4-1-2005

Holocene sea-level changes in the Indo-Pacific

Sarah A. Woodroffe
University of Durham

Benjamin P. Horton
University of Pennsylvania, bphorton@sas.upenn.edu

Postprint version. Published in *Journal of Asian Earth Sciences* , Volume 25, Issue 1, 2005, pages 29-43.
Publisher URL: http://www.sciencedirect.com/science/journal/03781127

Holocene sea-level changes in the Indo-Pacific

Abstract
Holocene sea-level reconstructions exist from many locations in the Indo-Pacific region. Despite being a large geographical region, the nature of Holocene sea-level change is broadly similar in all locations. Differences do exist, however, in the timing and magnitude of the Mid-Holocene High Stand (MHHS) and the nature of late Holocene sea level fall across the region. When the Indo-Pacific is subdivided into smaller regions, these discrepancies do not disappear, and in some cases the discrepancies are large within a single coastline.

It is clear from this analysis that the fundamental criteria to produce accurate local relative sea-level curves are hardly ever met. There are serious problems associated with the correct interpretation of sea-level indicators and their relationship to mean sea level, and with the quality of age determinations. A consistent methodology throughout the Indo-Pacific for the analysis of sea level data is lacking. Future sea-level analysis from far field locations must involve the application of a consistent methodology in order to allow meaningful comparison between studies. This should help to resolve the ongoing debate about the magnitude and timing of the Mid-Holocene High Stand, and the nature of late Holocene sea-level fall across the region.

Keywords
Holocene, Relative sea level, Steric expansion

Disciplines
Environmental Monitoring

Comments
Postprint version. Published in *Journal of Asian Earth Sciences* , Volume 25, Issue 1, 2005, pages 29-43.
Publisher URL: http://www.sciencedirect.com/science/journal/03781127

Holocene sea-level changes in the Indo-Pacific

S. A. Woodroffe [a] and B. P. Horton [b]

[a] Environmental Research Centre, Department of Geography, University of Durham, South Road, Durham, DH1 3LE, UK
[b] Department of Earth and Environmental Science, University of Pennsylvania, Philadelphia, PA, 19104-6316, USA

Abstract

Holocene sea-level reconstructions exist from many locations in the Indo-Pacific region. Despite being a large geographical region, the nature of Holocene sea-level change is broadly similar in all locations. Differences do exist, however, in the timing and magnitude of the Mid-Holocene High Stand (MHHS) and the nature of late Holocene sea level fall across the region. When the Indo-Pacific is subdivided into smaller regions, these discrepancies do not disappear, and in some cases the discrepancies are large within a single coastline.

It is clear from this analysis that the fundamental criteria to produce accurate local relative sea-level curves are hardly ever met. There are serious problems associated with the correct interpretation of sea-level indicators and their relationship to mean sea level, and with the quality of age determinations. A consistent methodology throughout the Indo-Pacific for the analysis of sea level data is lacking. Future sea-level analysis from far field locations must involve the application of a consistent methodology in order to allow meaningful comparison between studies. This should help to resolve the ongoing debate about the magnitude and timing of the Mid-Holocene High Stand, and the nature of late Holocene sea-level fall across the region.

1. Introduction

Oscillations between glacial and interglacial climate conditions during the Quaternary have been characterized by the transfer of immense volumes of water between ice sheets and the oceans (e.g. Broecker and Denton, 1989; Alley and Clark, 1999; McManus et al., 1999 and Lambeck et al., 2002). Since the latest of these oscillations, the Last Glacial Maximum (between about 30,000 and 19,000 years ago), approximately 50×10^6 km^3 of ice has melted from the land-based ice sheets, raising global sea level in regions distant from the major glaciation centres (far-field locations) by about 130 m (Fig. 1) (e.g. Lambeck et al., 2002). In contrast, relative sea levels have dropped by many hundreds of metres in regions once covered by the major ice sheets (near- and intermediate-field locations) as a consequence of the isostatic ‘rebound’ of the solid Earth following the melting of land-based ice (Shennan and Horton, 2002). Such rapid changes in sea level are part of a complex pattern of interactions among the oceans, ice sheets and solid earth, all of which have different response timescales. Geographical variability in Holocene sea-level change is well illustrated by Pirazzoli's (1991) atlas of sea-level curves and by geophysical model predictions (e.g. Clark et al., 1978; Peltier, 2002; Shennan et al., 2002; Lambeck et al., 2003 and Mitrovica, 2003). Clark et al. (1978) identified six different sea-level patterns which reflect a range of sea-level histories recorded at coasts which have emerged, submerged, or are in transitional areas and record a combination of both uplift and subsidence (Fig. 2). Although much contemporary research (e.g. Shennan and Horton, 2002) to test such theoretically derived models has focused on data sets from near- and intermediate-field locations (Zones I-II), it is widely recognised that 'far-field locations' (Zones III-VI) provide the best possible estimate of the 'eustatic function' (Clark et al., 1978; Yokoyama et al., 2000 and Peltier, 2002). Consequently, model derived reconstructions are increasingly recognising the importance of these areas to constrain and test their earth-ice models (eg. Fleming et al., 1998).

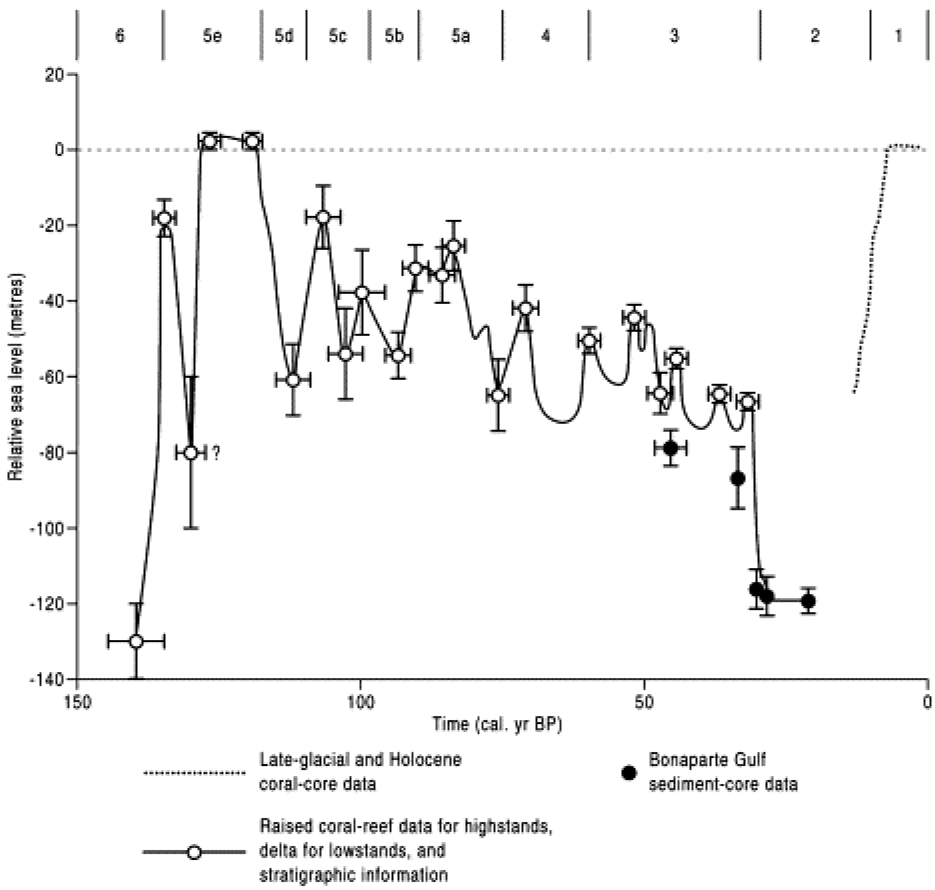

Fig. 1. The relative sea-level curve for the last glacial cycle for Huon Peninsula (Lambeck and Chappell, 2001) supplemented with observations from Bonaparte Gulf, Australia (Yokoyama et al., 2000). Error bars define the upper and lower limits (modified from Lambeck et al., 2002).

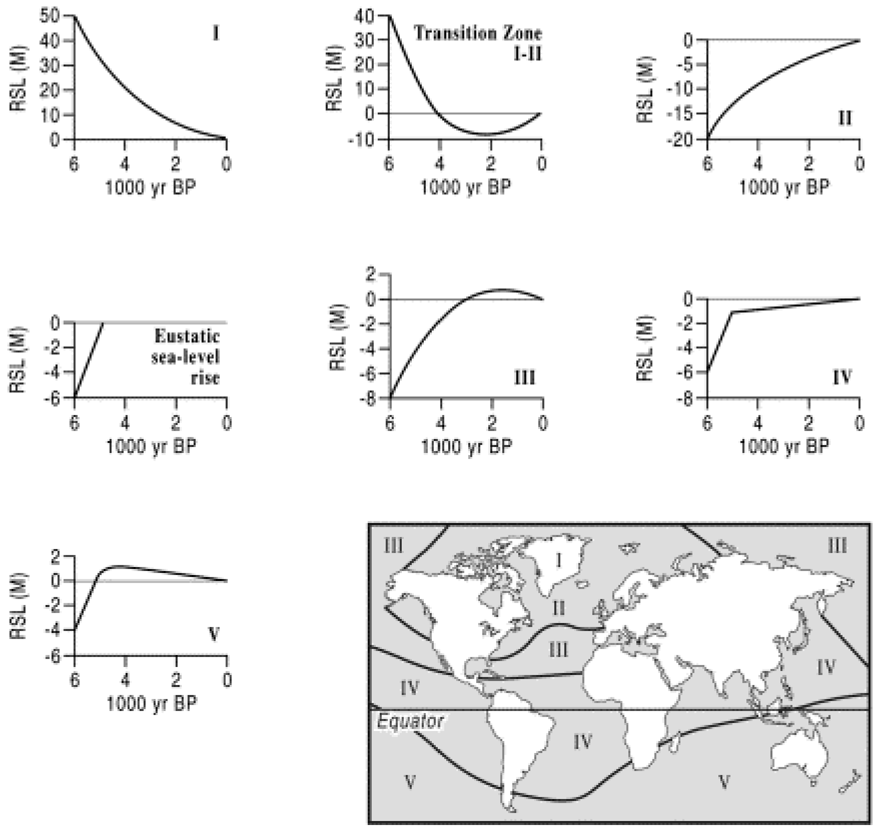

Fig. 2. Sea-level zones and typical relative sea-level curves deduced for each zone by Clark et al. (1978) under the assumption that no eustatic change has occurred since 5 ka BP.

2. Past sea-level changes

At their simplest changes in relative sea level (RSL) are a product of changes in oceanic and crustal variables. A change in RSL at any time or location is dependent upon a combination of sea surface level changes, any isostatic or tectonic changes, and local coastal processes.

The ocean, or eustatic level is influenced by three main groups of factors; the water volume in the oceans; the volume of the ocean basins; and the distribution of the water. The water in the ocean and the volume of glacial ice on land are in balance; when one increases the other decreases. This is known as *glacial eustasy* (first proposed by Suess, 1888). The ocean volume is also modulated by a number of other factors, such as the addition of juvenile water, storage of water in sediments, variations in the main hydrologic cycle changing continental lake volumes, cloudiness, and the evaporation/precipitation balance. A factor of high significance is the steric expansion/contraction of the water column. These changes are driven by temperature and, to a lesser degree, salinity. Steric changes are advocated as one of the most important factors in future sea-level estimates (e.g. Houghton et al., 2001). The basins of the oceans may change their level by crustal movements so that they increase and decrease in total volume. This is *tectono-eustasy*. In reality, this is merely deformation of the hypsographic curve, and is a slow process, which may only change sea level by 0.06 mm yr^{-1} (Mörner, 1996).

The ocean water level also changes as the result of the global distribution of oceanic water. This is known as *geoidal eustasy* (Mörner, 1976). The earth is not spherical, but broadly flattened at the poles and bulging at the equator and the ocean surface does not parallel that of the earth. Present geodetic sea level varies with respect to the earths centre by as much as 180 m. The geoid relief does not remain stable with time; it deforms vertically as well as horizontally. During the Last Glacial Maximum, the weight of continental ice sheets caused downward deformation of the crust, forcing sublithospheric flow away from the centres of load. A low latitude gravitational anomaly developed creating a high in the oceanic geoid. During the last glacial deglaciation, the continents viscoelastically rebounded causing the gravity anomaly to decay and the oceanic geoid to migrate from lower to higher latitudes. This geoidal process is called equatorial ocean syphoning (Mitrovica and Peltier, 1991). The RSL effects of equatorial ocean syphoning are only seen in the late Holocene (last 3000 years), where in low latitudes oceanic islands such as the Cocos Keeling Islands have experienced a fall in RSL (Woodroffe and McLean, 1990).

Global glacial isostatic adjustment is the process whereby the Earth's shape and gravitational field are modified in response to the large scale changes in surface mass load that have occurred due to the glaciation and deglaciation of the planetary surface (according to Archimedes principle). As shown by Daly (1934), the development of an ice sheet will result in subsidence beneath the ice mass (*glacio isostasy*), when deeper material flows away and a peripheral bulge is built around the ice margin. When the ice sheet melts, unloading occurs, resulting in uplift beneath the melted ice at rates which may locally reach 50–100 mm yr^{-1} (Pirazzoli, 1996); the marginal rim will consequently tend to subside and move towards the centre of the vanishing load. Furthermore, during deglaciation meltwater from the ice sheets produces a considerable load (of the order of 100 t m^2 for a sea-level rise of 100 m (Lambeck et al., 2002)) on the ocean floors so that the sea floor subsides (*hydro-isostasy*). Interest in studying the response of the planet to these Quaternary glacial cycles derives primarily from the fact that the geological, geophysical and astronomical data, which record them are of such high quality. Furthermore, these data are almost uniquely capable of providing firm constraints upon the viscoelastic properties of the ‘solid’ interior of the earth and a wide range of processes related to the internal dynamics of the climate system itself (e.g. Agassiz, 1840; Lambeck, et al., 2000; Peltier, 2002; Shennan et al., 2002 and Mitrovica, 2003). While most studies agree on the viscosity structure through the upper mantle and transition zone, there is considerable controversy about the average viscosity structure of the lower mantle. This is important in modelling the inundation of water into regions uncovered by grounded, retreating marine-based ice and ocean loading at evolving continental margins (e.g. Peltier, 1998; Mitrovica and Milne, 2002 and Peltier, 2002; Mitrovicia, 2003 Lambeck et al., 2003).

Although the interaction of eustatic and isostatic factors produces the general pattern of RSL changes, various factors operate at the coast and within an estuary that influence the registration of relative sea-level changes in the fossil, sedimentary record (Shennan and Horton, 2002). In tectonically active areas neotectonic deformation is common and may be due to stresses associated with the isostatic processes, but also due to compression, gliding, elastic rebound, faulting, folding or tilting of crustal blocks. Trends of vertical displacement of tectonic origin often appear to be continuous and gradual over time (Pirazzoli, 1996), but frequently consist of spasmodic movement, often suddenly at the time of earthquakes of great magnitude (eg. Zong et al., 2003). Other local scale factors include modifications in the tidal regime according to the coastal configuration and the relationship between the freshwater table and tide levels; (Shennan and Horton (2002) observe a change in the elevation of MHWST in relation to MTL through the mid to late Holocene in the Humber estuary, UK). Furthermore, changes in elevation of the sediment recording a past sea level since the time of deposition must be taken into account. These changes may be due to consolidation because of the accumulation of overlying sediments and land drainage.

3. Interpretation of former sea-levels

Palaeo sea-levels are reconstructed from a wide variety of environmental indicators including morphological and archaeological data, palaeosols and other lithostratigraphic change. To produce sea-level index points, which allow the construction of a sea-level envelope plotted in a time/ depth–elevation graph, each indicator must have an indicative meaning (Fig. 3). The indicative meaning of a coastal sample is the relationship of the local environment in which it accumulated to a contemporaneous reference tide level, with an associated error range (van de Plassche, 1986 and Horton et al., 2000). The indicative meaning can vary according to the type of evidence and it is commonly expressed in terms of an indicative range and a reference water level. The former is a vertical range within which the coastal sample can occur and the latter a water level to which the assemblage is assigned, for example, mean high water spring tide, mean tide level, etc.

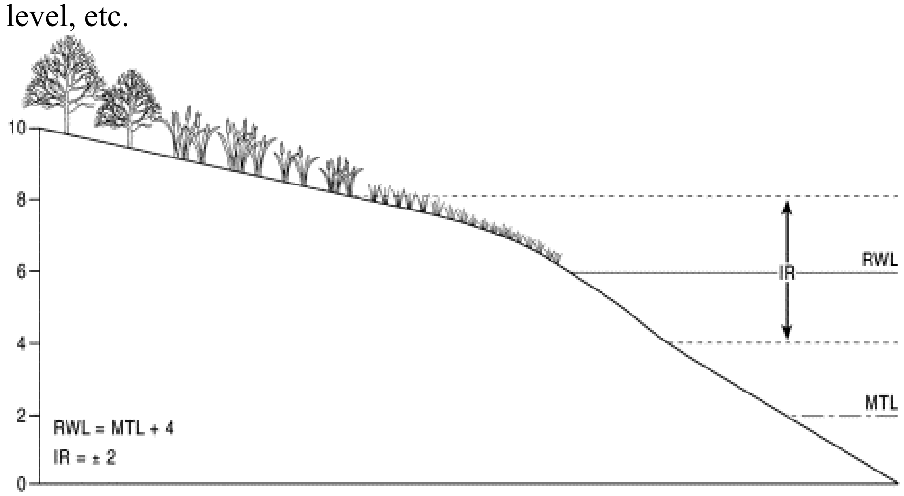

Fig. 3. Schematic diagram of the Indicative Meaning illustrating the how the Indicative Range (IR) and Reference Water Level (RWL) are derived from Mean Tide Level (MTL) (not drawn to scale).

By far the most widely used sea-level indicators in far-field locations are coral. There are many different types of coral, but only a few species are found in narrow elevation ranges (e.g. *Acropora palmata*, which lives within 5 m of the water surface (Lighty et al., 1982) and *Porites sp.* which can live near the surface and at depths up to 30 m (Done, 1982)). Corals in growth position are constrained by water depth, and most are only able to survive up to mean low water spring tides in unponded locations. They are readily dated using Uranium series disequilibrium and radiocarbon techniques. Unfortunately the lower depth limit of *Acropora* growth is relatively poorly constrained, leaving coral-based sea-level indicators with a large indicative range (c. ±5 m) (Blanchon and Shaw, 1995). Microatolls are a particular type of intertidal coral that live near low water and have an indicative range as low as 3 cm (e.g. Smithers and Woodroffe, 2000). They are seen as the most precise, geologically persistent and useful diagnostic sea-level indicator on coral reef systems.

Other widely used sea-level indicators include geomorphological features, such as palaeoshoreline notches, palaeoreef flats and beach deposits (e.g. Tija, 1996 and Dickinson, 2001). These indicators have wide indicative ranges, and large error terms associated with any reconstructions. Thus, for detailed Holocene sea-level reconstructions it is often the case that the error term applied to the sea-level indicator has the same magnitude as the RSL change under investigation. Fixed biological indicators are also used as sea-level indicators. These include rock clinging oyster beds and fossil calcareous tubeworm encrustations (e.g. Baker and Haworth, 2000).

With notable exceptions (e.g. Cann et al., 1993; Cann et al., 2000a; Cann et al., 2000b; Cann et al., 2002; Li et al., 2000 and Yokoyama et al., 2000), studies of microfossils and their relationship to RSL in far-field coastal environments of the Indo-Pacific are few. In contrast, many detailed studies of former sea levels in Northwest Europe and North America have used microfossils (e.g. diatoms, foraminifera, pollen) found within late and post-glacial sedimentary deposits (e.g. Horton et al., 2000 and Shennan and Horton, 2002). One of the few studies of sedimentary facies depicting late glacial and Holocene sea-level changes in a far-field location have been undertaken by Yokoyama et al. (2000) in the Bonaparte Gulf, northern Australia. They used sediment cores and their contained foraminifera and ostracods to infer changing environments of deposition through time, enabling sea-level reconstructions back to the last glacial maximum. Despite using tightly constrained foraminifera and ostracoda evidence, the reconstructions based on facies descriptions such as ‘shallow marine’ and ‘marginal marine’ have relatively wide indicative ranges (c.±1 m).

4. The late glacial and post-glacial period

It is widely implied that far-field locations have a spatially homogenous record of sea-level fluctuations. However, several diverse processes, such as gravitational effects causing movements in the geoid and meltwater loading of the ocean floors, tectonism, thermal expansion of ocean water and Antarctic mass balance changes, interplay with the overwhelming eustatic signal to create the relative sea-level record in far-field locations. Nevertheless, a small number of high-resolution RSL records have been used to approximate the Holocene glacio-eustatic signal (e.g. Fairbanks, 1989; Chappell and Polach, 1991; Blanchon and Shaw, 1995; Bard et al., 1996 and Yokoyama et al., 2000). In turn these records have been used in geophysical models to estimate the ice mass at the last glacial maximum and to tune models of earth rheology. The first global eustatic sea-level curve was generated from cored corals (*Acropora palmata*) in Barbados (Fairbanks, 1989). This record concentrates on the late glacial and early Holocene period, creating a eustatic record for the period 17,100–7800 BP. Unfortunately Barbados is situated on an accretionary prism between two oceanic plates, and therefore is not immune to problems of tectonic lithospheric movement over late glacial timescales (estimated uplift of 0.34 mm yr^{-1}). The major breakthrough of work in the Caribbean/Atlantic region on both cored coral and sediment cores was the discovery of three 'catastrophic rise events', or Meltwater Pulses (MWPs), between 14,200 and 7600 cal years BP. The largest rise (approximate magnitude of 13.5 m) occurred at 14,200 cal years BP (MWP 1a). This coincides with a period of ice sheet collapse (Heinrich layer 1) as defined in sediment cores from the north Atlantic, and predates the Younger Dryas cold event by about 1000 years (Fig. 4). However, no consensus on the source of the ice melt has been reached, with contributions from North America, Scandinavia and the Barents Sea, and Antarctica having been proposed (Blanchon and Shaw, 1995; Clark et al., 1996 and Clark et al., 2002).

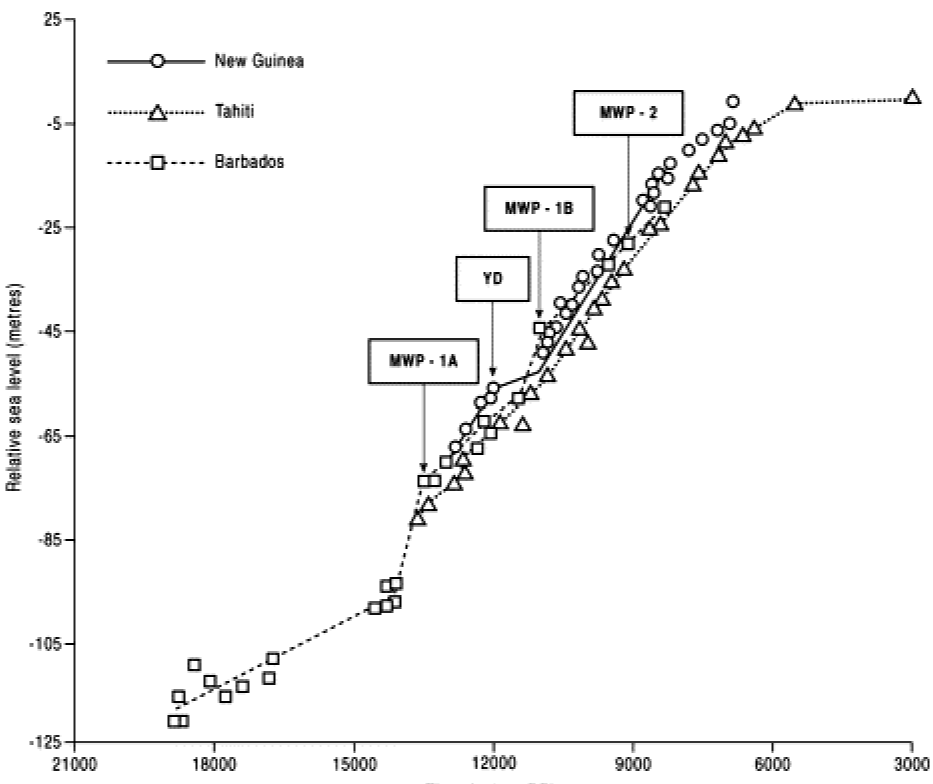

Fig. 4. Sea-level history reconstructed for long drill cores from Tahiti (triangles), Barbados (squares) and Papua New Guinea (dots). All radiocarbon dates have been corrected to calendar years and all data points have been appropriately corrected for uplift/subsidence.

Following this breakthrough in late glacial eustatic sea-level reconstruction, a coral coring project in Tahiti (Bard et al., 1996) produced a late glacial and early Holocene eustatic sea-level curve, which complements the record from Barbados. The Tahiti record has the advantage of not being contaminated by a complex tectonic component, being an oceanic atoll situated in the centre of the Pacific plate. However, Tahiti exists because of volcanic and tectonic activity in the past and is subsiding at a rate of 0.2 mm yr^{-1} (Bard et al., 1996). This new record shows a synchronous hiatus (MWP1a) with the Barbados record. However, the Tahiti record questions the global significance of the two other catastrophic rise events, MWP 1b at 11, 500 cal BP and MWP 2 at 7600 cal BP, both found in the Caribbean/north Atlantic records (Blanchon and Shaw, 1995).

A third global late glacial and early Holocene eustatic record has been inferred from raised coral terraces on the Huon Peninsula, Papua New Guinea (Chappell and Polach, 1991). The Huon Peninsula is tectonically active, being close to the edge of the Philippine plate, and is experiencing uplift at a rate of $\sim$1.9 mm yr^{-1}. By decoupling the tectonic and eustatic elements of RSL history in the area, a global eustatic sea-level curve has been developed. However, aseismic uplift on the Huon Peninsula is episodic, dominated by isolated, metre scale events with 1000 year reoccurrence intervals, which contaminates the record with a complex tectonic component.

The nature of mid to late Holocene global eustatic sea-level change is less well documented. Indeed Nunn (1998) refers to the last millennium as a ‘1000-year hiatus’ in sea-level research of the Pacific (Long, 2000). Emphasis switches from aiming to produce a globally valid eustatic sea-level curve to understanding local and regional influences on RSL reconstructions. Therefore, for the purposes of this review we have split the Indo-Pacific into six sub-regions; Southwest Indian Ocean, Northern Indian Ocean, Southeast Asia, Northern Asia, Australia and Pacific Ocean (Fig. 5 and Table 1).

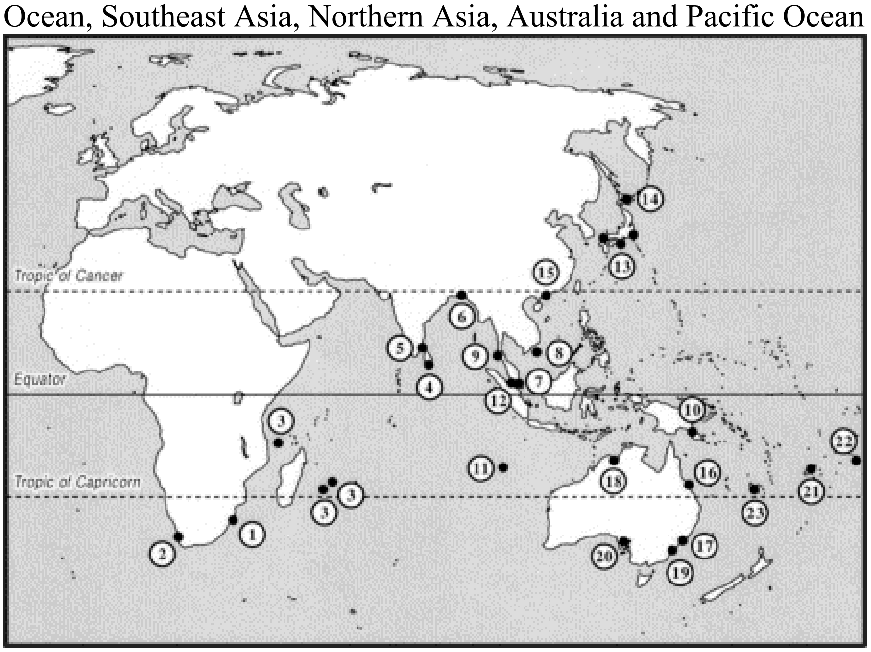

Fig. 5. Location of sea-level reconstructions in the Indo-Pacific region. Numbers refer to locations in Table 1.

Table 1
Details of sea-level reconstructions from the Indo-Pacific region (HS, high stand)

Indo-Pacific region	Location on Fig. 5	Location and authors	Shoreline indicators used and indicative range (if known)	Tectonic activity/ tidal range etc	Dating method	No. of Holocene high stands	Timing of mid-Holocene high stand	Magnitude of mid-Holocene high stand
SW Indian Ocean	1	South Africa (Ramsay and Cooper, 2002)	Beachrock (+/− 0.2 m), barnacles and attached oysters (+/− 0.5 m), estuarine fill sequences	Stable	Uranium series (beachrock)	2	1st HS-4650 ^{14}C yrs BP (4627 cal yrs BP) 2nd HS 1610 ^{14}C yrs BP (966 cal yrs BP)	1st HS-3.5 m above SL 2nd HS-1.5 m above SL
SW Indian Ocean	2	South Africa (Compton, 2001)	Saltmarsh organic material/molluscs	Stable, microtidal range	Radiocarbon	2	1st HS-6800 cal yrs BP 2nd HS-1300 cal yrs BP	1st HS-1 to 3 m above SL 2nd HS-0.5 to −1 m above SL
SW Indian Ocean	3	Mauritius, Reunion and Mayotte (Colonna et al., 1996; Camoin et al., 1997,)	Fringing coral reefs	Stable	Uranium series	0	SL stabilized at current levels 3 ka cal BP	No HS recorded-small atolls sinking through Holocene with ocean floor subsidence
Northern Indian Ocean	4	Sri Lanka (Katupotha and Fujiwara, 1988)	Coral and marine shells	Stable	Radiocarbon	2	1st HS-6170 to 5100 ^{14}C yrs BP (6485–5370 cal yrs BP) 2nd HS-3210 to 2330 ^{14}C yrs BP (2902–1558 cal yrs BP)	1st HS-1m above SL 2nd HS-1m (may not have lowered btwn 2 highs)
Northern Indian Ocean	5	East coast of India (Banjeree, 2000)	Beach ridges (+/− 0.5 m), *Porites* coral colony, intertidal shells	Stable	Radiocarbon (shells) Uranium series (coral)	2	1st HS-7300 to 5660 cal yrs BP 2nd HS-4300 to 2500 cal yrs BP	1st HS-3 m above SL 2nd HS-3m above SL
Northern Indian Ocean	6	Bangladesh (Islam and Tooley, 1999)	Sedimentary sequences and microfossils	Possible tectonic activity/glacio isostasy	Radiocarbon	0	No evidence for SL High Stand	No evidence for SL HS
SE Asia	7	Malay–Thai Peninsula (Tija, 1996)	Malaysia-Abrasion platforms, sea-level notches, oyster beds Thailand-peat, marine shells, mangrove wood	Stable assumes TL not changed	Radiocarbon	2 3	Malaysia-5000 and 2800 ^{14}C yrs BP (5131 and 2358 cal yrs BP) Thailand-6000, 4000 and 2700 ^{14}C yrs BP (6262, 3808 and 2198 cal yrs BP)	Malaysia-5 m and 3 m above SL Thailand-4 m, 2.5 and 2 m above SL
SE Asia	8	Sunda Shelf, Indonesia (Hanebuth et al., 2000)	Sediment facies from cores	Stable	Radiocarbon	N/A	Record from 12,000–14,000 cal yrs BP-shows MWP 1a (16 m SL rise in 300 yrs)	N/A
SE Asia	9	Phuket, S Thailand (Scoffin and Le Tissier, 1998)	Coral microatolls	Stable	Radiocarbon	1	Before 6000 cal yrs BP	1 m above SL
SE Asia	10	Huon Peninsula, P. New Guinea (Chappell and Polach, 1991)	Raised coral terraces	Tectonically active (emergence btwn 0.5–3 m/1000 yrs)	Uranium series	N/A	Late glacial/early Holocene record of SL rise up to 5800 ^{14}C yrs BP (6053 cal yrs BP)	N/A
SE Asia	11	Cocos Keeling Islands (Woodroffe and McLean, 1990)	Coral microatolls	Stable	Radiocarbon	1	Since about 3000 ^{14}C yrs BP (2600 cal yrs BP)	At least 0.5 m

Table 1 (*continued*)

Indo-Pacific region	Location on Fig. 5	Location and authors	Shoreline indicators used and indicative range (if known)	Tectonic activity/ tidal range etc	Dating method	No. of Holocene high stands	Timing of mid-Holocene high stand	Magnitude of mid-Holocene high stand
SE Asia	12	Strait of Malacca (Geyh and Kudrass, 1979)	Mangrove wood and peat (indicative meaning not accounted for)	Stable	Radiocarbon	1	Approx 4980 ^{14}C yrs BP (5089 cal yrs BP)	Between 2.5–5.8 m above SL
Northern Asia	14	Japan (Nakada et al., 1991)	Intertidal and subtidal shells, wood and peat	Tectonically active	Radiocarbon	1	HS at 600 cal yrs BP	5 m above SL
Northern Asia	15	Japan (Sawai et al., 2002)	Sedimentary sequences and microfossils	Tectonically active	Radiocarbon	N/A	3 falls in RSL since 3000 cal yrs BP	N/A
Northern Asia	16	Southeast China (Yim and Huang, 2002)	Fixed biological indicators (oyster beds)	Local monsoon activity	Radiocarbon	1	5140 ^{14}C yrs BP (5371 cal yrs BP)	No more than 2 m above SL
Australia	17	Central Great Barrier Reef, Australia (Beaman et al., 1994; Larcombe et al., 1995; Larcombe and Carter, 1998)	Oyster beds	Stable, some differential flexing of shelf over past 6000 yrs	Radiocarbon	1	Between 5660 and 4040 ^{14}C yrs BP (6000 and 4053 cal yrs BP)	1.65 m above SL
Australia	18	Eastern Australia (Flood and Frankel, 1989)	Intertidal tube worms	Stable	Radiocarbon	1	Some time before and after 3420 ^{14}C yrs BP (3274 cal yrs BP)	At least 1 m above SL
Australia	19	Bonaparte Gulf, NE Australia (Yokoyama et al., 2000)	Sedimentary facies and their contained microfossils	Stable	Radiocarbon	N/A	LGM and early late glacial record	N/A
Australia	20	New South Wales (Baker et al., 2001)	Fixed biological indicators (tubeworms, barnacles, oysters)	Stable	Radiocarbon	1	Around 3900 cal yrs BP	At least 1 m above SL
Pacific	21	Fiji (Nunn and Peltier, 2001)	Coral (Porites) microatolls and intertidal shells from raised beaches	Hotspot activity over Holocene timescales (subsidence)	Radiocarbon	1 or 2	1st HS-5650 to 3200 ^{14}C yrs BP (6021–2939 cal yrs BP) 2nd HS-6100 to 4550 and 3590–2800 ^{14}C yrs BP (6477–4678 and 3422–2485 cal yrs BP)	1st HS-1.35 to 1.5 m above SL 2nd HS-(i) 0.75–1.85 m, (ii) 0.90–2.46 m above SL
Pacific	22	Taihiti (Bard et al., 1996)	Coral (*Acropora*)	Slight subsidence (~0.2 mm/yr)	Uranium series	N/A	Record from 14,000–5000 cal yrs BP, shows MWP 1A	N/A
Pacific	23	W Central Equitorial Pacific (Grossman et al., 1998)	Coral, microatolls, beachrock, molluscs, peat	Hotspot activity over Holocene	Radiocarbon	1	4 ka ^{14}C yrs BP (3942 cal yrs BP)	1–2 m above SL

Radiocarbon ages calibrated to 2σ using the probability method in the program CALIB rev. 4.3, (Stuiver and Reimer, 1993), including corrections for the marine reservoir effect using the calibration data of Stuiver et al. (1998).

4.1. Southwest Indian Ocean

The southwest Indian Ocean includes the continental coastline of southeast Africa, the micro continent of Madagascar, and several oceanic atolls including Mayotte, Reunion, Mauritius and the Seychelles Field observations in these locations have utilized a range of geomorphological and biological indicators including beachrock, oyster beds, coral reefs and marine shells to reconstruct mid to late Holocene changes in RSL (e.g. Ramsay, 1995; Colonna et al., 1996; Camoin et al., 1997; Compton, 2001 and Ramsay and Cooper, 2002). The southwest Indian Ocean is tectonically stable, being situated towards the centre of the African plate and, therefore, observations from these locations should yield important information regarding glacio-eustasy and hydro-isostasy. Studies of Holocene sea-level change in South Africa have relied mainly on radiocarbon dating of fossil beachrock. Beachrock is a reliable sea-level indicator, assuming that the skeletal carbonate content is entirely contemporary. It forms today at the freshwater/saltwater interface at 0.1–0.2 m below mean sea level (Ramsay and Cooper, 2002). Ramsay (1995) produced a 9000 year record showing early Holocene RSL rise to a mid-Holocene high stand of +3.5 m at 4650 ^{14}C yrs BP with RSL subsequently falling below present levels, but also shows a secondary high stand at 1610 ^{14}C yrs BP (+1.5 m) before mean sea level is attained at 900 ^{14}C yrs BP. The sea-level observations are taken from a 180 km long stretch of coastline in eastern South Africa, thus reflecting regional RSL influences. However, the sea-level index points are not corrected for their indicative meaning. For example, some index points from the early Holocene use dated wood, which should be treated as limiting points, that is RSL must have been at or below the level at which they are found, however, they are included in the curve as standard index points. A second investigation in South Africa has concentrated on a saltmarsh lagoon on the southwest coast (Compton, 2001). This lagoon is situated on the Atlantic coast, where tectonic activity has occurred through the Holocene, and the RSL history reflects more Atlantic rather than Indian Ocean processes. The RSL record, derived from radiocarbon dated saltmarsh peats, agrees with Ramsay's two mid-late Holocene high stands theory, and uses potentially more precise indicators (Fig. 6). However, the indicative meaning of each indicator is not explicitly stated and all late Holocene sea-level fluctuations after the initial mid-Holocene peak of +3 m are within the elevation ranges of the indicators.

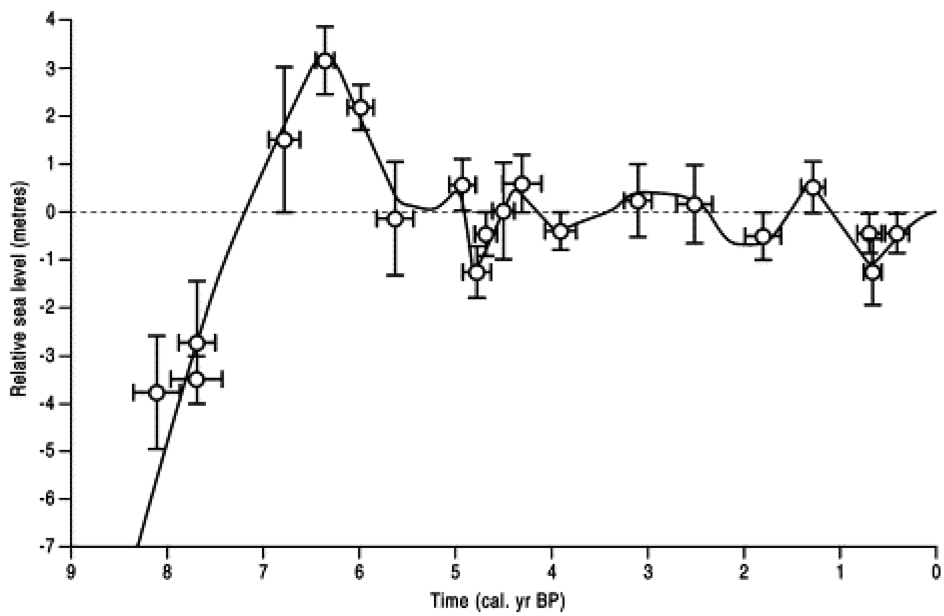

Fig. 6. Holocene sea-level fluctuations inferred from sea-level index points from the southern Langebaan Lagoon salt marsh, South Africa. Horizontal error bars refer to analytical uncertainty in radiocarbon age calibration (2σ range), and vertical error bars refer to uncertainty in sea level predictions derived from organic matter and shell material indicators (modified from Compton, 2001).

Holocene sea-level changes on the oceanic atolls of Mauritius, Reunion and Mayotte have been investigated to infer reef development histories and eustatic sea-level rise (e.g. Colonna et al., 1996 and Camoin et al., 1997). Drilled *Acropora* coral cores from the fringing reefs of all three islands demonstrate a continuous rise in eustatic sea level through the early and mid Holocene, stabilizing to current levels at 3000 ^{14}C yrs BP. There is no evidence of two Holocene high stands in these locations. However, on the microcontinent of Madagascar, in situ coral colonies have been dated between 3000 and 2000 ^{14}C yrs BP, 0.3–2.5 m above RSL, and at Farqhar Atoll in the Seychelles, coral limestones have been dated at 3640 ^{14}C yrs BP at +2 m (Camoin et al., 1997). This apparent diversity in the record may be due to the nature of these small oceanic atolls, which are affected by late-Holocene hydro-isostasy, causing them to react to eustatic sea-level change in the same way as the ocean floor. That is, they subside with the ocean floor, whereas sublithospheric flow below the sea floor is drawn to larger atolls and land masses (e.g. Seychelles, Madagascar, Africa) (Grossman et al., 1998). Larger land masses record a drop in RSL through the late Holocene while small oceanic atolls experience continued sea-level rise.

4.2. Northern Indian Ocean

The northern Indian Ocean region includes studies from the east coast of India, Bangladesh and Sri Lanka (Katupotha and Fujiwara, 1988; Islam and Tooley, 1999 and Banjeree, 2000) The Indian sub-continent is situated on the Indian plate, which in its centre is tectonically stable. Techniques used to reconstruct former sea levels here include dating coral and marine shells, beach ridges and sedimentary sequences and their contained microfossils. Banjeree (2000) studied regional RSL change on the east Indian coastline, utilizing beach ridges and exposed *Porites* coral colonies to indicate two mid-late Holocene high stands, the first peaking at 3 m above current mean sea-level at 7300 cal yrs BP, followed by a c. 2 m RSL fall, and a second pulse of minor RSL rise culminating at +3 m between 4300 and 2500 cal yrs BP. This study analyses evidence from sites covering over 150 km of coastline and, therefore, has a regional focus. The indicative meanings of these records are not explicitly taken into consideration when creating sea-level index points, and the contemporary reference level to which all fossil deposits are calibrated is low tide level rather than mean sea level, which precludes correlation with other studies unless the index points are re-calibrated.

Field data from Bangladesh using multi-proxy lithostratigraphical and biostratigraphical techniques (Islam and Tooley, 1999) shows a continuously rising RSL record through the Holocene. The field site in Bangladesh is situated on low-lying deltaic deposits, which are susceptible to long term subsidence, release of sediment and water from the catchment and anthropogenic activities.

4.3. Southeast Asia

This diverse region includes studies from the Strait of Malacca, Indonesia (Geyh and Kudrass, 1979), Malay–Thai Peninsula (Tija, 1996), Phuket, southeast Thailand (Scoffin and Le Tissier, 1998), Sunda Shelf (Hanebuth et al., 2000) and the Huon Peninsula, Papua New Guinea (Chappell and Polach, 1991). The Indonesian archipelago and its surrounding microcontinents (Papua New Guinea, Borneo, Peninsula Malaysia) have been subject to intense tectonic processes over the Holocene period. The region is a collision zone between the Eurasian, Indian, Philippine and Pacific plates and includes the active subduction zone and island arc of the Java trench, which covers a large part of southern Indonesia.

Holocene sea-level reconstructions in southeast Asia are limited and fragmentary. Tija (1996) studied abrasion platforms, sea-level notches and oyster beds in Peninsula Malaysia and produced over 130 radiocarbon-dated indicators. The sea-level curve, which proposes two Holocene high stands at 5000 and 2800 ^{14}C yrs BP, does not take into account the broad altitudinal range of the indicators, and the second high stand could easily be accounted for by error terms in the general RSL drop after 5000 BP (Fig. 7). Tija's (1996) sea-level reconstruction for Thailand, based mainly on peat, marine shells and mangrove wood, indicates potentially three mid/late Holocene high stands at 6000, 4000 and 2700 ^{14}C yrs BP, which he attributes to sea level lowering in a series of short periods of regression and transgression. Over 100 indicators were taken from around the Peninsula Coast, therefore reflecting regional rather than local influences; however, it is hard to reconcile this model of sea-level history with the scattered evidence that is presented. Scoffin and Le Tissier (1998) studied intertidal reef-flat corals (microatolls) for mid/late Holocene high stands at Phuket, Southern Thailand. The evidence here suggests a single Holocene high stand with a constant rate of RSL fall since 6000 cal yrs BP, and thus, contradicts conclusions drawn by Tija (1996). However, it should be noted that Scoffin and Le Tissier's (1998) research is from only one site with eleven dated samples.

Geyh and Kudrass (1979) conducted a study in the Strait of Malacca, Indonesia, which resulted in 33 directional (limiting) index points from ^{14}C dated fossil mangrove deposits. These index points, which are not given an indicative meaning, imply that Holocene sea level rose from below −12.8 to ∼1.2 m above present between 8000 and 6000 ^{14}C yrs BP, and between 5000 and 4000 ^{14}C yrs BP rose to its highest recorded level in southeast Asia, at ∼5.8 m above present (but this is a limiting index point, uncorrected for indicative meaning). The data does not show whether the late Holocene lowering of sea level was a steady or oscillatory process.

The Cocos Keeling Islands are a mid-oceanic atoll situated approximately 600 km southwest of Jakarta, Indonesia in the Indian Ocean. Their RSL history is recorded in massive coral and microatolls. The microfossil record produces a detailed late Holocene RSL history showing a single Holocene high stand after 3000 ^{14}C yrs BP of at least 0.5 m (Woodroffe and McLean, 1990).

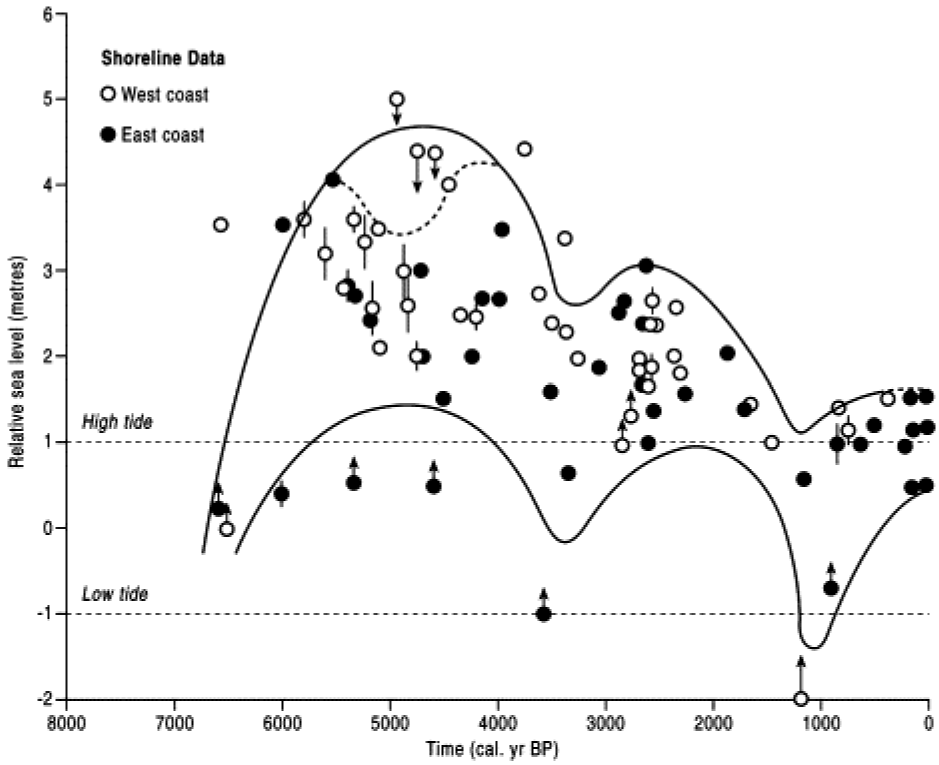

Fig. 7. Holocene sea-level envelope for Peninsula Malaysia (modified from Tija, 1996). Data points with arrows indicate directional (limiting) index points. Some index points have vertical error bars, where the vertical range of the sea-level indicator is understood. No age errors are considered, radiocarbon ages are plotted. The boundaries of the envelope are drawn midway between the extreme data points and the neighbouring points within the envelope (Tija, 1996).

4.4. Northern Asia

The areas grouped under this heading include Japan and southeast China, bordering the South China Sea Japan is forms part of a tectonically active island arc, which makes regional sea-level comparisons difficult due to local and regional crustal flexure and warping. Nakada et al. (1991) undertook RSL reconstructions from Osaka and Tokyo using previously published data and showed a significant spatial dependence to the amplitude of the mid-Holocene high stand. Fixed biological indicators (oyster beds) have been used at one site on the South China Sea coast to infer the amplitude and timing of the mid-Holocene high stand; no greater than 2 m above modern sea level at 5140 ^{14}C yrs BP (Yim and Huang, 2002). Further research using fixed biological indicators is needed from other locations of the coastline in order to constrain the amplitude of the mid-Holocene high stand and to better understand the effect of monsoon forcing on the RSL record.

4.5. Australia

Australia is a tectonically stable continent far from plate margins, situated at the centre of the Australian–Indian plate, and as such should preserve a sea level record dominated by eustatic processes Fixed biological indicators such as tubeworms, oyster beds, barnacles and corals have been used extensively along the east coast of Australia to infer Holocene sea-level movements (e.g. Beaman et al., 1994; Baker and Haworth, 2000 and Baker et al., 2001), and foraminifera have been used to infer relative estuarine/lagoonal and oceanic influences through the Holocene in the estuaries of South Australia (Cann et al., 1993; Cann et al., 2000a; Cann et al., 2000b and Cann et al., 2002). In the Northern Spencer Gulf the highest Holocene sea levels occurred at 7000 cal years BP, followed by continuous RSL fall through the late Holocene (Cann et al., 2000a). Studies from north New South Wales show a late Holocene high stand at +1.7 m at 3810 cal yrs BP (Baker et al., 2001). However, evidence for the mid-Holocene high stand in northeast Australia is much earlier, at 1.65 m above RSL between 5660–4040 ^{14}C yrs BP in Cleveland Bay, central Great Barrier Reef (Beaman et al., 1994).

The eastern coast of Australia is affected by differential crustal flexure over late Holocene timescales due to differences in the width of continental shelf and its impact on hydro-isostatic adjustment (Flood and Frankel, 1989 and Baker et al., 2001). This affects the timing of the mid-Holocene high stand along the eastern Australian coast. Fig. 8 illustrates that by fitting a polynomial trendline through sea-level index points created from fixed biological indicators (Baker et al., 2001) it is possible to draw the conclusion that two mid/late-Holocene high stands have occurred in this region, if age and altitude errors in the reconstructions are not taken into account. However, there is little justification for choosing an episodic late Holocene sea level trend using mathematical modelling which does not incorporate the range of potential values for each age/altitude index point.

Larcombe et al., 1995 and Larcombe and Carter, 1998 present evidence for an episodic early Holocene marine transgression across the central Great Barrier Reef shelf. The most contentious aspect is the evidence for a stillstand at −10 m at 8500 ^{14}C yrs BP, followed by a rapid fall in RSL to -17 m at 8200 ^{14}C yrs BP, and then a rapid rise in RSL to −5 m by 7800 ^{14}C yrs BP.

The database used to infer this sea-level trend incorporates evidence from a wide range of studies over a period of 25 years, by different researchers using different proxy indicators. The database makes no attempt to reclassify indicators for the inferred sea-level position at the time the indicator was lain down. Indicators from a wide geographical area (4° of latitude) are included in a single reconstruction, which would infer that differential crustal movement has not occurred over this wide area during the Holocene. Evidence from other studies (Hopley, 1975 and Chappell et al., 1983) show that up to 3 m of differential movement has occurred along this coastline over the past 6000 years.

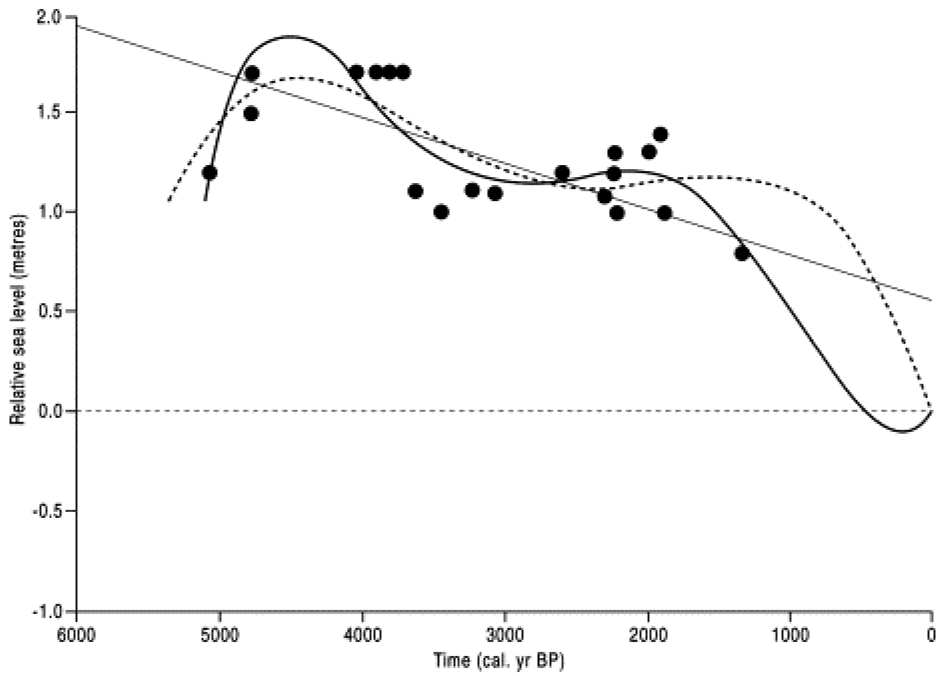

Fig. 8. A summary of linear and oscillating regression models from sea-level index points created using fixed biological indicators from SE Australia for the past 5000 years. Dotted line—4th order polynomial (r^2=0.69), smooth line—5th order polynomial (r^2=0.78) (after Baker et al., 2001).

4.6. Pacific Ocean

The Holocene RSL history of the Pacific Ocean basin is of prime importance for furthering our understanding of geoidal-eustasy, in particular Equatorial Ocean syphoning. A plethora of studies have emerged in recent years from the oceanic atolls of the Pacific, most noticeably the Tahiti coral record (Bard et al., 1996), and studies from Fiji, Kiribati, French Polynesia, Hawaii, Tonga, Marshall Islands and others (e.g. Grossman et al., 1998 and Nunn and Peltier, 2001). The vast majority of sea-level index points from the Pacific use massive coral species (*Acropora* and *Porites*), coral microatolls or beachrock as indicators, which each have their own limitations. In addition many, if not all, of the oceanic atolls of the Pacific are subject to tectonic movements over Holocene and even longer timescales. Many studies assume stability over the Holocene, whereas in fact slow, constant subsidence may be occurring. Mid/late Holocene sea-level records from the Pacific have been interpreted as evidence for a migrating geoid, through evidence of greater emergence in the central ocean basin interior during the mid/late Holocene, and most significant and recent RSL fall in the central equatorial Pacific (Mitrovica and Peltier, 1991 and Nunn and Peltier, 2001). In addition sea-level observations indicate that oceanic atolls less than 10 km in diameter are affected by hydro isostasy (Grossman et al., 1998).

5. Discussion

This review has illustrated the geographical variability of RSL reconstructions, and the variable quality of reconstructions across the Indo-Pacific region. It has also shown that evidence for the mid-Holocene high stand is variable in both magnitude and timing across the Indo-Pacific. In several locations limited evidence points towards a second, smaller magnitude high stand in the late Holocene (Table 1).

Field observations of mid-Holocene high stands provide essential constraints to geophysical models because model predictions of these high stands depend upon three large-scale, high amplitude, globally applicable mechanisms: ocean syphoning caused mainly by gravitational effects due to the collapse of peripheral forebulges, continental levering associated with local ocean loading, and on-going melting of global ice since the time of the high stand (magnitude and source). The first two mechanisms occur throughout the deglaciation process, with the mid-Holocene high stand marking a point in time when the eustatic signal reduces in magnitude so that these solid earth processes become apparent in the RSL record. Syphoning and continental levering are dependant on the viscosity structure and, thus, contribute to the sensitivity of different models. Since syphoning produces a more geographically uniform sea-level fall than the continental levering effect, the latter is likely to account for most of the site dependent variations of predicted high stands. Differences between sea-level observations and predicted high stand amplitudes are used to estimate the final mechanism; on-going melting of global ice since the time of the high stand (Mitrovica and Peltier, 1991; Peltier, 2002 and Mitrovica and Milne, 2002). If the observations are measured precisely, they may provide constraints on where the meltwater came from (northern or southern hemisphere), or at least the relative magnitudes.

Grossman et al. (1997) propose that a second, short-lived marine transgression in low and mid latitudes in the late Holocene may be due to the migration of the geoid. The behaviour of the migrating geoid might be as a series of oscillating waves flowing from its maximum position in the central ocean basin interior (the central equitorial Pacific), rather than as migration as a single peak. Low and mid latitudes might experience a slight, short-lived transgression and regression due to the interplay between geoidal eustasy and hydro isostasy during the late Holocene period when glacio-eustasy has ceased. In addition, Grossman et al. (1998) suggest that the geoidal anomoly may decay in the Pacific towards the Asian continent, directing a second transgression to this area only. This would fit with the evidence, which suggests that two Holocene high stands are only witnessed in areas of the Indo Pacific far from the central equitorial Pacific. However, this may just be a coincidence considering the poor quality of the empirical data.

This review of far field locations has raised the issue that the fundamental criteria to produce an accurate Holocene sea-level curve are hardly ever met. The accuracy or significance of each curve depends upon the number of sea level index points, the correct interpretation of its relation to the corresponding mean sea level, and the quality of age determinations. There are serious problems associated with the correct interpretation of the elevational relationship of the sea-level indicators to mean sea level. Different types of indicators have different degrees of precision, but this is often not acknowledged. Coral microatolls in some settings have a much higher degree of precision than other species such as *Porites* or *Acropora*, which have wide elevational ranges, but each study is considered accurate to the same level. Elevation error ranges are rarely included on sea-level curves. The indicative meaning and reference water level of each indicator is sometimes accounted for, sometimes not. Different workers assign different indicative meanings to the same indicator. It is common to use low tide level as a common datum rather than mean tide level, but this is not often stated in reports, precluding regional correlation of different studies.

The second serious form of error regards age determinations, in particular the calibration of radiocarbon dates, a necessary process because the assumption that the specific activity of the ^{14}C in the atmospheric carbon dioxide has been constant is invalid. The ^{14}C activity in the atmosphere and other reservoirs, and thus in the initial activity of the samples dated, has varied over time (e.g. Stuiver et al., 1998a and Stuiver et al., 1998b). A calibration dataset is necessary to convert conventional radiocarbon ages (^{14}C yr) into calibrated years (cal yr). However, many researchers do not calibrate their radiocarbon dates, or fail to state in their publications whether they have or not. If ages are calibrated, different researchers use different calibration programs, which use alternative calibration terms and produce different results. Similar uncertainties are found with the application of the Marine Reservoir Effect. Radiocarbon ages of samples formed in the ocean, such as shells, fish, marine mammals etc. are generally several hundred years older than their terrestrial counterparts. This apparent age difference is due to the large carbon reservoir of the oceans. A correction is necessary in order to compare marine and terrestrial samples, but because of complexities in ocean circulation the actual correction varies with location.

6. Conclusion

Thus, further sea-level analysis from far field locations must involve the application of the consistent methodology to the reconstruction of sea-level history, which was first formalised during International Geological Correlation Programme (IGCP) Project 61. This operated in the period 1979–1983, and has been a component of all subsequent IGCP Projects, especially Projects 200, 274, 367 and 437 (e.g. van de Plassche, 1986 and Shennan and Horton, 2002), however, this review has shown serious failings in its application. A new generation of sea-level records from far field locations will improve our understanding of the driving mechanisms behind Holocene sea-level change and coastal evolution over a range of spatial and temporal scales. This sea-level data could be meaningfully compared with the emerging high-precision palaeoenvironmental records from the ice sheets (cores), the oceans (e.g. corals, high-resolution sediment cores) and other terrestrial archives (e.g. peat bogs and loess sequences). Therefore, exploration of the implications of sea-level records for an understanding of existing terrestrial and oceanic records of Holocene environmental change, including the leads and lags associated with oceanic and terrestrial records (e.g. Visser et al., 2003) could occur. This provides an improved scientific background against which the role of humans as agents of coastal change can be appreciated.

Acknowledgements

This research was carried out while in receipt of a Natural Environment Research Council award (NER/S/C/2002/10581). Special acknowledgements are given to Operation Wallacea, and to S. Smithers and D. Smith for their valuable comments on the original version of this paper. The authors thank the cartography department at the Department of Geography, University of Durham for producing the figures, and to all members of the Environmental Research Centre, University of Durham for their help and advice. This paper is a contribution to IGCP project 437.

References

Agassiz, L., 1840. *Etudes sur les glaciers, privately published, Neuchtel.*

Alley, R.B. and Clark, P.U., 1999. The deglaciation of the northern hemisphere: a global perspective. *Annual Review of Earth and Planetary Science* **27**, pp. 149–182.

Baker, R.G.V. and Haworth, R.J., 2000. Smooth or oscillating late Holocene sea-level curve? evidence from cross-regional statistical regressions of fixed biological indicators. *Marine Geology* **163**, pp. 353–365.

Baker, R.G.V., Haworth, R.J. and Flood, P.G., 2001. Inter-tidal fixed indicators of former Holocene sea-levels in Australia: a summary of sites and a review of methods and models. *Quaternary International* **83-85**, pp. 257–273.

Banjeree, P.K., 2000. Holocene and Late Pleistocene relative sea level fluctuations along the east coast of India. *Marine Geology* **167**, pp. 243–260.

Bard, E., Hamelin, B., Arnold, M., Montaggioni, L., Cabioch, G., Faure, G. and Rougerie, F., 1996. Deglacial sea-level record from Tahiti corals and the timing of global meltwater discharge. *Nature* **382**, pp. 241–244.

Beaman, R., Larcombe, P. and Carter, R.M., 1994. New evidence for the Holocene sea-level high from the inner shelf, Central Great Barrier Reef, Australia. *Journal of Sedimentary Research* **A64** 4, pp. 881–885.

Blanchon, P. and Shaw, J., 1995. Reef drowning during the last deglaciation: evidence for catastrophic sea-level rise and ice-sheet collapse. *Geology* **23** 1, pp. 4–8.

Broecker, W.S. and Denton, G.H., 1989. The role of ocean–atmosphere reorganizations in glacial cycles. *Geochimica et Cosmochimica Acta* **53**, pp. 2465–2501.

Camoin, G.F., Colonna, M., Montaggioni, L.F., Casanova, J., Faure, G. and Thomassin, B.A., 1997. Holocene sea level changes and reef development in the southwestern Indian Ocean. *Coral Reefs* **16**, pp. 247–259.

Cann, J.H., Belperio, A.P., Gostin, V.A. and Rice, R.L., 1993. Contemporary benthic foraminifera in Gulf St. Vincent, South Australia, and a refined Late Pleistocene sea-level history. *Australian Journal of Earth Sciences* **40**, pp. 197–211.

Cann, J.H., Belperio, A.P. and Murray-Wallace, C.V., 2000. Late Quaternary palaeosealevels and palaeoenvironments inferred from foraminifera, northern Spencer Gulf, South Australia. *Journal of Foraminiferal Research* **30**, pp. 29–53.

Cann, J.H., Bourman, R.P. and Barnett, E.J., 2000. Holocene foraminifera as indicators of relative estuarine-lagoonal and oceanic influences in estuarine sediments of the River Murray, South Australia. *Quaternary Research* **53**, pp. 378–391.

Cann, J.H., Harvey, N., Barnett, E.J., Belperio, A.P. and Bourman, R.P., 2002. Foraminiferal biofacies ecosuccession and Holocene sealevels, Port Pirie, South Australia. *Marine Micropalaeontology* **44**, pp. 31–55.

Chappell, J. and Polach, H., 1991. Postglacial sea-level rise from a coral record at Huon Peninsula, Papua New Guinea. *Nature* **349**, pp. 147–149.

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

Clark, J.A., Farrell, W.E. and Peltier, W.R., 1978. Global changes in post glacial sea level: a numerical calculation. *Quaternary Research* **9**, pp. 265–287.

Clark, P.U., Alley, R.B., Keigwin, L.D., Licciardi, J.M., Johnsen, S.J. and Wang, H.X., 1996. Origin of the first global meltwater pulse following the last glacial maximum. *Paleoceanography* **11**, pp. 563–577.

Clark, P.U., Mitrovica, J.X., Milne, G.A. and Tamisiea, M.E., 2002. Sea-level fingerprinting as a direct test for the source of global meltwater pulse IA. *Science* **295**, pp. 2438–2441.

Colonna, M., Casanova, J., Dullo, W.C. and Camoin, G., 1996. Sea-level changes and d180 record for the past 34,000 years from Mayotte Reef, Indian Ocean. *Quaternary Research* **46**, pp. 335–339.

Compton, J.S., 2001. Holocene sea-level fluctuations inferred from the evolution of depositional environments of the southern Langebaan Lagoon salt marsh, South Africa. *The Holocene* **11** 4, pp. 395–405.

Daly, R.A., 1934. *The Changing World of the Ice Age*, Yale University Press, New Haven.

Dickinson, W.R., 2001. Paleoshoreline record of relative Holocene sea levels on Pacific islands. *Earth Science Reviews* **55**, pp. 191–234.

Done, T.J., 1982. Patterns in the distribution of coral communities across the central Great Barrier Reef shelf. *Coral Reefs* **1**, pp. 95–108.

Fairbanks, R.G., 1989. A 17,000-year glacio-eustatic sea level record: influence of glacial melting rates on the Younger Dryas event and deep-ocean circulation. *Nature* **342**, pp. 639–642.

Fleming, K., Yokoyama, Y., Lambeck, K., Chappell, J., Johnston, P. and Zwartz, D., 1998. Refining the eustatic sea-level curve since the Last Glacial Maximum using far- and intermediate-field sites. *Earth and Planetary Science Letters* **163** 1-4, pp. 327–342.

Flood, P.G. and Frankel, E., 1989. Late Holocene higher sea level indicators from eastern Australia. *Marine Geology* **90**, pp. 193–195.

Geyh, M.A. and Kudrass, H.-R., 1979. Sea-level changes during the late Pleistocene and Holocene in the Strait of Malacca. *Nature* **278**, pp. 441–443.

Grossman, E.E., Fletcher III, C.H. and Richmond, B.M., 1998. The Holocene sea-level high stand in the equatorial Pacific: analysis of the insular paleosea-level database. *Coral Reefs* **17**, pp. 309–327.

Hanebuth, T., Stattegger, K. and Grootes, P.M., 2000. Rapid flooding of the Sunda Shelf: a Late-Glacial sea-level record. *Science* **288**, pp. 1033–1035.

Hopley, D., 1975. Contrasting evidence for Holocene sea levels with special reference to the Bowen–Whitsunday area of Queensland. In: Douglas, I., Hobbs, J.E. and Pigram, J.J., Editors, 1975. *Geographical Essays in Honour of Gilbert J. Butland*, Department of Geography, University of New England, Armidale, pp. 51–84.

Horton, B.P., Edwards, R.J. and Lloyd, J.M., 2000. Implications of a microfossil transfer function in Holocene sea level studies. In: Shennan, I. and Andrews, J., Editors, 2000. *Holocene land-ocean interaction and environmental change around the North SeaGeological Society, London. Special Publications* **166**, pp. 275–288.

Houghton, J.T., Ding, Y., Griggs, D.J., Noguer, M., van der Linden, P.J. and Xiaosu, D., 2001. *Contribution of Working Group I to the third Assessment Report of the Intergovernmental Panel on Climate Change (IPCC)*, Cambridge University Press, Oxford.

Islam, M.S. and Tooley, M.J., 1999. Coastal and sea-level changes during the Holocene in Bangladesh. *Quaternary International* **55**, pp. 61–75.

Katupotha, J. and Fujiwara, K., 1988. Holocene sea level change on the Southwest and South coasts of Sri Lanka. *Palaeogeography, Palaeoclimatology, Palaeoecology* **68**, pp. 189–203.

Lambeck, K. and Chappell, J., 2001. Sea level change through the last glacial cycle. *Science* **292**, pp. 679–686.

Lambeck, K., Esat, T.M. and Potter, E.-K., 2002. Links between climate and sea levels for the past three million years. *Nature* **419**, pp. 199–206.

Lambeck, K., Purcell, A., Johnston, P., Nakada, M. and Yokoyama, Y., 2003. Water-load definition in the glacio–hydro–isostatic sea-level equation. *Quaternary Science Reviews* **22** 2-4, pp. 309–318.

Larcombe, P. and Carter, R.M., 1998. Sequence architecture during the Holocene transgression: an example from the Great Barrier Reef shelf, Australia. *Sedimentary Geology* **117**, pp. 97–121.

Larcombe, P., Carter, R.M., Dye, J., Gagan, M.K. and Johnson, D.P., 1995. New evidence for episodic post-glacial sea level rise, central Great Barrier Reef, Australia. *Marine Geology* **127**, pp. 1–44.

Li, L., Gallagher, S. and Finlayson, B., 2000. Foraminiferal response to Holocene environmental changes of a tidal estuary in Victoria, southeastern Australia. *Marine Micropalaeontology* **38**, pp. 229–246.

Lighty, R.G., Macintyre, I.G. and Stuckenrath, R., 1982. Acropora palmata reef framework: a reliable indicator of sea level in the Western Atlantic for the past 10,000 years. *Coral Reefs* **1**, pp. 125–130.

Long, A.J., 2000. Late Holocene sea-level change and climate. *Progress in Physical Geography* **24**, pp. 415–423.

McManus, J.F., Oppo, D.W. and Cullen, J.L., 1999. A 0.5-million-year record of millennial-scale climate variability in the North Atlantic. *Science* **283**, pp. 971–975.

Mitrovica, J.X., 2003. Recent controversies in predicting post-glacial sea-level change. *Quaternary Science Reviews* **22** 2-4, pp. 127–133.

Mitrovica, J.X. and Milne, G.A., 2002. On post-glacial sea level 1: general theory. *Geophysical Journal International* **133**, pp. 1–19.

Mitrovica, J.X. and Peltier, W.R., 1991. On postglacial geoid subsidence over the equitorial oceans. *Journal of Geophysical Research* **96** B12, pp. 20053–20071.

Mörner, N.-A., 1976. Eustasy and geoid changes. *Journal of Geology* **84**, pp. 123–151.

Mörner, N.-A., 1996. Rapid changes in coastal sea level. *Journal of Coastal Research* **12**, pp. 797–800.

Nakada, M., Yonekura, N. and Lambeck, K., 1991. Late Pleistocene and Holocene sea-level changes in Japan: implications for tectonic histories and mantle rheology. *Palaeogeography, Palaeoclimatology, Palaeoecology* **85**, pp. 107–122.

Nunn, P.D., 1998. Sea-Level changes over the past 1000 years in the Pacific. *Journal of Coastal Research* **14**, pp. 23–30.

Nunn, P.D. and Peltier, W.R., 2001. Far-field test of the ICE-4G model of global isostatic response to deglaciation using empirical and theoretical Holocene sea-level reconstructions for the Fiji Islands, Southwestern Pacific. *Quaternary Research* **55**, pp. 203–214.

Peltier, W.R., 1998. Postglacial variations in the level of the sea: implications for climate dynamics and solid-Earth geophysics. *Reviews of Geophysics* **36**, pp. 603–689.

Peltier, W.R., 2002. On eustatic sea level history: last Glacial Maximum to Holocene. *Quaternary Science Reviews* **21**, pp. 377–396.

Van de Plassche, O., Editor, 1986. *Sea-level research: a manual for the collection and evaluation of dataGeobooks, Norwich*.

Pirazzoli, P.A., 1991. *World atlas of Holocene sea-level changesOceanography Series* **vol. 58**, Elsevier, Amsterdam.

Pirazzoli, P.A., 1996. *Sea-level changes: the last 20 000 years*, Wiley, Chichester.

Ramsay, P.J., 1995. 9000 years of sea-level change along the southern African coastline. *Quaternary International* **31**, pp. 71–75.

Ramsay, P.J. and Cooper, J.A.G., 2002. Late Quaternary sea-level change in South Africa. *Quaternary Research* **57**, pp. 82–90.

Sawai, Y., Nasu, H. and Yasuda, Y., 2002. Fluctuations in relative sea-level during the past 3000 yr in the Onnetoh estuary, Hokkaido, northern Japan. *Journal of Quaternary Science* **17**, pp. 607–622.

Scoffin, T.P. and Le Tissier, M.D.A., 1998. Late Holocene sea level and reef-flat progradation, Phuket, South Thailand. *Coral Reefs* **17**, pp. 273–276.

Shennan, I. and Horton, B.P., 2002. Holocene land and sea-level changes in Great Britain. *Journal of Quaternary Science* **17**, pp. 511–526.

Shennan, I., Peltier, W.R., Drummond, R. and Horton, B.P., 2002. Global to local scale parameters determining relative sea-level changes and the post-glacial isostatic adjustment of Great Britain. *Quaternary Science Reviews* **21**, pp. 397–408.

Smithers, S.G. and Woodroffe, C.D., 2000. Microatolls as sea-level indicators on a mid-ocean atoll. *Marine Geology* **168**, pp. 61–78.

Stuiver, M. and Reimer, P.J., 1993. Extended ^{14}C database and revised CALIB radiocarbon calibration program. *Radiocarbon* **35**, pp. 215–230.

Stuiver, M., Reimer, P.J., Bard, E., Beck, J.W., Burr, G.S., Hughen, K.S., Kromer, B., McCormac, F.G., van der Plicht, J. and Spurk, M., 1998. INTCAL98 Radiocarbon Age Calibration, 24,000–0 cal BP. *Radiocarbon* **40**, pp. 1041–1083.

Stuiver, M., Reimer, P.J. and Braziunas, T.F., 1998. High precision radiocarbon age calibration for terrestrial and marine samples. *Radiocarbon* **40**, pp. 1127–1151.

Suess, E., 1888. *Das Anlitz der Erde, II: Die Meere der Erde. Wien*.

Tija, H.D., 1996. Sea-level changes in the tectonically stable Malay–Thai Peninsula. *Quaternary International* **31**, pp. 95–101.

Visser, K., Thunell, R. and Stott, L., 2003. Magnitude and timing of temperature change in the Indo-Pacific warm pool during deglaciation. *Nature* **421**, pp. 152–155.

Woodroffe, C.D. and McLean, R., 1990. Microatolls and recent sea-level change on coral atolls. *Nature* **344**, pp. 531–534.

Yim, W.W.-S. and Huang, G., 2002. Middle Holocene higher sea-level indicators from the South China coast. *Marine Geology* **182** 3-4, pp. 225–230.

Yokoyama, Y., Lambeck, K., De Deckker, P., Johnston, P. and Fifield, L.K., 2000. Timing of the Last Glacial Maximum from observed sea-level minima. *Nature* **406**, pp. 713–716.

Zong, Y., Shennan, I., Combellick, R.A., Hamilton, S.L. and Rutherford, M.M., 2003. Microfossil evidence for land movements associated with the AD 1994 Alaska earthquake. *The Holocene* **13**, pp. 7–20.