuart	420 ± 80	ANU5335	455 ± 155	610 – 300
		580 ± 90	SUA83	605 ± 90*	695 – 515
		1020 ± 100	ANU5334	908 ± 183	1090 – 725
		1210 ± 90	ANU5333	1135 ± 205	1340 – 930
		1275 ± 90	SUA82	1205 ± 155*	1360 –1050
		2745 ± 65	SUA82/2	2898 ±153*	3050 –2745
		1180 ± 100	ANU5332	1125 ± 205	1330 – 920
		1270 ± 100	ANU5331	1203 ± 158	1360 – 1045
		4040 ± 100	SUA81	4558 ± 288*	4845 – 4270
Woodroffe *et al.* (1985a)*	South Alligator River, Van Diemen Gulf	2120 ± 100*	ANUSS4064K	1630 ± 285	1915 – 1345
		1970 ± 70*	ANUSS4243K	1510 ± 190	1700 – 1320
Rhodes (1980)*	Pandanus transect	100 ± 85*	ANU1998	155 ± 155	310 – 0
		230 ± 80*	ANU1977	260 ± 260	520 – 0
		2070 ± 80*	ANU1827	2045 ± 295	2340 – 1750
		5380 ± 100*	ANU1691	6110 ± 270	6380 – 5840
	Karumba transect	630 ± 70*	ANU1745	618 ± 73	690 – 545
		1320 ± 80*	ANU1927	1220 ± 155	1375 – 1065
		1950 ± 75*	ANU1928	1920 ± 205	2125 – 1715
		2980 ± 60*	ANU1742	3108 ± 238	3345 – 2870
		3810 ± 105*	ANU1741	4248 ± 323	4570 – 3925
		5330 ± 95*	ANU1740C	6095 ± 270	6365 – 5825
		5540 ± 95*	ANU1740A	6223 ± 283	6505 – 5940
Chappell and Grindrod (1984)*	Princess Charlotte Bay	710 ± 70*	ANU2955	665 ± 55	720 – 610
		510 ± 80*	ANU2685	550 ± 105	655 – 445
		2170 ± 80*	ANU2687	2150 ± 210	2360 – 1940
		1900 ± 70*	ANU2689	1845 ± 125	1970 – 1720
		2460 ± 135*	ANU2458	2470 ± 320	2790 – 2150
		1670 ± 70*	ANU2954	1560 ± 165	1725 – 1395
		1540 ± 115*	ANU2327	1515 ± 190	1705 – 1325
		1960 ± 70*	ANU2949	1920 ± 185	2105 – 1735
		4030 ± 90*	ANU2694	4552 ± 289	4840 – 4263
Bird (1970)	Cairns Bay	5530 ± 130	GaK2676	5843 ± 248	6090 – 5595
Hopley and Murtha (1975)	Townsville	2350 ± 90	GaK1511	1903 ± 203	2105 – 1700
Cook and Polach (1973)	Charon Point, Broad Sound	4250 ± 125	ANU791	4215 ± 380	4595 – 3835
		3465 ± 110	ANU792	3215 ± 385	3600 – 2830
		2480 ± 100	ANU793	2033 ± 313	2345 – 1720

		1640 ± 100	ANU903A	1125 ± 230	1355 – 895
		1410 ± 90	ANU903B	885 ± 180	1065 – 705
		1430 ± 90	ANU903C	890 ± 180	1070 – 710
	Hoogly–Waverley Creek area, Broad Sound	5020 ± 90	ANU907	5278 ± 313	5590 – 4965
		3640 ± 80	ANU905	3430 ± 220	3650 – 3210
		2530 ± 70	ANU906	2045 ± 295	2340 – 1750
		2500 ± 70	ANU902	2038 ± 298	2335 – 1740
		1710 ± 70	ANU901	1198 ± 158	1355 – 1040
		1670 ± 70	ANU900	1143 ± 203	1345 – 940
		740 ± 70	ANU904	268 ± 268	535 – 0

Table 7. Chenier ridge ages from southern Australia. Sites are presented in order as they are found from west to east.

Publication	**Location**	**Chenier ridge ages (yr BP)**	**Lab codes**	**Corrected and calibrated ages (cal yr BP)**	**Chenier age range (cal yr BP)**
Short *et al.* (1989)	Tourville Bay	2010 ± 90	WK817	2025 ± 295	2320 – 1730
Belperio *et al.* (2002)	Franklin Harbour	2620 ± 85	WK762	2683 ± 293	2975 – 2390
		3230 ± 85	CS526	3495 ± 200	3695 – 3295
	Redcliff	3990 ± 105	ANU2335	4493 ± 333	4825 – 4160
		5060 ± 105	ANU2336	5830 ± 225	6055 – 5605
	Port Pirie	1320 ± 175	ANU10142	1270 ± 235	1505 – 1035
	Port Wakefield area	1910 ± 60	WK1123	1900 ± 180	2080 – 1720
		3340 ± 105	CS564	3605 ± 235	3840 – 3370
		3580 ± 95	CS563	3943 ± 298	4240 – 3645
		3350 ± 70	WK899	3655 ± 160	3815 – 3495
		1820 ± 60	WK895	1773 ± 153	1925 – 1620
		1820 ± 60	WK896	1773 ± 153	1925 – 1620
		3010 ± 85	WK897	3173 ± 318	3490 – 2855
	Port Gawler area	2880 ± 70	WK1116	3015 ± 195	3210 – 2820
		2070 ± 70	WK1115	2118 ± 203	2320 – 1915
		1790 ± 70	WK1114	1740 ± 170	1910 – 1570
		410 ± 70	WK1113	445 ± 100	545 – 345
		2020 ± 95	CS561	2030 ± 295	2325 – 1735
Thom *et al.* (1986)	Batemans Bay	2530 ± 100	SUA587	2043 ± 313	2355 – 1730

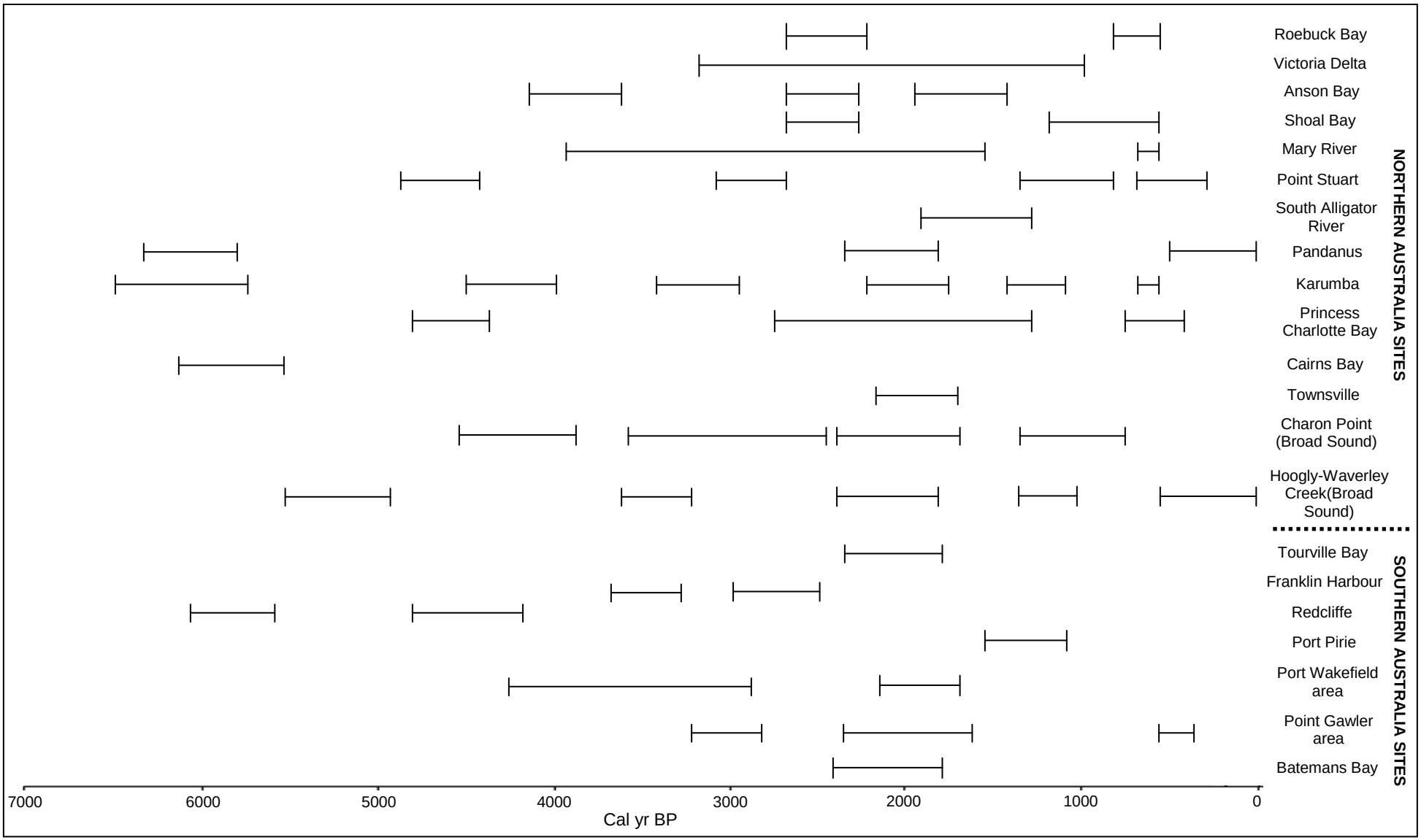

Figure 64. Diagrammatic representation of Australian chenier ridge age ranges used in the analysis. Specific values of each range are given in Appendix B. Sources are shown in Tables 6 and 7.

6.4.5 Results and evaluation of clumping hypotheses

Histograms of clump statistics generated under the null hypothesis, and occurrence values for the chenier age ranges for the Lees and Clements (1987), northern Australian and southern Australian datasets are shown in Figures 65, 66 and 67. The degree of clumping in each of the three datasets tested was measured by the test statistics shown.

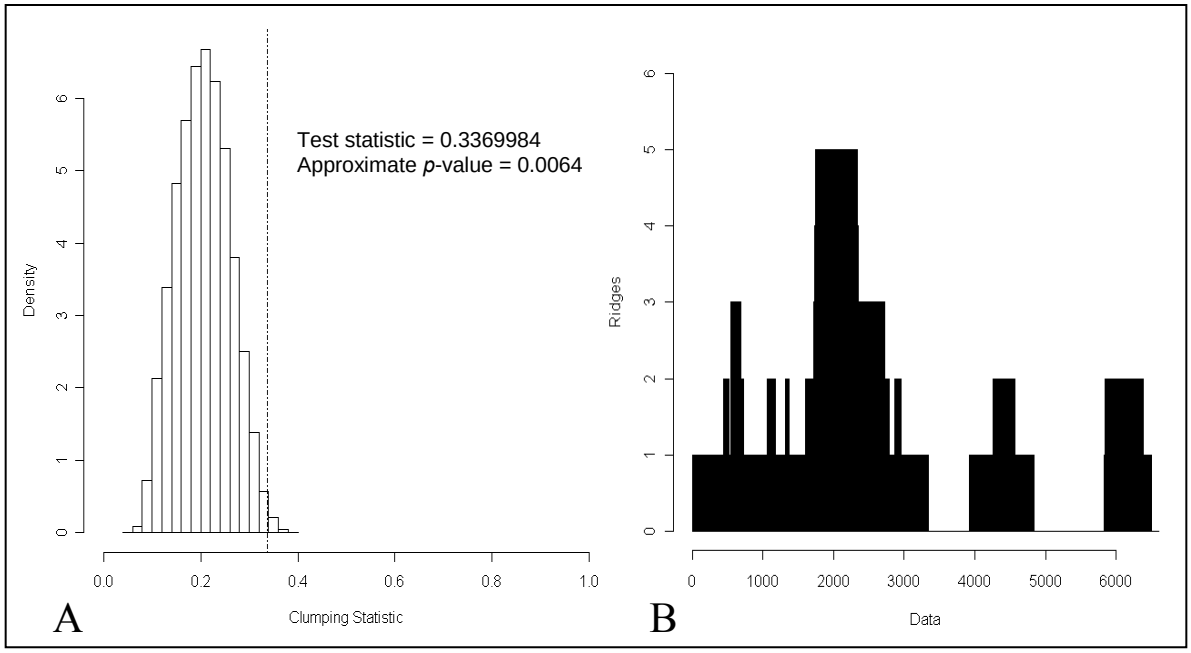

Figure 65. [A] Histogram of clump statistic values generated under the null hypothesis for chenier ridge ages from Lees and Clements (1987) sites. [B] Chenier age range occurrence along the 54 age brackets determined by the Lees and Clements (1987) combined chenier ridge age ranges.

The test statistic for the Lees and Clements (1987) data was a clumping statistic of 0.3369984. A clump statistic this high (or higher) occurred 320 times in 50 000 randomly generated datasets giving an approximated *p*-value of 0.0064. This low value provided strong evidence to reject the null hypothesis that chenier ridge development was affected solely by local events and accept the alternate hypothesis that wider scale events were at

least partly responsible. These results are similar to those obtained by Lees and Clements (1987) in that there is evidence that chenier ridge building was non-localised. The major peaks identified by Lees and Clements (1987) from 2800 to 1600 cal yr BP and from 400 to 800 cal yr BP, are also clearly visible in the occurrence histogram (Figure 65b), as well as minor earlier peaks.

Histograms of clump statistics generated under the null hypothesis, and occurrence values for the chenier ridge age ranges for the north and south Australian datasets are shown in Figures 66 and 67, respectively.

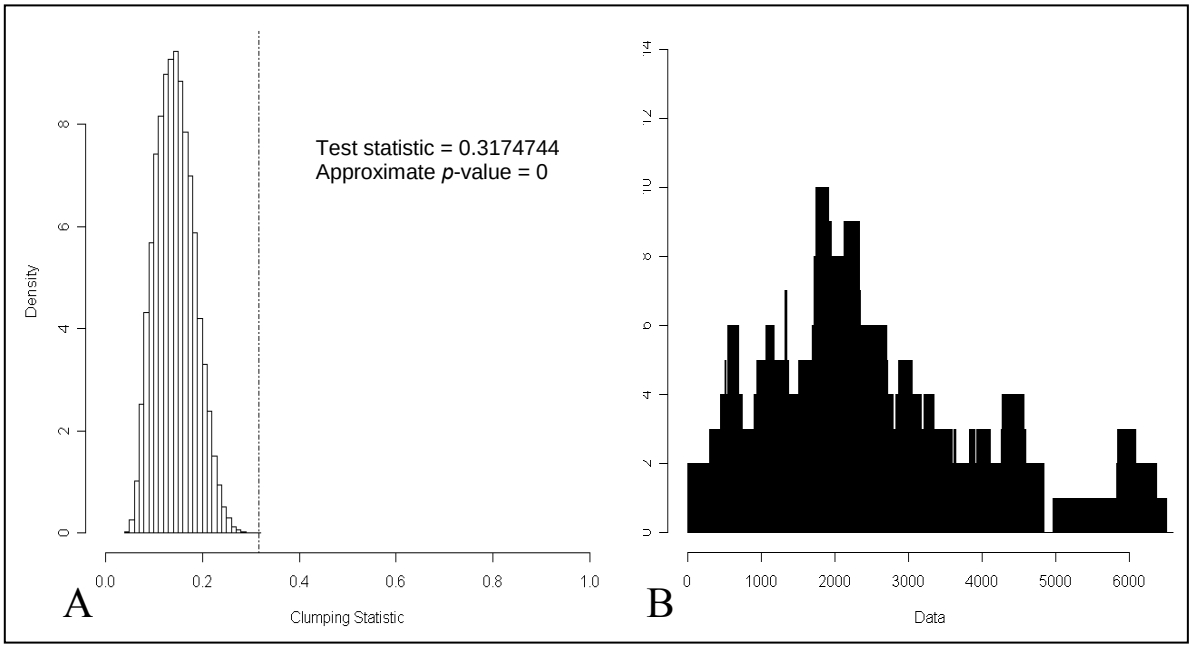

Figure 66. [A] Histogram of clump statistic values generated under the null hypothesis for chenier ridge ages from northern Australia. [B] Chenier age range occurrence along the 118 age brackets determined by the northern Australia combined chenier ridge age ranges shown in Figure 64.

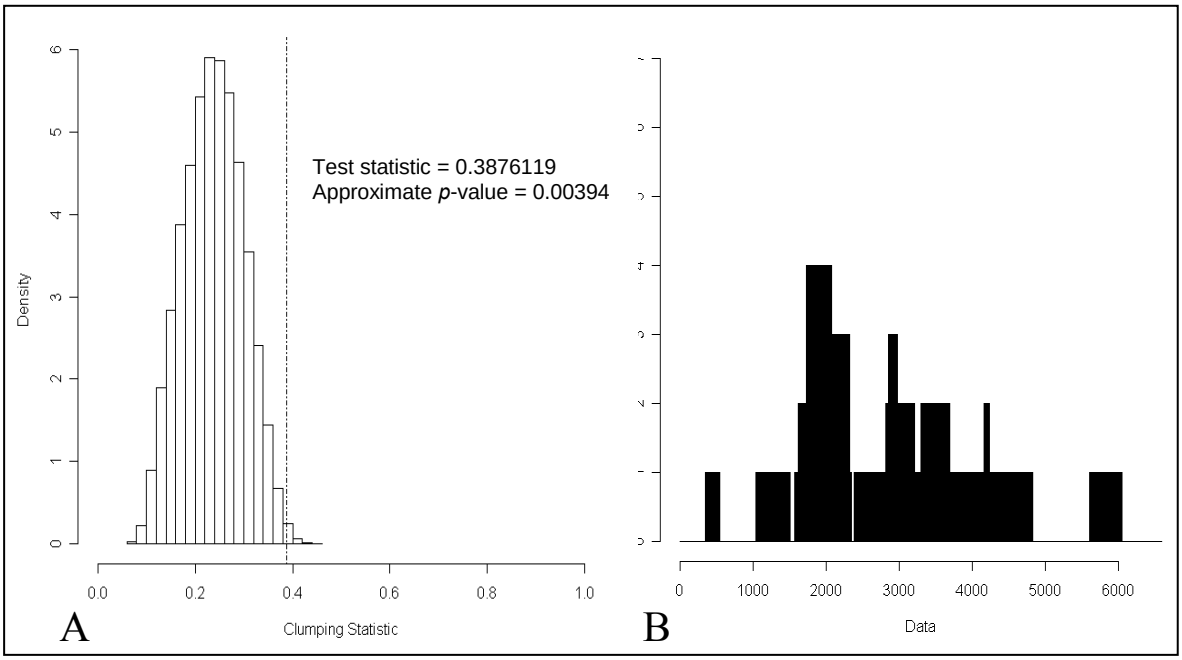

Figure 67.[A] Histogram of clump statistic values generated under the null hypothesis for chenier ridge ages from south Australia [B] Chenier age range occurrence along the 35 age brackets determined by the south Australian combined chenier ridge age ranges shown in Figure 49.

The test statistic of the northern Australian data was a clumping statistic of 0.3174744, and for the southern Australian data was a clumping statistic of 0.3876119. Clump statistics this high (or higher) occurred zero times in 50 000 randomly generated datasets (p = 0.00000) for the northern Australian data and 197 times (p = 0.00394) for the southern Australian data. In both cases, as with the Lees and Clements (1987) data, this provides strong evidence to reject the null hypothesis that chenier ridge development was affected solely by local events and accept the alternate hypothesis that wider scale events were at least partly responsible.

Although the clump test and randomisation procedure described did not allow the significance of individual chenier ridge building peaks to be evaluated (although this is an avenue for future research and is considered further in Chapter 7), the fact that distinct age

peaks clearly exist in the compilation of chenier ridge ages did allow a consideration of the processes responsible. Finally, there are potential issues in interpreting these data that should be considered.

6.4.6 Issues to consider in interpretation

Converting the chenier plain data from individual age ranges (Figure 64) to occurrence histograms (Figures 66b and 67b) was necessary for the purposes of clump testing. Although this served the tests well and allowed the clumping hypothesis to be successfully assessed, it involved a loss of information. This was important to consider when interpreting the results in terms of geomorphic events. For example, the histograms provided no information on which specific sites were forming chenier ridges. An issue related to this is derived from determining the age brackets using the start and end points of the individual age intervals (described in Section 6.4.2). Using 200 year age brackets, for example, meant chenier age ranges that were not overlapping, but within 200 years of each other, may be counted as overlapping (Figure 68). Using the current method of determining the age brackets by the start and end points of the age ranges (the exact method) meant that even if chenier ages were close together, they would not be counted as overlapping if they did not actually overlap.

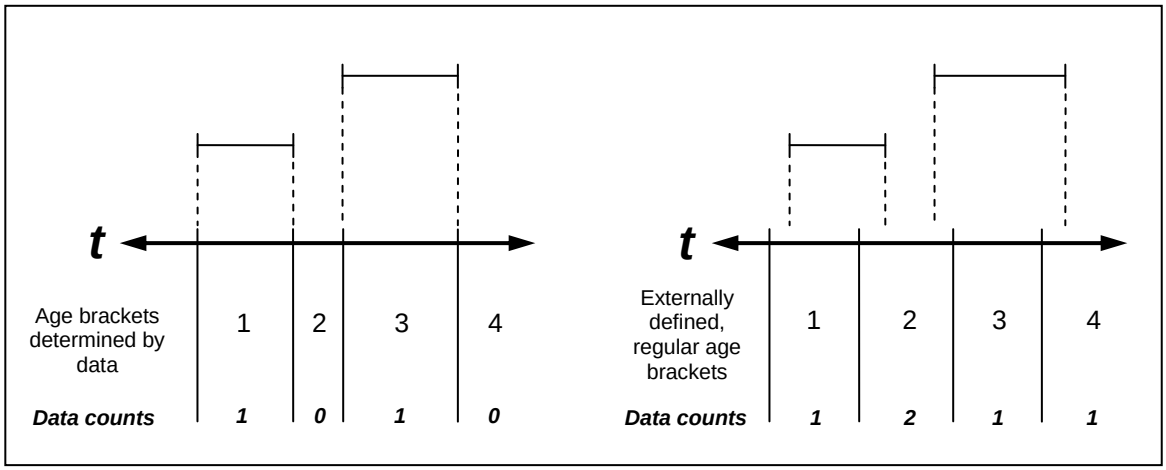

Figure 68. Conceptual diagram showing how the chenier ridge occurrence results can be affected by using either age brackets defined by the age ranges as in the new method (left hand side) or externally defined age ranges as in Lees and Clements (1987) (right hand side).

An example of how this affected the analysis can be seen by comparing the occurrence values for the Lees and Clements (1987) data to those generated using the exact method. Between 2000 and 2200 yr BP Lees and Cements (1987) found that chenier ridges formed at six out of six sites. The maximum using the exact method, however, was only five out of six sites. Although all six sites did have chenier ridges forming during the interval defined by the peak (between 2130 and 1750 cal yr BP), only five sites had ridges forming *at any one time*. In terms of the clump test, this provided a more stringent analysis and therefore more certainty that accepting the alternate hypothesis of non-localised events was correct. It could mean, however, that local and regional factors, which could add variability to the way in which a common event was recorded at every site, were not allowed for. These issues meant that although some interpretation was possible from the occurrence histograms, more information would be provided from a qualitative analysis of the age ranges across sites (i.e. Figure 64).

A final issue was related to radiocarbon dating of chenier ridges (Section 2.14). The potential to overestimate the ages of ridges (Section 2.14.2) means that theoretically, the radiocarbon age range obtained from any chenier ridge could be pushed forward in time to the „true" age. For the sake of the discussion, it was assumed that the error introduced by this was likely to exist, but did not prevent more fundamental relationships from being revealed.

6.4.7 Chenier ridge building peaks in the Australian data

In the north Australian data there appeared to be peaks of chenier ridge building around 620 cal yr BP (six sites forming chenier ridges simultaneously), 1120 cal yr BP (six simultaneous sites), 1850 cal yr BP (maximum 10 simultaneous sites) and 2230 cal yr BP (nine simultaneous sites). The final two of these could be considered sub-peaks of a more major phase of chenier ridge building lasting from around 2340 to 1700 cal yr BP. During this period 12 out of the 14 north Australian sites had chenier ridges forming, and the maximum number of sites with simultaneous chenier ridge building was 10. Other minor peaks occurred at around 2960 cal yr BP (five simultaneous sites), 4420 cal yr BP (four simultaneous sites) and 5960 cal yr BP (three simultaneous sites). The limited data from the south Australian sites appeared similar to this, with peaks around 1900 cal yr BP (four sites forming chenier ridges simultaneously) and possibly around 2910 cal yr BP (three simultaneous sites).

More generally, during the period between 2400 and 1300 cal yr BP there were chenier ridges formed at 19 out of the 21 sites that have been studied in Australia (Figure 64). Of

the two sites that did not have chenier ridge building occurring during this time, one was Cairns Bay where only the rear (oldest) ridge was dated, and Franklin Harbour in South Australia in which the record is much sparser at the outset. Widespread chenier ridge building has also occurred (albeit at a slightly lower rate) in the last 1000 years, with 11 out of the 21 Australian sites forming ridges during this time (Figure 64).

The timing of the occurrence peaks, as well as the more general periods of chenier ridge building were similar (but probably better constrained in time due to the use of more data) to those identified by Lees and Clements (1987) analysis (namely, a statistically significant period from 2800 to 1600 cal yr BP and possibly again between 800 and 400 cal yr BP).

6.5 Synthesis: Models for chenier plain development

Two important lessons can be synthesised from the results and discussion presented thus far. Firstly, it is likely that a model can be developed to explain the development of the Princess Charlotte Bay chenier plain that incorporates environmental controls not considered in previous studies. Second, the development of many chenier plains in Australia is likely to be influenced by common forcing factors operating on large spatial scales. These lessons are explained in the following sections.

6.5.1 Princess Charlotte Bay

The development of the Princess Charlotte Bay chenier plain was previously suggested to have resulted from an internal dynamic (Chappell and Grindrod, 1984; Section 2.5.1.8). While this concept may be useful to explaining the development of the chenier plain in Princess Charlotte Bay, and particularly in the eastern region investigated by Chappell and Grindrod (1984) a problem is that it is largely reliant on local factors. This reliance on local factors limits the applicability of the model to not only other areas of Princess Charlotte Bay (as is described in the following section), but also its ability to explain the development of other chenier plains, and consequently its ability to provide information about possible regional scale coastal geomorphological relationships.

The results of the reassessment of Australian chenier ridge ages in Section 6.4 suggest that chenier formation in Australia is unlikely to have been driven by localised processes. Furthermore, the apparent agreement between the ages obtained for the current transect and that examined by Chappell and Grindrod (1984) (Section 6.2.6) suggests at least some common chenier formation processes are likely to exist across the entire bay. The Princess Charlotte Bay chenier development model would be improved if it was capable of accounting for these observations. Specifically, wider scale processes need to be incorporated which allow common morphodynamic relationships across Princess Charlotte Bay, and possible other Australian chenier plains, to be recognised.

The following sections offer a critical assessment of the internal dynamic model in light of both new data and new palaeoenvironmental knowledge. Those considerations, as well as

the results of testing the six hypotheses for chenier plain development (Section 6.3) were then used to develop a new morphodynamic model for chenier plain development in Princess Charlotte Bay.

6.5.1.1 A critical assessment of the internal dynamic

Chappell and Grindrod (1984) identified three major contributing factors in the internal dynamic. These were (1) the effect of the mangrove fringe in focusing sedimentation by trapping shelly drifts and promoting mud deposition, (2) shell availability in source areas, thought to be affected by (3) the rate of previous muddy sedimentation (Section 2.5.1.8)

The first factor is the mangrove fringe. Although mangroves could certainly have assisted in focusing sedimentation at certain times, inspection of satellite imagery (discussed in Section 5.2.2) reveals that a significant mangrove fringe is not present at all locations. At the location examined by Chappell and Grindrod (1984), the mangrove fringe is between 100 m and 400 m wide, and in some locations up to 600 m. Moving westward along the shoreline, a fringe this wide persists until approximately 5 km beyond the mouth of the Normanby River, (approximately a third of the total distance along the shoreline). Moving further west, a discontinuous fringe between 100 m and 200 m wide persists until approximately 3.5 km beyond the mouth of the North Kennedy River (approximately two thirds of the distance along the shoreline). Beyond this point (including where the current study took place) there is no significant mangrove fringe present, and the only vegetation near the shoreline is that growing on the top and rear of the beach (e.g. Figures 16 – 23; Section 5.2.1). Allowing for the possibility that chenier ridge formation is not simultaneous

at all locations along the shoreline, if the mangrove fringe was an important factor in trapping chenier ridge sediment then we would expect to see a generally opposite trend in the extent of the incipient chenier forming at the shoreline as the one that just described for the mangrove fringe. This is not the case. While an incipient chenier is not visible at all locations along the shoreline, it is certainly visible in locations both with and without a mangrove fringe. The plausibility that the mangrove fringe would aid or play at least some role in trapping/focussing sediment notwithstanding, the current distribution of the fringe with respect to the extent of the incipient chenier ridge suggests that it is unlikely to be a *primary* controlling factor across the whole bay. Strong evidence for such a conclusion is also provided by differences in the nature of the underlying sediment between the ridge sequences in the eastern (mangrove facies) and western (grey intertidal muds) regions of the bay. These differences clearly show that the chenier ridges in the west have *not* formed in association with a significant mangrove fringe.

The second factor involved in the internal dynamic is shell availability in source areas. In the northeastern portion of the bay examined by Chappell and Grindrod (1984), the fact that chenier ridges are composed almost entirely of shell means that it is very likely shell availability has played a role there. Elsewhere in the bay, however, this role appears less likely. The concentration of shell material in beach and chenier ridge sediments appears to reduce in a westerly direction along the shoreline from their site, such that there is a mixture of quartz and shell in the chenier ridges in the area of the current, western, transect, and pure quartz forming the beaches on the western side (Frankel, 1974). This makes it very unlikely that shell availability is a major controlling factor across the entire bay.

The final factor involved in the internal dynamic is the „previous rate of muddy sedimentation". It is difficult to conceive of any other factor than climate fluctuations that would account for repeated fluctuations in the rate of muddy sedimentation through the late Holocene. As was discussed in Section 6.3.3, it appears that there is certainly scope for this to be included in a new morphodynamic model.

The above reasoning suggests that while the localised factors of mangrove distribution and shell availability may have contributed to the localised development of chenier ridges in the eastern region of the plain examined by Chappell and Grindrod (1984), they are unlikely to have directly controlled chenier development across the entire plain. A new morphodynamic model for chenier plain development in Princess Charlotte Bay is presented in the following section.

6.5.1.2 A new morphodynamic model

Inspection of imagery (Section 5.2.2), and comparisons between the results of Chappell and Grindrod (1984) and the current study (Section 6.2.6), indicates that while there are some broad similarities in ridge characteristics across the Princess Charlotte Bay chenier plain (primarily regarding timing of ridge building events), there are also differences in terms of the number, orientation, degree of preservation and dimensions of chenier ridges. This implies that there is likely to be considerable variation in the nature of morphodynamic processes across the bay. Put another way, it is likely that the processes that cause chenier ridge building have not operated at the same time, or at the same magnitude, at different locations across the bay. As was mentioned earlier, this observation highlights a

shortcoming of the concept of the internal dynamic: While it is useful to understand how processes in the *eastern* region of the plain may have interacted to result in chenier ridge building, it is not capable of explaining chenier formation across the entire bay.

In the sections that follow, a new model is presented for chenier ridge building in Princess Charlotte Bay. This model attempts to explain ridge building across the entire bay as a result of a number of environmental factors. The factors used include some of those that were tested as individual hypotheses (and not rejected) in Section 6.3, as well as some that contribute to the internal dynamic (Section 2.5.1.8), and is therefore a combination of models. The possibility of this was noted in Section 3.2.7. It is proposed that these forces form part of a „morphodynamic hierarchy", and that certain combinations of factors in this hierarchy have resulted in periods of optimal chenier ridge building conditions. Central to the concept of a morphodynamic hierarchy is that there are a number of environmental controls acting on different temporal and spatial scales. Those higher up in the hierarchy (e.g. first order and second order controls) effectively act as templates for those lower in the hierarchy, and have interacted to result in periods of „optimal chenier ridge building conditions". In this way, it is possible to explain both broad trends across the bay, as well as smaller scale differences in the number, orientation, degree of preservation, and dimensions of chenier ridges.

The internal dynamic, is itself a type of hierarchical model, in that it is primarily dependent on rate of previous sedimentation as the „underlying template" which influences shell availability. The influence of the mangrove fringe trapping sediment is then superimposed on these factors to result in periods of chenier ridge building. The new model presented is similar to this, and involves four orders of environmental controls.

6.5.1.3 The morphodynamic hierarchy

In order from first order to fourth order, the environmental controls that are included in the morphodynamic hierarchy are: (1) Holocene climate and sea level changes, (2) tropical cyclone activity, (3) mangrove distribution, and longshore currents, and (4) river channel changes. The concept of „optimal chenier ridge building conditions" is effectively the same as the concept of „cut and recover" periods put forward by Chappell and Grindrod (1984) in that they are the conditions under which chenier ridges are most likely to be built. While the nature of interacting environmental controls means that it is possible for chenier ridge building to occur outside of these „optimal" conditions (e.g. for ridge 11), it is nonetheless during these conditions that the majority of chenier ridges in the bay appear to have been built.

The first order controls: Holocene climate and sea level changes

Testing the hypotheses for climate fluctuations and sea level oscillations resulted in the conclusion that while neither of them appeared to have been *directly* responsible for determining the times at which chenier ridges were built, they appeared to have played an overall modulating role. A modulating role for climate change appears to be present because chenier ridges began forming in Princess Charlotte Bay around the time that climate became more variable (Section 6.3.3). A similar role appears to exist for sea level oscillations because chenier ridge building appears to have been more frequent during the periods of oscillation from as early as 6000 yr BP until around 2000 yr BP (i.e. when there

were stillstands or transgressive phases), while the phase of regression after 2000 yr BP has tended to have encouraged progradation and shoreline advancement (Section 6.3.6).

Considering climate fluctuations and sea level changes as the first order controls is therefore appropriate because they appear to have acted as a template upon which lower order controls have affected their influence. If true, then the wide spatial influence of sea level changes means it is likely that other chenier plains around Australia would also be subject to its influence (i.e. that it would also be a first order control for other chenier plains). The agreement in timing of chenier ridge building at chenier plains around Australia (Section 6.4.7) could therefore be due to the existence of such a control, as there are no other environmental forces capable of exerting a simultaneous influence at such a large spatial scale. This is considered further in following sections.

Out of all the models considered, climate fluctuations and sea level changes have the widest spatial influence. They are therefore likely to be responsible for the broad similarities in ridge characteristics (Section 5.2.2), and timings of ridge formation (Section 6.2.6) across the bay. Climate fluctuations and sea level changes alone, however, cannot account for other observations that sit below these broad trends. For example, the number of chenier ridges in the western region of the Princess Charlotte Bay plain far exceeds the number of sea level oscillations or climate fluctuations that are thought to have occurred since around 6000 yr BP. Furthermore, neither of these controls can account for the fact that during the late Holocene around twice as many chenier ridges were built in the west compared to the east (Section 6.2.6). It is proposed that the second order control is primarily responsible for these differences.

The second order control: tropical cyclone activity

The second order control in the morphodynamic hierarchy is tropical cyclone activity. As a control acting on the template provided by climate fluctuations and sea level changes, tropical cyclone activity is likely to have contributed to chenier ridge building in three main ways. Firstly, tropical cyclones have generated large waves which have allowed ridge crests to accrete to heights far beyond the reach of normal wave processes (Section 6.3.4). Secondly, the frequency at which large waves (generated by tropical cyclones) are hypothesised to have occurred in Princess Charlotte Bay may offer a reason as to why multiple ridges develop during broader phases of ridge building. Thirdly, such large waves would not have affected all parts of the shoreline equally, and this has contributed to differences across the bay. The first of these was explained in Section 6.3.4. The other two are explained below.

Aside from simply generating waves large enough to accrete chenier ridges, tropical cyclones may have played a role in the development of multiple ridges during phases of ridge building. Evidence for this possible role comes from the observation that, acknowledging a small possibility of circular logic from the records used, the frequency of ridge formation during phases of ridge building is similar to the return interval of major tropical cyclones (Section 6.3.4). As the grain size of the sediment making up the chenier ridges in Princess Charlotte Bay indicates that they were undoubtedly deposited by waves, the height to which ridges are able to accrete would be largely determined by the maximum height of wave run-up. This has been noted in previous studies linking the evolution of coarse-grained ridges to cyclone events (e.g. Forsyth, 2010). Once ridge crest reaches the

maximum height to which waves are able to transport sediment, ridge accretion ceases. Under a regime of constant sediment input and at times during which the optimal chenier building conditions are present, a new ridge would then be initiated on the seaward side of the larger ridge and the process would occur once again (and would continue until the optimal chenier ridge building conditions ceased, for example, following a sea level regression or pluvial climate phase).

The second way that tropical cyclone activity may have contributed to the development of the chenier plain comes from the fact that for a given cyclone event, there would be longshore variations in wave activity. For example, the positions of the current western transect and those examined by Chappell and Grindrod (1984) were shown in Figure 14 (Section 4.2.5). Although the western shoreline is by no means completely protected, its location relative to Bathurst Heads and the Flinders Islands means it would be offered a degree of protection (especially from cyclones occurring to the east and north east of Princess Charlotte Bay) that is not afforded to the western shoreline. This protection would have resulted in longshore variations in the degree to which optimal chenier building conditions were achieved, and may have contributed to the result that around twice as many ridges formed in the western region of Princess Charlotte Bay as in the east.

The concept of tropical cyclone activity superimposed on the broader influence of sea level changes and climate fluctuations can broadly explain the development of the chenier plain from when ridges began to be built at 4000 yr BP until ridges 6 and 7 were built at around 2000 yr BP. Depending on the evidence used this may extend to possibly until ridges 8 and 9 were built at around 1350 yr BP. The development of the most recent chenier ridge (11)

since around 820 cal yr BP may be due to a minor arid climatic phase (as postulated by Mayewski *et al.* (2004), or by a number of authors considering changes related to the little ice age or medieval warm period climate phases; Section 2.9.2.3), a minor still stand in sea level as suggested by Baker and Haworth‟s (2000b) oscillating curve from French Polynesia (Section 6.3.6), combined with a return to a more intense tropical cyclone regime (Forsyth, 2010).

The third order controls: longshore currents, mangrove distribution and shell availability

Superimposed on the first and second order controls are third order controls of longshore currents, mangrove distribution and shell availability. In contrast to the first order controls (that have largely provided the template under which either chenier ridge building or muddy progradation could occur), and second order controls (that have been responsible for actually building ridges), the third order controls may have influenced the longshore physical characteristics of chenier ridges, enhanced accretion at particular times, and influenced the composition of ridges (respectively). Third order controls are far more restricted in their spatial influence and contribute to spatial differences in ridge characteristics across the chenier plain.

The ridges on the western side of the Princess Charlotte Bay chenier plain appear to be more continuous and oriented in more parallel lines than those to the east (Section 5.2.2). A possible reason for this is that longshore currents are much stronger in this portion of the bay than elsewhere (Figure 14; Section 4.2.4), and they may act as a distribution mechanism for chenier ridge sediment as it is being accreted into ridges.

Although mangrove populations are unlikely to be necessary for ridge formation everywhere in the bay (Section 6.5.1.1), it is plausible that they have played a role hypothesised by Chappell and Grindrod (1984). A similar view was previously put forward by Grindrod (1985) who stated that although mangroves may promote sedimentation and sediment stability, their importance in this regard is minor in comparison to broader geomorphic processes which largely dictate coastline morphologies.

Finally, shell availability is likely to have played a role in determining the differences in chenier ridge composition around the bay (i.e. the change from predominantly shell in the east, to a mixture of shell and sand in the central areas, to pure quartz in the west and north).

The fourth order control: river channel migration

Although there is no evidence that the model of river channel *switching* (as applied in Louisiana or some chenier plains in China) has played a role in the construction of the chenier plain, the position of river mouths and changes through time is likely to have affected the current ridge distribution and degree of ridge preservation in Princess Charlotte Bay. Examples of this include the lack of preserved ridges through the central part of the bay, and localised features such as minor truncations and recurved spits (Section 5.2.2). The more continuous ridges in the west compared to the east may also be the result of more rivers existing in the east, which has increased the potential for reworking in that region.

6.5.2 Chenier plain development in Australia

The results of the reassessment of previous Australian chenier plain ages strongly suggested that chenier ridge building events in Australia are unlikely to have been driven by localised processes, and that wider scale drivers were at least partly responsible (Section 6.4.7). Although the exact timings of ridge building and the number of ridges built during the late Holocene clearly differs between sites to an extent (which is to be expected given differing morphosedimentary contexts and process climates) these differences would appear to be subordinate to other (common) drivers. Immediately, this would appear to fit nicely within the parameters of the proposed morphodynamic hierarchy: the wider scale common drivers being either of the first order controls (i.e. sea level changes or climate) and other more localised controls acting on these templates.

The ridge building peak that can be identified with most certainty, and one which is evident in data from both northern and southern Australian chenier plains, is the major phase of chenier ridge building lasting from around 2400 and 1300 cal yr BP (Section 6.4.7). During this period chenier ridges were built at 19 out of the 21 Australian sites, and 12 out of the 14 north Australian sites. While there are earlier peaks than this, these must be interpreted with caution due to the increasing possibility of overestimates as radiocarbon ages get older. While it is certainly possible that this major peak contains overestimates, the fact that it is evident in so many sites adds confidence that it represents an actual period of common ridge building, and would seem to be a reasonable position compared to the alternative situation in which overestimated ages happened to coincide at 19 of 21 sites spread around

the Australian coastline. To a lesser extent there also appears to be a more recent peak at around 600 cal yr BP, although this is not as evident in the South Australian data.

In the past these trends were explained in terms of climate fluctuations (Lees and Clements, 1987; Lees, 1992b). While climate fluctuations may be put forward as a possible explanation for ridge building events at particular (northern Australia) sites, however, it would be inappropriate to suggest that such fluctuations could be responsible for the broad peaks of Australian ridge building identified in the analysis. There are two reasons for this. Firstly, based on the Holocene climate reconstructions for northern Australia, the major phase of chenier ridge building lasting from 2400 and 1300 cal yr BP is around 1000 years later than the change to variable conditions, which occurred after 3500 yr BP (Section 2.9.2.1). Second, it is highly likely that the climate signal would differ greatly between sites (particularly northern and southern Australian sites).

As was proposed in the morphodynamic hierarchy for Princess Charlotte Bay, a possible explanation for the common peaks of ridge building is that late Holocene sea level changes have acted as a first order control to modulate other regional and local processes. Indeed, it is difficult to conceive of any other force apart from sea level changes that could be responsible for such widespread and simultaneous periods of chenier ridge building. If true, it suggests that all Australian chenier plains may be subject to a similar morphodynamic hierarchy, with common relationships derived from the first order controls, and differences derived from lower order controls acting on this template. This in turn may provide an avenue for the incorporation of chenier plains into general models of coastal behaviour around Australia.

Chapter Seven - Conclusion

7.1 New insights gained from this research

Prior to this research, although many chenier plains had been investigated across northern Australia, problems with dating methods and a lack of palaeoenvironmental records for comparison meant that knowledge gained from chenier plain studies was not able to be incorporated into general models of coastal behaviour around Australia. New knowledge and dating methods mean that these problems are now less significant, and chenier plains are once again a fruitful topic for research.

The primary aim of this research was to investigate the Holocene development of the Princess Charlotte Bay chenier plain, and the secondary aim was to consider the broader formation of several chenier plains in Australia and identify possible relationships. The research was undertaken using new data from a chenier ridge sequence on the western side of the Princess Charlotte Bay plain, data obtained from reassessing results of previous Australian chenier plain studies, and considering these in the context of new palaeoenvironmental knowledge.

Undertaking this research has yielded several new insights for studies of chenier plains and Australian coastal geomorphology. These are described in the following sections. The new insights relate to general models of chenier plain development in a worldwide context, the

Holocene development of the Princess Charlotte Bay chenier plain, and relationships between Australian chenier plains.

7.1.1 General models of chenier plain development

Many authors have emphasised the need for a fluctuating sediment supply in the development of chenier plains (e.g. Curray, 1969; Hoyt, 1969; Cook and Polach, 1973; Rhodes, 1982; Augustinus, 1989; Short, 1989). While this is indeed a contributing factor in many settings, it was shown that a more fundamental condition for the formation of chenier plains is a fine balance between „progradation" and „accretion" forces (Section 2.6). Alternations in this balance result in intervals of coarse sediment deposition and chenier ridge building (when progradation force < accretion force) and progradation of unconsolidated mudflat (when progradation force > accretion force). The seven models for chenier plain development (river channel switching, mud shoal migration, climate fluctuations, storm activity, human impacts, sea level oscillations, and an internal dynamic) represent different ways that this balance can be changed. There is no evidence that chenier ridges are emplaced by tsunami.

This new way of explaining the development of chenier plains can greatly simplify some past theories and classifications regarding chenier plains (e.g. Otvos and Price"s (1979) distinction between „bay head" and „bight coast" chenier plains), and was applied in the research design for this investigation (Chapter 3).

7.1.2 The development of the Princess Charlotte Bay chenier plain

The ages obtained from western region of Princess Charlotte Bay indicated that the ridge sequence began forming at around 4000 yr BP. In the ~1500 years following this, two more chenier ridges were built (ridges 2 and 3). Ridge 4 was built at around 2950 ± 550 yr BP. Ridge 5 was formed between 2100 ± 300 and 2200 ± 400 yr BP. Ridge 6 was built between 2270 ± 140 and 2200 ± 200 yr BP. Ridge 7 was built between 2230 ± 140 and 1900 ± 200 yr BP. Ridge 8 was built between 1420 ± 100 and 1300 ± 200 yr BP. Ridge 9 was built between 1380 ± 90 and 1300 ± 200 yr BP. The calibrated and corrected radiocarbon ages provided an upper limit of the age of ridge 11 at 820 ± 60 cal yr BP.

Three general phases of chenier ridge building were identified from these ages: (1) An early phase lasting from when the ridge sequence began to form (around 4000 yr BP) until around 2000 yr BP in which at least seven chenier ridges (1 to 7) were built; (2) a phase centred on approximately 1350 yr BP in which at least two ridges were built (8 and 9); and (3) a final phase since approximately 820 yr BP in which one final ridge (11) was built. It is likely that the system is presently in a phase of ridge building.

The timings of ridge building were broadly similar from both this (western) transect and the region studied by Grindrod (1983) and Chappell and Grindrod (1984). The major difference between the sequences was that around twice as many ridges appear to have formed in the west as in the east during the period since 4000 yr BP.

In considering a Holocene development model for the Princess Charlotte Bay chenier plain, it was concluded that while the concept of an internal dynamic may be useful, a problem is that it is largely reliant on local factors. This reliance limits the applicability of the model within Princess Charlotte Bay, to the development of other chenier plains, and consequently its ability to provide information about possible regional scale coastal geomorphological relationships. Testing the hypotheses for chenier plain development models revealed the likelihood that several environmental factors may have played a role in the development of the chenier plain. It was proposed that these forces form a morphodynamic hierarchy, and particular combinations have resulted in periods of „optimal chenier ridge building conditions". The higher order forces have the broadest influence and act as templates on which the lower order forces exert their influence. In this way, it was possible to explain both broad trends across the bay, as well as smaller scale differences in the number, orientation, degree of preservation and dimensions of chenier ridges.

In order from first order to fourth order, the environmental controls that were included in the morphodynamic hierarchy are: (1) Holocene climate and sea level changes, (2) tropical cyclone activity, (3) mangrove distribution, and longshore currents, and (4) river channel changes.

7.1.3 Relationships between Australian chenier plains

Reassessing ages from previous Australian chenier plain studies indicated that wider scale drivers were at least partly responsible for common ridge building events. The most significant of these events was a major phase of chenier ridge building lasting from around 2400 and 1300 cal yr BP. During this period chenier ridges were built at 19 out of the 21 Australian sites, and 12 out of the 14 north Australian sites. Although the exact timings of ridge building and the number of ridges built during the late Holocene clearly differed between sites to an extent, the likelihood of common, wider scale drivers appeared to fit within the parameters of the proposed morphodynamic hierarchy: the wider scale common drivers being either of the first order controls (i.e. sea level changes or climate) and other more localised controls acting on these templates.

In the past these trends were explained in terms of climate fluctuations (e.g. Lees and Clements, 1987). In light of new palaeoenvironmental knowledge, however, a more plausible explanation for common ridge building events is that sea level changes have acted as a first order control to modulate other regional and local processes. This opened up the possibility that all Australian chenier plains may be subject to a similar morphodynamic hierarchy, with common relationships derived from the first order controls, and differences derived from lower order controls acting on this template.

Although requiring further research, it is possible that these conclusions may provide an avenue for the incorporation of chenier plains into general models of coastal behaviour

around Australia as expressions of common geomorphological drivers acting in a morphodynamic hierarchy.

7.2 Future research directions

This research has yielded a number of avenues for future research. These are related to the use of OSL for dating chenier ridges, the status of chenier plains in general coastal geomorphological models in Australia, the record of coastal change that is preserved at Princess Charlotte Bay, and the statistical method used in the clump analysis of chenier ages.

Although the comparison was limited, the preliminary data suggested agreement between chenier ages obtained using OSL and radiocarbon methods. Given this agreement, and the theoretical considerations regarding the potential error introduced by radiocarbon, the possibility that OSL could be applied to yield more reliable chenier ridge ages should be explored in the future.

It was proposed that chenier plains might fit into general models of coastal geomorphology in Australia as expressions of common geomorphological drivers acting in a morphodynamic hierarchy. Although relationships were identified between many chenier plains around the Australian coastline that support this, the potential for overestimation when using radiocarbon ages precluded detailed interpretation in terms of specific peaks in the data. This could be overcome by having more certainty that chenier ridge ages were not

overestimates. This could potentially be gained by redating some chenier plains using OSL and comparing the ages to those obtained using radiocarbon. It is predicted that some older Holocene ages (i.e. those that have contributed to the pre 2400 yr BP peaks in the northern Australia analysis; Section 6.4.5) would prove to be overestimates. More certainty in chenier ridge ages would help to constrain the timings of the more recent peaks of chenier ridge formation by reducing the potential for overestimates and allow the differences between timings of chenier ridge formation between sites to be more thoroughly assessed. This would also allow an estimation of the „noise" introduced into each chenier plain record by lower order controls (i.e. local and regional influences), which would contribute to the differences observed between sites.

There are a number of avenues for further research in Princess Charlotte Bay. A question that was not addressed in this study was the reason for the apparent difference in the sediment composing chenier ridges 1–5 and 10 compared to ridges 6–9 and 11 (Section 5.4). In the absence of ages suggesting otherwise it would be reasonable to conclude that this difference was the result of ridges of different ages being subject to different amounts of weathering. The similarity of the ages (for example between ridges 5 and 6) indicates that this is not the case here. The reason may therefore lie in the *source* of sediment used to construct the ridges. Answering this question requires further research. Although not examined in detail, this investigation also revealed the existence of older deposits in Princess Charlotte Bay. The distance of these deposits inland on a progradational plain suggests that they are very old, and based on the ages obtained for the Holocene sequences they are almost certainly pre-Holocene. This in turn suggests that a very long and complex history of coastal change is recorded in Princess Charlotte Bay, and one that may extend

over several glacial periods. In a region where few good records of previous interglacial sea levels exist, the prospect that long term studies of coastal change could be undertaken in Princess Charlotte Bay is exciting.

Finally, the statistical method used in the clump test could be developed further to allow the significance of individual peaks of chenier ridge formation to be assessed. It may also be interesting to apply the procedure in a more global context and assess the relationship between chenier plains around the world. Although regional and local variations in processes would add considerable „noise" to such an analysis it may provide a unique opportunity to identify coastal features that may have developed primarily under a single global influence (i.e. eustatic sea level changes).

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Appendix A. GPS co-ordinates of locations mentioned in text

The specific coordinates of various features within Princess Charlotte Bay that are mentioned in Sections 5.3 and 5.4 are given in the table below.

Section	Figure and page	Feature	Latitude (S)	Longitude (E)
5.3	Figure 24, p 147; Figure 25, p148	Exposed palaeo surface in river bed	14°31’54.72’’	143°49’36.15’’
	Figure 25, p148	River bed sand sample location	14°29’56.43’’	143°49’30.92’’
	Figure 25, p148; Figure 27, p150	Rubbish pit	14°32’12.79’’	143°51’41.40’’
5.4	Figure 28, p153	Landward extent of transect	14°28’48.20’’	143°49’35.34’’
		Pivot 1*	14°28’17.82’’	143°50’12.19’’
		Pivot 2*	14°28’12.04’’	143°50’05.43’’
		End of transect at coastline	14°27’41.87’’	143°50’35.38’’
		Base station 1	14°28’48.21’’	143°49’35.42’’
		Base station 2	14°28’09.08’’	143°50’10.67’’
		Dating sample pit ridge 6	14°28’27.09’’	143°50’01.39’’
		Dating sample pit ridge 7	14°28’26.29’’	143°50’02.29’’
		Dating sample pit ridge 8	14°28’23.46’’	143°50’05.14’’
		Dating sample pit ridge 9	14°28’22.29’’	143°50’06.94’’

		Dating sample pit ridge 11	14°28’02.07’’	143°50’17.91’’

* Pivot points are numbered as they are encountered moving seawards along the transect.

Appendix B. Combined chenier ridge age ranges used in clump tests

The age ranges shown in the table below were represented graphically in Figure 64, Section 6.4.4. They were derived from the individual age ranges shown in Tables 6 and 7 in that section.

Author	Site	Age range(s) (cal yr BP)
O'Connor and Sullivan (1994)	Roebuck Bay	740 – 545
		2745 – 2110
Lees (1984, 1992)	Victoria Delta, Joseph Bonaparte Gulf	900 – 3180
Chappell (1993)	Daly River, Anson Bay	1960 – 1365
		2730 – 2125
		4120 – 3625
Hickey (1981)*	Shoal Bay	1175 – 540
		2720 – 2130
Mulrennan and Woodroffe (1998)	Mary River, near Point Stuart	695 – 515
		3900 – 1505
Lees (1987)*	Point Stuart	695 – 300
		1360 – 725
		3050 – 2745
		4845 – 4270
Woodroffe *et al.* (1985)*	South Alligator River, Van Diemen Gulf	1915 – 1320
Rhodes (1980)*	Pandanus transect	520 – 0
		2340 – 1750
		6380 – 5840
	Karumba transect	690 – 545
		1375 – 1065
		2125 – 1715
		3345 – 2870
		4570 – 3925
		6505 – 5825
Chappell and Grindrod (1984)*	Princess Charlotte Bay	720 – 445
		2790 – 1325
		4840 – 4263
Bird (1970)	Cairns Bay	6090 – 5595
Hopley and Murtha (1975)	Townsville	2105 – 1700
Cook and Polach (1973)	Charon Point, Broad Sound	1355 – 705
		2345 – 1700
		3600 – 2380
		4595 – 3835
	Hoogly-Waverley Creek area, Broad Sound	535 – 0
		1355 – 940
		2340 – 1740
		3650 – 3210
		5590 – 4965

Short *et al.* (1989)	Davenport Creek , Tourville Bay	2320 – 1730
Belperio *et al.* (2002)	Franklin Harbour	2975 – 2390
		3695 – 3295
	Redcliff	4825 – 4160
		6055 – 5605
	Port Pirie	1505 – 1035
	Port Wakefield area	2080 – 1620
		4240 – 2855
	Port Gawler area	545 – 345
		2325 – 1570
		3210 – 2820
Thom *et al.* (1986)	Batemans Bay	2355 – 1730