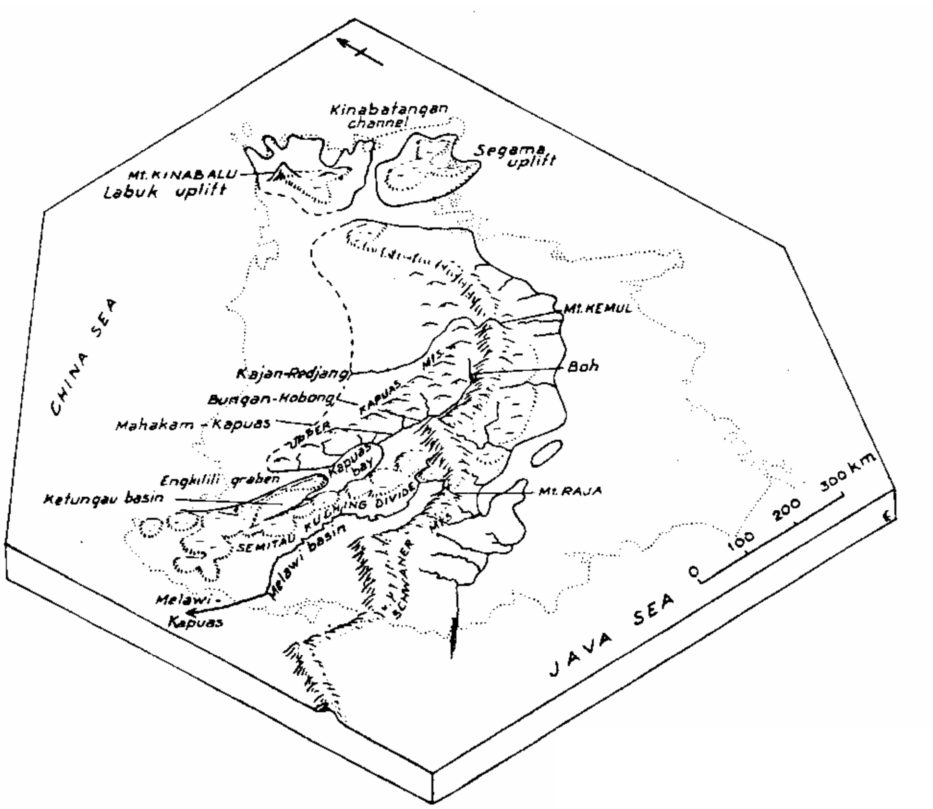
CHAPTER 3. THE PALAEOENVIRONMENTAL SCENE BASED ON DATA MINING

He thought the equator was unlucky-
its path lay mostly across oceans; it cut across two continents, true,
but it had no luck with Asia, which had pulled up out of the way.
Moreover it pressed down and squashed what it did manage to cross-
the tips of one or two things and some untidy islands.

The Defense, Vladimir Nabokov



Palaeogeographical reconstruction of Borneo in the lower Neogene by Smit-Sibinga (1953a)

3.1 GENERAL GEOLOGY AND TECTONIC FRAMEWORK

The region under consideration in this research is a tectonically very active area. Three converging lithospheric plates interact here: the India-Australian, Eurasian, and Pacific plates (Hall 1996). The present plate tectonic setting of Southeast Asia is dominated by active subduction zones and associated volcanic arc-trench systems that mark the western, southern and eastern margins of the region (Madon & Watts 1998). Figure 3.1 shows the main trenches, seas, Sunda Shelf, and uplands in SE Asia.

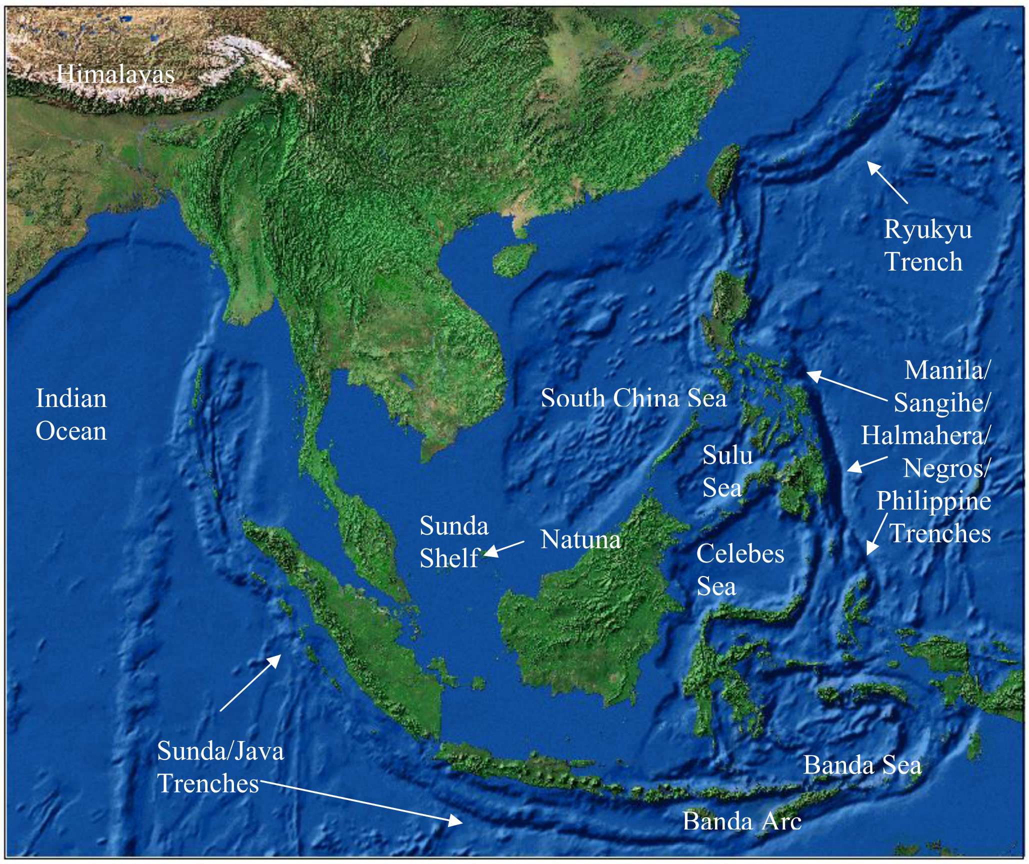


Figure 3.1. Worldsat Color Shaded Relief Image of SE Asia (from ESRI Data and Maps).

In SE Asia, during the Early Oligocene a pattern of rifting and subsidence, which began in the Late Eocene, continued with the opening of pull-apart basin in the regions of the South China Sea, Sumatra, and the West Java Sea. Most of the rifts contained large, often deep, freshwater lakes, which gradually filled with organic-rich muds or with fluvial sands during times of low lake level. By the beginning of the Early Miocene most of these rifts were filled, but subsidence continued over a wider area, further reducing the land area. At the beginning of the Miocene, a huge part of the Sunda region, from Vietnam to Natuna became submerged by a very shallow brackish-water sea. Further widespread transgressions occurred during the Early Miocene and earliest Middle Miocene, at which time the land area for SE Asia was at its minimum for the Tertiary (Morley 2000). This general overview shows that land forms in island SE Asia changed considerably during the Tertiary because of shifting coastlines.

Not only coastlines were shifting, but also global climates changed during the Tertiary. In SE Asia, these climate changes were strongly influences by the rising of the Himalayas and the Qinghai-Tibetan Plateau, which was the result of the collision between the Indian and Eurasian plates. By the Middle Miocene the Himalayas were probably over 3,000 m and the Qinghai-Tibetan Plateau more than 1,000 m a.s.l., although Zheng and Rutter (1998) report a somewhat lower elevation (500–1,000 m) for the latter at the start of the Pliocene. At the start of the Pleistocene, the Himalayan, Pamir, and Kunlun Mountains had attained an altitude of 4,600–6,350 m. Most authors agree that that most of the final uplift (2,000–3,000 m) took place during the Middle and Late Pleistocene (Ferguson 1993). The development of this massive mountainous area had a considerable influence on the climate in Asia.

In the following sections I will discuss the palaeogeographical and palaeoenvironmental changes that have taken place during the Late Tertiary and Quaternary.

3.2 PALAEOCLIMATE

General climate patterns of SE Asia

The present-day climate patterns in Sundaland and Wallacea are strongly influenced by their proximity to the equator. Here, barometric pressures are low and vary little over great distance. This causes sea winds to be light and evaporated water to rise almost vertically. The cooling of this mass of humid air causes saturation and abundant rains, which in freeing energy carries the mass of air to higher altitudes before it finally moves off to the north and south. This happens at the location of the Intertropical Convergence Zone (ITCZ). There it hits the jet streams before descending at about 30 ºN and S. When it reaches the surface of the earth, reheating pushes it towards the equator, thereby creating the trade winds; this system is called the Hadley Cell. Throughout the Late Mesozoic and Cenozoic these cells have remained approximately in the same place, between latitudes 10 and 40º N and S (Morley 2000).

During the northern hemisphere winter the region is dominated by NE and NW winds blowing from central Asia, which carry moisture-laden air towards to the ITCZ. At the ITCZ the air rises, resulting in heavy rains in (eastern) Indonesia and Australia (Wang et al. 1999). During the northern summer, two airstreams are important: SW winds from Africa and dry SE winds from Australia (Lockwood, 1976 in Maloney 1996) (see Figure 3.2a and b). All monsoon airstreams pick up moisture as they cross warm seas but the main periods of rainfall are associated with the movement of the Intertropical Convergence Zone during the time of advancing and retreating

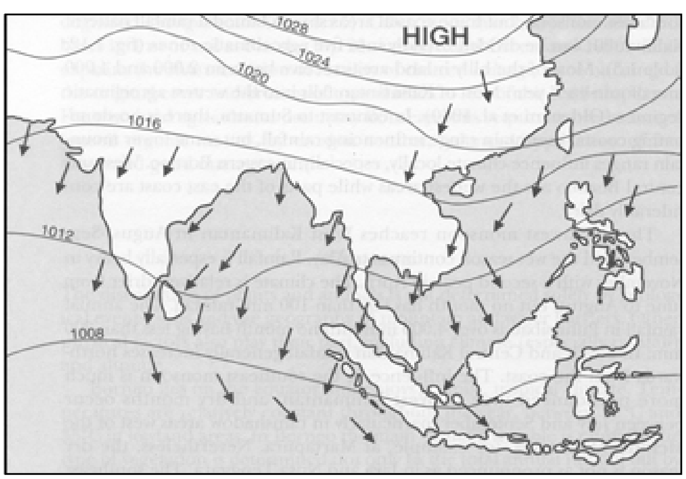


Figure 3.2a. Isobars and wind direction over Asia in the wet monsoon (November–April) (after MacKinnon et al. 1996).

monsoons. For eastern Indonesia this means that the relatively dry and cool southern monsoon results in a dry season from (May-) July to October (-November) (Wang et al. 1999). Mountainous areas in Sundaland, like the Sumatran Barisan range, can provide significant shelter from the force of these monsoons, and influence rainfall patterns (Maloney 1996). In Sumatra, this has lead to a much wetter western part of the island, compared to the drier and more seasonal east (Whitten et al. 2000). Similarly, west Borneo receives between 2,000 and 4,000 of rain each year, whereas the coastal areas of East Kalimantan and eastern Sabah are considerable drier (MacKinnon et al. 1996).

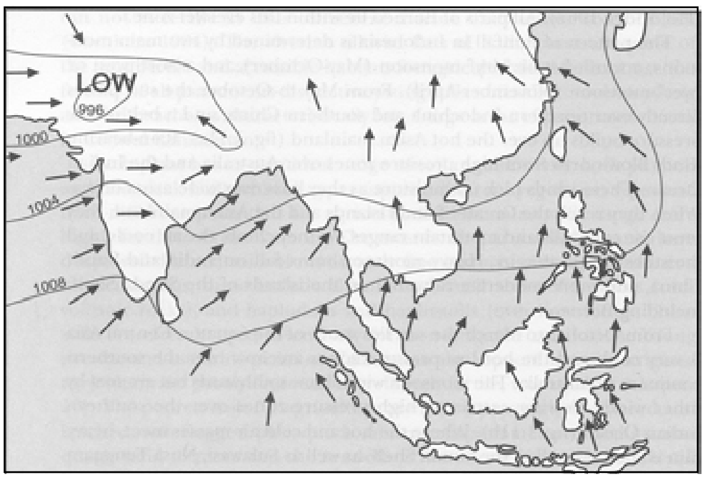


Figure 3.2b. Isobars and wind direction over Asia in the dry monsoon (May–October) (after MacKinnon et al. 1996).

These regional climate patterns have been influenced by changes in the global climate. Global climate variations appear to have been orchestrated by regular changes in the geometrical disposition of the Sun and the Earth, especially during the Quaternary. This theory of global climate change was developed by a Scotsman, James Croll, over 100 years ago and subsequently elaborated by Milankovitch, who developed, what became popularly known as the Milankovitch Theory (Lowe 2001). This theory suggests that the Earth's orbit varies over a cycle of ca. 100 Kya, the tilt of the Earth's axis varies over a cycle of 41 Kya, and the rotational movement of the Earth around its axis ("the Earth's wobble") varies over cycles with durations of 19 and 23 Kya. Milankovitch argued that these cycles working in combination drive long-term climatic variations on Earth (Lowe 2001). Research on both deep marine and long continental records indicates that the pulse of global climatic change has shifted in pace and strength during the Pliocene and Pleistocene (Prell, 1982; Ruddiman and

Raymo, 1988; Raymo and Ruddiman, 1992 all in Lowe 2001). Prior to around 2.8–2.6 Mya, the 19 to 23 Kya cycle predominated. The 41 Kya cycle became the dominant influence between ca. 2.6 and 0.8 Mya, and after this the later part of the Pleistocene became dominated by the 100 Kya cycle, while the amplitude of the cycle greatly increased during this period.

Other factors that have influenced regional climates include the aforementioned rise of the Himalayas and adjacent plateaus and the distribution of land and sea. The combination of these factors lead to the following general climatic changes during the Late Tertiary and Quaternary.

Late Tertiary and Quaternary temperature changes

During the Early Miocene, 23–16.5 Mya, the climate became a bit warmer and drier than the relatively cool Oligocene, culminating in the mid-Neogene climatic optimum at about 16 Mya (Nishimura & Suparka 1997). Low-latitude warming was reflected in re-expansion of tropical and sub-tropical types of forest zones, while relatively low sea-levels during the late Early Miocene (20–16.5 Mya) allowed extensive faunal exchange between Africa and Eurasia. The early Middle Miocene (around 16–13.5 Mya) marked the warmest part of the Neogene (Janis 1993), although Flower and Kennett (1994) reported that a major and permanent cooling step occurred between 14.8 and 14.1 Mya. In fact, Böhme (2003) found evidence that in Europe, the warm and humid climate peaked between 18 and 16.5 Mya, followed by two major seasonal phases at 16.3 to 15.7 Mya and 14.7 and 14.5 Mya which were accompanied by the immigration into Europe of dry-adapted taxa. The warm period ended abruptly between 14 and 13.5 Mya with a drop of 7 ºC in mean annual temperature. The later Miocene (from 13.5 Mya onward) exhibited a steady decline in temperatures and a continuation of the drying trend (Keller and Barron, 1983; Kennett, 1985; Miller and Fiarbanks, 1983 all in Janis 1993). Kennett (1968, in Flynn et al. 1991) documented marine carbon and oxygen events of global significance at ca. 6 Mya and 2.5 Mya. In addition, he inferred a marked ice-volume increase at about 3.5 Mya. The latter date is in line with estimates based on quantitative planktonic data from the western Pacific by Wang (1994). He reports three major cooling events in the Late Neogene, i.e. the Early Pliocene (4.7–3.5 Mya), the Late Pliocene (3.1–2.7 Mya), and the Early

Pleistocene (2.2–1.0). Not all of these, however, affected the tropical region to the same extent, and the last one was probably the most important in its irreversible, step-like change in sea surface temperatures, which established the glacial climate characteristic of the Late Pleistocene. Still, sea surface temperature changes in the tropics may have been minimal, and the subtropics were probably much more affected by temperature change.

The Pliocene ended with a long cold period at ca. 2.7 Mya. Another long, cold phase probably followed between 1.5 and 1.8 Mya, and then a very long phase of several glaciations started around 800 Kya and lasted until ca. 300 Kya (Beard et al. 1982; Singh & Srinivasan 1993). It has been suggested that the interglacial at Oxygen Isotope Stage 11 (423–326 Kya) was the most prolonged and possibly warmest interglacial of the last 500 Kyr (Zazo 1999). A cool, dry period likely occurred at the end of the Late Middle Pleistocene, around 190 to 180 Kya (Zheng & Lei 1999). The last deglaciation was not a smooth transition from one climate stage to another but rather occurred in a series of abrupt warming steps interrupted by sudden shifts towards near-glacial conditions. At least 24 abrupt oscillations, from cold stadial conditions to warm interstadial conditions and back to cold conditions again, characterize the interval between 110 and 14 Kya (Dansgaard et al., 1993 in Lowe 2001). Some of these irregular oscillations lasted only 1 to 3 Kya, while some of the cold-warm transitions occurred within a few decades (Alley et al., 1993 in Lowe 2001). The most prominent warming phase in the Late Pleistocene is the Bølling-Allerød warm period, which in the southern South China Sea is synchronous with a rapid 1°C increase in sea-surface temperatures between 14.7 and 14.5 Kya (Steinke et al. 2001) and the rapid flooding of the Sunda Shelf (Hanebuth et al. 2000). This period was followed by a change to near-glacial conditions, the so-called Younger Dryas period (Steinke et al. 2001). Maloney (1996) investigated evidence for the Younger Dryas event in SE Asia. During this event, which was initially described for Europe, the post-LGM deglaciation was replaced by a sudden return within a few decades to near glacial conditions, which then lasted for about 1 Kyr. Reliable pollen data are scarce but pollen diagrams from West Java and Sumatra suggest that there may have been two brief periods of a temperature decrease of 2.5–3 ºC, with a slightly warmer phase in between them between 11 and 10 Kya.

Monsoon fluctuations

For the reconstruction of palaeoenvironments in tropical Asia an understanding of monsoon fluctuations is important, probably even more than knowledge of absolute temperatures. Precipitation in the Sundaland area is derived chiefly from the northern (or boreal) monsoon, which becomes the summer monsoon when crossing the equator. The modern boreal winter monsoon transfers vapour and cold air from the marginal seas to the southern islands, and the monsoon fluctuations are responsible for the variation in precipitation over the SE Asian islands (Lim and Tuen, 1991 in Sun et al. 2000). Sun et al. (2000) suggested that, during the last glaciation, the strengthened northern monsoon absorbed moisture when crossing the sea and should have provided more precipitation to the islands south of the South China Sea, including Sumatra, Java, Borneo and Malaysia. Therefore, the sites without relief obstruction would have received rainfall from the moisture-laden airmass from north. All these places, located in the rain-shadow areas behind mountains, could not receive much moisture from the monsoon. Consequently, these sites probably experienced much drier conditions than those found today. In SE Asia this led to the following overall climate patterns during glacial periods: (1) the whole area of SE Asia was cool or very cool; (2) a dry area extended from India to New Guinea and northern Australia, via Southeast Asia; and (3) there was a narrow wet zone extending from northern India to south of Japan. In the northern winter of the LGM, a stronger easterly monsoon in the southern hemisphere prevailed in the equatorial area, while, on the other hand, a stronger winter monsoon prevailed in East Asia. This resulted in the areas of the Malay Peninsula, north Sumatra, and north Borneo becoming dry during the northern winter, while Java became drier during the northern summer because of a stronger southern hemisphere monsoon. The weaker summer monsoon from the Indian Ocean also caused drier conditions on Java and Sumatra (Urushibara-Yoshino & Yoshino 1997). In Java, where rainfall is controlled by tropical westerlies from the Indian Ocean (Eguchi, 1983 in Urushibara-Yoshino & Yoshino 1997), weaker westerlies during the LGM probably caused a longer dry season (Urushibara-Yoshino & Yoshino 1997). These findings are confirmed by detailed analysis of SE Asian deep sea and lake records that indicate a stronger winter monsoon during the LGM, moderate to weak summer and winter monsoons during the deglaciation and an enhanced summer monsoon in the Holocene (Huang et al. 1997). The monsoonal extremes alternated with the glacial-to-interglacial cycles during the last 200 Kyr, leading to continental aridity in subtropical South China and enhanced seasonality throughout the region during the glacials, and to extreme continental wetness in South China and low seasonality in the region during the interglacials (Wang et al. 1999; Jian et al. 2001). In addition to changes in the strength of the monsoon, the extent of land that the monsoonal winds needed to cross affected the amount of precipitation. Verstappen (1997) described how the extensive Sunda Shelf had an important effect on the Pleistocene climates of SE Asia as less evaporation and greater dryness occurred when it emerged.

3.3 SEA-LEVEL CHANGES

Below, a summary is provided of data on sea-level changes, which shows that these are not always in agreement. This may partly be because some data show global sea-level cycles, while other indicate changes at a regional or local scale. Geological processes on these different scales may have very different implications for the resulting change in relative sea-levels. Furthermore, some authors discuss general sea-level changes over long, medium, or short periods of time (first, second, third order curves, see, for example Lourens & Hilgen 1997). Thus within generally falling sea-levels, certain authors may recognize highstands while others who study the records in less detail may not mention them. This makes the incorporation of sea-level changes in the palaeoenvironmental reconstructions of doubtful value, and only when sufficiently detailed information was available were sea-level heights used in palaeoenvironmental mapping. Table 3.1 provides an overview of sea-level changes since the Early Miocene.

Period	***Global sea-level***	***Source***
25–16.5 Mya	major high sea-level periods	Baumann (1982)
21–16 Mya	generally high and mildly fluctuating	Haq et al. (1987)
17–15.2 Mya	cycles of rising and falling sea-level (between – 40 m and –130 m)	Vail et al. (1986, in Sitompul et al. 1992)
16.5–16.2 Mya	important low stand (down from +150 to +50 m a.s.l.)	Baumann (1982); Haq et al. (1987)
16.2–15.8 Mya	marine transgression reached maximum in early middle Miocene	Batchelor (1979); Anderson et al. (1993)
15–10.2 Mya	consistent fall from +130 m to +50 m (second order)	Haq et al. (1987)
14–9 Mya	very low sea-levels	Morley (1999)
13.6	very low sea-levels	several authors in (Lourens & Hilgen 1997)
13–10.5	very low sea-levels	Batchelor (1979)
12.8	very low sea-levels	several authors in (Lourens & Hilgen 1997)
12 Mya	important low stand	Bernor et al. year (in Jablonski 1993)
11.7 Mya	very low sea-levels	several authors in (Lourens & Hilgen 1997)
10.5 Mya	very low sea-levels (lower than ever before in Tertiary); -120 m a.s.l.	several authors in (Lourens & Hilgen 1997); Haq et al. (1987)
9.8–6.6 Mya	important low stand, generally rising sea-levels from ca. 0 m to + 30	Baumann (1982) Haq et al. (1987)
9.3 Mya	very low sea-levels	several authors in (Lourens & Hilgen

		1997)
8.5 Mya	maximum sea-level in Sumatra	Anderson et al. (1993)
6.6–2.8 Mya	major high sea-level periods	Baumann (1982)
6 Mya	mild fluctuation around the 0 m mark	Vail et al. (1986, in Sitompul et al. 1992)
5.8 Mya	very low sea-levels (-100 m) very low sea-levels (-50 m)	van den Bergh et al. (1996) Vail et al. (1986, in Sitompul et al. 1992)
5.3 Mya	relatively high sea-levels (-10 m)	van den Bergh et al. (1996)
5–4.2 Mya	very high sea-levels (+ 100 m)	Haq et al. (1987)
4.4 Mya	very high sea-levels (highest in period 12 Mya to present)	van den Bergh et al. (1996)
4.0–3.2 Mya	30–35 m higher than today	Krantz, 1991 (in Dowsett et al. 1994)
4.0–3.5 Mya	low sea levels (ca. –50 m a.s.l.)	Haq et al. (1987)
3.5–2.9 Mya	sea-levels up to 60 m higher than present	Haq et al. (1987)
3.0 Mya	about 25 m higher than today about 25 m higher then today	Krantz, 1991 (in Dowsett et al. 1994) Haywood et al. (2000)
2.8 Mya	significant drop in sea-level	Haq et al., 1987 in (Lourens & Hilgen 1997)
2.8–2.4 Mya	fluctuating increase in ice volume leading to fluctuating but generally lower sea-levels	
2.55–2.0 Mya	Sea-levels fluctuate in 41 Kya cycles, with peak eustatic water depth varying between +110±20 m and +25±20 m. Minimum peak eustatic water depths are ca +10 m	Naish (1997 in Pillans et al. 1998)
2.4 Mya	first major glaciation, sea-levels on average 70 m below present day levels. This does not show in Naish's (1997 in Pillans et al. 1998) sea-level curves.	van den Bergh et al. (1996)
1.7 Mya	significant drop in sea-level	Haq et al., 1987 in (Lourens & Hilgen 1997)
1.6 Mya	Sea-levels fluctuating between -40 and –100 m, somewhat lower than the period hereafter	van den Bergh et al. (1996)
1.4 Mya	sea-levels at ca. –30 m which was the highest in 600 Kya	van den Bergh et al. (1996)
1,0 Mya	highstand of +5 m above PDL	Shackleton et al. (1988 in Zazo 1999)
900 Kya	sea-levels fluctuating at generally relatively high levels between -90 and -30 m, or +22 m	van den Bergh et al. (1996) Zazo (1999)
800 Kya	Moderate fluctuation mode that existed since 2.4 Mya changes dramatically to high amplitude fluctuation with minimum sea-levels down to 170 below present-day level (PDL) and an average sea-level of around 90 m below PDL	van den Bergh et al. (1996)
660 Kya	sea-levels very low at -170 m	van den Bergh et al. (1996)

	significant drop in sea-level	Haq et al., 1987 in (Lourens & Hilgen 1997)
450 Kya	sea-level at -140 m	Chappell (1998) and Rohling et al. (1998)
400 Kya	sea-level possibly at +20 m, but as yet unconfirmed	Chappell (1998)
400 Kya	This may have been the first time that the sea-levels rose significantly above the -130 m continental shelf break and flooded the entire Sunda Shelf	Batchelor (1979)
340 Kya	glacial, sea-level at ca. -134 m	Rohling et al. (1998)
310 Kya	inter-glacial, sea-level at -50 m	Beard et al. (1982)
300–240 Kya	glacial; sea-levels at -130 m	Beard et al. (1982)
270 Kya	Glacial; sea-levels at 120 m	Rohling et al. (1998)
240–190 Kya	inter-glacial, sea-levels at ca. -10 m	Beard et al. (1982)
140 Kya	sea-level at –125 m	various authors in Pillans et al. (1998)
135 Kya	sea-level at -125 m	Rohling et al. (1998)
130–80 Kya	from glacial maximum into interglacial	Beard et al. (1982)
	During the Eemian global sea-level was at least three meters, and probably more than five meters higher that at present	Cuffey and Marshall (2000)
81 Kya	Sea-levels were already well below present levels and would pass the -50 m mark at 74 Kya	Gingele et al. (2002)
80 Kya	Sea-level at 14.6 m below present level for at least 4 Kya and probably longer	Kamaludin and Azmi (1997)
74 Kya	Sea-level dropped below -50 m, closing the Sunda Strait	Gingele et al. (2002)
80–30 Kya	Sea-level slowly falling from -20/-40 to -50/-100	van den Bergh et al. (1996)
59–50.2 Kya	sea-level low	Schönfeld and Kudrass (1993)
55 Kya	sea-level at –4 m off-shore Malaysia, lasting at least 2 Kya, but, data by Gingele et al. (2002) suggest that sea-levels did not go above –50 m from 74-12 Kya.	Kamaludin and Azmi (1997) Gingele et al. (2002)
29.5–18.3 Kya	lowest levels of the LGM	Schönfeld and Kudrass (1993)
22–21 Kya	sea-level at -120 m	Chappell (1998), Woodroffe (1993), Rohling et al. (1998)
12 Kya	Sea-level crosses the –50 m mark	Gingele et al. (2002)

Table 3.1. Global and regional sea-level estimates from literature

It needs to be noted that records of Pliocene and Early Pleistocene highstands are extremely scarce, possibly because the shorelines of this period were almost entirely eroded during the subsequent transgressions of the Middle Pleistocene, and also because the Early Pleistocene sea-level fluctuations had smaller amplitudes and frequencies than later Pleistocene fluctuations (Zazo 1999).

After the LGM, sea-levels initially rose relatively slowly with an average rate of ca. 6 m/Kya between 21 and 17 Kya. After that, for the next 10 Kya, the rate increased to 10 m/Kya. Significant departures from these average rates may have occurred at the time of the Younger Dryas and possibly around 14 Kya (Fleming et al. 1998). Throughout much of the region there is evidence that the sea-level rose to above its present level in mid-Holocene. The timing, magnitude and occurrence of this Holocene high sea-level is not, however, the same throughout the region (Woodroffe 1993). In NW Kalimantan, the post-glacial marine transgression culminated in the mid-Holocene at +3 m (Thomas et al. 1999). In the Tambelan Islands, Haile (1970) found that sea-levels were at least 0.3 m higher than today at 5,460±110 years BP, and 0.4 m at 5,270 ± 110 years BP, whereas in the Bunguran Islands, sea-levels were relatively lower by at least 0.7 m at 6,260 ± 120 years BP. In the southern Malacca Strait sea-level indicators for the time interval between 5 and 4 Kya have been found at +2.5 and + 5.8 m (Geyh et al. 1979). Tentative results from Singapore indicate that the Holocene Post-Glacial Marine Transgression reached present mean sea-level around 6.5–7 Kya, rose to nearly 3 m above present, and began to fall to present sea-level around 3 Kya or less. A comparison of the proposed sea-level curve of Singapore with that for Peninsular Malaysia indicates that the highest mid-Holocene sea-level may have been closer to +3 m rather than +5 m (Hesp et al. 1998). In southern Vietnam, maximum Holocene transgression occurred at around 5 Kya, with a sea-level of +3.5 m above the PDL (Ta et al. 2001). In south Central Java, the deposition of Holocene lagoonal-fluvial material suggests sea-levels that were ca. 6 m higher than PDL. It should be noted that all the above dates of Holocene sea level highs were obtained through ^{14}C dating; there is a need to calibrate these datings using other techniques.