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Post-glacial sea-level changes around the Australian margin: a review

Stephen E. Lewis
James Cook University

Craig R. Sloss
Queensland University of Technology, csloss@uow.edu.au

Colin V. Murray-Wallace
University of Wollongong, cwallace@uow.edu.au

Colin D. Woodroffe
University of Wollongong, colin@uow.edu.au

Scott G. Smithers
James Cook University

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Post-glacial sea-level changes around the Australian margin: a review

Abstract
It has been known since Rhodes Fairbridge's first attempt to establish a global pattern of Holocene sea-level change by combining evidence from Western Australia and from sites in the northern hemisphere that the details of sea-level history since the Last Glacial Maximum vary considerably across the globe. The Australian region is relatively stable tectonically and is situated in the 'far-field' of former ice sheets. It therefore preserves important records of post-glacial sea levels that are less complicated by neotectonics or glacio-isostatic adjustments. Accordingly, the relative sea-level record of this region is dominantly one of glacio-eustatic (ice equivalent) sea-level changes. The broader Australasian region has provided critical information on the nature of post-glacial sea level, including the termination of the Last Glacial Maximum when sea level was approximately 125 m lower than present around 21,000-19,000 years BP, and insights into meltwater pulse 1A between 14,600 and 14,300 cal. yr BP. Although most parts of the Australian continent reveals a high degree of tectonic stability, research conducted since the 1970s has shown that the timing and elevation of a Holocene highstand varies systematically around its margin. This is attributed primarily to variations in the timing of the response of the ocean basins and shallow continental shelves to the increased ocean volumes following ice-melt, including a process known as ocean siphoning (i.e. glacio-hydro-isostatic adjustment processes). Several seminal studies in the early 1980s produced important data sets from the Australasian region that have provided a solid foundation for more recent palaeo-sea-level research. This review revisits these key studies emphasising their continuing influence on Quaternary research and incorporates relatively recent investigations to interpret the nature of post-glacial sea-level change around Australia. These include a synthesis of research from the Northern Territory, Queensland, New South Wales, South Australia and Western Australia. A focus of these more recent studies has been the re-examination of: (1) the accuracy and reliability of different proxy sea-level indicators; (2) the rate and nature of post-glacial sea-level rise; (3) the evidence for timing, elevation, and duration of mid-Holocene highstands; and, (4) the notion of mid- to late Holocene sea-level oscillations, and their basis. Based on this synthesis of previous research, it is clear that estimates of past sea-surface elevation are a function of eustatic factors as well as morphodynamics of individual sites, the wide variety of proxy sea-level indicators used, their wide geographical range, and their indicative meaning. Some progress has been made in understanding the variability of the accuracy of proxy indicators in relation to their contemporary sea level, the inter-comparison of the variety of dating techniques used and the nuances of calibration of radiocarbon ages to sidereal years. These issues need to be thoroughly understood before proxy sea-level indicators can be incorporated into credible reconstructions of relative sea-level change at individual locations. Many of the issues, which challenged sea-level researchers in the latter part of the twentieth century, remain contentious today. Divergent opinions remain about: (1) exactly when sea level attained present levels following the most recent post-glacial marine transgression (PMT); (2) the elevation that sea-level reached during the Holocene sea-level highstand; (3) whether sea-level fell smoothly from a metre or more above its present level following the PMT; (4) whether sea level remained at these highstand levels for a considerable period before falling to its present position; or (5) whether it underwent a series of moderate oscillations during the Holocene highstand.

Keywords
australian, around, margin, changes, level, review, glacial, post, sea, GeoQuest

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Post-glacial sea-level changes around the Australian margin: A review

Stephen E. Lewis[1]*, Craig R. Sloss[2], Colin V. Murray-Wallace[3], Colin D. Woodroffe[3], Scott G. Smithers[4]

[1] Catchment to Reef Research Group, TropWATER, James Cook University, Townsville QLD 4811, Australia
[2] Earth, Environmental and Biological Sciences, Queensland University of Technology, GPO Box 2434, Brisbane, QLD 4001, Australia
[3] School of Earth & Environmental Sciences, University of Wollongong, Wollongong, NSW 2522, Australia
[4] School of Earth and Environmental Sciences, James Cook University, Townsville QLD 4811, Australia
* Corresponding author: phone: +61 7 4781 6629; fax: +61 7 4781 5589 email: stephen.lewis@jcu.edu.au

Abstract

It has been known since Rhodes Fairbridge's first attempt to establish a global pattern of Holocene sea-level change by combining evidence from Western Australia and from sites in the northern hemisphere that the details of sea-level history since the last glacial maximum vary considerably across the globe. The Australian region is relatively stable tectonically and is situated in the 'far-field' of former ice sheets. It therefore preserves important records of post-glacial sea levels that are less complicated by neotectonics or glacio-isostatic adjustments. Accordingly, the relative sea-level record of this region is dominantly one of glacio-eustatic (ice equivalent) sea-level changes. The broader Australasian region has provided critical information on the nature of post-glacial sea level, including the termination of the last glacial maximum when sea level was approximately 125 m lower than present around 21,000 to 19,000 years BP, and insights into meltwater pulse 1A between 14,600 and 14,300 cal. yr BP. Although most parts of the Australian continent reveals a high degree of tectonic stability, research conducted since the 1970's has shown that the timing and elevation of a Holocene highstand varies systematically around its margin. This is attributed primarily to variations in the timing of the response of the ocean basins and shallow continental shelves to the increased ocean volumes following ice-melt, including a process known as ocean siphoning (i.e. glacio-hydro-isostatic adjustment processes).

Several seminal studies in the early 1980's produced important datasets from the Australasian region that have provided a solid foundation for more recent palaeo-sea-level research. This review revisits these key studies emphasising their continuing influence on Quaternary research and incorporates relatively recent investigations to interpret the nature of post-glacial sea-level change around Australia. These include a synthesis of research from the Northern Territory, Queensland, New South Wales, South Australia and Western Australia. A focus of these more recent studies has been the re-examination of: 1) the accuracy and reliability of different proxy sea-level indicators; 2) the rate and nature of post-glacial sea-level rise; 3) the evidence for timing, elevation, and duration of mid-Holocene highstands; and, 4) the notion of mid to late-Holocene sea-level oscillations, and their basis.

Based on this synthesis of previous research it is clear that estimates of past sea-surface elevation are a function of eustatic factors as well as morphodynamics of individual sites, the wide variety of proxy sea-level indicators used, their wide geographical range, and their indicative meaning. Some progress has been made in understanding the variability of the accuracy of proxy indicators in relation to their contemporary sea level, the inter-comparison of the variety of dating techniques used and the nuances of calibration of radiocarbon ages to sidereal years. These issues need to be thoroughly understood before proxy sea-level indicators can be incorporated into credible reconstructions of relative sea-level change at individual locations. Many of the issues which challenged sea-level researchers in the latter part of the twentieth century remain contentious today. Divergent opinions remain about: 1) exactly when sea level attained present levels following the most recent post-glacial marine transgression (PMT); 2) the elevation that sea-level reached during the Holocene sea-level highstand; 3) whether sea-level fell smoothly from a metre or more above its present level following the PMT; 4) whether sea level remained at these highstand levels for a considerable period before falling to its present position; or 5) whether it underwent a series of moderate oscillations during the Holocene highstand.

Keywords: sea level; Australia; sea-level indicators; sea-level change; Holocene; post-glacial

1. Introduction

The relative tectonic stability of the Australian continent is due to its intra-plate setting. It is also far from the major ice sheets that covered large areas of the northern hemisphere continents during the Last Glacial Maximum (LGM) and has thus been unaffected by significant ice accumulation or the effects of glacio-isostatic rebound. Accordingly, the Australian margin is a suitable place to examine evidence of former shorelines dominated by glacio-eustatic sea-level changes with variable influences of hydro-isostasy (Bryant et al., 1992; Nakada and Lambeck, 1989; Murray-Wallace and Belperio, 1991). The relative tectonic stability and limited isostatic influence around the Australian coastal margin renders the region an important setting for reconstructing post-glacial glacio-eustatic (ice-equivalent) sea-level changes during the most recent marine transgression which followed the LGM (Ferland et al., 1995; Murray-Wallace et al., 1996, 2005). However, much of the Australian coast is bounded by a relatively shallow continental shelf of variable width that has experienced limited hydro- and sedimentary isostasy during the last glacial cycle which has had an influence in terms of the Holocene highstand (Larcombe et al., 1995; Baker and Haworth, 2000a; Belperio et al., 2002; Collins et al., 2006; Sloss et al., 2007; Lewis et al., 2008; Woodroffe, 2009). In 1983, David Hopley and colleagues published a synthesis of Holocene sea-level data from around the Australian mainland highlighting the key sites and studies up to that time (Hopley, 1983a). This review reflects on the continued influence of the key sites identified in Hopley's synthesis, and critically reviews the palaeo-sea level evidence presented in subsequent studies.

While the Australian continental margin reveals evidence for localised neotectonic uplift and subsidence (up to 10 m but typically < 2 m) based on the displacement of last interglacial (marine isotope sub-stage 5e ca. 125 ka) coastal successions that clearly relate to different geotectonic domains (Murray-Wallace, 2002), these rates of crustal movement have had negligible influence on resolving the general pattern of Holocene sea-level changes. For example, one of the more tectonically active regions in southern Australia has experienced uplift of approximately 0.7 m during the Holocene (Cann et al., 1999). This contrasts with adjacent plate-margin sites such as the Huon Peninsula in Papua New Guinea which has undergone rapid and apparently constant tectonic uplift of 20 – 30 m over the same period, where one of the most detailed records of relative sea-level changes for the past ~ 400 ka has been established (e.g. Chappell and Shackleton, 1986; Chappell et al., 1996).

The analysis in this review uses a similar approach to previous studies; it emphasises the continued significance of the previously published data (particularly age estimates and the reliability of proxy data to contemporary sea levels) and presents additional marine (corals) and estuarine (sedimentary and mangrove deposits) proxy indicators, together with other proxies now being examined using new techniques (e.g. fixed intertidal biological indicators or encrusting organisms). This work also examines the principal research trends over the past 25 years that recognised regional and local tectonic influences around the Australian margin, and regional variations in eustatic influences, leading to the reconstruction of sea-level histories on a more localised scale. Improved methods of dating enable a greater precision in the assignment of ages to fossil material, but the accuracy with which any inference about sea level can be made means that each sample must be interpreted in the context of the geomorphological setting in which it occurred. Indeed the review highlights the potential problems associated with the interpretation of sea-level indicators, most notably the assumptions that coastal boundary morphodynamic conditions have been constant through time.

The key sites with ongoing research on post-glacial sea-level changes examined in this review include (Fig. 1): The Huon Peninsula, Papua New Guinea (Chappell and Polach, 1991; Ota and Chappell, 1999) and the Joseph Bonaparte Gulf (Yokyama et al. 2000, 2001a) and its comparison with the Sunda Shelf (Hanebuth et al. 2000, 2009); Spencer Gulf and the Gulf of St Vincent (Belperio et al., 2002); eastern Australia (Baker et al., 2001a, b; Sloss et al., 2007; Lewis et al., 2008); and Western Australia (Searle and Woods, 1986; Baker et al., 2005; Collins et al., 2006). To accurately examine the sea-level histories following the LGM, this review synthesises available data for sea-level change around the Australian margin over the past 20,000 years. We focus on two separate intervals: 1) during the post-glacial transgression (20,000 to 7000 years BP); and 2) the mid-late Holocene (8000 years BP to present). We examine the key mechanisms that appear to explain sea-level change since the LGM at regional and local scales, recognising the legacy provided by the pioneering researchers and identifying remaining gaps in our understanding.

2. Background

The significance of both the relatively tectonically stable continent of Australia and the uplifted coast of Papua New Guinea for palaeo-sea-level investigations was realised by Rhodes W. Fairbridge. His studies in Western Australia in collaboration with Curt Teichert

(Teichert, 1950) identified a series of well-preserved sea-level indicators on the Western Australian coast at Point Peron, and offshore on Rottnest Island and the Abrolhos Islands. Fairbridge also realised the potential of the uplifted suite of reef terraces on the north coast of the Huon Peninsula (Fairbridge, 1960). In 1961, Fairbridge presented a thorough global synthesis and critique of sea level and climatic studies and painstakingly considered (and discounted) several alternative explanations for sea-level variability since the LGM including continental flexure, the ‘oscillating margin’ hypothesis, geodetic forcing and different forms of eustasy (listed below). He produced an integrated theory based on changes in ice volume (glacio-eustasy), tectonic influences (tectono-eustasy), sediment deposition in ocean basins (sedimento-eustasy) and other influences such as inputs from volcanic activity and thermo-expansion/contraction. Fairbridge (1961) also recognised the influence of local tectonic processes and crustal adjustments to water loading and unloading (hydro-isostasy) on sea-level records, although he did not attempt to quantify them in his work.

Fairbridge’s sea-level curve of global eustasy (Fairbridge, 1961), developed from a series of studies undertaken in the 1940’s and 1950’s with Teichert, propelled the evidence from emergent shorelines in Western Australia onto the world stage as key type localities for several highstands. At this time the prevailing view was that sea level had risen gradually up to present at a decelerating rate (e.g. Jelgersma and Pannekoek, 1960; Shepard, 1961). In light of that, Fairbridge included post-glacial sea-level evidence from the Bahamas and Florida positioned below present sea level. In Fairbridge’s eustatic curve, a series of highstands based on evidence from Australian sites were interspersed with lowstands based on evidence from sites in the Bahamas and Florida. It soon became apparent that these Atlantic and Pacific data sets could not be compiled into a single eustatic sea-level record. This was due to global and regional variations in the eustatic and isostatic response to changes in ice and ocean volumes, resulting in different relative sea-level histories (Bloom, 1977).

In Australia, Bruce Thom and John Chappell (1975) produced the first widely accepted sea-level envelope of post-glacial sea-level rise for Australia based on a compilation of radiocarbon ages from eastern Australia. Their study also suggested the possibility of mid-late Holocene emergence on some parts of the Australian coast, and recognised the significance of a stratigraphic record of prograded coastal plains that had accumulated over the past 6000 years. This was in contrast to the decelerating pattern of sea-level rise widely recognised in the North Atlantic, where present sea level was not attained until sometime between 4000 to 2000 yr BP (Thom and Chappell, 1975; Pirazzoli, 1991). Whether or not Holocene sea levels had been higher than present, particularly along the eastern Australian coast remained a contentious issue. A vigorous exchange of views had been expressed in the early volumes of Marine Geology, with evidence proposed in support of sea level having been above its present level observed in Queensland and Victoria (Gill and Hopley, 1972) and refuted along other sections of the coast (Thom et al., 1969, 1972).

In 1973, the Great Barrier Reef Expedition by the Royal Society of London and the Universities of Queensland, led by David Stoddart, undertook particularly extensive mapping and surveying of numerous reefs in the northern Great Barrier Reef. The results, published in 1978 in two volumes of the Philosophical Transactions of the Royal Society, provided unequivocal evidence for higher Holocene sea levels (McLean et al., 1978). It was as part of this expedition that the significance of coral microatolls as accurate sea-level indicators was realised (Scoffin and Stoddart, 1978). Shortly thereafter an extensive sequence of microatolls from the continental margin of north-east Queensland was described by John Chappell and colleagues (Chappell, 1983; Chappell et al., 1982, 1983). The two decades following saw concentrated coring and dating of a series of reefs which contributed to our understanding of the way in which the Great Barrier Reef had developed during the Holocene in response to sea-level variations. These are summarised by Hopley (1982) and updated in Hopley et al. (2007).

Stratigraphical and chronological studies of sand barrier and estuarine evolution along the south-east coasts of Australia by Bruce Thom and Peter Roy refined the understanding of sea-level history in this region (Roy and Thom, 1981; Thom, 1984a, b) and consolidated the earlier work of Thom and Chappell (1975). Thom and Roy (1983, 1985) subsequently plotted a sea-level envelope between the terrestrial (tree stumps, wood and charcoal) and marine (shells) palaeo-sea-level indicators recognising that these were 'directional' (relational) indicators, rather than providing an accurate (fixed) estimate of sea-level position. As a result, these sea-level data display a wide scatter of sample depth in relation to their contemporary sea level. The samples collected for radiocarbon dating were collected from locations extending from the Gold Coast in Queensland to Badger Head in Tasmania, representing an extensive geographical range (1,400 km in latitude) along Australia's eastern seaboard. The additional data incorporated by Thom and Roy (1983, 1985), notably the mangrove stump data, shifted the sea-level envelope to attain present levels much earlier than the previous work by Thom and Chappell (1975) thus showing that the rates of sea-level rise were faster than previously thought (we note that Thom and Roy also used 'environmentally corrected' ^{14}C ages for the marine reservoir effect in radiocarbon dating for their reconstruction). These studies laid the groundwork for the recognition that local sea levels are partly dependent on the geomorphological setting, and that geomorphological evolution of the coast may both preserve evidence of relative sea-level change and influence how other sea-level indicators preserve evidence of past sea-level changes. Nevertheless, the data allowed the construction of a broad envelope to constrain the position of early to mid-Holocene relative sea level for the east coast of Australia, indicating that sea level reached its present height by 6000 radiocarbon yr BP.

The focused research on sea level during the late 1970's and early 1980's enabled David Hopley and colleagues to compile a comprehensive data set of Holocene sea-level indicators across Australia in 1983 (see Hopley, 1983a, b, 1987; Hopley and Thom, 1983) with a further compilation by John Chappell in 1987 (Chappell, 1987). Critical issues identified in these compilations included the determination of when following the post-glacial marine transgression present sea level was reached, the elevation and duration of any mid-Holocene highstand, whether sea-level oscillations of 1 to 2 m lasting up to 1000 years have occurred since present sea level was initially reached, and understanding regional variations in Holocene relative sea-level histories around the Australian coastline and across other parts of the world (Hopley, 1983a, 1987; Chappell, 1987).

3. Sea-level indicators

A variety of proxy sea-level indicators has been used in previous studies and on-going research and a critical assessment of their indicative meaning is presented here. Interpreting past sea levels has involved the use of several different sea-level indicators to reconstruct sea-level change relative to a specific datum (commonly Australian Height Datum, which broadly approximates estimated present mean sea level (PMSL)). In some parts of the world, there is clear evidence that the sea was once higher, preserved as erosional notches cut into cliffs. Erosional features formed a substantial component of the evidence for higher sea level recognised by Fairbridge and others from Western Australia. However, constructional coasts, such as coral reefs, built by a framework of coral and with a range of secondary organisms, sedimentary and diagenetic processes, tend to preserve a wider range of potential indicators than erosional coasts. On other coasts, stratigraphical sequences of sedimentary facies (commonly dating intertidal and subtidal shells) have provided a first-order indication of transgressive or regressive tendency and function as directional indicators of previous limiting sea levels. Further sea-level evidence has been derived from beachrock, the remains of intertidal plants (mangroves), encrusting organisms (oysters, barnacles, tubeworms), foraminiferal assemblages and supratidal deposits, as well as intertidal erosional indicators (including notches, abrasion platforms and benches). In this section, we review the development, reliability, accuracy and limitations of the key proxy indicators that are commonly used to reconstruct palaeo-sea-level. A detailed review of sea-level indicators on coral reefs has recently been provided by Smithers (2011 and references therein) and other reviews including Hopley and Thom (1983), Davies and Montaggioni (1985), van de Plassche (1986), Pirazzoli, (1991, 1996) and Hopley et al. (2007) also provide excellent broader overviews. This section, therefore, provides only a brief summary of commonly used proxy indicators, and how they have been used in reconstructions in the Australian region. We examine the integrity and indicative meaning of each of these proxies and highlight their contribution to reconstructing relative sea levels around the Australian mainland. While the sea-level indicators described below have different indicative meanings relative to predicted tidal datums (e.g. corals: below or at mean low water springs; tubeworms: mean high water neaps), the difference between the elevation of the relict indicator and that of the highest or lowest living counterpart provides evidence of relative sea-level change compared to PMSL.

It is important to recognise that the indicators above preserve sea-level evidence that may be related to a level associated with specific coastal features or processes, but may never be directly tied to a surveyed and statistically derived tidal datum such as mean sea level (MSL). Biological indicators are mediated by the physiological response to environmental factors such as exposure, which may relate broadly to a tidal stage. However, tidal planes themselves are unlikely to be horizontal over even short distances, are variable in time and space due to interactions with local geomorphology, and open waters are rarely calm and still. These at-a-site factors which introduce variability and noise are less critical for assessments of rapidly rising post-glacial sea levels, but become pivotal in the interpretation of the details of sea-level histories after present sea level was initially reached. Indeed, a change in relative sea level as recorded by sea-level indicators can be the result of eustatic (ice volume change), isostatic (uplift or subsidence from changing ice/water loads on the continental crust), climatic (steric, wind forcing) or coastal morphodynamic (e.g. closing of a high energy window) changes.

The reliability with which indicators can be used in sea-level reconstructions is also partly dependent on their potential for accurate age-control (Sloss et al., in press). Dating can be performed by various methods including radiocarbon (via Accelerator Mass Spectrometry: AMS or conventional methods), U-series, Amino Acid Racemisation (AAR) and Optical Stimulated Luminescence (OSL). Radiocarbon (^{14}C) dating has unquestionably been the most widely applied numeric dating method in studies of post-glacial sea-level changes; however, ages need to be calibrated to account for secular atmospheric variations and the marine reservoir effects for marine samples (both global and regional; Sloss et al., in press). The ^{14}C ages presented in this review have been calibrated to sidereal years with 2 sigma standard deviations using the latest calibration programs and applying the currently recommended regional reservoir effect (ΔR) for each site (Stuiver et al., 2005; Weninger et al., 2011). Previous studies have shown that ΔR varies around different sections of the Australian coastline and also with the type of indicator (e.g. seagrass fibres, bivalves, corals); it is also likely to diverge from the open-water ΔR value in estuaries where there is significant dilution by fluvially-derived water with a different radiocarbon activity, as well as having varied considerably over time (Eisenhauer et al., 1993; Ulm, 2002; McGregor et al., 2008; Lewis et al., 2012). Insufficient data are currently available to account for these additional influences on ΔR, but compilation of age data and continued attention to disparities in the ΔR value promise to reduce discrepancies that arise from such issues. Before reliable radiocarbon calibration programs, 'environmentally corrected' ^{14}C ages where 450 years were subtracted from marine samples (commonly molluscs and corals) following the work of Gillespie and Polach (1979) to take account of the 'pre-inclusion' age of radiocarbon were commonly applied in sea-level reconstructions. The development of calibration datasets (e.g. Reimer et al., 2009) allows ^{14}C ages to be calibrated to sidereal years and this advance has shown that present sea level around Australia was reached somewhat earlier than previously thought (e.g. Sloss et al., 2007; Lewis et al., 2008). U-series has become the favoured method for the dating of corals as the ages do not require corrections for the marine reservoir effects and the analytical precision has greatly improved with the establishment of new analytical procedures (Yu and Zhao, 2010; Clark et al., 2012).

3.1 Coral reefs

Reef-building scleractinian corals flourish in shallow tropical waters, limited in their poleward extension by sea-surface temperatures (coral reef growth is mostly confined above the 18ºC isotherm), constrained in their upward growth by exposure at low water spring tide, and rare below water depths of 50 m where symbiotic photosynthetic algae receive insufficient light (Smithers, 2011). Individual fossil corals within Holocene coral reefs can be used as 'directional' sea-level indicators where relative sea-level position (compared with today) must have been higher at the time of coral growth (e.g. Veeh and Veevers, 1970). Individual massive or 'brain' corals grow at rates of about 1 to 2 cm per year, whereas many branching corals may extend 5-10 cm each year. Coral reefs, however, which are the aggregated outcome of a framework of corals and other calcifying reef organisms, tempered by erosion, accrete vertically at rates of just a few millimetres per year. Accordingly, they grew more slowly than rapidly rising sea levels following the LGM, with the result that the reef growth curve differs from the sea-level curve (Hopley 1986a; Eisenhauser et al., 1993). Three modes of coral reef response to sea-level rise have been identified: *keep-up* where a reef tracked sea-level rise; *catch-up* where a reef lagged behind the rising sea level but reached the sea surface after sea level had stabilised close to its present position; and *give-up* where a reef was drowned by rapid sea-level rise and remained as a submerged reef, commonly with an algal pavement over its surface (summarised in Hopley, 1982; Neumann and Macintyre, 1985; Davies and Montaggioni, 1985).

Recently it has become apparent that there are numerous drowned, *give-up* reefs in the Australian region. 'Drowned reefs' along the shelf edge of the Great Barrier Reef have been reported at depths between 50 and 70 m (Carter and Johnson, 1986; Harris and Davies, 1989; Beaman et al., 2008; Abbey et al., 2011; Webster et al., 2011; Yokoyama et al., 2011); these reefs are likely to provide important sea-level data, although they have yet to be dated. Drowned reefs at depths of 20 to 30 m have also been identified in the southern Gulf of Carpentaria, where they were unexpected because of relatively high turbidity and the fact that the area was a terrestrial environment with an inland lake for a significant part of the latter portion of the last glacial cycle (Chivas et al., 2001). Terrace reefal successions from the Cootamundra Shoals, north-western Australia ranging from 20 to 90 m depth have also been identified with radiocarbon ages of the 20 and 28 m terraces yielding 8460 ± 100 and 7630 ± 100 cal. yr BP, respectively (Flemming, 1986).

Much has been learnt about the internal structure of reefs from detailed studies of the tectonically-active margin of the Huon Peninsula, Papua New Guinea where coral reef terraces have been uplifted and preserved since at least 400,000 years ago (Chappell, 1980; Chappell et al., 1996; Pandolfi, 1996; Pandolfi et al., 2006). Where rates of tectonic uplift are well-constrained, these terrace successions reliably record previous interglacial and interstadial highstands (and possibly lowstands), but coring on the lowermost terrace has also provided important data on sea-level rise after the LGM (Chappell and Polach, 1991; Chappell et al., 1998; Ota and Chappell, 1999).

Submerged reefs may also be used as indicators where the rate of subsidence is known, and where reef flats and reef pavement deposits are located above the contemporary limit to coral growth they can indicate formerly higher sea levels (Hopley, 1986a; Matthews, 1990; Collins et al., 2006). At the Houtman Abrolhos Islands, Western Australia raised coral pavements were used to reconstruct Holocene sea levels with an accuracy of ±0.5 m (Collins et al., 2006), where they were interpreted to form during gradual sea-level fall (Eisenhauer et al., 1999). Corals grow prolifically in shallow water with a gradually rising sea, but may be almost absent if a mature reef flat becomes emergent because of a relative sea-level fall (Smithers et al., 2006).

3.2 Coral microatolls

Microatolls are disc-shaped coral colonies with dead tops and live sides. They form on reef flats and can grow up to several metres in diameter. Colony height varies with depth of the substrate, but for most microatolls used for sea-level research on the Great Barrier Reef it is < 1 m. As many as 23 coral genera have been observed to adopt a microatoll growth form on the Great Barrier Reef, but the most common microatolls are of *Porites* and *Faviid* spp. (Rosen, 1978; Scoffin and Stoddart, 1978; Stoddart and Scoffin, 1979). Massive *Porites* grow upwards in open water conditions until they reach an elevation close to mean low water springs (MLWS) or slightly higher, after which they cannot grow further vertically, but continue to extend laterally to form microatolls (Scoffin and Stoddart, 1978). Although the MLWS limit has been observed across much of the GBR for *Porites*, *Faviid* microatolls have been found up to 0.5 m higher (Perry and Smithers, 2011) and *Porites* microatolls on the Cocos (Keeling) Islands also grow to a higher level in open waters (between MLWS and mean low water neap tides; Smithers and Woodroffe, 2000). Microatolls can be locally moated behind rubble storm ridges (Hopley and Isdale, 1977), thus indicating a locally-elevated water level (discussed below), which can complicate sea-level interpretations where moated and open water microatolls are not easily distinguished.

Although flat-topped corals were described from Yule Point by Bird (1971), fossil microatolls elevated above their living counterparts were first recognised as high resolution sea-level indicators on the Great Barrier Reef by Scoffin and Stoddart (1978). Dated microatolls were an important component of the evidence used by McLean et al. (1978) to provide an estimate of a higher sea level (approximately +1.0 m) in the mid-Holocene. An estimate of when sea level first attained its modern elevation was given at 6310 ± 90 ^{14}C years for a microatoll on Fisher Island (6760 ± 230 cal. yr BP; McLean et al., 1978). For this particular study, the elevations of the fossil microatolls were measured conservatively to the current mean high neap tidal mark to account for possible 'moating effects'.

The most comprehensive sea-level reconstruction using coral micoatolls was performed by Chappell et al. (1983) where groups or 'fields' of raised fossil microatolls were measured relative to their modern counterparts on fringing reefs along the continental islands and mainland of the Great Barrier Reef ('moated' microatolls were excluded from the analysis). These data remain as one of the highest resolution records produced in Australia and have been used to 'calibrate' hydro-isostatic models and support the 'smooth-fall' hypothesis of Holocene sea levels following the onset of the highstand (e.g. Chappell et al., 1982). Baker and Haworth (2000b) statistically reanalysed Chappell et al.'s (1983) original data and suggested that either the linear fall trend or a seventh-order polynomial (oscillating) trend could fit the data equally well.

Yu and Zhao (2010) demonstrated the higher analytical precision of U-series methods by analysing 11 coral microatolls from Magnetic Island obtaining 2σ errors commonly within ±50 years. However, neither the studies of Chappell et al. (1983) nor those of Yu and Zhao (2010) report where on the microatoll surface the samples were collected for their analyses (i.e. collected from the outer rim, centre or elsewhere); this limits the interpretation of the timing of sea-level position. One of the most important considerations for sea-level studies is the precision to which elevation is measured and understanding the influence of other phenomena that may limit upward coral growth such as wave climate, ENSO and tidal patterns; such data on both spatial and temporal scales are rare. One such record exists from Christmas Island, Pacific Ocean, where detailed survey data of microatolls showed the influence of the geoid effect where living colonies from the southern section of the island were ~ 1 m lower than those on the northern side (Woodroffe et al., 2012).

While coral reef flats are typically restricted to low water springs tidal levels, shingle ramparts or algal rims on the reef flat can form moats or pools which effectively retain water above these tidal levels. These moats allow coral microatolls to grow to much higher elevations (the height largely dependent on tidal range), in some cases up to mean sea level (up to 2 m higher in macrotidal situations), but more commonly no higher than the low water neap tides (approximately 1 m higher than low water springs; Scoffin and Stoddart, 1978; Hopley, 1982, 1986a, 1987). Moating can be identified on reef flats from elevated pools at low water levels where coral colonies may display different surface morphology than their open water counterparts such as ‘smoother’ tops or terraces (Scoffin and Stoddart, 1978; Hopley, 1986a). Moating of fossil microatolls can be more difficult to identify. A tropical cyclone in the Great Barrier Reef in 1918 breached a shingle rampart on Holbourne Island killing the moated microatolls on the reef flat (Hopley and Isdale, 1977); without this prior knowledge these microatolls may have been interpreted as representing a higher sea level. Chappell et al. (1983) suspected some microatolls dated in their study were moated based on these microatolls being younger than those found offshore. These microatolls were excluded from their sea-level curve. Importantly, whereas coral microatolls can overestimate sea levels due to moating, it is not possible for them to underestimate the relative sea-level position; therefore other palaeo-sea-level indicators that suggest lower sea levels than the microatolls need to be treated with caution.

In microtidal settings outside the cyclone belt, moating of microatolls is less likely. Extensive surveys of several hundred *Porites* microatolls on the Cocos (Keeling) Islands revealed that their upper surfaces are constrained between mean low water springs and mean low water neaps range (Smithers and Woodroffe, 2000; Smithers, 2011). Open water microatoll ‘fields’ grow within a ±10 cm elevation range. They can be influenced by wave climate and tidal range but the upper surface of modern microatolls is clearly constrained by a tidal plane, and the former upper surface of fossil microatolls represents one of the more reliable indicators of the tidal level at which they were confined (Woodroffe and McLean, 1990; Smithers and Woodroffe, 2001).

3.3 Encrusting organisms

Encrusting organisms, such as oysters, tubeworms and barnacles (also referred to as fixed biological indicators; Chappell, 1987), are confined to a restricted range within the intertidal zone on rocky shorelines and have been used as sea-level indicators (Baker and Haworth, 1997; Lewis et al., 2008). For example, encrusting intertidal organisms may be buried or submerged as a consequence of rapid sea-level rise, but are more likely to be preserved under a falling sea level. Although their potential as sea-level indicators has long been recognised (Smith, 1978; Bird, 1988) and their indicative relationships to tidal levels at a particular site have been relatively well-established (e.g. Endean et al., 1956; Dakin, 1987), ages of elevated relict deposits were not obtained until 1988 and into the 1990s (Playford, 1988; Flood and Frankel, 1989; Beaman et al., 1994; Baker and Haworth, 1997). Subsequently, researchers have used relict oyster bed, barnacle and tubeworm deposits in sea-level reconstructions along the east and west coasts of Australia to delineate the magnitude and duration of the mid-Holocene highstand (Beaman et al., 1994; Baker and Haworth, 1997, 2000a; Baker et al., 2001a, b, 2005; Lewis et al., 2008).

In north-eastern Australia, the oysters *Saccostrea cuccullata* and *S. echinata* (also referred to as *Crassostrea amasa*) form large beds/visors of up to 1 m width in their upper ~20 cm vertical growth range (approximately mean high water neaps). This upper limit can be traced horizontally to within ±0.15 m across an embayment with a 2 km headland, although in areas of wave run-up deposits can occur at higher elevations in crevices or with changing wave exposure. The nature of these fossil oyster bed deposits, which can accumulate over periods of up to 1000 years, suggest that sea levels were stable or fluctuated around a common position for lengthy periods in the mid-Holocene; this evidence is difficult to reconcile with the smoothly falling sea levels inferred from sequences of coral microatolls and other prograding accretionary deposits (Beaman et al., 1994; Lewis et al., 2008). Fossil barnacles of *Octomeris brunnea* and *Tetraclitella* sp. from north-eastern Australia (Beaman et al., 1994; Higley, 2000) clearly have potential as indicators of past water levels in this region, although detailed studies are required before their full significance can be determined (Hopley et al., 2007).

Serpulid tubeworms (commonly *Galeolaria caespitosa*) are constrained between the low and mid-tide range (up to high water neap); their use as sea-level indicators was described by Bird (1988). The transitional boundary between tubeworm and the barnacles *Chamaesipho tasmanica* and *Tesseropora rosea* forms a distinct marker indicative of a former tidal level reported to have a reproducible elevation range within ±10 cm; where this transitional boundary is not observed, an accuracy of ±25 cm has been reported (Baker and Haworth, 1997, 2000a; Baker et al., 2001a, b). Elevation of this tubeworm-barnacle boundary can vary under different exposure regimes (Baker and Haworth, 1997). *G. caespitosa* typically grow in sheltered environments and two separate growth morphologies have been recognised which have been linked to their vertical growth range (Baker and Haworth, 2000a). However, tubeworm and barnacle deposits can occupy a much broader depth envelope in regions of larger tidal range or higher wave exposure (e.g. Laborel and Laborel-Deguen, 1996), diminishing their potential use for high-resolution reconstructions of past sea-level changes (Sloss et al., 2007). Indeed, relict tubeworms from Wasp Head and Clear Point, New South Wales, a region with relatively higher wave/swell exposure, were reported at +2.7 to 3.0 m above the modern assemblage (Baker et al., 2001b). Baker et al. (2001b) rejected these deposits as reliable indicators of former sea level, arguing that the high elevation was an artefact of the energetic wave regime at the sample location. Recognition of wave exposure as a possible confounding factor is important, as is acknowledgement that a key assumption when using these indicators to examine past sea-level changes is that wave conditions have not significantly changed over the period of investigation (see Goodwin, 2003).

Variations in the upper growth limit of these organisms at a single location is another issue that constrains the use of encrusting organisms, implying that they are more likely to preserve evidence of local variations in past conditions at that site rather than indicating a regional sea-level trajectory. This is particularly so as encrusting organisms appear to be best preserved in sheltered settings where local effects can be important, and less well preserved on open water settings where they can be directly compared with their modern counterparts (e.g. Scoffin, 1977; Hopley et al., 2007; Perry and Smithers, 2011; Smithers, 2011). However, suitable growth patterns for oyster deposits can sometimes be distinguished in the ancient record based on their morphology such as those that form a thick (> 20 cm) bed/visor on a horizontal plane with clear separation of visor and wave splash zone (Lewis, 2005). As most relict oyster bed deposits are preserved in caves and crevices, they may record a relatively higher sea level compared with more open-water indicators such as microatolls (Lewis et al., 2008; Smithers, 2011). Nevertheless, these deposits can accumulate over long periods and their continued presence at similar elevations across large parts of eastern, northern and Western Australia provides evidence for a prolonged highstand over the mid-late Holocene. Moreover, changes in the ecology, morphology and elevation of encrusting organisms (e.g. Baker and

Haworth, 2000a; Baker et al., 2001b; Wright, 2011) are likely to provide important evidence for changes in coastal boundary conditions such as water level, wave exposure and climate. Important data have been gathered on both the geographical and elevation ranges of oyster and barnacle species surveyed to tidal datum along the Great Barrier Reef coastline and offshore islands extending from Cooktown to Gladstone (Wright, 2011).

3.4 Stratigraphical sequences

Distinctive boundaries between different sedimentary facies can be used to decipher past sea levels where the depositional development of different facies are related to sedimentary processes dominant at particular elevations with respect to sea level. Such facies relationships have been used to establish relative changes in sea level in sedimentary settings including raised beach/barrier systems preserved in coastal plains (Burne, 1982; Belperio et al., 1984, 2002; Searle and Woods, 1986), the sedimentary infill of incised coastal valleys (e.g. Jones et al., 1979; Gagan et al., 1994; Sloss et al., 2005, 2006, 2007, 2011), and in sediment cores from the continental shelf that preserve sequences of marine (onlap) and terrestrial (offlap) facies related to inundation and emergence (e.g. Larcombe and Carter, 1998). Different materials have been dated within these deposits, including mangrove wood, intertidal and subtidal molluscs, charcoal, peat, the remains of seagrass fibres and organic rich muds. In some cases, changes in sediment facies within sediment cores can be directly related to intertidal deposition and processes linked to a specific tidal datum or depth. For example, Belperio et al. (1983, 1984, 2002) produced detailed sea-level reconstructions for different sections of Spencer Gulf and the Gulf of St Vincent, South Australia (partitioned according to distance from the continental margin) using established relationships between transitions from subtidal seagrass facies to intertidal sand flat facies, and from sand flat to mangrove facies (material included seagrass, mangrove and molluscs). The work of Bruce Thom, Peter Roy and colleagues on the geomorphology and sedimentology of back barrier deposits and drowned river valley successions along the coasts of southern Queensland and New South Wales also allowed different depositional environments to be identified and related to sea-level position (Martin, 1972; Macphail, 1974; Roy et al., 1980; Roy and Thom, 1981; Thom et al., 1981, 1992; Thom, 1984b). These data on the stratigraphic sequences in New South Wales provided important information on the post-glacial sea-level rise and when present sea level was achieved. Moreover, the studies provided insights on the various responses of these coastal deposits that could be interpreted in relation to local levels of the sea, including tidal planes.

Searle and Woods (1986) used the interface between sublittoral sands and beach sediments ('swash zone' facies) to reconstruct sea-level changes along the southern Western Australian coastline. In their reconstruction, they used radiocarbon ages from well-preserved 'delicate' bivalves (*Donax* and *Paphies* sp.); these bivalves are easily weathered and broken and so well-preserved materials were considered to indicate little or no reworking (Searle and Woods, 1986). A transgressive sandsheet formed as rising seas breached remnants of Last Interglacial barriers during the most recent post-glacial marine transgression (ca. 12,000 – 7000 cal. yr BP) is widely encountered in coastal embayments in eastern Australia (e.g. Sloss et al., 2006, 2007), where it extends up to near present sea level. The transgressive deposit is characterised by a shell-rich mix of marine and estuarine molluscan fauna and represents reworked intertidal sand flats, tidal channel sands and flood tide delta sands (Sloss et al., 2006, 2007, 2011). Recognition of this and overlying facies provides another example of the application of facies associations and stratigraphy as a sea-level indicator. However, such deposits accumulate over a relatively long time period (ca. 2,000 years), can only provide a time-averaged assessment of sea level at the time of deposition, and may contain reworked molluscs which complicate interpretations of the age of the deposits (Sloss et al., 2005, 2007). As for the biological indicators, a key assumption in the interpretation is that coastal boundary conditions such as wave exposure, tidal range and sediment supply have remained relatively constant as the deposit has formed (see Chappell and Thom, 1986).

3.5 Mangrove deposits

Many of the numerous estuaries along the tropical and sub-tropical coasts of Australia contain extensive mangrove forests. Transgressive mangrove muds also occur beneath the deltaic plains or marginal estuarine deposits which record post-glacial inundation of former valleys by rising seas. Such mangrove deposits are relatively good indicators of this rapid sea-level rise (e.g. high preservation potential) and they are generally restricted to the upper part of the intertidal zone. Several researchers have used organic clays, peats, roots, wood fragments and *in situ* stumps that are associated with mangroves to indicate Holocene sea-level change where mangroves occur (e.g. Jennings, 1975; Belperio, 1979; Jones et al., 1979; Thom and Roy, 1983, 1985; Grindrod and Rhodes, 1984; Woodroffe et al., 1987; Woodroffe, 1988; Sloss et al., 2007). The accuracy with which mangrove deposits can be used to define past water level is likely to be less in areas of large tidal range, as mangroves grow from mean sea level up to highest astronomical tide (HAT), with some isolated stands of mangrove persisting even above this level in a few locations. Mangrove wood and the fibrous remains of the roots of *Rhizophora* are distinctive but these organic remains are highly compressible, and the muds within which they occur can be compacted (Woodroffe, 1988; Grindrod et al., 1999). It is therefore important to choose basal samples that overlie bedrock or consolidated pre-Holocene deposits if the evidence is to be used in reconstructing the final stages of post-glacial sea-level rise (Woodroffe et al., 1987, 1989).

Where intertidal mangrove stumps can be detected, they are particularly effective sea-level indicators. Jones et al. (1979) describe the stumps of *Avicennia* exposed by storm erosion on a beachface at Bulli in the Illawarra, and an age on these remains (7720 ± 260 cal. yr BP) continues to support the argument for early attainment of present sea level in New South Wales (Young et al., 1993; Sloss et al., 2007). However, it is noted that this is only one datum and accordingly should be viewed with caution until the early age of the onset of the Holocene highstand can be confidently replicated. Roy and Crawford (1981) obtained a comparatively younger radiocarbon age (7030 ± 280 cal. yr BP) from a fossil mangrove stump (*Avicennia marina*) from Kurnell Peninsula, New South Wales at PMSL. Mangrove stumps, together with carbonate nodules containing the remains of mud lobsters (*Thalassina*) crop out on erosional banks along many of the estuaries of northern and north-western Australia, and are associated with 'big swamp' mangrove deposits that record a phase of widespread mangrove forest development between 7900 and 5800 cal. yr BP as sea level stabilised (Woodroffe et al., 1985).

Mangrove deposits are good indicators of rapid sea-level rise or when sea level attained a particular elevation, but are less useful indicators of sea-level stillstands or slow transgression because: they can grow in the upper part of the intertidal zone from around mean sea level to the supratidal zone, and their roots penetrate into the muddy sediments for up to as much as 1 m (Bunt et al., 1985; Woodroffe, 1988; Smithers, 2011). Mangrove sediments are also prone to post-depositional compaction (Chappell, 1987; Beaman et al., 1994; Larcombe et al., 1995); and the intrusion of younger rootlets presents contamination issues for radiocarbon ages (Beaman et al., 1994; Hopley et al., 2007; Smithers, 2011). Despite these potential problems, mangrove deposits have been used to indicate that mid-Holocene sea levels have not been significantly higher than 1 m above PMSL in several locations including Western Australia, Northern Territory and Queensland (e.g. Jennings, 1975; Belperio, 1979; Woodroffe et al., 1987). These deposits appear to underestimate sea-level elevations that have been established with other sea-level indicators (e.g. microatolls and encrusting organisms) that suggest sea levels were >1 m above PMSL, a result likely due to compaction and erosion of mangrove deposits and/or modifications of the sea-level signal as the coastal setting has evolved over the late Holocene (Chappell et al., 1983; Beaman et al., 1994; Baker et al., 2001b). Chappell and Grindrod (1984) discuss this issue for Princess Charlotte Bay, where the chenier plain and underlying mangrove sediments indicate a gradual progradation of the coast with no detectable change in sea level, at odds with evidence of emergence on nearby offshore islands. In this, as in other locations in northern Australia, it is likely that morphodynamic changes in the landscape result in local alteration of tidal prism and consequently evidence records water levels locally (Chappell and Thom, 1986).

3.6 Supratidal deposits

Supratidal deposits include sedimentary facies or organic remains that have been deposited above the highest astronomical tide (HAT). Supratidal chenier ridges deposited over mangrove muds provide an example, where the boundary between the two deposits has been used to indicate a sea-level position in several studies. This relationship was applied on a chenier sequence near Karumba, Gulf of Carpentaria, to establish a mid-Holocene highstand of ~ 2.5 m above PMSL followed by a gradual sea-level fall (Rhodes et al., 1980; Rhodes, 1982; Chappell et al., 1982). At Broad Sound on the central Queensland coast, chenier deposits have similarly been examined to infer sea-level changes across the eastern and western sides of the sound (Cook and Polach, 1973; Cook and Mayo, 1977).

Chenier ridges are wave-built landforms comprising generally relatively coarse sediments and commonly abundant accumulations of marine shells. Despite numerous investigations, there remains insufficient evidence to demonstrate that chenier bases define a reproducible datum level, or that they occur above the high tide plane. In Princess Charlotte Bay, it was argued that the sedimentary deposits were compacted and thus particular ridges underestimated sea level (Chappell and Grindrod, 1984; Chappell, 1987). Moreover, chenier ridges are ‘complex response systems’ strongly influenced by changes in storm frequency, sediment supply and shell production rates, which may confound the production and interpretation of a reliable sea-level signal (Chappell and Thom, 1986).

Other supratidal indicators include terrestrial materials such as freshwater peats, organic-rich muds, wood fragments and *in situ* tree stumps. They provide a ‘directional’ sea-level indicator, implying that sea level was lower than the elevation at which they occur at a particular point in time, but they do not indicate how far below. In some cases, these indicators can be used in conjunction with intertidal/estuarine sediments that bury or overlie them. For example, *in situ* tree stumps that are covered by estuarine muds have been used to define when sea level first reached a particular elevation (Sloss et al., 2007). Supratidal deposits are prone to compaction, reworking and diagenesis, and thus are generally not relied on to reconstruct accurate accounts of former sea-level position (Larcombe et al., 1995).

3.7 Beachrock

Beachrock comprises cemented beach sands and is usually found as clearly bedded and dipping outcrops of consolidated or moderately consolidated beach sediment, reflecting the gradient of the beach. The details of beachrock formation have been debated in the literature for several decades, but it is now generally conceded that it can form through several different processes (McLean, 2011, and references therein). It is largely agreed that beachrock forms in the intertidal zone when unconsolidated sediments become lithified by aragonitic and/or calcitic cements, preserving the internal fabric of the beach. Outcrops can be over 3 m in thickness in areas with relatively high tidal ranges such as on the Great Barrier Reef or exposed to large waves (Hopley, 1986b; McLean, 2011). While beachrock may provide a relatively constrained sea-level indicator in microtidal environments, the exact uppermost limit of formation is difficult to determine, particularly where the tide range is large. Therefore there is both considerable variability and uncertainty as to what tidal level, if any, the upper limit of beachrock indicates (Hopley, 1986b, 1987; Hopley et al., 2007). Based on beachrock occurrence above assumed sea-level limits, Kelletat (2006) asserted that beachrock forms in supra-tidal environments well above the highest water levels due to splashing and spraying, however Knight (2007) dismissed this conclusion arguing that beachrock forms in the beachface proper by well-established processes (described in McLean, 2011).

Hopley (1971, 1975, 1980) interpreted raised beachrock preserved on continental islands in north-eastern Australia, as evidence that sea level had been higher, reporting deposits 3 to 4 m above the current mean high water springs tide level (thought to represent the upper limit of cementation). This original estimate of elevation was later revised down to 3 m above the present HAT level (Hopley, 1983b, 1986b). Assuming that this elevation documents the peak highstand sea level, it must be noted that it is still 1.0 to 1.5 m higher than microatoll (highest ‘unmoated’ microatoll is +1.45 m) and oyster bed (+1.65 m: Beaman et al., 1994) data from the Great Barrier Reef.

The use of beachrock as a sea-level indicator is also problematic because it is difficult to determine a precise age on its formation. The constituent grains that formed the beach sand before lithification are almost certain to be of widely differing ages and pre-dating the cementation event. Accordingly, in principle it would be preferable to directly date the cements instead. Well-cemented beachrock may have undergone several diagenetic phases (Vousdoukas et al., 2007), and consequently the cements from the one deposit can generate different ages to any biogenic carbonate (Desruelles et al., 2009). While the ages of beachrock cements provide a minimum timing of deposition, the analysis of shell and coral material in beachrock provides a maximum age (Hopley, 1983b, 1986b), and this could span a wide range, for example shells on modern beaches can be up to 8000 years old (Donner and Junger, 1981) or possibly even reworked from previous interglacial successions (Murray-Wallace and Belperio, 1994). Due to the issues related to uncertainties relative to sea level at the time of formation, mixed age of biogenic carbonates and post-depositional diagenetic processes, beachrock should only be used as a sea-level indicator with caution and is only acceptable if the elevations agree with other recognised indicators (Smithers, 2011).

3.8 Foraminifera

Distinct changes in microfossils, particularly foraminiferal assemblages, have long been used for sea-level reconstructions on salt marshes in temperate regions around the world, most recently using complex statistical ‘transfer functions’ (Gehrels, 2000; Gehrels et al., 2001; Woodroffe, 2009, and references therein). The use of foraminifera in this way requires detailed studies on the distribution and zonation of foraminifera in modern environments (Haslett, 2007), on the basis of which it may then be possible to interpret changes in assemblages in sediment cores in terms of past sea-level adjustments (e.g. Haslett et al., 2010). Although microfossil studies have been conducted in southern Australia to reconstruct the characteristics of the post-glacial marine transgression (Belperio et al., 1988; Cann et al., 1988, 1993, 2002, 2006; Wang and Chappell, 2001), it is only recently that the ‘transfer function’ approach has been adopted in the Australasian region to decipher sea-level changes over the past few millennia (Southall et al., 2006; Woodroffe, 2009; Callard et al., 2011; Gehrels et al., 2012).

In the Australian context, these techniques have been extended from temperate saltmarsh systems to tropical mangrove environments, particularly those associated with the Great Barrier Reef (Horton et al., 2003, 2007; Woodroffe et al., 2005). Adopting this approach, Woodroffe (2009) produced a sea-level reconstruction for Cleveland Bay in north Queensland based on benthic foraminifera. The results indicate a prolonged mid-late Holocene highstand with sea level reaching elevations as high as 2.8 m above PMSL. These elevations conflict with evidence for a less pronounced highstand using other sea-level indicators for the region, such as coral microatolls and oyster beds. The taphonomy and elevational accuracy with which these microfossils can be used as sea-level indicators has been questioned as they can clearly be transported and redeposited (Smithers, 2011; Perry and Smithers, 2011). Recent studies suggest that the foraminiferal transfer function may be inappropriate in tropical environments due to complex mixing, bioturbation and degradation processes that affect preservation potential and modify assemblage composition (Berkeley et al., 2009a, b). A prograded coastal plain can provide a sedimentary record indicating little overall change in sea level over the mid-late Holocene (e.g. Thom and Roy, 1985), but intertidal microfossils (foraminifera, mangrove pollen) may preserve a more confused history because of compaction, contamination or diagenetic changes (oxidation of near-surface sediments and the development of potential acid-sulphate soils). The full potential of these methods needs to be further examined in a broader range of environments around Australia, but the approach appears to be most suitable for low-energy microtidal environments outside of the tropics (Callard et al., 2011; Gehrels et al., 2012).

3.9 Intertidal erosional indicators

In many parts of the world, past sea-level highstands are best indicated by erosional evidence. Intertidal erosional indicators are typically carved into softer sedimentary rocks (Pirazzoli, 1996; Smithers, 2011). Indeed, there has been a long history in Australia using erosional rock platforms as evidence for higher sea level around the Australian mainland (reviewed in Langford-Smith and Thom, 1969). Raised and submerged notches in Western Australia were used in Fairbridge's (1961) sea-level compilation and subsequent researchers have analysed relict tubeworm deposits preserved within notches on Rottnest Island (Playford, 1988; Baker et al., 2005), presuming that these deposits provide a minimum age for notch formation. Although notches can provide visually persuasive evidence of former sea-level positions, they can rarely be dated accurately.

3.10 Summary

The broad coastal plains, sand barrier systems and estuaries of southern Australia, and the coral reefs and extensive deltaic-estuarine plains of northern Australia have developed during the mid and late Holocene when the sea has been relatively close to its present level. These prograding stratigraphical sequences provide a first-order indication of the accumulation of shoreline deposits, and they contrast with the coastal landforms that have developed in parts of the world that have not experienced such relative sea-level stability, either because of vertical tectonic movements or various isostatic influences on relative sea-level changes.

Particular sea-level indicators are more likely to be formed, and preserved, under different sea-level scenarios; some are dependent on suitable accommodation space. Despite the limitations of geographical distribution and the patchy spread of evidence, as well as the uncertainties in indicative meanings that may be associated with proxy sea-level indicators, they all clearly have a role in indicating aspects of past sea levels. The accuracy and precision of each, and the reconciliation of the apparent discrepancies between them, are the key challenges for those who seek a more regional insight into past sea-level change, as well as those who are attempting to develop histories of local coastal landform development. The following section provides examples of how different sea-level indicators have been used at specific locations around Australia. It demonstrates how no one proxy sea-level indicator or specific site contains enough data over a sufficient interval of the post-glacial to provide evidence for a continuous sea-level history. Accordingly, in an attempt to provide a comprehensive review of sea-level research around the Australian margin, we have split the sections into two specific intervals of time (20,000 to 7000 cal. yr BP and 8000 cal. yr BP to present) with the latter period further divided to reflect geographical distribution.

4 Post-glacial sea-level rise (Last Glacial Maximum to 7000 cal. yr BP)

The first estimate for the LGM sea-level lowstand in the Australian region was produced by van Andel and Veevers (1967) who reported a -130 m sea level at ~18,000 years BP based on shell material from the Sahul Shelf. Subsequently, a sea level of -175 to -150 m at ~17,000 years BP was reported for the southern Great Barrier Reef (offshore One Tree Island) based on a submerged dredged *Galaxea clavus* coral colony (the modern range of this species can occur up to 75 m below sea level) and a lithified sample interpreted as ‘beachrock’ (Veeh and Veevers, 1970). Phipps (1970) found evidence in New South Wales for sea levels at -130 m around 18,000 years BP. Lowstand sediments, containing alternating successions of finegrained mixed quartz-carbonate sand and densely packed mollusc-dominated sediments with the shallow-water molluscs *Pecten fumatus* and *Placamen placidium* commonly represented, were identified in vibrocores collected from the outer continental shelf off New South Wales in depths between 123 and 152 m below PMSL, in broad agreement with depth suggested by the earlier evidence. AAR dating of the shallow water shells indicates that these successions represent deposition during the past three glacial maxima (Ferland et al., 1995; Murray-Wallace et al., 1996, 2005), confirming that sea levels during these maxima were at least 120 m below PMSL (Murray-Wallace et al., 2005).

A more detailed record from cores collected in the Joseph Bonaparte Gulf by Yokoyama et al. (2000, 2001a) places sea level at -124 m (± 4 m) at ca. 20,000 years BP. While some disagreement remains on the precise timing of the LGM (Peltier, 2002; Peltier and Fairbanks, 2006; De Deckker and Yokoyama, 2009), the magnitude of the sea-level lowstand at -120 to -130 m in the Australasian region has generally been agreed upon (e.g. Hopley and Thom, 1983; Chappell, 1987; Lambeck and Chappell, 2001; Murray-Wallace et al., 2005) and is consistent with glacio-eustatic ice sheet models (e.g. Lambeck and Nakada, 1990; Peltier and Fairbanks, 2006) and evidence collected from the Sunda Shelf (Hanebuth et al., 2000, 2009). Forthcoming results from IODP expedition 325 to the Great Barrier Reef are likely to further resolve the timing and depth of sea level at the LGM, and the nature of subsequent post-glacial sea-level rise (Webster et al., 2011; Yokoyama et al., 2011).

The key records of post-glacial sea-level rise from the Australian sites (Fig. 2) as well as from Barbados and Tahiti show remarkable agreement when corrected for tectonic influences (e.g. Lambeck and Chappell, 2001; Woodroffe, 2003). Although the broad pattern of eustatic post-glacial sea-level rise has been independently supported by these other studies, there continue to be further refinements to the detail, particularly associated with the relatively fast rise during meltwater pulse 1A (and possibly 1B and 2), and a slowed rate of rise during the Younger Dryas. Ooid deposits formed in shallow marine environments in the southern Great Barrier Reef place sea level at approximately 100 m below PMSL at 16,800 years BP just prior to meltwater pulse 1A (Yokoyama et al., 2006). Meltwater pulse 1A is particularly well-represented in the Sunda Shelf record (detected largely using mangrove facies). Sea level rose about 16 m in only 300 years between 14,300 to 14,600 years BP (Hanebuth et al., 2000, 2009). Following this time, the coral core data from the uplifted reef platforms on Huon Peninsula, Papua New Guinea, provide a relatively complete record from 13,000 to 7000 years BP showing a continuous rise over this period (Chappell and Polach, 1991; Edwards et al., 1993; Ota and Chappell, 1999). This period is also partially covered by other datasets from locations including Christchurch, New Zealand (Gibb, 1986), Great Barrier Reef (Larcombe et al., 1995; Larcombe and Carter, 1998), New South Wales (Sloss et al., 2007), southern Australia (Belperio et al., 2002; Cann et al., 2006) and Western Australia (e.g. the Abrolhos Islands, Eisenhauser et al., 1993) (Fig. 2).

In 1982 the Sirius Expedition, led by Nic Flemming, surveyed near Cootmaundra Shoal, northwest of Bathurst Island and 240 km from Darwin, looking for archaeological evidence of early human colonisation of this part of north-western Australia. Terraces at 90, 59, 48, 34 and 25 m depth were identified on the van Diemen Rise (van Andel and Veevers, 1967; Chappell and Thom, 1977), from which reefal carbonate shoals arise, with Cootamundra Shoal itself the shallowest at 16 m depth. The shoals presently support scattered live *Turbinaria, Platygyra* and *Porites* corals, but beneath a cemented algal crust, sticks of branching *Acropora* were recovered. Radiocarbon ages of 8460 ± 100 years BP (ANU3259) from 28 m and 7630 ± 100 years BP (ANU3257) from 20 m are broadly consistent with the anticipated sea-level curve (Flemming, 1986), although a further 14 unpublished ages were also obtained. Coral rubble was recovered to depths of at least 60 m, and inferred to 80 m, underlain by incised siliceous sandstone. A prominent terrace level was noted at 24-28 m below PMSL.

Detailed palaeoecological and sedimentary investigations reveal that rising post-glacial sea levels breached the Arafura Sill ca. 12,200 cal. yr BP with full marine conditions in the Gulf of Carpentaria established by 10,500 cal. yr BP (Chivas et al., 2001; Yokoyama et al., 2001b; Reeves et al., 2008). U-series ages of the drowned coral reefs at depths of 20-30 m in the southern Gulf of Carpentaria indicate that reef growth commenced between 10,500 and 9500 cal. yr BP associated with marine inundation into the Gulf and continued to flourish until ca. 7000 yr BP (Harris et al., 2008). Similar extensive fossil reefs in 30-50 m water depth have been described from around Lord Howe Island, at present the southernmost limit to reef growth in the Pacific Ocean, with radiocarbon and U-series ages from the same period between 9100 and 7300 cal. yr BP (Woodroffe et al., 2010).

Some researchers have argued for a highly episodic post-glacial sea-level rise (up to nine pauses) using data from the Great Barrier Reef and New Zealand (Carter and Johnson, 1986;

Carter et al., 1986; Larcombe et al., 1995; Larcombe and Carter, 1998). These data (including undated abrasion platforms/shorelines/drowned reefs and dated sea-level indicators) have been interpreted to represent several stillstands or even a 6 m regression (centred around 8200 years BP) during this period. A large oscillation around this time was shown in the sea-level curve proposed by Larcombe et al. (1995); this oscillation/regression was originally suggested as a possibility by Thom and Chappell (1975). Harris (1999; see also Harris and Davies, 1989) rejected this claim highlighting that there is no evidence for a 6 m sea-level fall in ice sheet models or in the marine $\delta^{18}O$ record, and Hopley et al. (2007) offered another critique arguing that the elevation and error terms associated with the data were too large to achieve the precision necessary to identify the proposed oscillation. There is an increasing body of evidence from around the world that there may have been a rapid rise at around 8200 years BP. An inflection occurs in a sea-level reconstruction from Singapore (Bird et al., 2007) which appears to coincide with the back-stepping identified in several southeast Asian deltas at that time (Hori and Saito, 2007). Such a rapid sea-level rise might explain the apparent drowning of submerged reefs around Lord Howe Island, and even those in the Gulf of Carpentaria (Harris et al., 2007, 2008; Woodroffe et al., 2010). However, although this phase of reef drowning would be consistent with such an inflection in the sea-level curve (meltwater pulse 2), there is no direct evidence for a sea-level stillstand or regression during this period in the latest available data (Fig. 2).

5 Mid to late Holocene sea-level changes (~8000 cal. yr BP to present)

Since the earlier sea-level compilations by Hopley (1983a, 1987) and Chappell (1987), new research has provided additional insights into Holocene sea-level variability around Australia. While there seems to be a general consensus about when sea level attained its modern level and the magnitude of the mid-Holocene highstand (albeit with some remaining discrepancies), there appears to be regional variability and conflicting evidence regarding the nature of late Holocene sea-level fall. In this section we divide the Australian mainland into several regions and review the key sea-level evidence and interpretations from the mid-late Holocene for each region.

5.1. Northern Australia (Gulf of Carpentaria/Northern Territory)

The principal mid to late Holocene records in northern Australia include studies focused on the South Alligator River using mangrove sediments (Woodroffe et al., 1985, 1987, 1989) and at Karumba using chenier deposits (Chappell et al., 1982; Rhodes et al., 1980; Rhodes,

1982) (Fig. 1). Mangrove evidence from the South Alligator River indicates that the transgression culminated in widespread mangrove forests, termed the 'big swamp' when sea level reached its present level around 7400 ± 200 cal. yr BP. The mangroves within this macrotidal setting extend across a vertical elevation of 3 m in the upper intertidal zone (Woodroffe et al., 1987; Fig. 3), and the big swamp facies contains no evidence that the sea was higher than present during the Holocene. Oxidation of the upper 1 – 2 m of the estuarine plains may have destroyed mangrove material, although there does appear to be a transition from mangrove pollen to pollen of grasses and sedges in the upper sections of some cores (Woodroffe et al., 1985).

In contrast, the chenier data from Karumba, in the southern Gulf of Carpentaria, indicate that sea level was much higher (approx. + 2.5 m) ca. 6400 cal. yr BP before falling smoothly to present level after 1000 cal. years BP (original ^{14}C ages in Rhodes, 1980; Chappell et al., 1982; Rhodes et al., 1982). This record of considerably higher sea level in the mid-Holocene provided one of the key datasets for modelling of hydro-isostatic adjustment in northern Queensland to explain the differences in shoreline elevation around the mainland (Chappell et al., 1982, 1983). We note that the reliability of these storm deposited sediments as precise palaeo-sea-level indicators is questionable (see section 3.6). Emergence along the southern and eastern shores of the Gulf of Carpentaria is further supported by studies of the deltas of the McArthur and Gilbert Rivers (Woodroffe and Chappell, 1993; Jones et al. 2003), and younger radiocarbon ages (ca.1500 to 3300 cal. yr BP) on beachrock indicating a sea level of +1 to 2 m near the McArthur River (Nott, 1996).

5.2 Queensland (east coast)

The Great Barrier Reef extends from Torres Strait, where only a few preliminary ages on microatolls have been published (Woodroffe et al., 2000), south to Lady Elliot Island (distance of 2,300 km), which comprises a sequence of ridges showing little variation in the pattern of accumulation over the past 5000 years (Chivas et al., 1986). This is a particularly wide area with large variability in the width of the continental shelf, and there are likely to be substantial variations in relative sea-level change across this region which cannot be covered in a single compilation (i.e. the importance of regionally/locally specific curves: Woodroffe, 2009; Lambeck et al., 2010). Geophysical modelling indicates a variable pattern of mid-Holocene sea-level change across the region (Chappell et al., 1982; Nakada and Lambeck, 1989). The time at which different reefs reached sea level varies from the outer reefs to the mainland coast (Hopley et al., 2007). In an early interpretation, Hopley (1984) invoked a high-energy window, whereby landforms on the mainland coast contained evidence deposited during high water levels associated with storm waves whose development was not constrained by the outer reefs at that stage. In the late Holocene, when the outer reefs had grown to sea level, that 'window' was closed, introducing a factor other than sea level to which indicators have responded. There are relatively few useful data available from southern Queensland, where fossil shell and coral suggest emergence of at least approximately 0.5 m at 5670 ± 190 cal. yr BP (original ^{14}C age in Flood, 1983). It is clear that there is a need for further research to continue to refine the sea-level history of the Great Barrier Reef and southern Queensland.

More recent studies have synthesised Holocene sea-level data from along the Queensland coastline. Larcombe et al. (1995) compiled all available sea-level data (excluding beachrock) from the central Great Barrier Reef (Cairns to Whitsunday Islands) over the past ca. 13,500 cal. yr BP. In summary, a synthesis of these data indicate that sea level approached its present position (within 3 m) by 8170 ± 200 cal. yr BP (^{14}C data from Hinchinbrook Island originally reported in Grindrod and Rhodes, 1984). This is consistent with Woodroffe's (2009) conclusion that sea-level reached modern levels between 8000 and 6200 cal. yr BP. Lewis et al. (2008) compiled the sea level data which they considered had the most reliably established and precise indicative meaning (e.g. intertidal sea-level indicators such as coral microatoll and encrusting organisms) from eastern Australian (Torres Strait to New South Wales, with a concentration of data from Queensland) over the past 7000 cal. yr BP. Based on a relict barnacle deposit from Magnetic Island sea level continued to rise to approximately +1 m by 7380 ± 240 cal. yr BP (Higley, 2000; Lewis et al., 2008) which is supported by ages from fossil coral microatolls (+0.7 to 0.9 m) from Magnetic Island (7012 ± 22 cal. yr BP: Yu and Zhao, 2010) and Orpheus Island (7530 ± 360 cal. yr BP: original ^{14}C data in Zwartz, 1995).

The Queensland sea-level indicator evidence supports conflicting estimates for the elevation of the mid-Holocene highstand. Sea-level estimates based on coral microatoll evidence places relative sea level at +1.3 to 1.5 m between 6770 and 5750 cal. yr BP (original ages provided in Chappell et al., 1983; Yu and Zhao, 2010), in good accord with that derived from oyster bed data at +1.6 m between 6280 and 5720 cal. yr BP (Beaman et al., 1994; Higley, 2000; Lewis et al., 2008). However, beachrock data suggest water levels of up to 2 to 3 m above present sometime between 6450 and 3050 cal. yr BP (Hopley, 1980, 1983b, 1986b) and the highstand suggested by foraminiferal transfer function analyses is +2.8 to 3.3 m between 4270 and 3580 cal. yr BP (Woodroffe, 2009) (Fig. 4). While a higher mid-Holocene sea level (> +2 m) cannot be completely ruled out, a +1.0 to 1.5 m highstand is currently the most accepted elevation for this region (Hopley et al., 2007; Lewis et al., 2008; Perry and Smithers, 2011). The most recent assessment indicates that microatolls and encrusting organisms are more accurate sea-level indicators than beachrock or foraminifera (Perry and Smithers, 2011).

There are conflicting interpretations of the mid to late Holocene sea-level fall in Queensland. The earliest interpretation using microatoll data suggested that sea-level fell smoothly to present levels (Chappell et al., 1983; Chappell, 1983). More recent data compilations have indicated a sustained highstand followed by a pronounced fall after 2000 cal. yr BP (Lewis et al., 2008; Woodroffe, 2009). Lewis et al. (2008) noted the possibility of up to two oscillations of about 1 m amplitude at approximately 4800 and 3000 cal. yr BP. Such marked oscillations have been discounted as an artefact of data interpolation (i.e. data gaps used to suggest oscillations and the validity of specific data points; Perry and Smithers, 2011). Perry and Smithers (2011) argue that the coral microatoll evidence should be given the most weight in sea-level reconstructions and contend that the smoothly-falling sea-level model remains the most plausible explanation for mid-late Holocene variability on this section of the Queensland coast. In contrast, the relict oyster bed deposits, which can accumulate for up to 1000 years in relatively sheltered locations, indicate that sea levels remained stable for long periods and provide evidence for a stepped sea-level fall (Lewis, unpublished data). The disparity between sites is only partly attributable to the regional pattern of sea-level change, as it is clear from geophysical modelling that there has been a variable hydro-isostatic response over the region. Further anomalies can be ascribed to the fact that the indicators record different tidal or supratidal levels, and the fact that there have been local changes in these levels as the coast evolves.

5.3 South-eastern Australia

The most recent compilation of Holocene sea-level data from south-eastern Australia presents many newly derived radiocarbon and AAR ages and a synthesis of previously published data from studies undertaken in the 1970's and 1980's (Sloss et al., 2004, 2007). A wider range of proxy sea-level indicators was considered, including molluscs from transgressive and estuarine sedimentary successions and *in situ* tree and mangrove stumps. The latter, in particular, imply that modern sea level was attained between 7900 and 7700 cal. yr BP (Jones et al., 1979; Sloss et al., 2007). Although this age is more narrowly constrained than the range estimated from the Queensland data (between ca. 8200 – 7500 cal. yr BP), and is broadly consistent (if not earlier) with the inferred ages for other parts of Australia, its validity rests on a single radiocarbon measurement on mangrove wood previously published by Jones et al. (1979) and the time-averaged age of the transgressive facies that extend to present sea level (Sloss et al., 2007).

The compilation by Sloss et al. (2007) indicates that sea-level reached a highstand of +1.0 to 1.5 m between 7700 and 7400 cal. yr BP, and remained at this elevation until 2000 cal. yr BP when sea level gradually fell to its present position (Fig. 5). The +1.0 to 1.5 m magnitude is marked by encrusting organisms/fixed biological indicators (tubeworms), molluscs and organic-rich muds and peats; this result is comparable with earlier reconstructions that suggest sea levels were no higher (or within ± 1 m) than present (e.g. Thom et al., 1969; Thom and Chappell, 1975; Thom and Roy, 1983, 1985). The higher mid-Holocene sea level suggested by Sloss et al. (2007) is also consistent with evidence from raised coral/boulder deposits and emerged carbonate mud deposits from Lord Howe Island that indicate mid-Holocene sea levels as high as +1.0 to 1.5 m (Woodroffe et al., 1995).

The prolonged mid-Holocene sea-level highstand followed by a fall to present levels (with the possibility of oscillations) proposed by Baker et al. (2001a, b) conflicts with the smooth-fall model that remains the favoured interpretation of Queensland data. The data comprising many ages on tubeworms reported by Baker and Haworth (2000a; Baker et al. 2001a, b) are central to their arguments. Baker and Haworth (2000b) suggest that there have been a series of sea-level oscillations based on statistical analysis of their data. They interpret up to three sea-level oscillations within the order of 1 m based on biostratigraphical relationships (i.e. assemblage changes etc). Sloss et al. (2007) and Perry and Smithers (2011) questioned the quality of the tubeworm and other encrusting organisms as accurate sea-level indicators, and suggested the sea-level oscillations more probably represent the adjustments of intertidal species to variations in coastal exposure and/or variable wave and climate conditions during the Holocene, or may simply be an artefact of missing data. This led Sloss et al. (2007) to conclude that the culmination of the Holocene marine transgression was followed by a sea-level highstand that lasted until about 2000 years ago, followed by a relatively slow and smooth regression of sea-level from +1.5 m to present level. Donner and Junger (1981) found evidence for a stable sea level with ‘no detectable fluctuations’ for the past 3200 years using radiocarbon ages of mollusc beds within low energy back barrier deposits from the Cullendulla Creek area (Bateman’s Bay), New South Wales. Seams of heavy minerals (ilmenite, rutile, zircon) within a prograding beachridge at Cudgen, New South Wales at or above the high water mark also provide evidence for a stable sea level over the period of sediment accumulation (Thom and Roy, 1985). The debate as to whether a smoothly falling sea level, a stable and prolonged highstand, or a highstand punctuated by oscillations occurred is currently unresolved and will continue until the precise indicative meaning and quality of existing data are better understood or new data are acquired (Baker et al., 2001a, b; Sloss et al., 2007; Perry and Smithers, 2011). In any case, the documented changes in the ecology, morphology, elevation and $\delta^{18}O$ composition of tubeworm deposits (Baker and Haworth, 2000a; Baker et al., 2001a, b) are likely to provide important insights into changing water level, climate, wave exposure, ENSO and Pacific Decadal Oscillation. Indeed, there is potential that such indicators are recording short-term transient spikes from ENSO-related steric influences that are not recorded by ‘slow response systems’ such as mangrove deposits.

5.4 South Australia

The most comprehensive sea-level compilation for South Australia was derived largely from a series of sediment cores and backhoe excavations along Spencer Gulf and the Gulf of St Vincent by Belperio et al. (2002). Transitions in sedimentary facies (seagrass, sandflat, mangrove, samphire and chenier ridge) coinciding with various tidal levels were recognised, measured relative to present sea-level position, and dated. As with the other sections of the Australian coast, the dataset has been compiled from over 25 years of research and contains an overview of the earlier studies, particularly those by Burne (1982), Belperio (1993), Belperio et al. (1983, 1984), Short et al., (1986), and Harvey et al. (1999). The data show that present sea level was reached between 8000 and 7500 cal. yr BP (original ^{14}C data in Belperio et al., 2002 and also in Bowman and Harvey, 1986).

The elevation of the mid-Holocene highstand was variable along Spencer Gulf and the Gulf of St Vincent with higher sea levels (up to + 3 m) being recorded at the northern-most sites (Redcliff) farthest from the continental shelf-edge. Lower highstands (+ 1 m) were recorded at the more southern locations (Port Lincoln; Belperio et al., 2002; Fig. 6). While the earlier work in this region by Belperio and colleagues indicated that recent neotectonic activity within the Torrens Hinge Zone adjacent to the Flinders Ranges explained the highstand and late Holocene sea-level fall (Belperio et al., 1983, 1984), this interpretation has since changed to account for the influence of hydro-isostasy (Chappell, 1987; Belperio et al., 2002). In fact, the dataset provides some of the strongest evidence for hydro-isostasy in the Australian region where it closely matches the modelled outputs (Nakada and Lambeck, 1989; Lambeck and Nakada, 1990).

The ‘earlier’ sea-level curves for Spencer Gulf suggested either a stepped or smooth sea-level fall over the mid-late Holocene (Burne, 1982; Belperio et al., 1983, 1984). Earlier evidence of a +3.8 m sea-level highstand for Spencer Gulf (Burne, 1982; Belperio et al., 1983) has since been revised down to +3 m (Belperio et al., 2002). Belperio et al. (2002) used only the most reliable data, conducted a more detailed statistical analysis, and highlighted the importance of separating the data to reflect geographical location. They showed that the mid-late Holocene sea-level fall could be explained by either smooth-fall (linear trend) or an oscillating fall (polynomial fit); however, the polynomial analysis favoured ‘gaps’ in the data rather than genuine points and, based on the simple progradational coastal sedimentary deposits, they favoured the smooth-fall as the most likely option (Belperio et al., 2002).

5.5 Western Australia

The sites that Fairbridge (1961) originally described in Western Australia remain important in investigations of Holocene sea-level changes; however, more recent research has included coral core records from the Houtman Abrolhos Islands (Eisenhauer et al., 1993) and reef-flat pavements (Collins et al., 2006). Other palaeo-sea level records from Western Australia include ‘swash zone’ facies from coastal plain deposits in the Perth Basin (Semeniuk, 1985, 1996; Searle and Woods, 1986), and intertidal tubeworm deposits largely from Rottnest Island (Playford, 1988; Baker et al., 2005). The coral core data indicate that sea level was within 2 m of its modern level by 7100 ± 70 cal. yr BP (Eisenhauer et al., 1993) while swash zone deposits place sea level at modern levels by 8015 ± 230 cal. yr BP (original ^{14}C age in Semeniuk, 1996). The time lag between the coral data and the other indicators is unsurprising as subtidal corals can lag sea level by long periods (e.g. Hopley, 1986a; Hopley et al., 2007). Unless the corals have formed microatoll or pavement morphology, they may be from a large depth range and should only be considered as directional sea-level indicators.

The reported magnitude of the mid-Holocene highstand is variable along the Western Australian coastline, with estimates of +2.0 to 3.5 m at Becher/Rockingham, Perth Basin (swash zone facies: Semeniuk, 1985; Searle and Woods, 1986; Searle et al., 1988) and at Rottnest Island (tubeworms: Baker et al., 2001b, 2005), +2.05 m at the Houtman Abrolhos Islands (coral pavements: Collins et al., 2006), and +1.5 m in Shark Bay (shell material: Logan et al., 1970). However, the precise indicative meaning for most of these indicators is unresolved. Nevertheless, there is a degree of concordance between the tubeworm data from Rottnest Island (+2.2 m: Baker et al., 2005) and coral pavement data from the Houtman Abrolhos Islands (+2.0 m: Collins et al., 2006), which may provide an upper constraint on the elevation of the mid-Holocene highstand (Fig. 7).

The three most complete records of mid to late Holocene sea-level change in Western Australia are from swash zone facies (Searle and Woods, 1986), tubeworm deposits (Baker et al., 2005) and coral pavements (Collins et al., 2006) (Fig. 7). The swash zone facies and coral pavements are interpreted to represent a smooth sea-level fall from +2.0 to 2.5 m to present levels after 1000 cal. yr BP (Searle and Woods, 1986; Collins et al., 2006). In contrast, the tubeworm deposits have been interpreted to indicate an oscillating sea level (Baker et al., 2005). However, these conflicting interpretations may simply reflect different statistical treatment of the data (see Belperio et al., 2002). Semeniuk and Searle (1986) reported completely different sea-level reconstructions (i.e. no highstand, smooth-fall, highstand then fall) for three sites separated only by 170 km within the Perth Basin. These discrepancies were thought be a function of local tectonism, although Baker et al. (2005) reported relatively consistent sea-level heights (approximately +1.5 m) from tubeworm deposits along the south-western coastline. The accuracy of swash zone facies as sea-level indicators needs to be resolved as their uncertainties may be too large to produce high quality reconstructions. Wyrwoll et al. (1995) argued that these swash zone deposits are influenced by coastal morphodynamic conditions (in particular hydrodynamic controls) that can change over time and are also variable even across this particular stretch of coastline.

In northern Western Australia, mangrove muds that were deposited during the mid-Holocene ‘big swamp’ phase also marked the time sea level stabilised around its present level (Woodroffe et al., 1985). Similar exposures of mangrove stumps were described from the tidal flats flanking the Ord River (Wright et al., 1972), but they remain undated. The emerged tidal flat deposits of the Ord River estuary is interpreted to have formed during a period where the adjacent embayment would have been much deeper (prior to the development of the Ord Delta and the progradation of coastal sediments) and as such the tidal amplitude/wavelength would have been much greater at that time (Wright et al., 1972). The stratigraphy and sedimentology of the tidal flats flanking King Sound (tidal range > 12 m) have been described in detail by Semeniuk (1980a, b, 1981, 1982). His descriptions included reference to mangrove muds with abundant organic remains including *in situ* stumps, which he assigned to the Christine Point Clay. In early studies this clay was considered Pleistocene in age. However, radiocarbon ages extending back to 8415 ± 1740 cal. yr BP (original ^{14}C age in Jennings, 1975) on similar muds in the adjacent Fitzroy River estuary suggest a Holocene age. Radiocarbon ages on wood fragments back to 7630 ± 200 cal. yr BP from Cambridge Gulf, adjacent to the Ord River estuary (originally reported in Thom et al., 1975) also suggest a mid-Holocene heritage.

5.6 Other locations

Detailed time-series observations of Holocene sea-level changes in Victoria or Tasmania are rare. Gill (1983) reported highstands around +2.0 m in Victoria, and Haworth et al. (2002) described a relict tubeworm deposit at +1.5 m at Wilson's Promontory dated at 5570 ± 250 cal. yr BP (original ^{14}C age in Haworth et al., 2002). The emerged shell bed evidence from the Gippsland Lakes presented by Gill (1983; Gill and Hopley, 1972) has since been explained by Thom (1984b) to reflect locally elevated water levels due to changing morphodynamics in the region (i.e. prior to the closure of the lagoon). Indeed, Thom (1984b) showed that the morphostratigraphical relationships of coastal sand barrier deposits in the Gippsland area are strongly influenced by changing energy conditions (wind and waves), sediment supply and sea level. For example, washover deposits become less common (or cease) in the stratigraphical record as the shorelines prograde seaward.

Bowden and Colhoun (1984) suggested that there was no clear evidence of a mid-Holocene highstand in Tasmania but Baker et al. (2001b) inferred a +0.5 m sea level from a relict tubeworm deposit at King Island dated at 2400 ± 330 cal. yr BP, and Colhoun (1983) cautiously suggested that shell material in a raised gravel beach deposit on Flinders island with an age of 3450 ± 400 cal. yr BP may indicate a highstand at +1.8 m. However, foraminiferal evidence (two dates: 4620 ± 330 and 6170 ± 280 cal. yr BP) from south-eastern Tasmania suggest that sea levels in the mid-Holocene were no higher than present (Gehrels et al., 2012). Murray-Wallace and Goede (1995) reported radiocarbon ages on the intertidal mollusc *Katelysia* sp. from back-barrier lagoon and embayment fill successions on Flinders Island. A *Katelysia* specimen from Cameron Inlet at + 0.5 m AHD yielded an age of 3600 ±

110 cal. yr BP (SUA-3001). Additional radiocarbon ages on *Katelysia* sp. ranging between 5580 ± 80 (SUA-3010) and 2060 ± 100 (SUA-3015) were based on specimens collected at AHD suggesting that sea level was relatively stable over this period. Jack Davies reported seaward sloping raised beach ridge sequences from Tasmania as evidence for higher Holocene sea level , but the ages and exact elevations were not detailed (Davies, 1959, 1961).

5.7 Summary

The evidence for higher Holocene sea level that Fairbridge described from Western Australia appeared from early radiocarbon ages to have occurred at different times within the past few millennia. His sea-level observations plotted with results from elsewhere in the world, produced a so-called global eustatic curve in the form of a succession of oscillations of sea-level rising above and falling below the PMSL. Recognising the geographical differences between the pattern of decelerating relative sea level experienced in the Caribbean and their 'Australian sea-level curve', Thom and Chappell (1975) decoupled areas with very different glacio-isostatic behaviour, and emphasised that the oscillations were not real, but an artefact of compiling data from a wide geographical range and differing relative sea-level histories. There has been a resurgence of the view that sea level has oscillated over a smaller amplitude during the mid-late Holocene (see Woodroffe and Horton, 2005; Baker and Haworth, 2000b; Lewis et al., 2008). Baker et al. (2005) specifically revisited Fairbridge's sites from Western Australia and with newly acquired radiocarbon ages from tubeworms, demonstrated that an oscillating curve could be fitted to their available data. A smoothly falling interpretation is considered preferable for dated reef-flat evidence from the Abrolhos Islands and also from other parts of Western Australia (Searle and Woods, 1986; Collins et al., 2006).

The 'early' sea-level studies used uncorrected (or 'environmentally corrected'; i.e. marine reservoir effect) ^{14}C ages; these suggested that sea level attained its modern elevation between 5500 and 6500 years BP (Hopley, 1983a). It has become increasingly clear based on calibrated ^{14}C, U-series and AAR ages that sea level reached present levels somewhat earlier, between 7500 and 8000 cal. yr BP.

Despite the reports of a +3 m mid-Holocene sea level by Fairbridge (1961 and references therein) in Western Australia, early sea-level research focused on whether Holocene sea levels were higher than present with research concentrated along the east coast of Australia (e.g. Hails, 1965; Thom et al., 1969, 1972; Gill and Hopley, 1972; Cook and Polach, 1973;

Belperio, 1979; Jones et al., 1979; Hopley, 1980). The sea-level evidence at this time was largely centred around relatively low-resolution indicators (beachrock, raised shell beds, freshwater peats, mangrove deposits) until the studies from the Great Barrier Reef Expedition and John Chappell's (1983) work on coral microatolls confirmed a +1.0 to 1.5 m mid-Holocene sea level on the inner shelf in north-east Queensland. Whereas most studies now recognise that sea levels around most parts of mainland Australia in the mid-Holocene reached between 1 and 2 m above present levels (e.g. Baker and Haworth, 2000a; Baker et al., 2005; Sloss et al., 2007; Lewis et al., 2008; Perry and Smithers, 2011), studies purporting a higher level between 2 and 3 m cannot be discounted entirely (e.g. Hopley 1971, 1975, 1978, 1980; Searle and Woods, 1986; Woodroffe, 2009). While some of this variation in magnitude may be explained by the geographical variability in the hydro-isostatic response around the Australian margin, further research is required to resolve the uncertainties associated with the various proxy sea-level indicators used (i.e. beachrock, swash zone facies and foraminiferal transfer function) before they can be fully integrated into sea-level reconstructions. In particular, these indicators need to be avoided where their elevations do not concur with the more established coral microatolls and encrusting organisms at the same location. The Gulfs in Southern Australia record magnitudes up to + 3 m (Belperio et al., 2002) and similar magnitudes have also been reported for Western Australia (e.g. Searle and Woods, 1986). Such variations in elevation and timing of changes in sea-level trends reveal that it is inappropriate to derive a single Holocene sea-level curve for the entire coastal margin of Australia and that there is regional variability in sea-level histories due to different isostatic influences, antecedent geomorphology of coastal margins, shelf characteristics, and oceanographic and climatic controls. Accordingly, the regional variability in sea-level histories as well as the subtle variations in the indicative meanings of different sea-level indicators in different settings and locations validates the construction of regional sea-level envelopes rather than a single sea-level curve.

The nature of the mid-late Holocene sea-level fall around mainland Australia is perhaps the most contentious issue remaining to be resolved. Proponents are split between a smoothly-falling sea level, a prolonged and stable mid-late Holocene highstand followed by a later fall, and an oscillating mid-late Holocene sea level. In essence, the debate is centred on two different types of sea-level indicators that include those that are part of accretionary landforms, such as prograding coastal and estuarine sedimentary successions, or coral reefs, coral microatolls on reef flats, coral pavements and unconformities in coastal plain successions; these indicators are more likely to record a relative sea-level fall. The other indicators include encrusting organisms, where numerous ages on scattered remnant deposits are suited to record a prolonged highstand with possible superimposed intermittent oscillations.

Clearly, continuous location-specific sea-level reconstructions based on a range of different types of sea-level indicators are highly desirable, and are more likely to credibly resolve relative sea-level history around the Australian margin. Indeed, indicators that record a regional pattern related to hydro-isostasy must be separated from those that record changes in local water levels produced by changes in coastal geomorphology over time. However, such diverse multi-proxy records remain elusive, and may remain so. Such a reconstruction is problematic because open water coral microatoll data for the past ca. 4000 years is sparse for north-eastern Australia (Lewis et al., 2008), probably reflecting the age structure of host reefs and the limited accommodation space for unmoated microatoll establishment once mature reef flats have been established (Smithers et al., 2006; Perry and Smithers, 2011). The encrusting organisms (particularly tubeworms) are well represented over this period (see Lewis et al., 2008) and indicate a prolonged sea-level highstand. While clear geographical transitions in biostratigraphical assemblages coupled with $\delta^{18}O$ fluctuations in the tubeworm data as well as some shifts in vertical elevation undoubtedly coincide with climatic and environmental changes (e.g. sea temperature and salinity), it remains disputed whether these changes represent genuine sea-level oscillations (see Lambeck et al., 2010; Perry and Smithers, 2011). Furthermore, oscillations in the tubeworm data as postulated by Baker et al. (2001a, 2005) could be the result of changing wave exposure regimes over the Holocene (Goodwin, 2003, Sloss et al., 2007) or reflect specific climate/ocean interaction (e.g. ENSO and steric effect of the East Australian Current) which are known to influence the elevations of such deposits (Laborel and Laborel-Deguen, 1996). The larger scale oscillations (~1 m) proposed by Baker and Haworth (2000b; Baker et al., 2001a, b, 2005) using their original tubeworm data (and re-assessed in Sloss et al., 2007) need to be discriminated from statistical artefacts where ‘gaps’ in the data are weighted towards oscillations (e.g. Belperio et al., 2002; Perry and Smithers, 2011).

6 Modelling Holocene sea-level changes

Recognition that various locations around the world have different relative Holocene sea-level histories and that there is no single ‘global’ solution (e.g. the work of the International

Geological Correlation Programme; Bloom, 1977; Hopley, 1983a), prompted other explanations for varying sea-level patterns including tectonic and isostatic influences. A key breakthrough came when geophysical modellers revisited an early hypothesis that the land could subside and rebound with increased and decreased ice and water loads (Walcott, 1972; Clark et al., 1978; Clark and Lingle, 1979); these models provided a powerful tool to explain relative sea-level changes around the world. Walcott (1972) highlighted the different responses that would be expected based on the location of former ice sheets coupled with sea-level indicator data from around the world. Further models to predict sea-level curves for more specific locations were developed by Chappell (1974), Clark et al. (1978), Clark and Lingle (1979) and Nakiboglu et al. (1983) where the contribution of hydro-isostasy for locations in the 'far-field' of former ice sheets (relevant for most parts of Australasia) were better quantified. However, the contribution of hydro-isostatic factors along the Australian mainland were not quantified until Chappell et al. (1982) and later Nakada and Lambeck (1989) and Lambeck and Nakada (1990) completed specific model predictions for this region. In the following discussion, we review these models (and later revisions) noting specific comparisons with the sea-level indicator data as well as examining other potential influences of mid-late Holocene sea-level changes.

6.1 Hydro-isostatic adjustments

The concept of hydro-isostasy relates to the viscoelastic response or flexure of the continental crust and mantle to varying water loads. Thus the key parameters (that are most difficult to quantify) within the models include mantle viscosity and lithospheric thickness as well as the 'ice' model such as the timing of meltwater influx to ocean basins and when it reduced or ceased (Lambeck and Nakada, 1990). In fact, the glacio-hydro-isostatic model of Lambeck and Nakada (1990) originally predicted that sea level reached its present position between 9000 and 7000 cal. yr BP which is in accord with the present concensus as discussed previously in this review. The hydro-isostatic model provides insights into the melting of different ice sheets and the volumes required to produce sea-level changes of the magnitude observed in the post-glacial sea-level rise. The Lambeck and Nakada (1990) model was the first to examine changes at far-field sites that combined the regional (water loading on shelf) and global data (i.e. glacio-isostatic changes).

The predictions of a smooth sea-level fall for the far-field sites stemmed from both observational evidence as well as the input data based on a negligible contribution from ice melting after 7000 to 6000 cal. yr BP. In particular, Chappell et al. (1982) 'calibrated' the hydro-isostatic model to observational data from Karumba (chenier ridges) and the Great Barrier Reef (coral microatolls) by matching the 'smooth-fall' response apparent in these indicators. They also introduced the concept of a 'hinge' response where the continental shelf margin down-warps to accommodate lower relative sea levels on the outer shelf and a corresponding upward flexure occurs in coastal areas, producing a relatively higher sea level on the inner shelf. The slow deformation of the crust-mantle system in the continental shelf and ocean margin areas following the post-glacial transgression results in a smooth sea-level fall along coastlines that are located adjacent to relatively wide shelves. Variations in the shelf width and depth result in variable mid-late Holocene sea-level histories; for example the highstands at Cairns, Queensland where the shelf width is around 80 km is much lower than the southern end of the Great Barrier Reef where the shelf extends to ~ 300 km offshore. Coral microatoll (up to +0.95 m: Chappell et al., 1983) and oyster bed (+1.2 m: Higley, 2000; Lewis et al., 2008) data from Princess Charlotte Bay (north of Cairns) show lower elevations compared with their southern counterparts (+1.4 and 1.6 m, respectively). The smooth sea-level fall from highstands of varying magnitudes depending on shelf width is also supported by the dataset from Spencer Gulf and the Gulf of St Vincent, South Australia, where the most inland locations coincide with the highest emergence (Belperio et al., 2002). We note, however, that some researchers are less convinced that the magnitude of the mid-Holocene highstand always varies with the width of the eastern Australian continental shelf (see Haworth et al., 2002).

Interestingly, adjustments to the timing of the meltwater contribution (i.e. continuing after 6000 to 7000 years BP) in the hydro-isostatic models will produce a prolonged mid-Holocene sea-level highstand (see Thom and Chappell, 1978; Lambeck, 2002; Peltier, 2002). Goodwin (1998) suggested that changes in Antarctic ice melt volumes during the Holocene could have also potentially affected late-Holocene sea-level fall. Indeed, the continued influx of meltwater was postulated by Sloss et al. (2007) and Woodroffe (2009) to account for the prolonged sea-level highstand observed in their datasets from eastern Australia.

While the hydro-isostatic modelling appears to yield outputs that agree with the regional datasets (e.g. Searle and Woods, 1986; Belperio et al., 2002; Collins et al., 2006), there are also discrepancies and anomalies in some regions. For example, mid-Holocene highstands on Rottnest Island and southern Australia (Haworth et al., 2002; Baker et al., 2005), and possibly also Lord Howe Island (Woodroffe et al., 1995) do not match the predicted hydro-isostatic response (Nakada and Lambeck, 1989; Lambeck and Nakada, 1990; Lambeck, 2002). Local tectonic influences have been implied for south-western Australia including Rottnest Island (e.g. Semeniuk and Searle, 1986; Playford, 1988; Nakada and Lambeck, 1989; Lambeck and Nakada, 1990), although these have been questioned by Baker et al. (2005) who found similar magnitude highstands along the Western Australian coastline and at Rottnest Island. Moreover, regional tectonic uplift along southern Australia and Tasmania (see Bryant, 1992; Murray-Wallace and Goede, 1995; Murray-Wallace, 2002) may be superimposed on the hydro-isostatic influence (e.g. Lambeck and Nakada, 1990). Indeed, Baker et al. (2001) and Haworth et al. (2002) found higher sea levels of +0.5 m at King Island and +1.5 m at Wilson's Promontory, both locations where hydro-isostatic modelling suggests there should be no highstand.

At this stage, there are inadequate observational spatial and temporal datasets (termed 'fragmentary') to satisfactorily inform the hydro-isostatic and neotectonic models (Lambeck and Nakada, 1990; Lambeck, 2002). While many valuable location-specific datasets have been produced, many of which have been used to 'calibrate' the model, a continuous high-resolution reconstruction remains elusive. Only when the apparent discrepancies between the 'prograding and accretionary' indicators (smooth-fall) and the encrusting organisms (prolonged highstand) are reconciled and the precise indicative meaning of included indicators is resolved can modelling be performed to account for the addition of extra meltwater during the late Holocene.

6.2 Other influences

It has been postulated from those in favour of possible ~1 m sea-level oscillations in the mid-late Holocene that they were produced by the break-up of ice sheets in either the north Atlantic ('Bond Cycles') or Antarctica (e.g. Baker et al., 2005; Lewis et al., 2008). Edmund Gill was another supporter of eustatic mid-late Holocene sea-level changes, and while he acknowledged the possible influence of hydro-isostasy he was a strong advocate for oscillating sea levels concluding his 1983 paper with "It is no longer a question of whether oscillations occurred, but what the oscillations were" (Gill, 1983, p. 63). Other studies have dismissed larger (~ 1 m) oscillations but consider the possibility of smaller-scale fluctuations related to climate forcing such as ENSO, Pacific Decadal Oscillation and changes in wave climate (e.g. Chappell, 1987; Goodwin, 2003; Sloss et al., 2007). These smaller scale changes in sea level would be superimposed on the eustatic and isotactic sea-level history and thus add complications in using proxy sea-level indicators that may be influenced by such ocean-atmospheric influences.

'Recent' tectonic movements have also been invoked to explain variations in the magnitude of the mid-Holocene highstand along the Queensland coast (e.g. Hopley, 1975) and in South Australia (Belperio et al., 1983, 1984), although latter work has shown that the magnitude variability in the South Australian evidence is due to hydro-isostasy (Belperio et al., 2002). Research on chenier plain deposits at Broad Sound, Queensland highlighted the possibility of local tectonic movements, where deposits on the western side of the Sound suggest no mid-Holocene highstand (Cook and Polach, 1973) but those on the eastern side suggest a highstand perhaps as large as 4-6 m above PMSL (Cook and Mayo, 1977).

Bryant (1992) reviewed the variable sea-level highstands of the last interglacial (based on the analysis of Murray-Wallace and Belperio, 1991) and mid-Holocene around Australia and found that there was possible down warping of northern Australia and up warping along the southern edge of the continent (including Tasmania). Most of the east coast of New South Wales and west coast of Western Australia were classed as relatively stable. Byrant (1992) argued this tectonic influence also explained the apparent variability in mid-Holocene sea-level maxima across Australia and contrasted the findings of the Lambeck and Nakada (1990) models which suggested higher sea levels to the north of the continent. However, Nott (1996) subsequently discovered evidence of a raised (approximately +3 m) fossil *in situ* reef believed to be of last interglacial age (although diagenesis prevented reliable dating of the deposit), suggesting that subsidence of the northern Australia margin to the degree predicted by Bryant (1992) is unlikely.

Another possibility that would influence the reliability of sea-level indicators is historical changes in the tidal range. The most precise sea-level indicators grow to a specific sea-level datum such as coral microatolls (approximately mean low water springs) and encrusting organisms (approximately high water neaps). As such, a change in tidal range could account for discrepancies between these indicators. Cook and Mayo (1977) demonstrated that there had been a change in the tidal range at Broad Sound of close to 1 m in the past 6000 years, although the biggest absolute vertical changes are likely to occur where tidal ranges are large (e.g. Broad Sound has up to a 9 m tidal range).

7 Conclusions

Fairbridge's pioneering research led, not to a global eustatic curve as he had anticipated, but to the recognition that the pattern of relative sea-level change in the Australian region differed from that observed in the Atlantic. A series of seminal sea-level studies were undertaken in the following 25 years. The stabilisation of sea level close to its present elevation in the mid-Holocene set the scene for the detailed reconstructions that were undertaken at different locations around the Australian mainland. The comprehensive reviews by Hopley (1983a) and Thom (1984a) summarise the state of knowledge about sea level and geomorphology around the Australian coast at that time. These early studies provide a solid foundation that remains the basis for current data compilations of sea-level studies in Australia over the ensuing 25 years (Hopley, 1982; Hopley et al., 2007). However, additional carefully selected and measured indicator data from specific locations are required to further refine the nature of mid-late Holocene sea-level changes at various locations.

Significant progress has been made to extend sea-level research in regard to the post-glacial sea-level rise, including research on the Huon Peninsula (e.g. Chappell and Polach, 1991), Joseph Bonaparte Gulf (Yokoyama et al., 2000, 2001a) and the Sunda Shelf (Hanebuth et al., 2000, 2009). Preliminary regional compilations and syntheses of sea-level data have been produced for north-eastern Australia (Larcombe et al., 1995; Lewis et al., 2008), New South Wales (Sloss et al., 2007) and South Australia (Belperio et al., 2002). There is broad agreement about a sea-level envelope encompassing a range of evidence within a broad swath of data variability, but there remains active debate about the specific details of post-glacial, and particularly Holocene sea-level behaviour at any one site. Establishing broader regional patterns of relative sea-level change has long been a goal of those researchers who have compiled datasets, but it is clear that sea level has behaved differently at different localities around our coast. The calibration of ^{14}C ages has allowed previous datasets to be directly aligned by accounting for global and regional marine reservoir effects and places sea level attaining modern elevations between 7500 and 8000 cal. yr BP around the coastal margin of Australia. Depending on the indicators of preference, sea level either fell smoothly from a +1 to 2 m highstand or remained at these levels for a considerable period before falling or oscillating to its present position.

Clearly, the selection of sea-level data for a synoptic assessment of sea-level change requires a thorough understanding of the limitations and integrity of each data point as accurate palaeo-sea-level indicators. However, the inherent uncertainty of the data can be somewhat difficult to estimate from the previous studies (i.e. the precision and accuracy of the elevation measurement to tidal datum, whether the sample was *in situ*). A way forward may be the development of systematic criteria in the selection of data points to ensure the quality of such compilations. It is recommended that sea-level studies in the future should ground truth indicators to sea-level datum using laser levelling technology, high precision GPS or satellite altimetry. Moreover, the key to quality sea-level reconstructions is a complete understanding of the indicative meaning, limitations and reproducibility of each indicator. The influence of tidal range and variability, and short-term climatic fluctuations (such as ENSO) on the various sea-level indicators also needs to be understood so that short-term and long-term deviations can be discriminated.

Great progress has been made since Fairbridge's initial compilation that has improved the knowledge of post-glacial sea-level change around the Australian continent, but there is still much to be resolved. A clearer understanding of past sea-level changes and their causes is urgently needed to better inform our ability to forecast future changes. A concerted effort is required, through the compilation of existing data, renewed fieldwork, dating analysis and modelling to address the issues of whether there have been oscillations of the sea surface and if so, of what magnitude. The pattern and rate of fall from the Holocene highstand to modern levels, and of the contributions of the various factors to this change, both global 'eustatic' or 'steric' components and local geophysical, tectonic and land instability issues also need to be addressed (see Belperio, 1993).

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Figure captions

Figure 1. Map of key sea-level sites around the Australasian region.

Figure 2. Summary of data showing the post-glacial sea-level rise for the Australasian region. The envelope is drawn to capture intertidal indicators and the zone between the terrestrial and marine directional indicators. Sites include New Zealand (NZ: e.g. Gibb, 1986), north-west shelf (NW shelf: e.g. Yokoyama et al., 2000, 2001a; James et al., 2004), Huon Peninsula (Huon: e.g. Chappell and Polach, 1991; Edwards et al., 1993; Ota et al., 1993; Chappell et al., 1996), Queensland (QLD: e.g. Larcombe et al., 1995), Sunda Shelf (Sunda: Hanebuth et al., 2000, 2009), Western Australia (WA: e.g. Einsenhauer et al., 1993; Semeniuk, 1985, 1996), Northern Territory (NT: Woodroffe et al., 1987), South Australia (SA: Belperio et al., 2002) and New South Wales (NSW: Sloss et al., 2007). The following vertical errors have been assigned to the data: ±3 m for the intertidal (Inter.) indicators, +10, -1 for the marine indicators and +1, -10 for the terrestrial (Terr.) indicators. Note that meltwater pulse 1A (1A) is well-represented in the Sunda Shelf dataset.

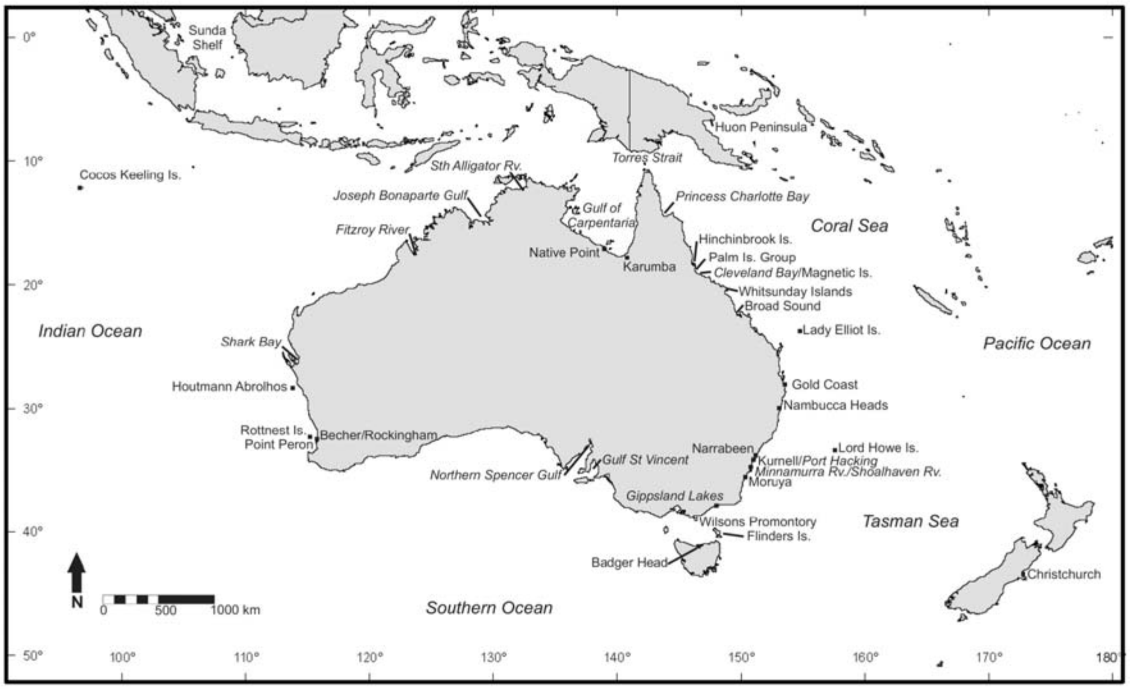
Figure 3. Data from mangrove material from the South Alligator River, Northern Territory (calibrated ^{14}C ages from Woodroffe et al., 1987).

Figure 4. Summary of sea-level data for the Queensland region (a). Indicators include barnacles (Beaman et al., 1994; Higley, 2000), beachrock (Hopley, 1980), foraminiferal transfer function (Woodroffe, 2009), mangroves (Larcombe et al., 1995), coral microatolls (Chappell et al., 1983) and oyster beds (Beaman et al., 1994; Higley, 2000; Lewis et al., 2008). Note the clear offset between the microatolls, barnacles and oysters compared with the beachrock and foraminifera data. The data fit within a tighter envelope when only the most reliable indicators are considered where the elevations can be directly measured to the modern counterparts (b; barnacles, microatolls and oyster beds).

Figure 5. Summary of key sea-level data from New South Wales (compiled in Sloss et al., 2007).

Figure 6. A selection of data from Spencer Gulf, South Australia to highlight the difference in sea-level magnitudes between Port Lincoln (PL), Redcliff (Red) and Port Pirie (PP) that varies with distance from the continental shelf (calibrated ^{14}C ages from Belperio et al., 2002).

Figure 7. Summary of the key sea-level data from Western Australia including barnacles, tubeworms (Baker et al., 2005), swash zone deposits (Searle and Woods, 1986; Searle et al., 1988) and coral pavements (Collins et al., 2006).



Sunda Shelf
Huon Peninsula
Torres Strait
Cocos Keeling Is.
Sth Alligator Rv.
Joseph Bonaparte Gulf
Gulf of Carpentaria
Princess Charlotte Bay
Coral Sea
Fitzroy River
Native Point
Karumba
Hinchinbrook Is.
Palm Is. Group
Cleveland Bay/Magnetic Is.
Whitsunday Islands
Broad Sound
Lady Elliot Is.
Indian Ocean
Shark Bay
Pacific Ocean
Houtmann Abrolhos
Gold Coast
Nambucca Heads
Rottnest Is.
Point Peron
Becher/Rockingham
Narrabeen
Lord Howe Is.
Gulf St Vincent
Kurnell/Port Hacking
Minnamurra Rv./Shoalhaven Rv.
Northern Spencer Gulf
Moruya
Gippsland Lakes
Wilsons Promontory
Tasman Sea
Flinders Is.
Badger Head
Christchurch
N
0
500
1000 km
Southern Ocean
0°
10°
20°
30°
40°
50°
100°
110°
120°
130°
140°
150°
160°
170°
180°

Figure 1.

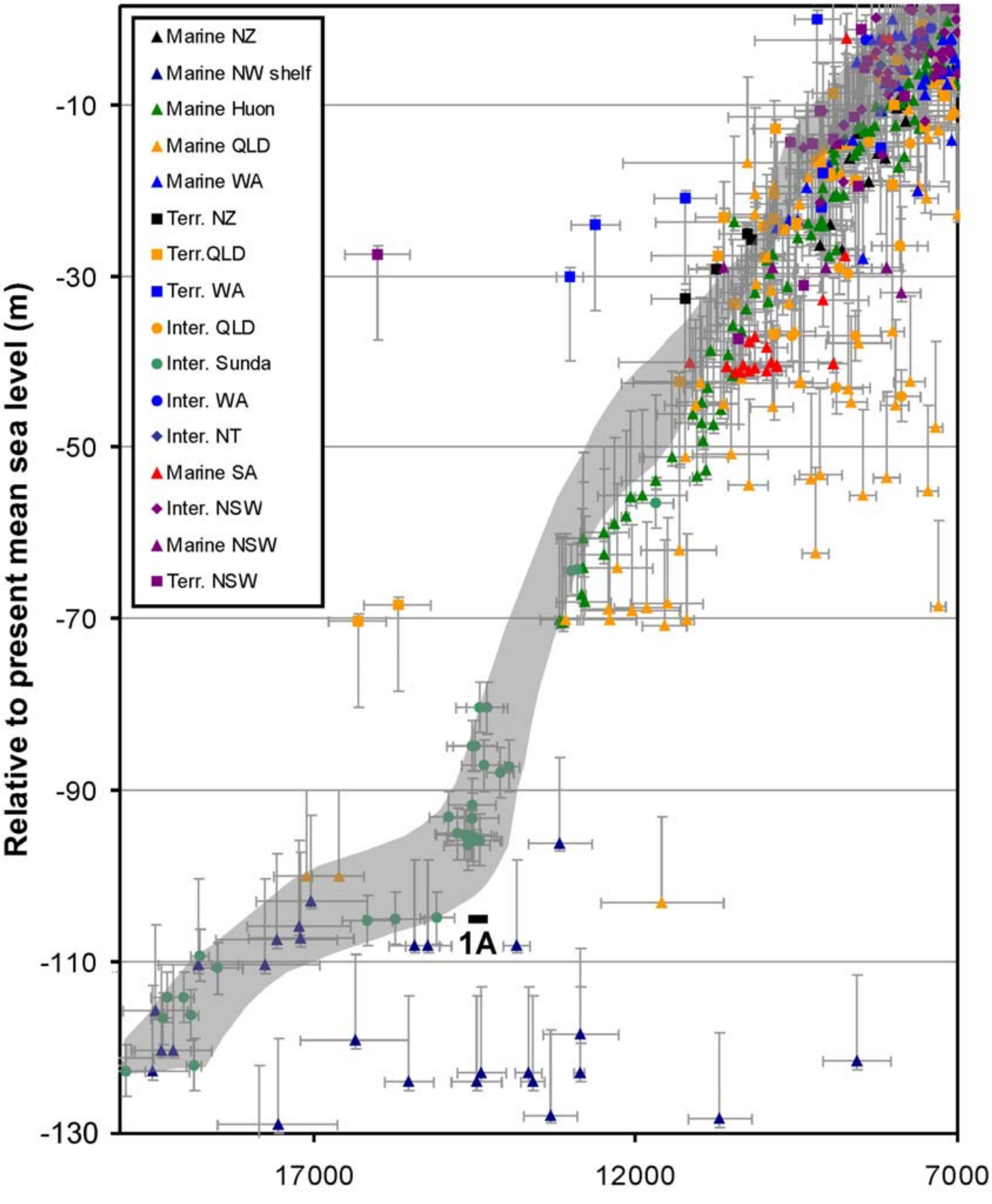
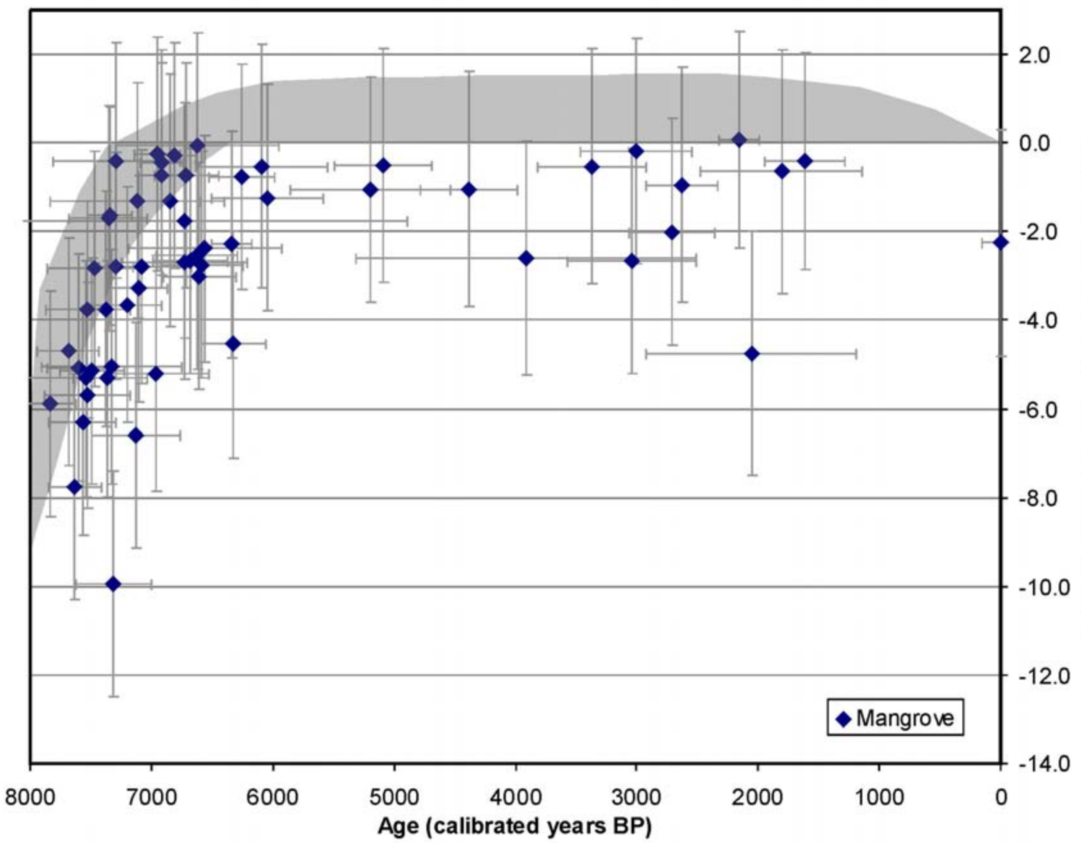

Marine NZ
Marine NW shelf
Marine Huon
Marine QLD
Marine WA
Terr. NZ
Terr.QLD
Terr. WA
Inter. QLD
Inter. Sunda
Inter. WA
Inter. NT
Marine SA
Inter. NSW
Marine NSW
Terr. NSW
Relative to present mean sea level (m)
-10
-30
-50
-70
-90
-110
-130
17000
12000
7000
Age (calibrated years BP)
1A

Figure 2.



2.0
0.0
-2.0
-4.0
-6.0
-8.0
-10.0
-12.0
-14.0
Relative to present mean sea level (m)
Mangrove
8000
7000
6000
5000
4000
3000
2000
1000
0
Age (calibrated years BP)

Figure 3.

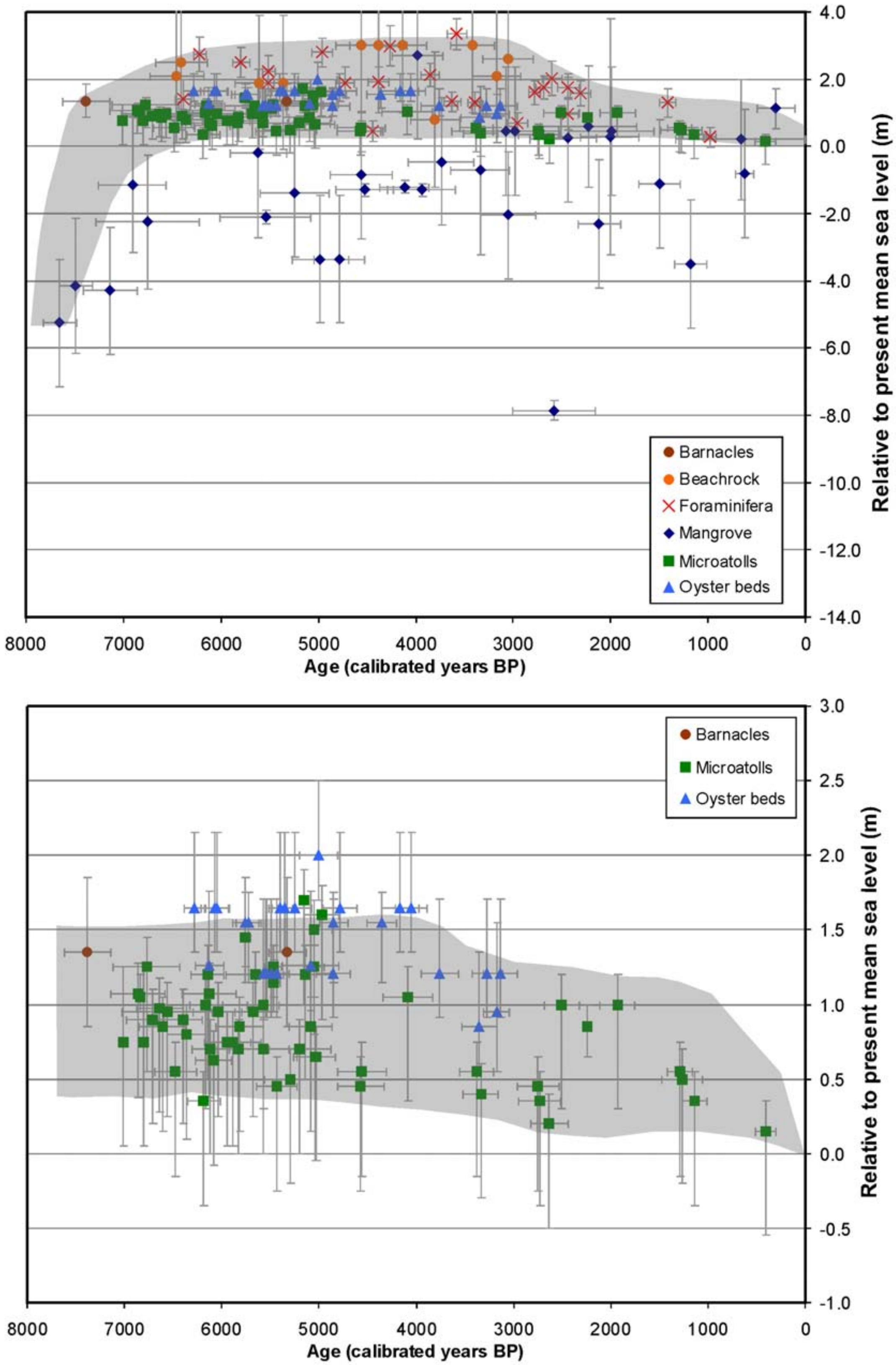

Barnacles
Beachrock
Foraminifera
Mangrove
Microatolls
Oyster beds
Relative to present mean sea level (m)
Age (calibrated years BP)
4.0
2.0
0.0
-2.0
-4.0
-6.0
-8.0
-10.0
-12.0
-14.0
8000
7000
6000
5000
4000
3000
2000
1000
0
Barnacles
Microatolls
Oyster beds
Relative to present mean sea level (m)
Age (calibrated years BP)
3.0
2.5
2.0
1.5
1.0
0.5
0.0
-0.5
-1.0
8000
7000
6000
5000
4000
3000
2000
1000
0

Figure 4.

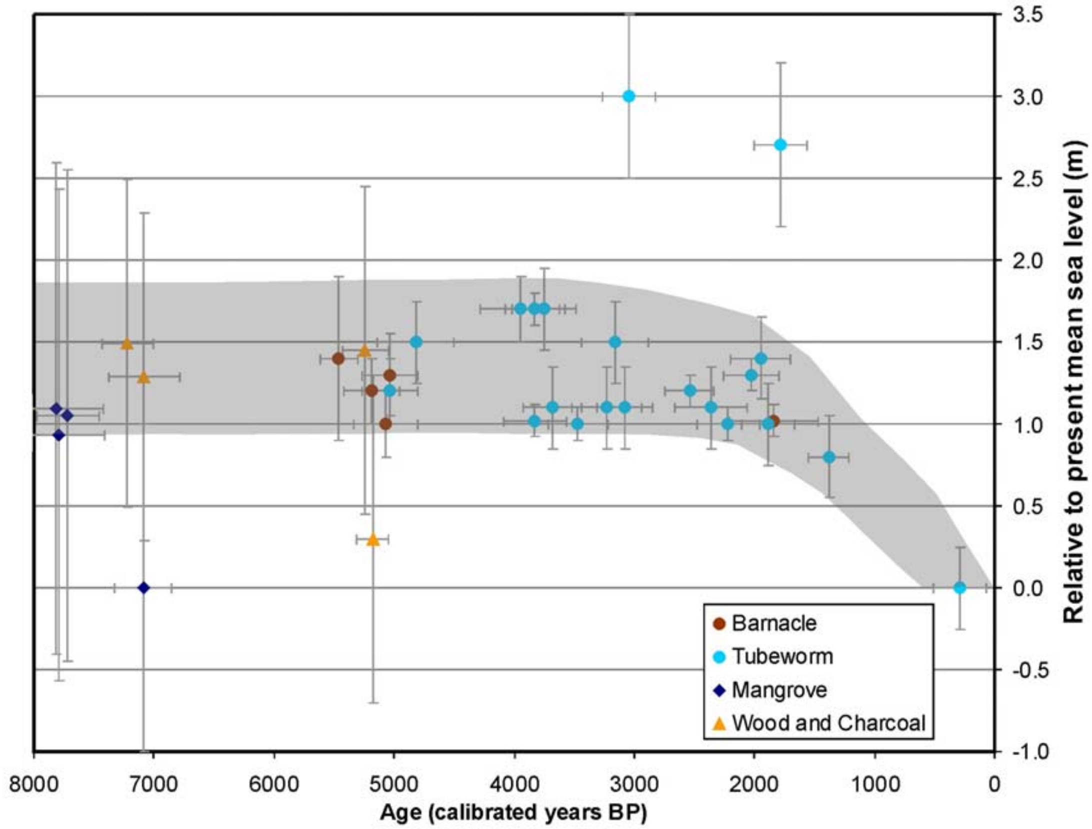

3.5
3.0
2.5
2.0
1.5
1.0
0.5
0.0
-0.5
-1.0
Relative to present mean sea level (m)
8000
7000
6000
5000
4000
3000
2000
1000
0
Age (calibrated years BP)
Barnacle
Tubeworm
Mangrove
Wood and Charcoal

Figure 5.

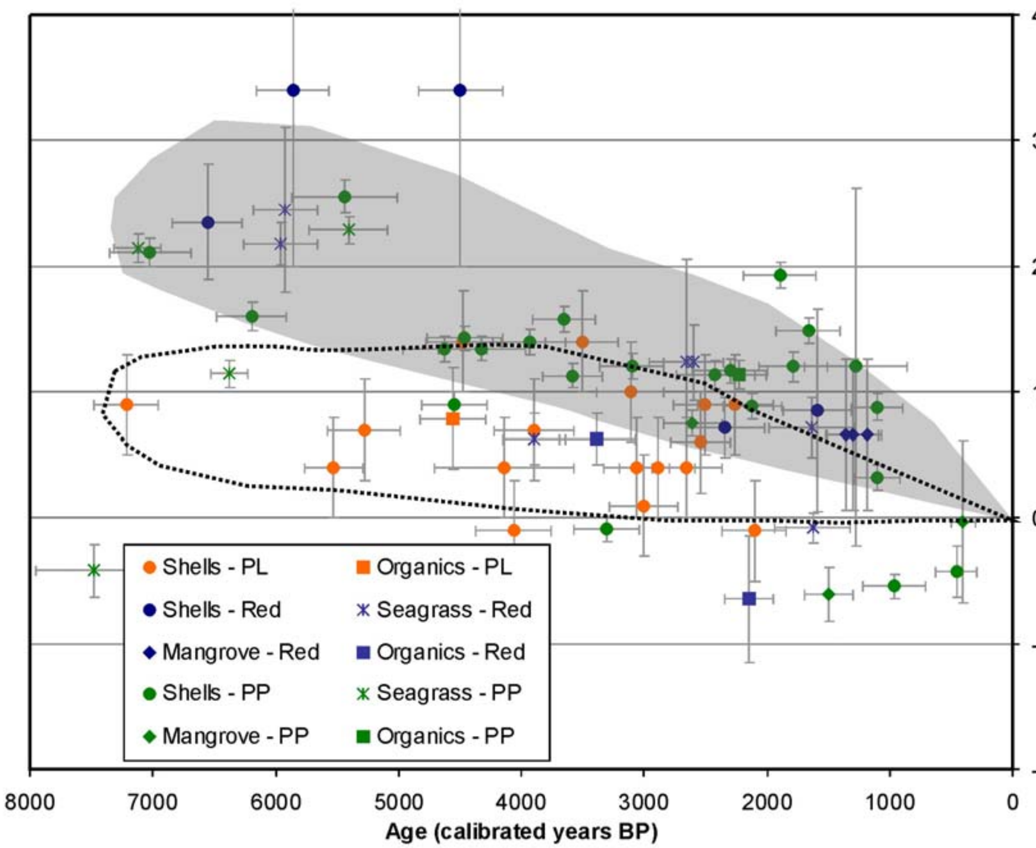
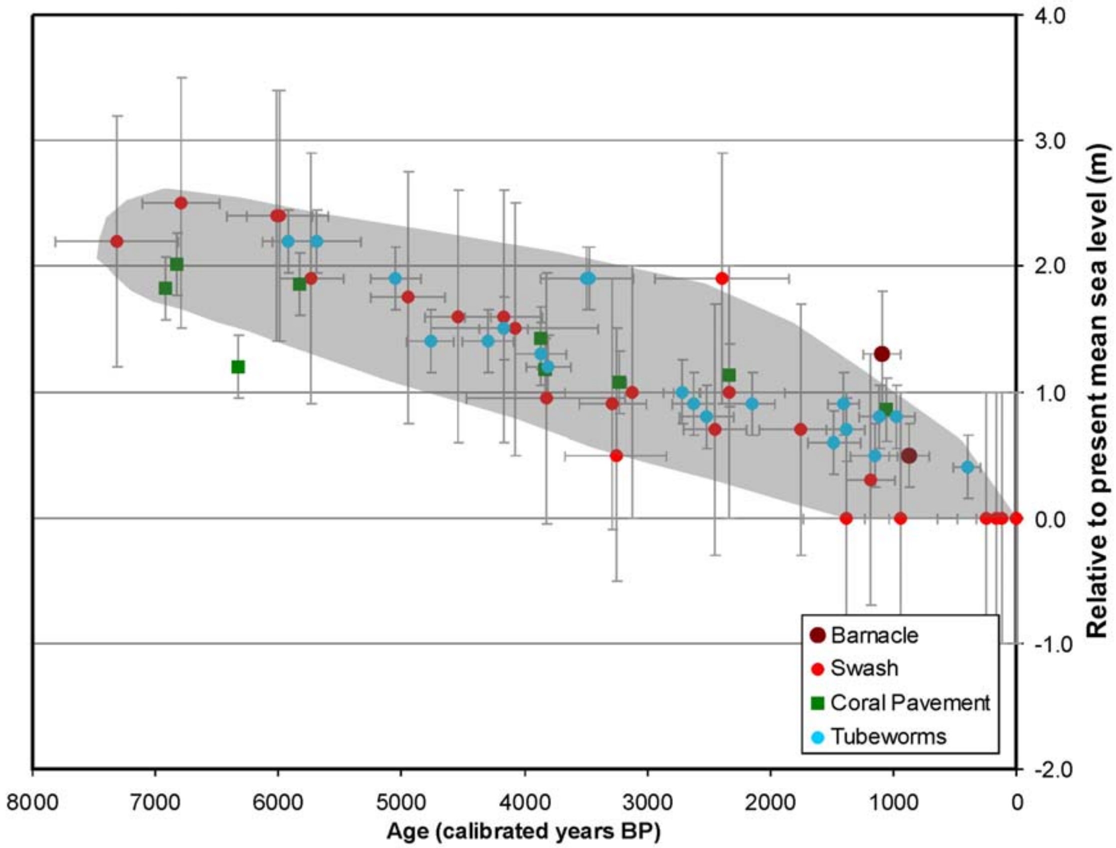

Relative to present mean sea level (m)
4.0
3.0
2.0
1.0
0.0
-1.0
-2.0
8000
7000
6000
5000
4000
3000
2000
1000
0
Age (calibrated years BP)
Shells - PL
Organics - PL
Shells - Red
Seagrass - Red
Mangrove - Red
Organics - Red
Shells - PP
Seagrass - PP
Mangrove - PP
Organics - PP

Figure 6.



2492
2493 Figure 7.