12-1-2005

Holocene sea levels and palaeoenvironments, Malay-Thai Peninsula, southeast Asia

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

P. L. Gibbard
University of Cambridge

G. M. Milne
University of Durham

R. J. Morley
Palynova Ltd

C. Purintavaragul
Prince of Songkla University

See next page for additional authors

Postprint verson. Published in *The Holocene*, Volume 15, Number 8, December 2005, pages 1199-1213.
Publisher URL: http://dx.doi.org/10.1191/0959683605hl891rp

Holocene sea levels and palaeoenvironments, Malay-Thai Peninsula, southeast Asia

Abstract

Sedimentological and palynological investigations of Great Songkhla Lakes, east coast of the Malay-Thai Peninsula, Southeast Asia, reveal sedimentary sequences rich in palynomorph assemblages dominated by pollen of mangroves and freshwater swamps. Compared with other regions in Southeast Asia the assemblages are of relatively low diversity. Geochronological data indicate that the Great Songkhla Lakes record one of the earliest mangrove environments in Southeast Asia (8420–8190 cal. yr BP), which are subsequently replaced by a freshwater swamp at 7880–7680 cal. yr BP owing to the decline of marine influence. Sea-level observations from Great Songkhla Lakes and other areas of the Malay-Thai Peninsula reveal an upward trend of Holocene relative sea level from a minimum of – 22 m at 9700–9250 cal. yr BP to a mid-Holocene high stand of 4850–4450 cal. yr BP, which equates to a rise of c. 5.5 mm/yr. The sea-level fall from the high stand is steady at c. – 1.1 mm/yr. Geophysical modelling shows that hydroisostasy contributes a significant spatial variation to the sea-level signal between some site locations (3–4 m during the mid-Holocene), indicating that it is not correct to construct a single relative sea-level history for the Malay-Thai Peninsula.

Keywords

LAKE SEDIMENTOLOGY, PALYNOLOGY, RELATIVE SEA LEVEL, GEOPHYSICAL MODELLING, HYDRO-ISOSTASY, MANGROVE, SWAMP, ARCHAEOLOGY, MID-HOLOCENE HIGH STAND, THAILAND

Comments

Postprint verson. Published in *The Holocene*, Volume 15, Number 8, December 2005, pages 1199-1213.
Publisher URL: http://dx.doi.org/10.1191/0959683605hl891rp

Author(s)

Benjamin P. Horton, P. L. Gibbard, G. M. Milne, R. J. Morley, C. Purintavaragul, and J. M. Stargardt

Holocene sea levels and palaeoenvironments, Malay-Thai Peninsula, southeast Asia

B. P. HORTON. Sea Level Research Laboratory, Department of Earth and Environmental Science, University of Pennsylvania, Philadelphia, PA 19104-6316, USA. Tel: 215 573 5388; Fax: 215 898 0964; E-mail: bphorton@sas.upenn.edu.

P. L. GIBBARD. Godwin Institute for Quaternary Research, Department of Geography, University of Cambridge, Downing Place, Cambridge CB2 3EN, UK.

G. M. MILNE Department of Earth Science, University of Durham, South Road, Durham, DH1 3LE, UK.

R. J. MORLEY. Palynova, Vila Indah Pajajaran, Jalan Kertarajasa No 12A, Bogor 16153, Indonesia.

C. PURINTAVARAGUL, Department of Biology, Faculty of Science, Prince of Songkla University, Hat Yai, Songkhla 90112, Thailand.

J. M. STARGARDT. Sidney Sussex College and Quaternary Palaeoenvironments Group, Department of Geography, University of Cambridge, Downing Place, Cambridge CB2 3EN, UK.

Abstract

We reveal through sedimentological and palynological investigations of Great Songkhla Lakes, east coast of the Malay-Thai Peninsula, Southeast Asia, sedimentary sequences consisting of very rich palynomorph assemblages dominated by pollen of mangroves and freshwater swamps, but compared with other regions in Southeast Asia are of relatively low diversity. Geochronological data indicate that the Great Songkhla Lakes record one of the earliest mangrove environments in Southeast Asia (8420 to 8190 cal yrs BP), which are subsequently replaced by a freshwater swamp at 7880 to 7680 cal yrs BP due to the decline of marine influence.

Sea-level observations from Great Songkhla Lakes and other areas of the Malay-Thai Peninsula reveal an upward trend of Holocene relative sea level from a minimum of -22 m at 9700 - 9250 cal yrs BP to a mid-Holocene high stand of 4850 - 4450 cal yrs BP, which equates to a rise of c. 5.5 mmyr^{-1}. The sea-level fall from the high stand is steady at c. -1.1 mmyr^{-1}. Geophysical modelling shows that hydro-isostasy contributes a significant spatial variation to the sea-level signal between some site locations (3-4 m during the mid-

Holocene), indicating that it is not correct to construct a single relative sea-level history for the Malay-Thai Peninsula.

Keywords: sedimentology, palynology, relative sea level, geophysical modelling, hydro-isostasy, mangrove, archaeology, mid-Holocene high stand

Introduction

There has been a fascination with the palaeoecology of Southeast Asia for some time (e.g. Whitmore, 1975, 1994, 1998; Morley, 1981; Collins *et al.,* 1991; Van de Kaars, 1991, 1997, 2001; Woodroffe, 1993; Dam *et al.,* 2001; Hope, 2001; White *et al.,* 2004). The region is a key tropical biodiversity hotspot because it is considered a geological and biological anomaly (Wallace, 1869; Daws and Fujita, 1999) and is located at a zone of transition between the two distinct faunas associated with the Asian and Australian continents. Furthermore, there is a pressing need for high quality Holocene environmental evidence from the region to better understand global climate change (Street-Perrott and Perrott, 1990; Dam *et al.,* 2001; Gupta *et al.,* 2003; Visser *et al.,* 2003; Dickens *et al.,* 2004). Southeast Asia has strong through-flows from the warm western Pacific Ocean to the Indian Ocean, which is a major source of heat for the globe (Hope, 2001). Moreover, the region is critically located in relation to monsoon activity and El Niño variability (Cobb *et al.,* 2003; Gupta *et al.,* 2003).

With respect to sea-level studies, considerable research (e.g. Shennan and Horton, 2002) has focused on datasets from regions once covered by major ice sheets (near- and intermediate-field sites). However, it is widely recognised that regions distant from the major glaciation centres (far-field sites) that are tectonically stable will provide the best possible estimate of the 'eustatic function' (Clark *et al.,* 1978; Yokoyama *et al.,* 2000; Peltier, 2002; Woodroffe and Horton, 2005 in press), which provides information on the transfer of immense amounts of water between two of its largest reservoirs on Earth: the ice sheets; and the oceans (Lambeck *et al.,* 2002). Despite the accepted importance of Holocene sea levels and environmental change to understanding the region and its potential role in climate forcing and human society, research into most aspects of Holocene history of Southeast Asia has been limited (e.g. Nunn, 1998; Hanebuth *et al.,* 2001; Penny, 2001).

The Great Songkhla Lakes of Songkhla-Pattalung-Nakorn Si Thammarat Provinces lie on the east coast of the Malay-Thai Peninsula, Southeast Asia between 6˚90’ to 7˚60’N and 106˚5’ to 106˚20’E (Figure 1). This significant location forms part of the famous Kra Ecotone of the Isthmus, and is at the centre of the shallow Gulf of Thailand-South China Sea area described by the coastlines of Thailand, Cambodia, Vietnam, the north coast of Borneo, eastern Indonesia and the west coast of Palawan Island (Philippines). It is of known regional significance as it frontiers two climatic and forest zones (Whitmore, 1994, 1998), and contains an exceptional concentration of archaeological sites from 6000-600 cal yrs BP and became a primary zone of settlement in the early historic period, 2000 - 1400 cal yrs BP (Stargardt, 1983, 1998, 2001). The lagoons of the Great Songkhla Lakes are separated from the sea by a long and substantial beach-barrier system: the Satingpra Peninsula (Stargardt, 1983).

Background to the study area

The Indonesian archipelago of Southeast Asia has been subject to intense tectonic processes through the Holocene. The region is a collision zone between the Eurasian, Indian, Australian, 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. However, the Malay-Thai Peninsula is found within a geologically stable region known as Sundaland, which has experienced very slow vertical crustal movements (less than 0.1 $mmyr^{-1}$; Tjia, 1996).

The Malay-Thai Peninsula of Southeast Asia lies in the humid tropics. The Satingpra Peninsula lies in a rain shadow for the southwest monsoon. Thus most of its rainfall comes from the northeast monsoon in October to February. Its annual rainfall is 1200 – 1500 mm per year. The southern, western and northern perimeters of the Songkhla Lakes, by contrast, average 1500 – 2000 mm per year, receiving rainfall from both the southwest and northeast monsoons. Tides in the area are semi-diurnal and microtidal, with mean spring tidal ranges of 1.90m (Table 1. Admiralty Tide Tables, 2003).

Methods

We investigated satellite imagery, air photographs, archaeological and topographical maps of the Great Songkhla Lakes of Songkhla-Pattalung-Nakorn

Si Thammarat Provinces to determine the key areas for field examination. Subsequently, we decided to investigate the isolation of the small Thale Noi, found north of the major lagoon system (Figure 1), to understand more specifically the development of the lake basin system and to attempt to build a sequence of events. Extensive field examination over a period of six weeks included open sections, boreholes at critical localities and study of landforms from the ground and air. We undertook sampling from cores with a series of samples being collected for both sedimentological and palynological analysis.

The analysis of fossil pollen and spore assemblages is a sensitive technique for the reconstruction of past plant communities (e.g. Moore *et al.,* 1991), sea-level history (e.g. Horton *et al.,* 2004) and coastal zone palaeogeography (e.g. Willard *et al.,* 2003). We have used pollen analysis to assign the organic units within sediment cores to depositional wetland environments in order to reconstruct the palaeogeography of the coastal zone. We follow the standard procedures outlined in Moore *et al.* (1991) for pollen preparation and identification.

Approximately 300 pollen and spores have been counted from each sample. Freshwater algae, marine algae and charred graminoid cuticle fragments were logged in addition to this sum. Assemblages are of three types: a) fern spores dominant with minor mangrove pollen/spores and freshwater-derived pollen; b) mangrove pollen/spores dominant with subordinate freshwater-derived fern spores, freshwater-derived pollen, marine and freshwater algae; and c) freshwater-derived fern spores dominant with subordinate mangrove pollen/spores and freshwater-derived pollen. To illustrate vegetation change within both brackish and freshwater settings and to emphasize both local and regional vegetation change, 'total freshwater-derived pollen and spores' has been used as the 'main' pollen sum, with mangrove pollen/spores being presented in terms of 'total freshwater-derived and total mangrove pollen/spores'. Freshwater algae is presented in terms of 'total freshwater-derived pollen and spores and total freshwater algae'.

Palynomorph concentration has been estimated by using an exotic 'spike'. A single Lycopodium clavatum tablet was added to each l cm sample. This resulted in assemblages yielding between 10% and 50% of the spike, Palynomorph concentrations vary between 3000 and 48000 per cm^3 in clastic, and from 42000 to 60000 per cm^3 in organic lithologies.

Mangrove associations are interpreted in terms of salinity zones by reference to surface sample studies from mangrove and tidal swamps from the southern Malay-Thai Peninsula (Morley, 1999; Kamaludin, 2001). Peat swamp vegetation is interpreted by reference to the phasic communities of Anderson (1963, 1983) and by reference to surface sample studies from peat swamps in Brunei (Anderson and Muller, 1975) and Central Kalimantan (Morley, 1981).

We followed a consistent approach to reconstruct sea-level history. This involves sedimentological and palynological analysis at the site scale to establish the relationship between sedimentation, water table height and sea-level changes. For age/altitude analysis, a sea-level index point must have an indicative meaning. The indicative meaning of a coastal sample is the relationship of the local environment in which it accumulated to a contemporaneous reference tide level (van de Plassche, 1986). The indicative meaning can vary according to the type of coastal sample 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, such as, highest astronomical tide (HAT) and mean high high water (MHHW). Relative sea level is calculated as altitude minus the reference water level. The RSL error range is calculated as the square root of the sum of squares of altitudinal error, sample thickness, tide level error and indicative range (given as a maximum).

The method for estimating the indicative meaning from temperate coastlines was developed during the International Geological Correlation Programme (IGCP) Project 61, and has been a component of all subsequent IGCP Projects, especially Projects 200, 274, 367 and 437 (e.g. van de Plassche, 1986; Shennan and Horton, 2002). However, there is little information for tropical coastlines (Zong, 2004). Thus, we have compiled numerous results from the local area and beyond (Anderson, 1963, 1983; Anderson and Muller, 1975;

Morley, 1991, 1999; Semeniuk, 1980; Grindrod, 1988; Kamaludin, 1989, 1993, 2001; Sa and Hock, 1993; Bunt, 1999; Cantera *et al.,* 1999; Horton *et al.,* 2003, 2005 in press; Woodroffe *et al.,* in press) to make conservative estimates of the indicative meanings of commonly dated materials. For example, where pollen data can demonstrate a gradual transition from an organic freshwater deposit to a clastic mangrove sediment, the contact is validated as a sea-level index point which formed between HAT and MHHW (Table 2). Where the palynological data for the sample indicate a palaeoenvironment that is not clearly related to a fossil tide level, the sample is classified as a “limiting date”. Such samples may indicate a water table controlled by tide levels or they may be independent (i.e. above, tide levels). Therefore, limiting ages have a wide indicative range but help constrain sea-level reconstructions because sea level must have been at or below the level indicated by a “limiting age” (Shennan and Horton, 2002).

Palaeoenvironmental analysis

Sedimentology

The sedimentary sequences of Thale Noi lake basin clearly contain four major sedimentary units (Figure 2). Detailed analyses of Core TN-3 (the longest of the cores obtained) show the basal unit (7.00 m – 6.75 m) comprises a hard, uncompressible, light brown clay with sandy horizons (Figure 2). This is overlain by highly tenacious organic grey clay with an increase in dark detritus mud between 6.70 m and 6.31 m, which has an organic content greater than 50%.

The upper part of the organic grey clay between 4.18 m and 3.03 m also shows an increase in organic matter, notably wood and herbaceous detritus. The clay is overlain at 3.03 m by a dark brown fibrous peat. This unit is rich in coarse plant material and, although variable, comprises 70% to 75% organic matter by mass. Within this unit there is a marked decline in organic matter to below 50% at 1.25 m. It is not clear from the analyses or from the sediments themselves whether this represents a true interruption in sedimentation or whether it results from inwash of inorganic matter as a consequence of flooding. Finally, there is a fourth stratigraphic unit from about 0.27 m to the ground surface. Here there is an alteration of inwashed detritus mud and fibrous peat that is shown in the analyses by a marked increase in the silicate component accompanied by a sharp decline in organic sedimentation to about 55% by mass.

Palynology

Palynological analysis of Core TN-3 shows that the basal light brown clay is dominated by abundant monolete fern spores with rare pollen of freshwater taxa and of freshwater algae such as *Concentricystes circulus* (Figure 2). The pollen suggests a freshwater depositional setting, and most probably reflects a palaeosol. The low miospore concentration (3400 specimens per cm^3) probably reflects palynomorph degradation within the soil profile (Figure 3).

The overlying organic grey clay is dominated by abundant *Rhizophora*-type pollen and high palynomorph concentrations (27700 per cm^3), which suggest the presence of a mangrove swamp with salinities in the order of 5 - 20. However, there are subtle changes of environment within this stratigraphic unit. Between 6.07m and 5.77m, the presence of marine dinoflagellates, such as *Spiniferites* and *Polysphaeridium* suggests an increase in saline influence (10-15). The lithology at these horizons are well-sorted fine silts, and a setting within the distal fringe of a mangrove swamp, or in a tidal flat adjacent to mangroves, is most probable.

The presence of common *Nypa* pollen and *Acrostichum* spores suggests a return to a back-mangrove setting with salinities in the order of 5 between 5.77 m and 5.17 m. This sample provides the best evidence in the profile for a brackish swamp dominated by the mangrove palm *Nypa fruticans*. Between 5.17 m and 4.27 m, abundant *Rhizophora*-type pollen with very high palynomorph concentrations of 48500 per cm^3 suggest a depositional setting within a mangrove swamp dominated by *Rhizophora* or possibly *Bruguiera*, or possibly a tidal flat adjacent to mangroves. With the presence of dinoflagellate cysts, this is probably the most marine-influenced section of the profile, with salinities in the order of 15-20. Towards the stratigraphic boundary between the organic clay and overlying peat, superabundant levels of *Rhizophora*-type pollen (up to 95%) and extremely high palynomorph concentrations (over 163000 per cm^3) are shown. These assemblages suggest a very widespread mangrove swamp at the sample location.

With reference to regional vegetation, the climate and vegetation disturbance within the organic grey clay between 6.67 m and 4.03 m show that Gramineae pollen and charred Gramineae cuticle fragments are persistent elements, which could have been derived from grass swamps growing behind the mangrove belt, or from dry land vegetation. The presence of charred grass cuticles suggests burning of grasslands, either by human or natural causes (Stargardt, 1983, 1998, 2001). Scattered *Pinus* pollen through this interval suggests some seasonal dry land vegetation on a regional/supraregional basis. Pollen of rainforest trees is scattered throughout the section, indicating that rain forests were present in the area from 8500 cal yrs BP, but few comments can be made about their composition, other than that they include the dipterocarps *Shorea* and *Dipterocarpus*, *Burseraceae* and the forest palm *Eugeissona*. The first three species remain ecologically dominant and economically significant up to the present (Stargardt, 1983, 1998, 2001; Stargardt and Purintavaragul, 1999).

The upper section of the organic grey clay yields reduced percentages of Gramineae pollen and charred cuticle compared to the section underlying. This may be due to the swamping effect of mangrove pollen. However, the regular presence of pollen of *Flagellaria indica* suggests a possible change to more oligotrophic soils.

The penultimate stratigraphic unit (brown fibrous peat) is dominated by pollen indicative of freshwater conditions. The lower section of the peat is wholly dominated by spores of the bole climber *Stenochlaena palustris* with a virtual absence of mangrove or other pollen or spores. *Stenochlaena palustris* is common on disturbed freshwater swamps with sudden and severe disruption of the vegetation. Mangrove swamps, dominated by *Rhizophora* or *Bruguiera*, perhaps festooned with the climbing fern *Lygodium scandens*, are indicated for the centre of the peat from sample 205 cm, with freshwater swamps with *Stenochlaena palustris* in close proximity. Salinities were likely to be in the order of 5. *Casuarina* pollen is common in this sample, possibly suggesting the widespread occurrence of *Casuarina* on a nearby beach ridge. Pollen samples from the upper section of the peat yield common pollen of peat swamp trees, such as *Campnosperrna*, together with pollen of the swamp palm *Zalacca*, indicating the presence of vegetation transitional to true peat swamp forest. This association compares favourably with the *Campnosperma/Cyrtostachys/Zalacca* subassociation of Anderson (1983). The sample is rich in *Nepenthes* pollen tetrads, suggesting that the insectivorous pitcher plant *Nepenthes* was prolific in this vegetation. Due to the dominance of locally sourced pollen and spores, no comments can be made regarding regional vegetation from the pollen samples of the brown fibrous peat.

The fourth stratigraphic unit yields common *Palaquium* pollen, and several other peat swamp elements, suggesting the development of true peat swamp forest (Phasic community 1 of Anderson 1963). Peat swamp elements are greatly reduced in this sample, and the abundance of Gramineae pollen suggests the removal of peat swamps, and their replacement by grass swamp which is subject to burning, as indicated by the common occurrence of charred Gramineae cuticle fragments.

Sea-level index points

We used the sedimentological and palynological data from Thale Noi Core 3 (TN-3) to evaluate the status of the radiocarbon dated contacts as index points for the reconstruction of sea-level history. The core provides two sea-level index points and one limiting data point. The contacts between the basal clay and organic clay (-4.47 m above MSL), and the organic clay and fibrous peat (-0.76 m above MSL) of Core TN-3 have been dated to 7535 ± 55 BP (8420 to 8190 cal yrs BP) and 6980 ± 50 BP (7880 to 7680 cal yrs BP), respectively (Table 3). The contact between basal clay and organic clay represents a transgressive sea-level index point formed between HAT and MHHW during a positive tendency of sea-level movement as pollen and dinoflagellates indicators at the bottom of the organic clay and above show its deposition within intertidal or mangrove environments. The contact of the organic clay and fibrous peat signifies a regressive index point formed between HAT and MHHW during a negative sea-level tendency as mangrove species are replaced by pollen indicative of freshwater conditions. The limiting data point from the upper contact of the fibrous peat (-1.47 m above MSL) has been dated to 2435 ± 50 BP (2720 to 2350 cal yrs BP). The limiting date marks the change in environment from freshwater swamp to a true peat swamp forest (Anderson, 1963).

Holocene environmental change

Palynological analysis of the Thale Noi Core 3 reveals the presence of very rich palynomorph assemblages (Figure 3) dominated by pollen of mangroves and freshwater swamps but, compared with other regions in Southeast Asia, of relatively low diversity, with only 121 palynomorph taxa being recorded from the 22 samples analysed. At more equatorial latitudes, this number of taxa would be expected in a single sample from a shallow marine environment (e.g. Anderson 1963, 1983; Morley, 1981; Sa and Hock, 1993; van der Kaars *et al.,* 1991, 2000; van der Kaars, 2001). The basal stratigraphic unit of Thale Noi Core 3 probably reflects a palaeosol formed in a freshwater depositional setting. Kamaludin (1989, 1993) studied the Kuala Kurau, western coast of the Malay-Thai Peninsula and suggested that the palaeosol marked the Pleistocene-Holocene boundary. At Thale Noi, a lithological succession has formed over the soil surface following an increase in marine influence. The overlying organic unit is dominated by mangrove pollen (up to 95% of the assemblage) and is dated to 8420 to 8190 cal yrs BP. This is one of the earliest recorded mangrove environments in Southeast Asia (Somboon and Thiramongkol, 1992). Streif (1979), Bosche (1988) and Kamaludin (1989, 1993) dated the appearance of mangrove environments on the western coast of the Malay-Thai Peninsula to 7010 – 6630, 6300 – 5700 and 5920 – 5600 cal yrs BP, respectively.

Furthermore, a study of lowland swamp dynamics in Borneo (Morley, 1981) indicates the occurrence of mangrove conditions from 5400 cal yrs BP. In Northern Australia, detailed studies of estuarine plains (e.g. Woodroffe, 1993; Woodroffe *et al.,* 1985; Crowley *et al.,* 1990; Crowley and Gagan, 1995; Larcombe *et al.,* 1995) documented a transgressive phase as sea-level rose, with widespread mangrove swamps from 7850 cal yrs BP. The pollen record from Core TN-3 suggests that the representation of mangroves was previously much greater than that seen along this coastline today, which was also noted by Stargardt (1983, 1998, 2001) and Stargardt, and Purintavaragul (1999). The main dominants in these swamps were fairly typical of the Southeast Asian mangrove flora, being dominated by *Rhizophoraceae.* In contrast the present mangrove flora is somewhat restricted, and is dominated by *Sonneratia caseolaris*.

The mangroves of Thale Noi were replaced by a freshwater swamp at 7880 to 7680 cal yrs BP due to the decline of marine influence. Crowley and Gagan (1995) suggest that in regions of high rainfall, such as the Malay-Thai Peninsula, high freshwater input combined with lower rates of relative sea-level rise prevents the development of hypersaline soils and permits the direct replacement of mangrove forests with nonhalophytic communities. This contrasts with areas of lower rainfall, for instance Northern Australia (less than 1200 $mmyr^{-1}$), where along prograding shores away from riverine influence, increasing salinization caused by irregular tidal inundation has led to the replacement of mangroves by saltmarsh (Grindrod, 1985; Crowley and Gagan, 1995).

The freshwater swamp of Thale Noi grades into the early stage of true peat swamp forest (2720 to 2350 cal yrs BP) before burning resulted in the swamp forest's destruction. Interestingly, there is no sedimentological and palaeontological evidence for the development of *Melaleuca* swamp or the freshwater herbaceous swamp communities which characterise the swamp forests in the region today. The absence of evidence for the development of the present swamp communities may reflect the cessation of peat formation prior to their establishment. The non-representation of present-day communities in the pollen profile may also be due to low pollen productivity or poor preservation.

The development of this basin has determined the pattern and date of human settlement, while later the lake system itself seems to have been modified by human activities (Stargardt 1983, 1998). The presence of charred grass cuticles and charcoal through the mangrove, freshwater swamp and true peat swamp, indicates that fire has long been a feature of the dynamics of the environment of the Malay-Thai Peninsula. However, considering *Homo sapiens* is likely to have been present in Southeast Asia for at least 60000 years (the oldest inferred age; Dam *et al.,* 2001), it cannot be established that fires would have occurred naturally in the forests. With respect to the local study area, Songkhla-Pattalung-Nakorn Si Thammarat Provinces, scattered prehistoric settlements occurred *c.* 4500 cal yrs BP in calcareous karstic rock shelters (e.g. the Khao Rang Kiat group) along the eastern edge of the central uplands of the Isthmus (Stargardt, 1983, 1998; Stargardt and Purintavaragul, 1999). Humans then moved eastwards, following the changes in the lakes, probably to exploit fishing and aquatic plants (Purintavaragul, 1988, 1999). The excavation of channels through the ridge barrier occurred in the early/mid-historic period (1500-600 BP) (Stargardt. 1983, 1998), and thus it seems highly likely that the extensive burning and disturbance of the true peat swamp forest in the lagoonal system may have been a consequence of this human intervention.

There is no evidence for climatic change through the profile, but the evidence for grassland vegetation, which has been subject to burning, throughout the profile, and the presence of scattered *Pinus* pollen in the lower part of the core suggests that seasonal climates were characteristic for the whole of the period represented by the profile (Boyd and McGrath, 2001).

Sea-level reconstructions from Malay-Thai Peninsula

The most important feature of Holocene sea-level records from far field locations, such as the Malay-Thai Peninsula, is the magnitude and timing of the mid Holocene highstand. This highstand results from the interplay between glacio- and hydro-isostasy, and geoidal eustasy, and differs in timing and magnitude across the Indo-Pacific. In addition, evidence in selected locations across Southeast Asia points towards a second, smaller magnitude late

Holocene highstand occurring; this is not predicted by current geophysical models (e.g. Peltier *et al.,* 2002).

Holocene sea-level studies in the Malay-Thai Peninsula include various local studies such as the Straits of Malacca (Streif, 1979; Geyh *et al.,* 1979; Hesp *et al.,* 1998), Phuket, Southwest Thailand (Scoffin and Le Tissier, 1998), and Malay Peninsula (Tjia, 1996; Kamaludin, 2001). The study by Geyh *et al.* (1979) conducted on 33 ^{14}C dated fossil mangrove deposits in the Straits of Malacca, Indonesia, shows that Holocene sea level rose from -13 m to 5 m above present between 8000 and 4000 ^{14}C yrs BP. The data do not show whether the late Holocene sea level lowering was a steady or oscillatory process. Tjia (1996) compiled a database for a variety of published and unpublished index points for the Malay-Thai Peninsula. The sea-level curve for the Malay Peninsula implies two Holocene highstands at 5000 and 2800 ^{14}C yrs BP, where as his sea-level reconstruction for Thailand indicates potentially three mid/late Holocene highstands at 6000, 4000 and 2700 ^{14}C yrs BP. However, these studies do not take into account the broad altitudinal range of the indicators and it is hard to reconcile this model of sea-level history with the scattered evidence that is presented. Scoffin and Le Tissier (1998) studied massive *Porites* coral evidence for mid/late Holocene highstands at Phuket, southwest Thailand. The evidence here suggests a single Holocene highstand with a constant rate of RSL fall since 6000 cal yrs BP, and thus, contradicts conclusions drawn by Tjia (1996).

However, it should be noted that the Scoffin and Le Tissier (1998) research is based on only one site with eleven dated samples.

Kamaludin (2001) studied two contrasting locations from the west and east coasts of the Malay Peninsula to produce a series of index points, which indicate a single highstand of over 3m at *c*. 6000 cal yrs BP. Furthermore, the study suggested differences in sea-level history between the west and east coasts. However, this is based on only seven index points.

The sea-level data from Malay-Thai Peninsula locations do not meet the fundamental criteria to produce an accurate sea-level curve. The accuracy or significance of each curve depends upon the number of data points (e.g. Hesp *et al*., 1998; Scoffin and le Tissier, 1998; Kamaludin, 2001), the correct interpretation of their relation to the corresponding mean sea level, and the quality of age determinations. There are serious problems associated with the correct interpretation of the altitudinal relationship of the sea-level indicators to sea level. Different types of indicators have different degrees of precision, but these are often not acknowledged (e.g. Tjia, 1996). Altitude error ranges are rarely included on sea-level curves (e.g. Geyh *et al.,* 1979; Scoffin and Le Tissier, 1979). Furthermore, low tide level has been used as a common datum rather than mean tide level (e.g. Scoffin and le Tissier, 1998), precluding regional correlation of different studies.

The second serious form of error concerns age determinations, in particular the necessary calibration of radiocarbon ages (Yim, 1999). Nevertheless, all the sea-level studies from the Southeast Asian region, except Kamaludin (2001), only present radiocarbon ages. Thus, we have compiled a new sea-level database for the Malay-Thai Peninsula, applying the consistent methods developed during IGCP projects to data to extract accurate sea-level index points from different palaeo-coastal environments to constrain Holocene sea levels (Appendix). The greater part of the previous dataset was rejected (c. 70%) owing to missing information (a fault which could conceivably be corrected with further research), or to uncertainty over the reliability of their relationship to a past sea-level. This is particularly the case for sub-tidal indicators.

Figure 4 shows an age-altitude plot of the validated index points, including the data from this study, for the Malay-Thai Peninsula. It comprises 81 sea-level index points quantitatively related to a past tide level together with an error estimate, and a further 12 data points that provide limits on the maximum altitude of the contemporary local sea level (Figure 5, Appendix). The plot of all index points from Southeast Asia confirms the upward trend of Holocene relative sea level to a mid-Holocene high stand and subsequent sea-level fall to present, typical of a far-field area. Relative sea-level rises from an observed minimum of -22.15 m ± 0.55 m at 9700 - 9250 cal yrs BP to a maximum of 4.87 m ± 0.57 m at 4850 - 4450 cal yrs BP, which equates to a presumed average rate of sea-level rise of *c*. 5.5 $mmyr^{-1}$. The sea-level fall from the high stand is steady (*c*. -1.1 $mmyr^{-1}$), with no evidence of a second high stand. However, observations from the Straits of Malacca indicates that relative sea-level was below present between 760 and 1095 cal yrs BP. These observations conflict with geophysical model predictions (see below) and invite further analysis.

Predictions of glaciation-induced sea-level changes are shown in Figure 5 for the seven locations across the region considered in this study. Each of the seven locations was defined as the average position for a cluster of closely grouped core locations (see Figure 6). The predictions are based on a spherically symmetric, self-gravitating and compressible Maxwell visco-elastic Earth model. The elastic and density structure of the model are taken from seismic constraints (Dziewonski and Anderson, 1981) and the viscous structure is defined in three layers comprising a very high viscosity outer shell, or lithosphere, with thickness 100 km, a viscosity of 5×10^{20} Pa^{-1} s^{-1} beneath this region to a depth of 670 km (defining the sub-lithosphere upper mantle region) and a viscosity of 10^{22} Pas^{-1} from 670 km to the core-mantle boundary (defining the lower mantle region). This viscosity structure is broadly compatible with that inferred in a number of recent studies (e.g. Mitrovica and Forte, 1997; Lambeck *et al.,* 1998; Kaufmann and Lambeck, 2003). The ice model adopted to generate the predictions shown in Figure 5 is based on the ICE-3G deglaciation history (Tushingham and Peltier, 1991). A number of significant revisions have been made to this model, including the addition of a glaciation phase and tuning of the model to produce a good fit to far-field sea-level records (see Bassett *et* *al.,* 2004 for more details). The predictions shown in Figure 5 are based on a revised sea-level equation that incorporates new improvements such as migrating shorelines, perturbations in Earth rotation and an accurate treatment of the water loading in near-field regions during times of ice retreat (e.g. Milne, 2002).

The top-left frame in Figure 5 shows that significant scatter (3-4 m during the mid-Holocene) is predicted between some of the locations (e.g. compare predictions for sites C and F). This scatter is dominated by the hydro-isostasy signal associated with the local ocean loading from the Last Glacial Maximum to the mid-Holocene. The spatial variation in the sea-level signal across the region is shown in Figure 6 at the time of the predicted highstand, 6000 cal yrs BP. The results show a variation in highstand amplitude of c. 7 m across the region. The dominant influence of hydro-isostasy is clear from the relatively large gradients predicted perpendicular to the shorelines of major land masses. The gradient is associated with the hydro-isostatic process know as continental levering (e.g. Clark *et al.,* 1978) which results in an upwarping of continental areas relative to the ocean basins causing elevated sea-level highstands at sites located furthest inland towards continental areas (e.g. sites A, E and F in Figure 6). The model predictions illustrated in Figures 5 and 6 capture the general temporal and spatial trend of the observations reasonably well. The model has not been tuned to fit the data and so there are significant misfits. In particular, the results show that the model over-predicts sea level between c.

8000 and 4000 cal yrs BP. This discrepancy is relatively consistent for most sites considered, suggesting that some alteration of the ice model is required to increase the melt rate, and thus lower the predicted sea levels during this period. (Note that the relatively large misfit for site A could be a consequence of the sediment loading associated with the Chao Phraya Delta). We are currently performing a more detailed modelling analysis to determine an optimal Earth and ice model for these data. The results of this work will be reported in a future article.

The modelling results show that hydro-isostasy is a contributor to the scatter illustrated in Figure 4. A number of other processes are also potential contributors to the spatial variation evident in the data. For example, tidal range variations and sediment compaction cannot be ruled out. Changes in tidal range during the time period under consideration have the potential to introduce altitudinal errors into sea-level reconstructions; however, such changes are likely to be less significant in areas of low tidal ranges such as the Malay-Thai Peninsula (Admiralty Tide Tables, 2003). In contrast, the majority of sediments employed as index points have been subjected to post-depositional displacement in altitude largely due to compaction of the underlying peat, clay and silt (Shaw and Cemen, 1999). Shennan and Horton (2002) have reviewed some of the literature on this topic and stressed that compaction of deposits with a high sand fraction is very low whilst compaction of peat may be as high as 90% by volume. Despite this, the problem of compaction of sediments within sea-level research has not been fully resolved.

Conclusion

We investigated the sedimentology and palynology of the Great Songkhla Lakes, east coast of the Malay-Thai Peninsula, to reveal the occurrence of very rich palynomorph assemblages, which are dominated by pollen of mangroves and freshwater swamps, but compared to other regions in Southeast Asia are of relatively low diversity. The sedimentary sequences reveal four major sedimentary units. The basal unit comprised a hard, uncompressible, light brown clay with sandy horizons reflecting a palaeosol formed in a freshwater depositional setting that marks the Pleistocene-Holocene boundary. The subsequent sedimentary unit, a highly tenacious organic grey clay dominated by mangrove pollen, is one of the earliest recorded mangrove environments in Southeast Asia (8420 to 8190 cal yrs BP). It spans the major climatic instability event which occurred around 8200 years ago (e.g. Barber *et al*., 1999). The mangroves were replaced by a freshwater swamp, composed of a dark brown fibrous peat, at 7880 to 7680 cal yrs BP as the result of decline of the marine influence. The fourth stratigraphic unit indicates that the freshwater swamp grades into the early stage of true peat swamp forest (2720 to 2350 cal yrs BP), before burning resulted in the swamp forest’s destruction.

Following IGCP methods, we produce two sea-level index points and one limiting data point, which when combined with other verified sea-level observations (81 index points and 12 limiting data in total) from the Malay-Thai Peninsula reveal an upward trend of Holocene relative sea level from a minimum of -22 m at 9700 - 9250 cal yrs BP to a mid-Holocene high stand of 5 m at 4850 - 4450 cal yrs BP, which equates to a rise of *c*. 5.5 $mmyr^{-1}$. The sea-level fall from the high stand is steady at *c*. -1.1 $mmyr^{-1}$. Geophysical modelling shows that hydro-isostasy can account for some of the spatial variation in sea level across the region considered. These results suggest that it is not valid to construct a single (spatially invariant) sea-level history for this region.

Word Count (Excluding references, tables and figure captions) = 5994

Acknowledgements

Although the work in this investigation is shared, J. Stargardt and P.L. Gibbard were principally responsible for the geology and the field work, together with R.J. Morley and C. Purintavarangul who investigated the palaeobotany. B.P Horton and G.M. Milne undertook the sea-level synthesis and the sea-level modelling. All the authors gratefully acknowledge the considerable field assistance of Professor A.L. Vietti Bitencourt (Unisinos, Brazil) and Santi Morley (Palynova). We also thank NERC for the support of a small research grant and for funding the radiocarbon dates, the National Research Council of Thailand for permission to carry out the fieldwork and The Faculty of Science, Prince of

Songkhla University, Thailand and the Reconnaissance Unit, Royal Thai Airforce, Songkhla Squadron for scientific cooperation.

References

Admiralty Tide Tables. 2003: Indian Ocean and South China Sea, Volume 3. Hydrographer to the Navy, Admiralty Hydrogr. Dep. 357 pp.

Anderson, J. A. R. 1963: The flora of the peat swamp forests of Sarawak and Brunei, including a catalogue of all recorded species of flowering plants, ferns and fern allies. Gardens Bulletin, Singapore, 20, 131-238.

Anderson, J. A. R. 1983: The tropical peat swamps of Western Malaysia. *In:* Gore, A. J. P. (ed.) Ecosystems of the world 4B, 181-199, Elsevier, New York.

Anderson, J. A. R. and Muller, J. 1975: Palynological study of a Holocene peat and a Miocene coal deposit from N.W. Borneo. Review of Palaeobotany and Palynology, 19, 291-351.

Barber, D.C., Dyke, A.; Hillaire-Marcel, C.. Jennings, A.E.; Andrews, J.T., Kerwin, M.W., Bilodeau, G., McNeely, R., Southon, J., Morehead, M.D., Gagnon, J.-M. 1999: Forcing of the cold event of 8,200 years ago by catastrophic drainage of Laurentide lakes. Nature, 400, 344-348.

Bassett, S. E., Milne, G. A., Mitrovica, J. X. and Clark, P. U. 2004: Modelling far-field sea-level data to infer the source of Meltwater Pulse IA. Earth and Planetary Science Letters, submitted.

Bosche, J. H. A. 1988: The Quaternary deposits in the coastal plains of Southern Perak, Peninsula Malaysia. Geological Survey of Malaysia, QG/1 of 1988.

Boyd, W. E. and McGrath, R. J. 2001: The geoarchaeology of prehistoric ditched sites of the upper Mae Nam Mun Valley, NE Thailand: Late Holocene vegetation history. Palaeogeography, Palaeoclimatology, Palaeoecology, 171, 307-328.

Bronk Ramsey C. 1995: Radiocarbon Calibration and Analysis of Stratigraphy: The OxCal Program. Radiocarbon, 37, 425-430.

Bronk Ramsey C. 1998: Probability and Dating. Radiocarbon, 40, 461-474.

Bunt, J. S. 1999: Overlap in mangrove species zonal patterns: some methods of analysis. Mangroves and Salt Marshes, 3, 155-164.

Cantera, J. R., Thomassin, B. A. and Arnaud, P. M. 1999: Faunal zonation and assemblages in Pacific Columbian mangroves. Hydrobiologia, 413, 17-33.

Chaimanee, N., Tiyapirach, S. and Jongkanjanasoonthorn, Y. 1985: Quaternary Geology of Satting Pra and Khao Chaison areas. Geological Survey D, M. R. Bangkok, 59 pp.

Chaimanee, N., Jongkanjanasoonthorn, Y., Kenvises, S. and Theerarungsikul, N. 1985: Quaternary Geology of Sanamchai and Pattalung area. Geological Survey D, M. R. Bangkok, 61 pp.

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

Cobb, K. M., Charles, C. D., Cheng, H. and Edwards, L. 2003: El Nino/Southern Oscillation and tropical Pacific climate during the last millennium. Nature, 424, 271-276.

Collins, N. M., Sayer, J. A. and Whitmore, T. C. 1991: The conservation atlas of tropical forests: Asia and the Pacific. London, Macmillan.

Crowley, G. M. and Gagan, M. K. 1995: Holocene evolution of coastal wetlands in wet-tropical northeastern Australia. Holocene, 5, 385-389.

Crowley, G. M., Anderson, P., Kershaw, A. P. and Grindrod, J. 1990: Palynology of a Holocene marine transgressive sequence, lower Mulgrave River valley, north-east Queensland. Australian Journal of Ecology, 15, 231-240.

Dam, R. A. C., van der Kaars, S. and Kershaw, A. P. 2001: Quaternary environmental change in the Indonesian region. Palaeogeography, Palaeoclimatology, Palaeoecology, 171, 91-95.

Daws, G. and Fujita, M. 1999: Archipelago: the Islands of Indonesia: from the Nineteenth Century Discoveries of Alfred Russell Wallace to the Fate of Forests and Reefs in the Twenty-first Century. University of California Press, Berkeley, 254 pp.

Dickens, A. F., Gelinas, Y., Masiello, C. A., Wakeham, S. and Hedges, J. I. 2004: Reburial of fossil organic carbon in marine sediments. Nature, 427, 336-339.

Dziewonski, A. M. and Anderson, D. L. 1981: Preliminary Reference Earth Model (PREM). Physics of the Earth and Planetary Interiors, 25, 297-356.

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

Gupta, A. K., Anderson, D. M. and Overpeck, J. T. 2003: Abrupt changes in the Asian southwest monsoon during the Holocene and their links to the North Atlantic Ocean, Nature, 421, 354-357

Grindrod, J. 1985: The palynology of mangroves on a prograded shore, Princess Charlotte Bay, North Queensland, Australia. Journal of Biogeography, 12, 323-348.

Grindrod, J., 1988: The palynology of Holocene mangrove and saltmarsh sediments, particularly in northern Australia. Review of Palaeobotany and Palynology, 55, 229-245.

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

Hesp, P. A., Hung, C. C., Hilton, M., Ming, C. L. and Turner, I. M. 1998: A first tentative Holocene sea-level curve for Singapore. Journal of Coastal Research, 14, 308-314.

Hope, G. 2001: Environmental change in the Late Pleistocene and later Holocene at Wanda site, Soroako, South Sulawesi, Indonesia. Palaeogeography, Palaeoclimatology, Palaeoecology, 171, 129-145.

Horton, B. P., Innes, J. B., Shennan, I. and Lloyd, J. M. 2004: A contribution to the history of coastal change in East Norfolk: palaeoenvironmental data from Somerton and Winterton Holmes. Proceedings of the Geologists' Association.

Horton, B. P., Thomson, K., Woodroffe, S. E., Whittaker, J. E., and Wright, M. W. 2005 in press: Contemporary foraminiferal distributions, Wakatobi National Park, Southeast Sulawesi, Indonesia. Journal of Foraminiferal Research.

Horton, B. P., Larcombe, P., Woodroffe, S. E., Whittaker, J. E., Wright, M. W. and Wynn, C. 2003: Contemporary foraminiferal distributions of the Great Barrier Reef coastline, Australia: implications for sea-level reconstructions. Marine Geology, 3320, 1-19.

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

Lambeck, K., Smither, C. and Johnston, P. 1998: Sea-level change, glacial rebound and mantle viscosity for northern Europe. Geophysical Journal International, 134, 102-144.

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, 1-44.

Kamaludin, H., 1989: Significance of palynology in Late Quaternary sediments in Peninsula Malaysia. Geological Society Malaysian Bulletin, 23, 199-215.

Kamaludin, H., 1993: The changing mangrove shorelines in Kuala Kurau. Peninsula Malaysia. *In:* Woodroffe, C. D. (ed.) Late Quaternary evolution of coastal and lowland riverine plains of Southeast Asia and northern Australia, 187-197, Sedimentary Geology, 83.

Kamaludin, H., 2001: Holocene sea level changes in Kelang and Kuantan, Peninsula Malaysia. PhD thesis, University of Durham, 290 pp.

Kaufmann, G. and Lambeck, K. 2002: Glacial isostatic adjustment and the radial viscosity profile from inverse modelling. Journal of Geophysical Research, 107, 2280.

Milne, G. A. 2002: Recent advances in predicting glaciation-induced sea-level changes and their impact on model applications. *In:* Mitrovica, J. X. and Vermeersen, L. L. A. (eds) Glacial isostatic adjustment and the earth system: sea level, crustal deformation, gravity and rotation, 157-176, American Geophysical Union, Washington (D. C.).

Mitrovica, J. X. and Forte, A. M. 1997: Radial profile of mantle viscosity: results from the joint inversion of convection and postglacial rebound observables. Journal of Geophysical Research, 102, 2751-2769.

Moore, P. D., Webb, J. A. and Collinson, M. E. 1991: Pollen Analysis. Blackwell, Oxford.

Morley, R. J. 1981: Development and vegetation dynamics of a lowland ombrogenous swamp in Kalimantan Tengah, Indonesia, Journal of Biogeography, 8, 383-404.

Morley, R. J. 1999: Origin and Evolution of Tropical Rain Forests. John Wiley and Sons, London, 380 pp.

Nunn, P.D., 1998: Sea-Level changes over the past 1000 years in the Pacific. J. Coastal Res. 14, 23-30.

Penny, D. 2001: A 40,000 year palynology record from north-east Thailand; implications for biogeography and palaeoenvironmental reconstruction. Palaeogeography, Palaeoclimatology, Palaeoecology, 171, 97-128.

Peltier, W.R. 2002: On eustatic sea level history: Last Glacial Maximum to Holocene. Quaternary Science Reviews, 21, 377-396.

Peltier, W.R., Shennan, I., Drummond, R. and Horton, B. P. 2002: On the post-glacial isostatic adjustment of the British Isles and the shallow visco-elastic structure of the Earth. Geophysical Journal International, 148, 443-475.

Purintavaragul, C. 1988: Seasonal appearances of aquatic macrophytes in Kukut, Songkhla Lake. Proceedings of the Workshop on Biology, Kasetsart University.

Purintavaragul, C. 1990: The seasonal appearances and some environmental effects of aquatic weeds. Proceedings of the Symposium on Managements of Aquatic Weeds, Bogor.

Sa, T. T and Hock, L. C. 1993: Impacts of sea level rise on the mangroves of Peninsula Malaysia. Malaysian Journal of Tropical Geography, 24, 57-72.

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

Semeniuk, V. 1980: Mangrove zonation along an eroding coastline in king Sound, north-western Australia. Journal of Ecology, 68, 789-812.

Shaw, J. and Ceman, J. 1999: Salt-marsh aggradation in responses to late-Holocene sea-level rise at Amherst Point, Nova Scotia, Canada. The Holocene, 9, 439-451.

Shennan, I. and Horton, B. P. 2002: Relative sea-level changes and crustal movements of the UK. Journal of Quaternary Science, 16, 511-526.

Sinsakul, S. 1990: Evidence of sea level changes in the coastal areas of Thailand: a review. Workshop on global environmental change, Chiang Mai, Thailand.

Sinsakul, S. 1992: Evidence of quaternary sea level changes in the coastal areas of Thailand: a review. Journal of Southeast Asian Earth Science, 7, 23-37.

Somboon, J. R. P. 1988: Palaeontological study of the recent marine sediments in the lower central plain, Thailand. Journal of Southeast Asian Earth Science, 2, 201-210.

Somboon, J. R. P and Thiramongkol, N. 1992: Holocene highstand shoreline of the Chao Phyraya delta, Thailand. Journal of Southeast Asian Earth Science, 7, 53-60.

Southon, J., Kashgarian, M., Fontugne, M., Metivier, B. and Yim, W. W-S. 2002: Marine reservoir corrections for the Indian Ocean and Southeast Asia. Radiocarbon, 44,167-180.

Stargardt, J. M. 1983: Satingpra I, the environmental and economic archaeology of south Thailand, Oxford & Singapore, British Archaeological Reports [BAR] & Institute of Southeast Asian Studies [ISEAS], 332 pp.

Stargardt, J. M. 1998: Earth Rice and water: reading landscape as record on the early history of the Satingpra Peninsula, south Thailand, 4th -14th century. *In:* Grove, R. (ed.) Nature and the Orient; the environmental history of South and South East Asia, 121-183, Oxford and Delhi, Oxford University Press.

Stargardt, J. M. 2001: Behind the Shadows: archaeological data on two-way sea-trade between Quanzhou and Satingpra, South Thailand, 10th-14th century. *In:* Schottenhammer, A. (ed.) In Emporium to the World: Quanzhou and Maritime Trade, 10th-14th century, 309-393, Leiden, E. J. Brill.

Stargardt, J. M. and Purintavaragul, C. 1999: 'Three forest types and long-term environmental change on the Kra Ecotone, South Thailand. Royal Society South East Asian Rainforest Research Programme Newsletter.

Street-Perrott, F. A. and Perrott, R. A., 1990: Abrupt climate fluctuations in the tropics: the influence of the Atlantic Ocean circulation. Nature, 343, 607-612.

Streif, H. 1979: Holocene sea-level changes in the Straits of Malacca. Proceedings of the 1978 International Symposium on coastal evolution in the Quaternary, Sao Paolo, Brazil, 552-572.

Stuiver M. and Braziunas, T.F. 1993: 14C Ages of Marine Samples to 10,000 BC. Radiocarbon, 35, 137-189.

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

Thiramongkol, 1984: Geomorphology of the lower Central Plain Thailand: Proceedings of the 3rd meeting of the working group of river and coastal plain, 103-117.

Tjia, H. D., Fujii, S. and Kigoshi, K. 1983: Holocene shorelines of Tioman Island in the South China Sea. Geologie en Mijnbouw, 62, 599-604.

Tjia, H.D., 1996: Sea-level changes in the tectonically stable Malay-Thai Peninsula. Quaternary International, 31, 95-101.

Tiyapunte, S. and Theerarungsikul, N. 1988: The geology of map sheet Amphoe Nongjik and Changwat Pattani. Geological Survey Report D.M.R., 71 pp.

Tushingham, A. M. and Peltier, W. R. 1991: ICE-3G: A new global model of late Pleistocene deglaciation based on geophysical predictions of postglacial relative sea level change. Journal of Geophysical Research, 96, 4497-4523.

Van de Plassche, O. 1986: Sea-level research: a manual for the collection and evaluation of data. Geobooks, Norwich, 617 pp.

Van der Kaars, S. 1991: Palynology of eastern Indonesian marine piston-corres: a Late Quaternary vegeational and climatic record for Australia. Palaeogeography, Palaeoclimatology, Palaeoecology, 85, 239-302.

Van der Kaars, S. 2001: Pollen distribution in marine sediments from the south-eastern Indonesian waters. Palaeogeography, Palaeoclimatology, Palaeoecology, 171, 341-361.

Van der Kaars, S., Wang, X., Kershaw, P., Guichard, F., Arifin Setiabudi, D. 2000: A Late Quaternary palaeoecology record from the Banda Sea, Indonesia: patterns of vegetation, climate and biomass burning in Indonesia and northern Australia. Palaeogeography, Palaeoclimatology, Palaeoecology, 155, 135-153.

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

Wallace, A.R., [1869] 1962: The Malay Archipelago: The Land of the Orang-utan and the Bird of Paradise. A Narrative of Travel with Studies of Man and Nature: Macmillian, London. Reprint. Dover Publications, New York, 638 pp.

White, J. C., Penny, D., Kealhofer, L. and Maloney, B. 2004: Vegetation changes from the late Pleistocene through the Holocene from three areas of archaeological significance in Thailand. Quaternary International, 113, 111-132.

Whitmore, T. C. 1975: Tropical Rainforests of the Far East (with a chapter on solis by C. P. Burnham). Clarendon Press, Oxford, 282 pp.

Whitmore, T. C. 1994: Biogeographical Evolution of the Malay Archipelago, (Oxford Monographs on Biogeography, No 4). Clarendon Press, Oxford, 147 pp.

Whitmore, T. C. 1998: An introduction to tropical rain forests. Oxford University Press, Oxford, 226 pp.

Willard, D. A., Cronin, T. M. and Verardo, S. 2003: Late Holocene climate and ecosystem history form Chesapeake Bay sediment cores, USA. The Holocene, 13, 201-214.

Woodroffe, C. D. 1993: Late Quaternary evolution of coastal and lowland riverine plains of Southeast Asia and northern Australia. Sedimentary Geology, 83, 163-175.

Woodroffe, C. D., Thom, B. G. and Chappell, J. 1985: Development of widespread mangrove swamps in mid-Holocene times in northern Australia. Nature, 317, 711-713.

Woodroffe, S. A. and Horton, B. P. 2005 in press: Late and post glacial sea-level changes of the Indo-Pacific: a review. Journal of Asian Earth Science.

Woodroffe, S. A., Horton B. P. and Larcombe, P. in press: Contemporary intertidal foraminifera distributions of Cleveland Bay, Central Great Barrier Reef shelf, Australia: implications for sea-level reconstructions. Journal of Foraminiferal Research.

Yim, W. W-S. 1999: Radiocarbon dating and the reconstruction of late Quaternary sea-level changes in Hong Kong. Quaternary International, 55, 77-91

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, 713-716.

Zong, Y., 2004: Mid-Holocene sea-level highstand along the Southeast Coast of China. Quaternary International, 117, 55-67.

Tables

Highest Astronomical Tides (m above MSL)	Mean High High Water (m above MSL)	Mean Low High Water (m above MSL)	Mean High Low Water (m above MSL)	Mean Low Low Water (m above MSL)	Lowest Astronomical Tides (m above MSL)
1.9	0.90	0.10	-0.20	-1.00	-2.00

Table 1 Tide levels for Songkhla station, Thailand. Chart datum = 1.90 m below MSL (Admiralty Tide Tables, 2003)

Dated Material	Indicative Range	Reference Water Level
Mangrove peat directly above or below clastic marine deposit	MHHW - MHLW	(MHHW + MHLW)/2
Mangrove peat directly above or below freshwater deposit	HAT- MHHW	(HAT + MHHW)/2

Table 2 Indicative range and reference water level for commonly dated materials. The indicative range (given as a maximum) is the most probable vertical range in which the sample occurs. The reference water level is given as a mathematical expression of tidal parameters.

	Depth (m)	Altitude (m MSL)	Laboratory code	14C age ± 1σ	δ13C	Calibrated age yrs BP	Reference water level	RSL (m MSL)
Thale Noi, TN-3	2.96	- 2.56	SRR-6574	6980 ± 50	-29.9	7880-7680	(MHHW + HAT)/2	-3.96 ± 0.58
Thale Noi, TN-3	6.67	- 5.27	SRR-6575	7535 ± 55	-28.2	8420-8190	(MHHW + HAT)/2	-6.67 ± 0.57

Table 3 Summary of sea-level index points and limiting data from the Thale Noi, Core TN-3. All radiocarbon assays are based on peats or organic muds as indicated in the figures. All assays are calibrated using OxCal Program, Version 3.8 (Bronk Ramsey, 1995, 1998), using the 95% confidence limits for the probability option. The reference water level is given as a mathematical expression of tidal parameters.

Appendix

Location	Site	Latitude	Longitude	14C age ± 1σ	Calibrated age yrs BP.	RSL (m MSL)	Reference
Index							
A	Chao Phraya Delta	13°30'N	100°30'E	2250±110	2750-1900	1.35±0.50	Somboon & Thiramongkol, 1992
A	Chao Phraya Delta	13°30'N	100°30'E	2250±110	2750-1900	1.60±0.50	Sinsakul, 1992
A	Chao Phraya Delta	13°30'N	100°30'E	3670±125	4450-3650	2.55±0.50	Somboon & Thiramongkol, 1992
A	Chao Phraya Delta	13°45'N	100°15'E	5670±135	6800-6150	0.85±0.50	Somboon & Thiramongkol, 1992
A	Chao Phraya Delta	14°25'N	100°30'E	5830±310	7450-5950	1.65±1.10	Somboon & Thiramongkol, 1992
A	Chao Phraya Delta	13°30'N	101°30'E	6270±280	7750-6450	-0.82±0.67	Somboon & Thiramongkol, 1992
A	Chao Phraya Delta	14°00'N	100°30'E	6490±135	7700-7000	-1.15±0.50	Somboon & Thiramongkol, 1992
A	Chao Phraya Delta	13°30'N	100°30'E	6490±135	8050-7500	-0.90±0.50	Sinsakul, 1992
A	Chao Phraya Delta	13°45'N	100°00'E	6560±140	7700-7150	2.35±0.50	Somboon & Thiramongkol, 1992
A	Chao Phraya Delta	13°30'N	100°00'E	6630±190	7950-7050	-0.40±1.35	Somboon & Thiramongkol, 1992
A	Chao Phraya Delta	13°55'N	100°30'E	6680±140	7800-7250	-0.55±0.50	Somboon & Thiramongkol, 1992
A	Chao Phraya Delta	13°30'N	100°30'E	6680±140	7800-7250	-0.30±0.50	Sinsakul, 1992

A	Chao Phraya Delta	14˚00'N	101˚00'E	6760±145	7950-7300	0.55±0.50	Somboon & Thiramongkol, 1992
A	Chao Phraya Delta	13˚30'N	100˚30'E	6760±140	7950-7300	0.80±0.50	Sinsakul, 1992
A	Chao Phraya Delta	14˚00'N	100˚45'E	6830±140	7950-7400	0.85±0.50	Somboon & Thiramongkol, 1992
A	Chao Phraya Delta	13˚30'N	100˚30'E	6830±140	7950-7300	1.10±0.50	Sinsakul, 1992
A	Chao Phraya Delta	14˚20'N	100˚00'E	7060±220	8350-7500	-1.90±0.65	Somboon & Thiramongkol, 1992
A	Chao Phraya Delta	14˚20'N	101˚00'E	7190±150	8350-7650	-3.80±0.75	Somboon & Thiramongkol, 1992
A	Chao Phraya Delta	14˚20'N	100˚00'E	7300±180	8450-7700	-1.35±0.70	Somboon & Thiramongkol, 1992
A	Chao Phraya Delta	13˚30'N	101˚00'E	7440±150	8550-7900	-0.65±0.50	Somboon & Thiramongkol, 1992
A	Chao Phraya Delta	13˚30'N	100˚30'E	7500±35	8390-8190	-8.15±0.50	Somboon & Thiramongkol, 1992
A	Chao Phraya Delta	13˚30'N	100˚30'E	7500±35	8390-8190	-7.90±0.50	Sinsakul, 1992
A	Chao Phraya Delta	14˚20'N	101˚10'E	7630±170	9000-8000	-3.95±0.70	Somboon & Thiramongkol, 1992
A	Chao Phraya Delta	14˚00'N	100˚30'E	7800±40	8650-8430	-10.15±0.50	Somboon & Thiramongkol, 1992
B	Prachuab Kirikhan[#]	11˚45'N	99˚45'E	3290±125	3200-2400	1.50±0.70	Thiramongkol, 1986
B	Prachuab Kirikhan[#]	11˚45'N	99˚45'E	4180±160	4400-3400	2.00±0.70	Sinsakul, 1990
C	Phuket	7˚45'N	98˚25'E	2210±57	2350-2060	1.34±0.50	Scoffin & Le Tissier, 1998
C	Phuket	7˚45'N	98˚25'E	2425±57	2720-2340	1.45±0.50	Scoffin & Le Tissier, 1998
C	Phuket	7˚45'N	98˚25'E	3230±57	3590-3350	1.61±0.50	Scoffin & Le Tissier, 1998

C	Phuket	7˚45'N	98˚25'E	3345±61	3720-3440	1.47±0.50	Scoffin & Le Tissier, 1998
C	Phuket	7˚45'N	98˚25'E	4100±61	4830-4500	1.75±0.50	Scoffin & Le Tissier, 1998
C	Phuket	7˚45'N	98˚25'E	4200±57	4860-4570	1.82±0.50	Scoffin & Le Tissier, 1998
C	Phuket	7˚45'N	98˚25'E	4235±57	4880-4570	1.78±0.50	Scoffin & Le Tissier, 1998
C	Phuket	7˚45'N	98˚25'E	4830±57	5670-5450	1.82±0.50	Scoffin & Le Tissier, 1998
C	Phuket	7˚45'N	98˚25'E	5260±57	6180-5910	1.88±0.50	Scoffin & Le Tissier, 1998
C	Phuket	7˚45'N	98˚25'E	5880±57	6810-6540	1.95±0.50	Scoffin & Le Tissier, 1998
C	Phuket	7˚45'N	98˚25'E	5945±57	6910-6640	1.97±0.50	Scoffin & Le Tissier, 1998
D	Pattalung	7˚35'N	100˚10'E	6900±180	8150-7400	-4.70±0.50	Chaimanee *et al.,* 1985
D	Pattalung	7˚35'N	100˚10'E	7720±180	9050-8150	-4.30±0.50	Chaimanee *et al.,* 1985
D	Pattani	6˚50'N	100˚15'E	2920±280	3850-2350	1.10±0.50	Tiyapunte & Theerarungsikul, -1988
D	Satting Pra	7˚30'N	100˚10'E	6300±140	7500-6800	-1.40±0.50	Chaimanee *et al.,* 1984
D	Satting Pra	7˚30'N	100˚10'E	6600±150	7750-7200	-0.70±0.50	Chaimanee *et al.,* 1984
D	Thale Noi, TN3	7˚45'N	100˚10'E	6980±50	7880-7680	-3.96±0.58	This paper
D	Thale Noi, TN3	7˚45'N	100˚10'E	7535±50	8420-8190	-6.67±0.57	This paper
E	Kelang	03˚00'N	101˚15'E	4045±49	4650-4410	2.90±0.30	Kamaludin, 2001
E	Kelang	03˚00'N	101˚15'E	4073±86	4850-4300	2.71±0.20	Kamaludin, 2001

E	Kelang	03˚00'N	101˚15'E	5270±47	6180-5920	3.08±0.20	Kamaludin, 2001
E	Kelang	03˚00'N	101˚15'E	5331±46	6210-5990	3.44±0.10	Kamaludin, 2001
E	Kelang	03˚00'N	101˚15'E	5349±65	6290-5990	3.09±0.30	Kamaludin, 2001
E	Kelang	03˚00'N	101˚15'E	5556±47	6450-6280	3.43±0.20	Kamaludin, 2001
F	Straits of Malacca	2˚31'N	101˚46'E	1055±85	1180-760	-0.70±0.55	Geyh *et al.,* 1979
F	Straits of Malacca	2˚25'N	101˚58'E	1145±80	1270-920	-0.61±0.51	Geyh *et al.,* 1979
F	Straits of Malacca	1˚37'N	103˚25'E	4040±100	4850-4250	4.87±0.57	Geyh *et al.,* 1979
F	Straits of Malacca	2˚1'N	102˚47'E	4135±90	4850-4420	3.97±0.57	Geyh *et al.,* 1979
F	Straits of Malacca	1˚48'N	103˚15'E	4275±70	4990-4570	3.83±0.68	Geyh *et al.,* 1979
F	Straits of Malacca	2˚19'N	102˚5'E	5975±75	7010-6630	0.18±0.58	Geyh *et al.,* 1979
F	Straits of Malacca	1˚32'N	103˚28'E	6260±75	7330-6940	-0.35±0.52	Geyh *et al.,* 1979
F	Straits of Malacca	1˚32'N	103˚28'E	6610±70	7620-7410	-0.13±0.58	Geyh *et al.,* 1979
F	Straits of Malacca	1˚37'N	103˚25'E	6945±120	7980-7570	-1.43±0.58	Geyh *et al.,* 1979
F	Straits of Malacca	2˚22'N	102˚0'E	6985±180	8200-7500	-3.75±0.55	Geyh *et al.,* 1979
F	Straits of Malacca	2˚10'N	102˚36'E	7015±80	7970-7670	-3.73±0.67	Geyh *et al.,* 1979
F	Straits of Malacca	2˚25'N	101˚56'E	7175±70	8170-7840	-6.05±0.55	Geyh *et al.,* 1979
F	Straits of Malacca	1˚59'N	102˚40'E	7340±100	8360-7960	-6.05±0.55	Geyh *et al.,* 1979

F	Straits of Malacca	2˚25'N	101˚56'E	7440±175	8170-7840	-6.83±0.53	Geyh *et al.*, 1979
F	Straits of Malacca	2˚25'N	101˚56'E	7560±135	8360-7960	-7.58±0.53	Geyh *et al.*, 1979
F	Straits of Malacca	2˚12'N	102˚18'E	7985±220	9550-8350	-13.75±0.60	Geyh *et al.*, 1979
F	Straits of Malacca	1˚51'N	102˚9'E	8490±85	9700-9250	-22.15±0.55	Geyh *et al.*, 1979
F	Straits of Malacca#	1˚20'N	103˚45'E	7190±70	7580-7120	-5.60±0.65	Hesp *et al.*, 1998
F	Straits of Malacca#	1˚20'N	103˚45'E	7310±80	7700-7210	-8.80±0.65	Hesp *et al.*, 1998
F	Straits of Malacca	1˚20'N	103˚45'E	7690±50	8060-7580	-9.60±0.65	Hesp *et al.*, 1998
F	Straits of Malacca	1˚20'N	103˚45'E	7790±60	8850-8400	-10.45±0.95	Hesp *et al.*, 1998
F	Straits of Malacca	1˚20'N	103˚45'E	6870±100	7350-6750	-3.81±0.90	Hesp *et al.*, 1998
F	Straits of Malacca	1˚20'N	103˚45'E	7060±150	7190-6900	-4.54±0.65	Hesp *et al.*, 1998
F	Straits of Malacca	1˚20'N	103˚45'E	5720±60	6670-6400	1.11±0.85	Hesp *et al.*, 1998
F	Straits of Malacca	1˚20'N	103˚45'E	3530±80	4000-3630	2.16±0.85	Hesp *et al.*, 1998
F	Straits of Malacca	1˚20'N	103˚45'E	5870±70	6810-6490	1.36±0.85	Hesp *et al.*, 1998
G	Kuantan	03˚00'N	103˚20'E	3967±43	4530-4280	1.24±0.10	Kamaludin, 2001
G	Tioman Island#	02˚45'N	104˚15'E	1900±90	1450-850	1.10±0.70	Tjia *et al.*, 1983
G	Tioman Island#	02˚45'N	104˚15'E	2370±120	2050-1300	1.60±0.70	Tjia *et al.*, 1983
G	Tioman Island#	02˚45'N	104˚15'E	3160±170	3150-2150	0.90±0.70	Tjia *et al.*, 1983

G	Tioman Island[#]	02˚45'N	104˚15'E	4160±150	4400-3400	1.70±0.70	Tjia *et al.,* 1983
Limiting							
A	Chao Phraya Delta	13˚30'N	100˚30'E	7440±150	8550-7900	-0.40±1.00	Somboon, 1988
A	Chao Phraya Delta	13˚30'N	100˚30'E	6560±140	7700-7150	2.60±1.00	Sinsakul, 1992
A	Chao Phraya Delta	14˚00'N	101˚00'E	7300±40	8180-8010	6.85±1.00	Somboon & Thiramongkol, 1992
A	Chao Phraya Delta	13˚45'N	101˚30'E	7310±110	8360-7930	6.70±2.50	Somboon & Thiramongkol, 1992
A	Chao Phraya Delta	13˚55'N	101˚30'E	7570±170	8850-7950	5.20±1.65	Somboon & Thiramongkol, 1992
A	Chao Phraya Delta	13˚55'N	101˚35'E	6010±320	7650-6150	6.55±1.80	Somboon & Thiramongkol, 1992
A	Chao Phraya Delta	14˚20'N	100˚00'E	7600±190	9000-7950	5.23±1.35	Somboon & Thiramongkol, 1992
A	Chao Phraya Delta	14˚00'N	100˚00'E	7310±180	8450-7750	0.00±1.65	Somboon & Thiramongkol, 1992
D	Thale Noi, TN3	7˚45'N	100˚10'E	2435±50	2720-2350	2.37±0.90	This paper
D	Pattani	6˚50'N	105˚15'E	6630±130	7700-7260	5.30±1.00	Tiyapunte & Theerarungsikul, 1988
D	Pattani	6˚50'N	105˚15'E	7730±120	9000-8200	1.70±1.00	Tiyapunte & Theerarungsikul, 1988
D	Pattani	6˚50'N	105˚15'E	6010±200	7350-6400	3.10±1.00	Tiyapunte & Theerarungsikul, 1988

Summary of sea-level index points and limiting data from Malay-Thai Peninsula. Locations A to G are shown in figure 5 and 6. All assays are calibrated using OxCal Program, Version 3.8 (Bronk Ramsey, 1995, 1998), using the 95% confidence limits for the probability option. All radiocarbon assays are based on peats, wood or organic muds except those indicated by #. These assays are based on shells and have been calibrated using the calibration curve modelled for the oceans (Stuiver *et al.,* 1998) offset for local variations using Delta_R corrections (Stuiver and Braziunas, 1993; Southon *et al*., 2002).

Figures

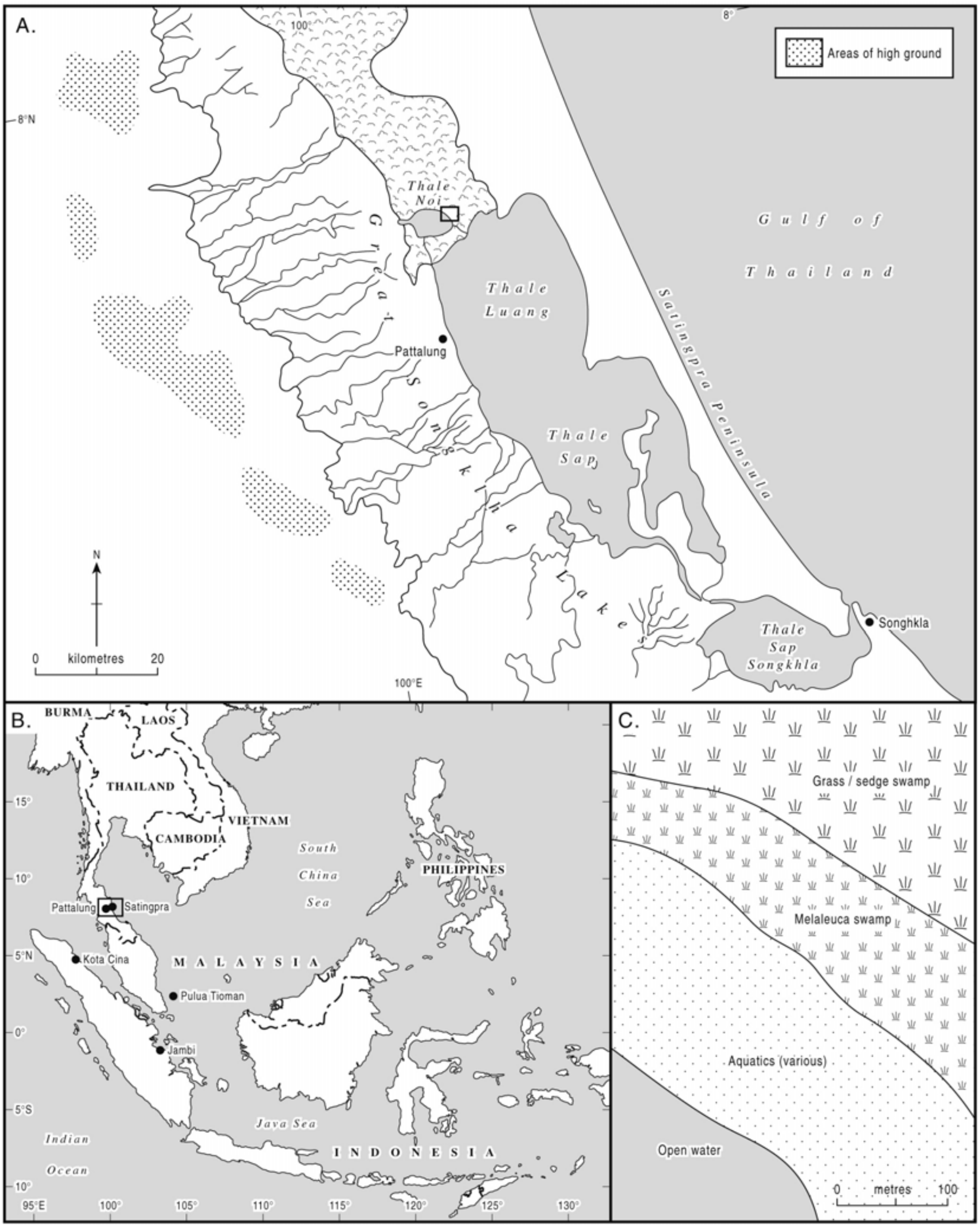

Figure 1 Site location map of (A) Great Songkhla Lakes of Songkhla-Pattalung, (B) the Malay-Thai Peninsula, Southeast Asia. (C) The borehole locations and (D) the stratigraphy of Thale Noi are shown.

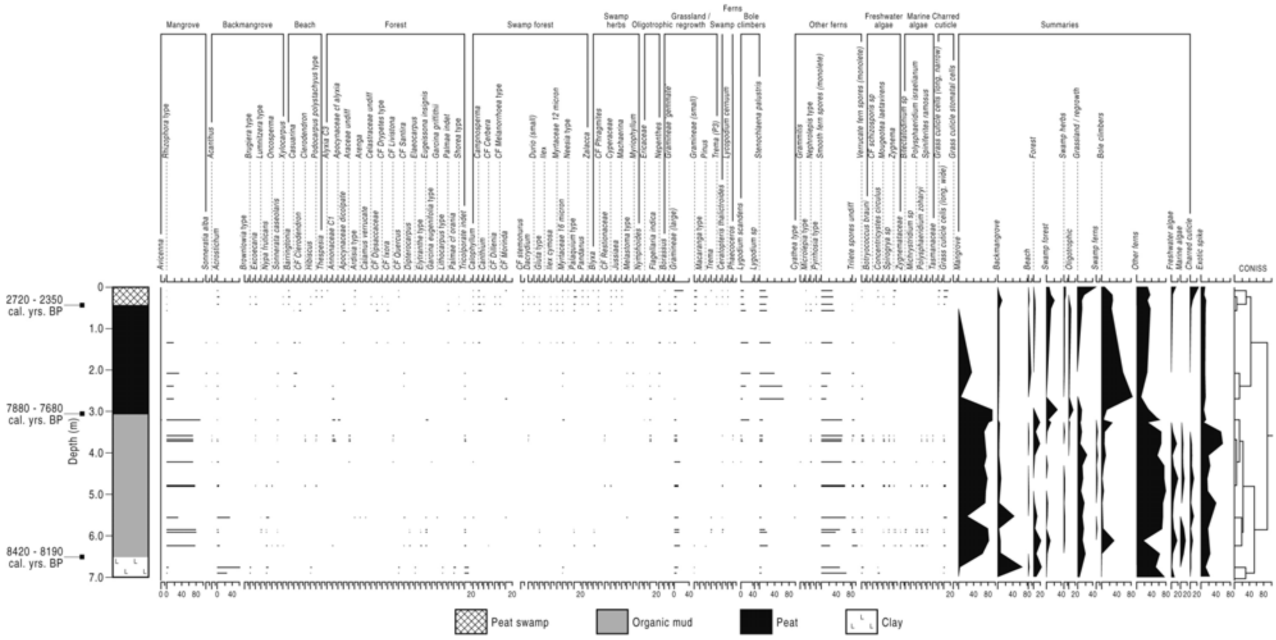

Figure 2 Thale Noi Core 3 (TN-3) microfossil diagram. Total freshwater-derived pollen and spores has been used as the main pollen sum, with mangrove pollen/spores being presented in terms of total freshwater-derived and total mangrove pollen/spores; freshwater algae being presented in terms of total freshwater-derived pollen and spores and total freshwater algae. Calibrated radiocarbon ages and depth (m) down-core shown to the left of the lithology column.

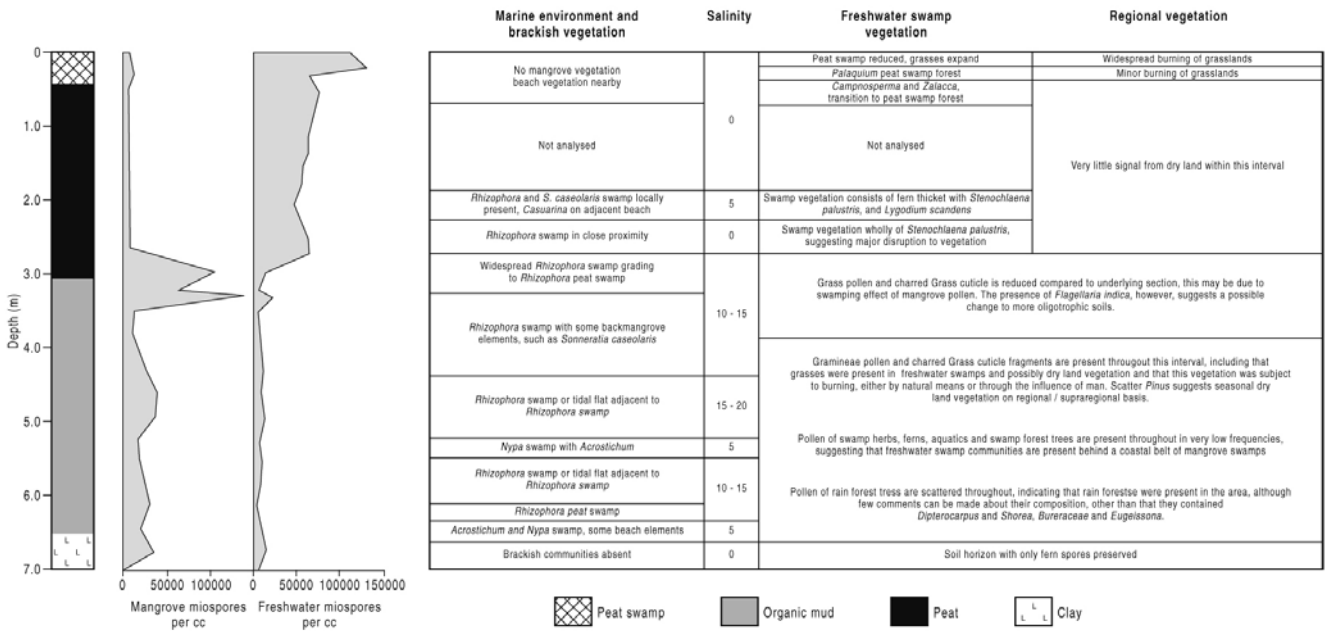

Figure 3 Thale Noi Core 3 (TN-3) palaeoenvironmental interpretation based on sedimentological and palynological data. Palynomorph concentrations are shown.

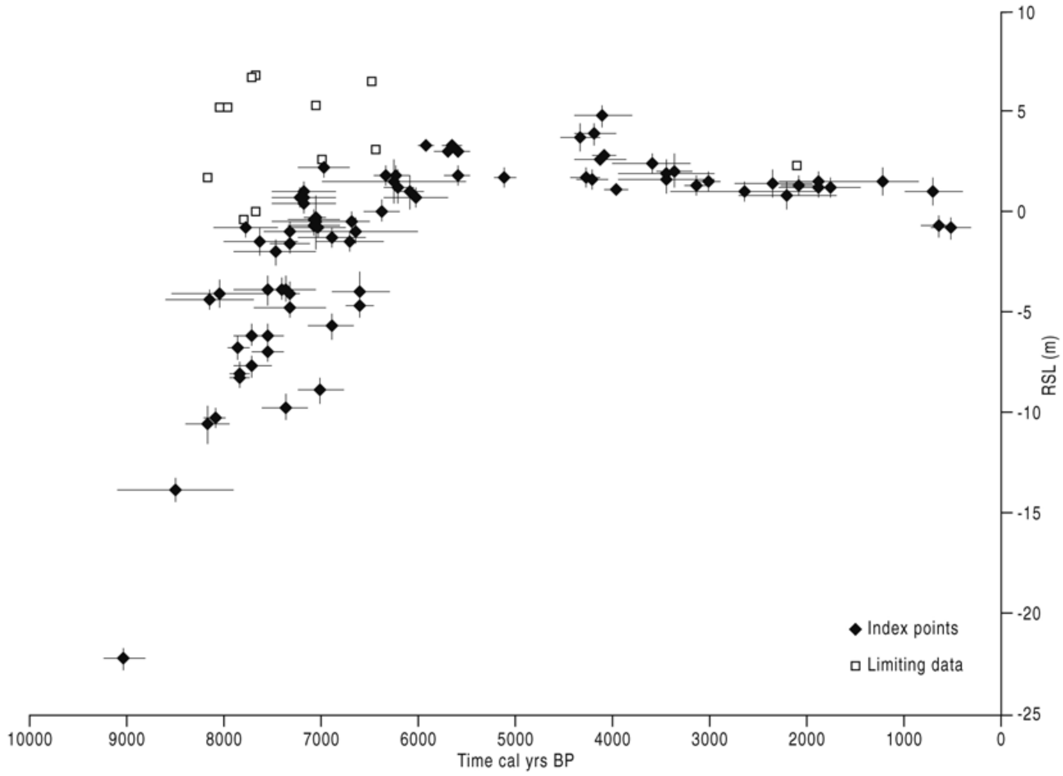

Figure 4 Malay-Thai Peninsula relative sea-level index points

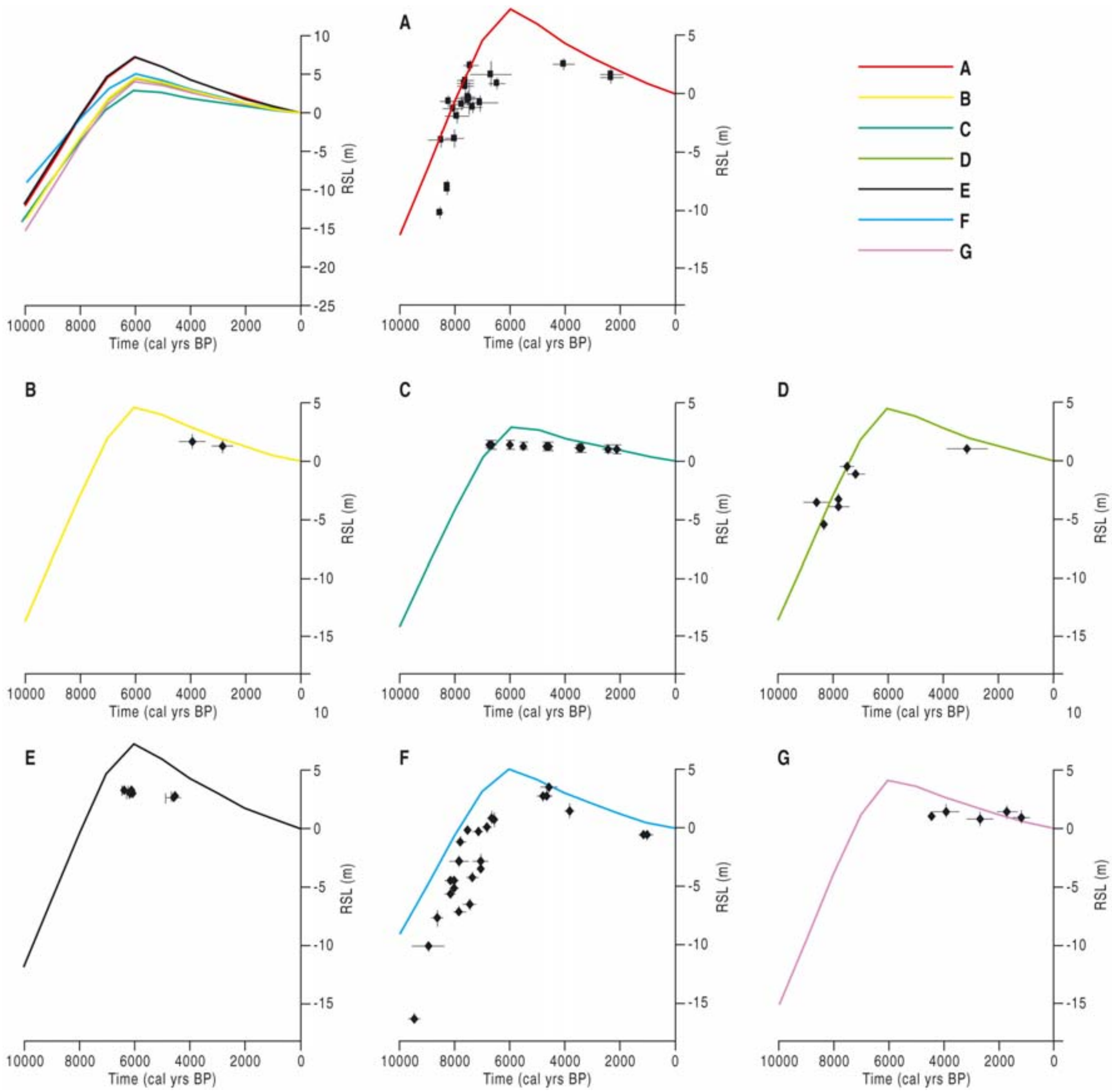

Figure 5 Geophysical model predictions and observations for regions A-G of Malay-Thai Peninsula. Top left-hand graph shows sea-level predictions at the seven locations (A-G) indicated in Figure 6. The remaining graphs show sea-level predictions (solid lines) and observations for each location (A-G). Details of the sea-level model are provided in the main text.

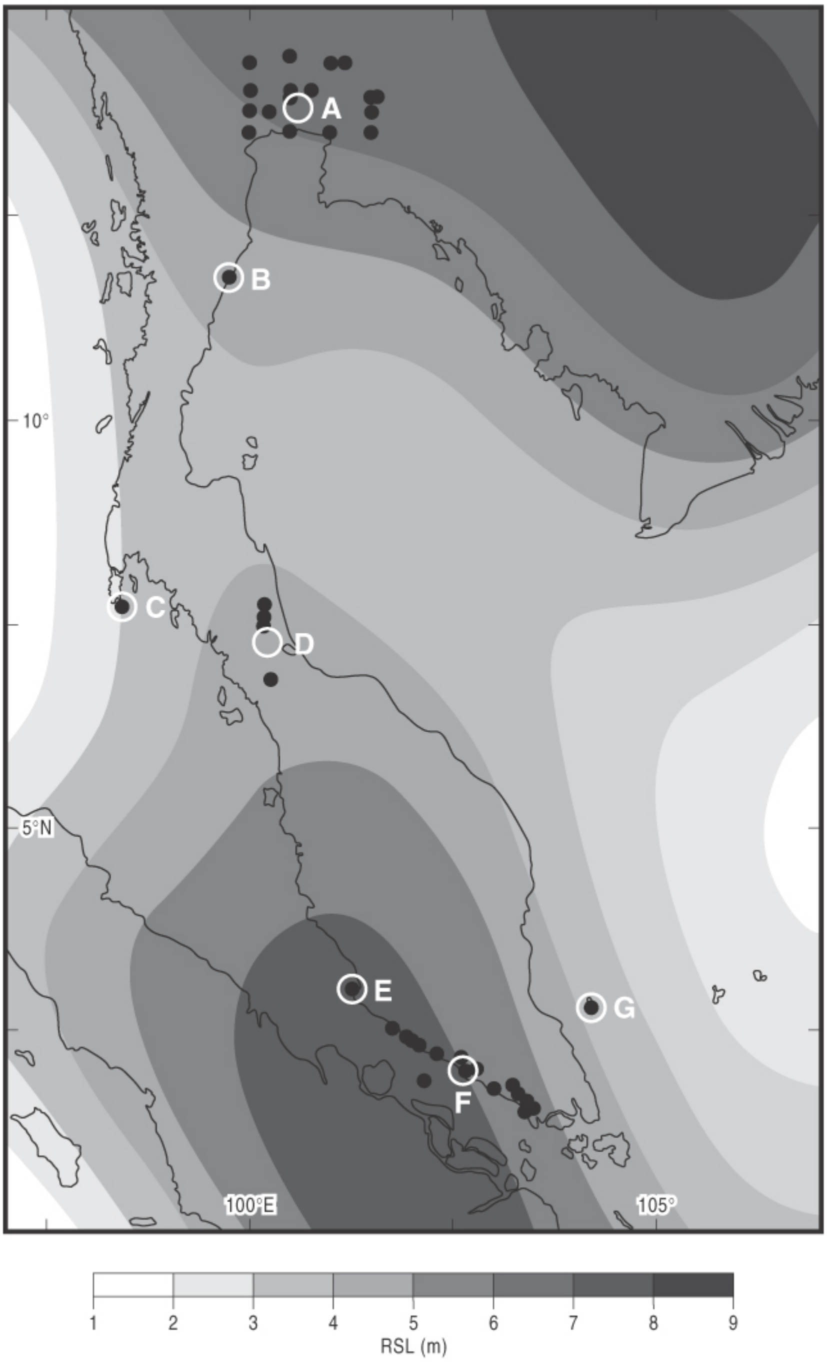

Figure 6 Map showing geophysical model predictions of mean sea-level for Malay-Thai Peninsula at 6000 cal yrs BP. The black dots illustrate the locations of individual cores used to reconstruct ancient sea levels. The white circles (labelled A-G) show the site positions adopted in generating the sea-level

predictions shown in Figure 5. These positions were calculated by taking the average location of the cores clustered around a given area. The predictions show a relatively large differential signal of c. 7 m across the region at this time. This variation is driven, mainly, by hydro-isostasy which results in the larger land masses being tilted upwards due to the post Last Glacial Maximum sea-level rise and consequent water loading of the adjacent ocean areas.