1995

Late tertiary and quaternary geology of the East Java Basin, Indonesia

Susilohadi
University of Wollongong

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LATE TERTIARY AND QUATERNARY GEOLOGY
OF THE EAST JAVA BASIN, INDONESIA

A thesis submitted in (partial) fulfilment of the
requirements for the award of the degree of

DOCTOR OF PHILOSOPHY

from

THE UNIVERSITY OF WOLLONGONG
NEW SOUTH WALES, AUSTRALIA

by

SUSILOHADI
(Ir. ITB)

School of Geosciences
1995
ABSTRACT

The East Java Basin is classified as a retro-arc basin, situated on the southeastern margin of the Sunda Shelf, Indonesia. A detailed Late Tertiary and Quaternary sedimentary evaluation of this basin is based primarily on seismic stratigraphic interpretation, combined with onshore outcrop studies of equivalent geological units. The primary aim of this study was to assess the sedimentation patterns and sedimentological history in the East Java Basin.

During the Middle Miocene and earlier, the northern part of the East Java Basin was strongly controlled by northeast-trending half graben systems which occurred along sutures between the earlier Late Cretaceous-Early Tertiary subduction systems. Little is known about the basin configuration in the southern part of the basin prior to the Middle Miocene.

Since the Late Miocene, east-trending anticlinal zones developed and were superimposed on the previous northeast-trending structures. The anticlinal Rembang Zone, which extends from the Blora area to Madura Island, became a dominant structure which divided the East Java Basin into two major basinal (synclinal) areas. In the north, a basinal area occurred between the Karimunjawa and Bawean Arches and the Rembang Zone. In the south, a basin developed between the Rembang Zone and the chain of volcanism along the median line of Java. Peak sedimentation in these east-trending basins occurred during the Late Miocene and Pliocene. The sedimentation patterns, such as local thickening, in the north basinal area (southern Java Sea) suggest that the basin development ceased in the Early/Middle Pleistocene. In contrast, rapid basin subsidence still occurred in the south basinal area at least until the late Middle Pleistocene, when extensive east-trending folding accompanied by diapiric movement occurred to form the anticlinal Kendeng Zone.

Middle Miocene and earlier sedimentation have not been comprehensively studied in the East Java Basin, particularly because of the paucity of outcrop and well data. The present seismic stratigraphic study of the onshore Rembang Zone revealed that the Middle Miocene sedimentation (Tawun Formation) patterns were strongly controlled by half graben systems, the southern extension of the northeast-trending structures indicated earlier. These structures still partly controlled the Late Miocene and Pliocene sedimentation in the Rembang Zone and the southern Java Sea.
Hemipelagic marl deposition (Wonocolo Formation) occurred in the southern part of the East Java Basin (onshore area) during a sea level highstand in the early Late Miocene, but little is known about equivalent deposition in the southern Java Sea at this time. The marl deposition was contemporaneous with the deposition of volcaniclastic turbidites (Kerek Formation) derived from Late Miocene volcanism along the median line of Java. This volcanism appears to have ceased in the late Late Miocene. A shallow shelfal facies (Ledok Formation) occurred in the southern part of the Rembang Zone (onshore area) which southward, grades laterally into the marl of the Kalibeng Formation.

During the Pliocene, the sedimentation was confined within the northern and southern basinal areas. Widespread pelagic marly deposition occurred during a prolonged highstand of sea level in the Pliocene which reflects a global eustatic sea level rise recognised by Haq et al. (1988). In the southern basinal area, the deposit is represented by the Mundu (seismic unit PL1), Kalibeng and Atasangin Formations respectively; whereas in the northern basinal area (southern Java Sea) the deposit is represented by seismic units JP1 and JP2. A relative sea level fall occurred in the Late Pliocene and led to the development of an anoxic lacustrine deposit (Lidah Formation) in the southern basinal area.

During the Quaternary, the seismic stratigraphic study indicates that sediment stratal patterns were highly controlled by frequent relative sea level changes, and sedimentation was still confined within the two major basinal areas indicated earlier. In the southern basinal area, seismic sections clearly display an intertonguing between volcaniclastics derived from the Quaternary volcanism along the median line of Java, and shallow marine sediments derived from the exposed Rembang Zone in the north. Field observations indicate that volcaniclastic deposits are coarse-grained, whereas on seismic sections from Madura Strait these deposits are associated with relatively steep depositional slopes (of alluvial fan/fan delta systems). The seismic stratigraphic study in Madura Strait indicates that the depositional timing of these deposits was strongly controlled by the relative sea level changes in this area.

In the southern Java Sea (northern basinal area), the basin filled during the Pliocene as revealed from the seismic stratigraphic study, giving a flat basin topography at the end of the Pliocene. Nine thin seismic subunits are recognised in this area and are believed to represent at least nine Quaternary sea level fluctuations. Each subunit is relatively thin, and tends to be distributed widely because of deposition on a relatively flat lying area. The seismic
characters of these subunits are very similar, with subparallel reflection or almost reflection free patterns representing marine deposits, topped by extensive fluvial channelling. This repetitive succession represents highstand and lowstand of sea level respectively.

The integration of seismic sequence stratigraphic analysis and outcrop sedimentological studies has provided a comprehensive view of the East Java Basin on the timing of deposition and deformation of the basin fill, mode of deposition and sedimentation patterns within the sub-basins, and the influence of sea level changes on stratal patterns.
ACKNOWLEDGMENTS

I gratefully acknowledge the Australian Government through the Australian Agency for International Development (AusAID) for providing a scholarship for this research.

I would like to express my deep appreciation to my supervisors Dr L.E.A. Jones and Dr C. Murray-Wallace, and also to Associate Professor B.G. Jones for their help, encouragement and guidance during the study.

I am also indebted to the former Heads of the Geology Department, Wollongong University, Dr A.C. Cook and Associate Professor A.J. Wright, for their constructive suggestion to do this research. Special gratitude also goes to all academic, administrative and technical staff of the department, particularly Ms B. McGoldrick, Mr M. Perkins, Mr D.A. Carrie, Mr A.M. Depers and Ms P. Williamson.

Special thanks are forwarded to the former Director of the Marine Geological Institute of Indonesia, Mr Ismail Usna, and the present Director Mr Aswan Yasin, who permitted the author to do this research. Special appreciation is also forwarded to my colleagues in this institute: Ms K.T. Dewi for micropalaeontological analysis, and Messrs S. Tjokrosapoetro, B. Dwiyanto, A. Masduki, A. Setiabudhi, P. Astjario, S. Lubis and B. Dharmawan for their support and suggestion.

Special thanks are also addressed to the management of Pertamina, PPT Migas and Lemigas for the permission to use the geological, geophysical and well data from the Java Sea and Cepu area. The author would like to acknowledge Drs N. Hasjim, A. Muin and S. Soeka; Messrs T. Ratkolo, A. Sutrisman, N. Suparjono, and S. Musliki for their help in searching for some references in these institutes.

Many thanks are also forwarded to my post-graduate colleagues, including Messrs Surono, Y. Kusumahbrata, A. Pujobroto, Moradian, Soltani, A. Mandile, A. Wahib, D. Dharmayanti, Destini, H. Sobir, Prasetyo, A. Hamedi and others, for lively discussions during this research. Finally, I am thankful to my wife and daughter, Kresna and Arum, and the rest of my family for their support.
Except where otherwise acknowledged, this thesis represents the author’s original research and the material included has not been submitted for a higher degree to any other institution.

Susilohadi
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CHAPTER ONE
INTRODUCTION

The East Java Basin is located on the southeastern margin of the Sunda Shelf, an extensive shelf which comprises the whole of western Indonesia. The basin is classified as a retro-arc basin (Dickinson, 1977) because of its position behind a volcanic arc in a convergent plate tectonic setting. Like other retro-arc basins in Indonesia, this basin is geologically and economically important. The East Java Basin is an important petroleum province and has been the subject of geological investigations since the end of the 19th century. Besides petroleum-related geological investigations in this area, important research has also been undertaken in the fields of biostratigraphy and Quaternary anthropology. These interests have highlighted the necessity to continuously assess the region’s geological evolution in the light of new geological concepts.

The development of geological concepts and ideas has been extremely rapid during the last decade, particularly after the introduction of (seismic) sequence stratigraphy (Vail et al., 1977; Posamentier et al., 1988; Posamentier & Vail, 1988). Application of these ideas is world wide, but the numerous applications concentrate on shelf margin settings rather than retro-arc basins. Since the introduction of these concepts, very few attempts have been made to apply them to the East Java Basin and these few were aimed at specific problems in limited areas of the basin (e.g. Yulianto, 1993). The present study critically examines the East Java Basin based on seismic sequence stratigraphic concepts. Within the limits of the available data, this study investigates the Late Tertiary and Quaternary strata in almost the whole East Java Basin with a major emphasis on seismic interpretation.

1.1 PREVIOUS INVESTIGATIONS IN THE EAST JAVA BASIN

a. Pre 1940s

Before the 1940s, geological exploration was carried out primarily by Dutch geologists. In the inland area, two institutes with different interests were involved. In the northern part of this area (the anticlinal Rembang Zone; see Chapter 2.5 for an explanation), the geological exploration, which began in 1887, was carried out by the Dortsche Petroleum Maatschappij and then continued by the Bataafsche Petroleum Maatschappij (BPM). Both of these Dutch oil
companies found some oil fields shortly after exploration commenced, but the results of their study were proprietary and not publicly available (van Bemmelen, 1949).

Geological exploration in the southern part of the inland area (the anticlinal Kendeng Zone; see Chapter 2.5 for an explanation) was carried out by the Geological Survey of the Netherlands Indies, and involved geological mapping and stratigraphic evaluation of Tertiary and Quaternary strata. Some East Java geological mapping reports are considerably detailed (e.g. Duyfjes, 1938a,b,c,d). Most of the work of the Geological Survey of the Netherlands Indies was summarised by van Bemmelen (1949). Until recent years this summary remained the main source of information on the geology of East Java.

Quaternary geological studies in East Java began with the finding of mammalian faunas and hominid fossils by Dutch geologists in the late 19th century, as reported by Duyfjes (1936) and de Vos and Sondaar (1982). These faunas occur in three main Quaternary stratigraphic units, the Pucangan, Kabuh and Notopuro Formations. Examination of the relationships between Quaternary deposits and sea level changes in East Java is timely, as the present study suggests that the extent of these stratigraphic units is largely determined by Quaternary sea level changes.

Previous geological investigations in the East Java offshore area were carried out mostly as part of a regional study. The early work is credited to Earle (1845) who noted that the Sunda Shelf, on which the present study area is located, has an extensive, flat sea floor. This idea was further explored by Molengraaff (1921) who suggested that the Sunda Shelf is a submerged peneplain resulting from eustatic sea level change. In the 1930s the Snellius Expedition I, a large scale oceanographical exploration, was conducted in Indonesian waters. The expedition included bathymetric profiling, the results of which prompted Kuenen (1935b; 1950) to suggest the occurrence of several distinguishable drainage patterns across the Sunda Shelf.

b. Post 1940s

After the 1940s (post-Dutch colonisation), geological investigations in East Java were carried out by various geological institutes. The Bataafsche Petroleum Maatschappij (BPM) concession and its exploration activities were taken over by Lemigas (a national oil institute),
and the Geological Survey of Indonesia continued the work of the Geological Survey of the Netherlands Indies.

In the late 1960s and early 1970s, Lemigas and BEICIP (Bureau d’Etudes Industrielles et de Cooperation de l’Institut Francais du Petrole) conducted a regional review of the geology of East Java. Unfortunately, most of the results were proprietary and only a few parts were published as a general review of the geology of East Java (e.g. De Genevraye & Samuel, 1972; Soetantri et al., 1973).

One of the significant achievements of geological investigations in the East Java onshore area after the 1940s was the application of planktonic foraminiferal biostratigraphic zonation for unravelling a chronology of deposition during the Neogene (e.g. Bolli, 1966; Pringgoprawiro & Baharuddin, 1979; van Gorsel & Troelstra, 1981; Pringgoprawiro, 1983; Hasjim, 1987). This application was made possible by the fact that the Neogene strata are rich in foraminifers and highly diversified faunal assemblages, and more importantly because of the increasing interest in using planktonic foraminifers in biostratigraphic correlation since the mid 1950s (Bandy, 1964). Although the chronology of the Neogene strata is now better understood, sedimentation patterns in sub-basins of the East Java onshore and offshore areas are still poorly understood.

Quaternary geological investigations in East Java have been carried out by at least two institutes, Institut Teknologi Bandung (affiliated university) and the Geological Research and Development Centre (a subsidiary of the Geological Survey of Indonesia). Their activities were directed towards accurate dating of sedimentary horizons which contain hominid remains (Sartono et al., 1981; Jacob, 1978; Hyodo et al., 1992, 1993), rather than the relationship between Quaternary sedimentation patterns and sea level changes.

In the late 1960s and early 1970s, regional marine geological and geophysical surveys were carried out in the Sunda Shelf by the Woods Hole Oceanographic Institution as a contribution to the United Nations ECAFE-CCOP program. This shelf was studied as part of reconnaissance mapping of the continental margin structure off eastern Asia (Ben-Avraham & Emery, 1973). The study included seismic reflection and refraction profiling, gravity, magnetics and bathymetry. The results of the Java Sea part were published in two papers by Emery et al. (1972) and Ben-Avraham and Emery (1973). Emery et al. (1972) corroborated
the results of the previous surveys by Dutch geologists. They agreed that many fluvial cut-and-fill structures occur in the uppermost 50 m of the Java Sea floor, and have attributed these to subaerial erosion of the region as a consequence of sea level lowstands during the Pleistocene glacial maxima. A study by Ben-Avraham and Emery (1973) stressed the regional structure of the Tertiary basement that underlies the Sunda Shelf rather than the sedimentation pattern within the basins on this shelf.

Since the mid 1960s some major oil companies have been granted concessions in the Java Sea and Madura Strait. Although numerous geological and geophysical studies have been undertaken (Tanner & Kennett, 1972), the results are seldom available. Only a few papers have been published (e.g. Najoan, 1972; Sudiro et al., 1973; Kenyon, 1977; Bishop, 1980). The regional stratigraphy of the Java Sea was discussed by Najoan (1972), while Sudiro et al. (1973) concentrated on the Tertiary basement structural configuration of this area. Bishop (1980) discussed the relationship between structure, stratigraphy and hydrocarbon prospects in the eastern part of the Java Sea. They agree that northeast-trending structures controlled the Early and Middle Tertiary sedimentation patterns, but no further evaluation has been made of the Late Tertiary and Quaternary sedimentation.

Today, offshore geological research is mainly carried out by the Marine Geological Institute (a subsidiary of the Geological Survey of Indonesia). Its activities are centred on sea floor mapping, including the sedimentary strata extending a few hundred metres below the surface (Quaternary and recent sediments). However, until now, no comprehensive study has been undertaken using these data.

1.2 AIMS OF STUDY

The initial aim was to investigate the distribution of the Late Tertiary and Quaternary sediments of the offshore East Java Basin (Madura Strait and southern Java Sea), beyond the level of understanding reached by early workers through inferences based on the geology of the inland areas and shallow soundings. Preliminary investigations highlighted the necessity to include the onshore area to provide a comprehensive overview of the East Java Basin.
In general, the study concerns three different parts of the East Java Basin: the onshore (the Rembang and Kendeng Zones), Madura Strait and the southern Java Sea. The main objectives of the study include:

1. a critical reassessment of the sedimentation history of the onshore East Java Basin during the Middle to Late Tertiary, focusing on the relationship between the northern (Rembang Zone) and southern (Kendeng Zone) anticlinal zones;

2. a reinterpretation of the onshore Quaternary succession in the light of extensive new data, determined from shallow seismic investigations carried out in Madura Strait, which forms the eastern offshore extension of the East Java onshore area. The seismic interpretation forms the major part of this study;

3. Seismic stratigraphic interpretation of the onshore portion of the East Java Basin and its relationship to tectonism and global sea level changes; and

4. Seismic stratigraphic interpretation of the southern Java Sea to assess the sedimentary sequences extending from the Rembang Zone onto the margin of the stable Sunda Shelf.

1.3 AREA OF STUDY

Geographically the study area is located between longitudes 111° and 114° East and latitudes 5°55’ and 7°40’ South, situated on the southeastern part of the Sunda Shelf (Fig. 1.1). It includes the northeastern part of East Java, Madura Island, Madura Strait and the southeastern part of the Java Sea. For the present study emphasising seismic sequence stratigraphic interpretation, the geophysical data coverage (particularly offshore seismic data) is relatively comprehensive in this area and includes almost the whole East Java Basin.

The study area which experiences year round maximum temperatures between 29 and 32°C, is about 600 km distance from the Indonesian capital city, Jakarta, and can be reached by plane to Surabaya airport or by public transport. During the offshore surveys, the research vessel used three main ports in the study area, Surabaya, Probolinggo, and Tuban, to load and unload research materials.

The onshore part of the study area consists of low hills commonly covered by teak forests, and plains which are mostly covered by alluvium of two major rivers, the Solo and Brantas Rivers. These rivers originate in the mountainous area of Quaternary volcanoes along the longitudinal line of Java, and traverse toward the Java Sea and Madura Strait. Rapid rural
development in East Java has made transportation within the study area easy, as almost all remote areas are linked by public transport. However, as in other tropical areas, geological outcrops are rare, which has been the major constraint during this study. Geological outcrops are mostly located in river banks and quarries. Geological field work is impractical during the rainy season (southern hemisphere summer, November to February), as flooding caused by both major river systems is the main hindrance.

1.4 ORGANISATION OF THESIS

This thesis comprises eight chapters. The geological discussion begins with a review of the geology of the study area within a regional geological framework, which is presented in Chapter Two. Chapter Three is a review of the stratigraphy of the study area, focussed particularly on the onshore part of the study area. The sedimentology and sequence stratigraphic discussions are presented in Chapters Four, Five, Six and Seven. These four chapters are arranged so that each chapter is concerned with a different geographic area. This arrangement is also necessary because of the different methods used. Multifold seismic sections combined with some geological observations were used in the study on the northern part of East Java (Chapter Four), while the study on the southern part of East Java (Chapter Five) was carried out based on outcrop observations and laboratory work only. Chapters Six and Seven, which discuss the sequence stratigraphy of Madura Strait and the Java Sea respectively, rely heavily on shallow seismic profiles. The final chapter, Chapter Eight, concludes and summarises the discussion presented in the preceding chapters.
CHAPTER TWO
GEOLOGICAL SETTING OF THE EAST JAVA BASIN

2.1 INTRODUCTION

This chapter discusses and summarises the geological setting of the East Java Basin. The discussion concentrates on the tectonic setting in terms of plate tectonic concepts, regional geological structures and regional depositional history. The aim of the discussion is a regional review of the geological processes that have controlled the development of the East Java Basin since the Early Tertiary.

2.2 TECTONIC SETTING

The East Java Basin is located on the southeastern margin of the Sunda Shelf. This shelf constitutes most of the western Indonesian region and includes the Malay Peninsula, Sumatra, Kalimantan, Java and the region covered by sea between these islands (Fig. 2.1). The plate tectonic setting of this region has been described in great detail by Hamilton (1977, 1979, 1988). He reported that the western Indonesian region has experienced a complex geological history resulting from the convergence of the Eurasian plate and the Indian-Australian plate (Fig. 2.1). Hamilton (1977, 1979) and Baumann (1982) recognised six major morphologic elements in western Indonesia resulting from this convergence (Fig. 2.1 & 2.2). These are: the Sunda (Java) trench, accretionary wedge, fore-arc ridge, fore-arc basin, active volcanic arc and back-arc/retro-arc basin (Fig. 2.2a). These elements can be traced from the southern Andaman Sea southeastward to the Banda Sea area. To the south of Java the fore-arc ridge and basin are submerged (Hamilton, 1977; 1979; Fig. 2.2a). The East Java Basin has in fact developed as a retro-arc basin to the north of the active volcanic arc in central Java (Fig. 2.3).

The earliest evidence for basin development should correlate with the commencement of the subduction system in western Indonesia. Hamilton (1977) suggested that subduction commenced as early as Late Oligocene. Eocene and Oligocene sediments near the Eurasian plate margin are dominated by non-volcanic facies deposited in a stable continental shelf setting. This opinion was supported by Baumann et al. (1972) and de Genevraye and Samuel (1972). They found that the oldest volcanic facies in Java are the Late Oligocene to Middle
Miocene Jampang Formation in West Java, and the Middle Miocene to Late Miocene Kerek Formation in Central and East Java.

Dickinson (1977) suggested that in a back-arc/retro-arc basin, evolution of the basin is controlled by regional downward flexure of the plate margin in response to local tectonic loading at the plate edges, and sedimentation keeps pace with subsidence. This is apparently true for the East Java Basin where thick and relatively continuous sedimentation has occurred. A report by Lemigas/BEICIP (1969) indicated that more than 5000 m of sediment has been deposited in this basin since the Early Tertiary and drill holes have never reached basement.

The present study is an attempt to evaluate the depositional styles in the East Java retro-arc basin, and focuses on the Late Tertiary and Quaternary strata where data are more available. This study also evaluates the role of tectonics and sea level change on the depositional stratal pattern in this setting. The wide coverage of the data used allows a comprehensive study from the retro-arc basin to shelfal area.

2.3 REGIONAL GEOLOGICAL STRUCTURES

Most workers agree that the stress caused by the northward motion of the Indian-Australian plate has been responsible for most of the structural deformation in eastern Java and the surroundings (Sudiro et al., 1973; Situmorang et al., 1976; Bishop, 1980; Baumann, 1982).

Faults are the most prominent structural elements on the Sunda Shelf (Ben-Avraham & Emery, 1973). In the eastern Java Sea (Fig. 2.2a & 2.3) the most prominent faults trend northeastward. According to Koesoemadinata and Pulunggono (1971) these faults may have originated as dextral strike-slip faults, but some differential vertical movement must have also occurred.

A number of northeast-trending structural highs occur in the eastern Java Sea (Figs 2.2a & 2.3) and are suspected to be the remnants of an older arc system (Ben-Avraham & Emery, 1973; Katili, 1989). These structural highs extend to southeastern Kalimantan where serpen tinised ultrabasic rocks as well as radiolarian cherts are exposed, indicating a previous subduction complex. Early Tertiary sedimentation is in fact confined within the depression between these highs (Situmorang et al., 1976; Bishop, 1980; Baumann, 1982).
In eastern Java, the most prominent structure is the east-west half graben along the median line of Java, parallel to the plate margin as shown on Figure 2.2b (its suture is now occupied by the Quaternary volcanic deposits). This feature is very well displayed by the regional Bouguer gravity mesh diagram (Fig. 2.4). The data for this diagram were digitised from the gravity maps produced from regional gravity surveys conducted by the Bataafsche Petroleum Maatschappij (pre 1940s) and by the Geological Survey of Indonesia (post 1940s). In this diagram the Bouguer anomaly shows, at a regional scale, a progressive decrease in a southerly direction from about +40 mgal to about -50 mgal. This is interpreted as the result of a progressive southward deepening of basement. According to Ben-Avraham and Emery (1973) the gravity minimum is approximately coincident with the thickest sediments in the East Java Basin.

Superimposed on this southward regional trend are a number of local gravity highs. On the northern half of the diagram these highs are linearly oriented in a general east-west direction. It is suggested that these anomalies result from vertical differential movement along east trending faults in the basement. The seismic sections interpreted during this study reveal that these positive anomalies correspond to Tertiary topographic highs.

On the southernmost part of the diagram the gravity highs are less oriented. They are interpreted as the expression of magmatic activity in this region. Interpretation of the free-air gravity anomaly over the volcanoes in Java by Ben-Avraham and Emery (1973) has indicated the presence of a mass excess as expressed by an increase in value of up to +150 mgal.

2.4 REGIONAL DEPOSITIONAL EVOLUTION

According to van Bemmelen (1949) the oldest stratigraphic unit on the Sunda Shelf is Palaeozoic, consisting of crystalline schists of unknown age that are products of sedimentary deposits altered by regional metamorphism. These rocks have a very restricted distribution, mainly on Sumatra, Kalimantan and the Malay Peninsula.

In Java, Early Tertiary sedimentary deposits are not widely observed. Outcrops of Early Tertiary lithologies are found only in West and Central Java (Fig. 2.5), while in East Java early Tertiary rocks are known only from well data.
In West Java, early Tertiary sequences are represented by about 1900 m of strata dominated by paralic to neritic facies (Lemigas/BEICIP, 1969; Kusumahbrata, 1994). In Lokulo, Central Java, the Eocene sedimentary sequence unconformably overlies pre-Tertiary rocks. The sequence consists of polymict conglomerate and sandstone, with carbonaceous material at the base and sandstone alternating with shale, marl and lenses of foraminiferal limestone in the upper portion. In Jiwo Hill, Central Java, the Eocene strata unconformably overlie pre-Tertiary metamorphic rocks. Two units can be differentiated: conglomerate, marl, sandstone and limestone of the Gunung Wungkal Formation and foraminiferal limestone and marl of the Gamping Formation which overlies the Gunung Wungkal Formation (van Bemmelen, 1949).

Ben-Avraham and Emery (1973) noted that the Tertiary basins on the Sunda Shelf did not originate simultaneously. Koesoemadinata (1969) argued that the sedimentary sequences within these basins are similar, an idea that has been supported by Baumann (1982). As in the case of the study area, where the basin is fragmented into smaller basins, the similarity of sedimentary sequences appears to have been strongly controlled by relative sea level changes.

Baumann (1982) recognised five sedimentation cycles in Sumatra and Java during the Cainozoic. Each of these cycles commenced with a transgression, and ended with a phase of volcanism and tectonism with the duration of the cycles being up to 10 Ma. These cycles are: Middle Eocene to Early Oligocene, Late Oligocene to Early Miocene, late Early Miocene to Middle Miocene, Middle to Late Miocene, and Pliocene to Recent. Figure 2.6 displays the tentative palaeogeography within the first four cycles as summarised by Baumann (1982). This figure illustrates that sedimentation mostly occurred along the rim of the Sunda Shelf where tectonic disturbance as well as volcanic activity has been dominant. The core of the shelf was relatively stable and has been a major source of sediment, in addition to the volcanic arc (Baumann, 1982).

The most widespread lithologic unit on the southern Sunda Shelf is an Early Miocene reefal limestone. These deposits were followed by a regressive phase during which mudstone was deposited in the deeper parts of the western Java Sea basins, while mudstone and carbonate rocks accumulated in the eastern Java Sea. This regressive phase in the eastern Java Sea was associated with uplift and folding (Ben-Avraham & Emery, 1973).
During the Plio-Pleistocene, marine sedimentation continued over the Sunda Shelf and gradually changed the morphology of the basin to a relatively flat plain (Ben-Avraham & Emery, 1973). The present seismic study in Madura Strait and the eastern Java Sea by the author has confirmed this opinion (see Chapters 6 & 7).

Magmatic activity on the Sunda Shelf occurred throughout the Tertiary and Quaternary. Volcanism occurs along the rim of this shelf and was thought to be associated with major tectonic events by van Bemmelen (1949), which the modern plate tectonic concepts would suggest relate to the plate convergence between the Indian-Australian plate and Eurasian plate (Hamilton, 1977; 1979).

2.5 PHYSIOGRAPHY

The Sunda Shelf has an extensive flat floor, bordered by a volcanic arc along the southern and western margins. On this shelf, water depth is generally shallower than 100 m, in contrast with the rather complex and irregular bathymetry of the surrounding deep-ocean floor. The flatness of the sea bottom in the Sunda Shelf was first reported by Earle (1845). After the first Snellius Expedition in the 1930s Kuenen (1950) suggested the presence of several drainage systems which have been submerged during the present highstand of sea level (Fig. 2.7). This finding is very well confirmed by the present seismic stratigraphic study in the Java Sea (see Chapter 7).

Van Bemmelen (1949) described the physiography of Java in detail (see Fig. 2.8). These physiographic features have mostly resulted from orogenic movements during the Tertiary and Quaternary. In East Java, where the study area is located, there are two prominent zones of anticlinoria, the Rembag and Kendeng Zones.

The Rembang Zone has been recognised as an oil-rich zone in East Java. This zone consists of a number of roughly east-west trending anticlinoria, alternating with alluvial plains. The Rembang Zone has an average width of 50 km and a height of less than 500 m above sea level. From the Tuban-Paciran area, the folds plunge westward to the northwest of Purwodadi (Fig. 2.8). In some places the Rembang Zone was eroded to expose Miocene strata. In its north-eastern half, this zone is characterised by the presence of a Plio-Pleistocene reefal limestone that overlies the Middle Miocene formations.
The Kendeng Zone is separated from the Rembang Zone by the depression of the Randublatung Zone (Fig. 2.8). Little is known about the geology of the Randublatung Zone due to limited outcrop and well data. As in the Rembang Zone, the Kendeng Zone is also characterised by a number of east-trending anticlinoria formed during the Middle Pleistocene. The height, width and fold complexity of the Kendeng Zone gradually decrease eastward, and it disappears under alluvial deposits in the Mojokerto area (Fig. 2.8). The anticlines are generally eroded down to marly and clayey sequences of Late Miocene age. The southern limit of the Kendeng Zone is a depression which has been filled with the Quaternary volcanic products of the Solo Zone.

The character of the sedimentary units in the Kendeng Zone is considerably different from those in the Rembang Zone. The differences are mostly expressed by their composition, which indicates that there were two different sources of sedimentary detritus during the Tertiary and Quaternary. Volcaniclastic and volcanogenic fragments are common in the sequences from the Kendeng Zone, while continentally-derived sediment is common in the Rembang Zone. The Bouguer gravity diagram (Fig. 2.4) indicates that the gravity minimum coincides with the Kendeng Zone. As Ben-Avraham and Emery (1973) pointed out, this gravity minimum coincides with the thickest sedimentary sequence, and continuous basin subsidence would have occurred in the Kendeng Zone.

2.6 SUMMARY

The East Java Basin situated on the Sunda Shelf is considered to represent a retro-arc basin. This type of basin is commonly characterised by thick sediment accumulations (Dickinson, 1977), especially towards the arc.

The collision between the Eurasian and Indian-Australian plates has been responsible for most of the structural deformation in and around the East Java Basin. The resulting structures largely controlled the Tertiary and Quaternary sedimentation pattern in this region. Two prominent structures are displayed: the east-west basement depression running along the median line of Java, and the northeast-southwest graben in the Java Sea between Java and Kalimantan. In East Java the east-west depression zone contains two anticlinorial zones, the Rembang and Kendeng Zones.
The oldest sedimentary strata (Early Tertiary) in the Java area are underlain by a pre-Tertiary basement of metamorphic rocks which crop out in several places in Java. In East Java the oldest recognised Tertiary sequence is Early Miocene. The Tertiary sedimentary sequence in the East Java Basin is similar to the other parts of the Sunda Shelf, and was deposited in five periods: Middle Eocene to Early Oligocene, Late Oligocene to Early Miocene, late Early Miocene to Middle Miocene, Middle Miocene to Late Miocene and Pliocene to Recent (Baumann, 1982).
3.1 INTRODUCTION

In this chapter, a review of the stratigraphy of the Middle Tertiary to Quaternary sedimentary sequences in East Java is presented. The discussion is based on previous investigations in this area combined with some field observations and laboratory analyses. In this chapter, the discussion is confined to the lithofacies variation and relationship between lithofacies. The results of biostratigraphic, climatostratigraphic and magnetostratigraphic studies by previous investigators are also discussed to evaluate the lithofacies analysis in a chronological context. The sedimentological aspects of the Middle Tertiary to Quaternary strata are discussed in Chapters Four and Five. Chapters Three, Four and Five provide a geological framework for understanding the geology of Madura Strait and the Java Sea which will be discussed in Chapters Six and Seven respectively.

3.2 LITHOSTRATIGRAPHY

The depositional environments of formations in the Rembang and Kendeng Zones range widely from open marine to tidal flat, with fluvial environments also developed in the Kendeng Zone. The Rembang Zone sequence, particularly during the Middle Miocene, was strongly influenced by continentally derived sediment. In contrast, the Kendeng Zone succession right up to the Quaternary has been influenced by volcaniclastic detritus. Volcanic activity has been present in the vicinity of the Kendeng Zone since the Middle Miocene (van Bemmelen, 1949; de Genevraye & Samuel, 1972). The lithostratigraphic and lateral facies variations of these formations are described in the following sections.

Detailed geological mapping of East Java was carried out mostly by the Geological Survey of the Netherlands Indies in the 1930s. In the Rembang Zone where petroleum is abundant, geological studies were also carried out by the Bataafsche Petroleum Maatschappij (a pre-1940s Dutch petroleum company). Two different stratigraphic subdivisions evolved, particularly for the Rembang Zone. The stratigraphic subdivision by Trooster (1937; Table 3.1) was commonly used by investigators from the petroleum industries, while the subdivision
in Table 3.2 by van Bemmelen (1949) was formally used by the Geological Survey of the Netherlands Indies. As these stratigraphic subdivisions came before formal stratigraphic nomenclature was established, the terminology used is mixed between lithofacies characters and geographic terms.

Pringgoprawiro (1983) proposed a revision of the terminology used by previous authors to follow the formal stratigraphic nomenclature (Table 3.3). Pringgoprawiro's (1983) proposal was basically a renaming of the earlier terminology. The ages of the lithological formations were redefined and fine-tuned based on his biostratigraphic studies and those of other workers, but no significant re-organisation of lithofacies was proposed. However, there is still a tendency for individual workers to establish their own stratigraphic subdivisions without supporting evidence (e.g. Ardhana, 1993).

The lithostratigraphic discussion presented in this chapter follows the stratigraphic subdivision in Table 3.4. This stratigraphic subdivision has been modified from Pringgoprawiro (1983; see Table 3.3 and Fig. 3.1) with some modifications based on the biostratigraphic studies by Bolli (1966) and Blow (1969) for the Bojonegoro-1 well, by van Gorsel and Troelstra (1981) in the Solo River section, and by the author in the area north of Kabuh. These modifications are particularly concerned with the timing of the Tawun, Atasangin and Sonde Formations. The validity of some formations will be discussed based on the results from the present biostratigraphic and sedimentological studies by the author. It has been a tradition to separate the stratigraphic discussion for the Rembang and Kendeng Zones. Discussion presented in Chapter Two has indicated the presence of different tectonic and basement behaviour for these two zones as well as different sedimentary sources.

During the Middle Miocene to Quaternary seven major lithostratigraphic units developed in the Rembang Zone: the Tawun, Bulu, Wonocolo, Ledok, Mundu, Paciran and Lidah Formations. Nine lithostratigraphic units have been recognised in the Kendeng Zone during the same and younger time intervals, these are: the Kerek, Banyak, Kalibeng, Atasangin, Sonde, Lidah, Pucangan, Kabuh and Notopuro Formations. The Lidah Formation in the Kendeng and Rembang Zones is equivalent, since it was deposited as the result of regional base levelling at the end of the Pliocene.
3.2.1 Rembang Zone

a. Tawun Formation

Sediments of the Tawun Formation were deposited during the Middle Miocene, and the stratotype is in Tawun-5 well (Bolli, 1966; Lemigas/BEICIP, 1969; Pringgoprawiro, 1983). This formation is widely distributed along the Rembang Zone (Fig. 3.1) and Madura Island, mostly cropping out in the cores of anticlines. The formation may reach 1500 m in thickness and is typified by brownish grey mudstone with interbeds of calcareous sandstone. The upper part of the formation is organic-rich and often contains thin beds of lignite. This part is also associated with a mappable quartz sandstone (Ngrayong Member) and relatively thin (up to 1 m) orbitoidal (larger foraminifers) limestone beds. The Ngrayong Member has been the main petroleum reservoir rock in the Rembang Zone.

The age of this formation was determined mainly on the basis of species of Orbitoididae which are abundant particularly in the upper part. These species are: *Lepidocyclina atuberculata, Lepidocyclina ephippioides, Lepidocyclina sumatrensis, Lepidocyclina nipponica, Miogypsinoides bantamensis* and *Cycloclypeus* spp. (Pringgoprawiro, 1983). Based on the planktonic foraminiferal study by Bolli (1966) for the Bojonegoro-1 well, Blow (1969) indicated that the Tawun Formation is underlain by a surface of unconformity (early Middle Miocene unconformity). The planktonic foraminiferal zone N10, and part of zones N9 and N11, are missing in this well.

In the past the environment of deposition of the Tawun Formation was interpreted as paralic (van Bemmelen, 1949; Lemigas/BEICIP, 1969; Pringgoprawiro, 1983). A recent study by Ardhana (1993) indicated that the depositional facies varied laterally, particularly in the upper part, from paralic to hemipelagic and was strongly controlled by the palaeogeography. In the areas north of Cepu and Bojonegoro (Fig. 3.1) the facies are mostly paralic. Hemipelagic facies occur in areas such as south of Cepu and Bojonegoro. The present seismic stratigraphic study (Chapter 4) indicates that the upper part of the Tawun Formation formed under regressive conditions.

b. Bulu Formation

Sediments of the Bulu Formation were deposited on the paralic facies of the Tawun Formation in the late Middle Miocene. It occurs mainly between Cepu and Rembang (Fig. 3.1) and consists predominantly of reefal limestone with some interbeds of quartz sandstone
and calcarenitic limestone with abundant fragments of benthonic foraminifers and red algae. The age determination was based on the presence of the benthonic foraminiferal (Orbitoididae) species: *Lepidocyclina angulosa*, *Lepidocyclina sumatrensis*, *Cycloclypeus annulatus*, *Cycloclypeus indopacificus*, *Lepidocyclina* spp. which are associated with a middle neritic environment (Pringgoprawiro, 1983). This formation attains a thickness of 250 m.

The nature of contact of this formation with the underlying Tawun Formation is debatable, mainly because of lack of palaeontological controls near the boundary of the formations. Lemigas/BEICIP (1969) reported that local erosion occurred at the top of the Tawun Formation. In contrast, based on field observations in the Bulu area, Pringgoprawiro (1983) argued that the Bulu and Tawun Formations are conformable. The present seismic stratigraphic study (Chapter 4) suggests that this formation was only locally developed (on the high area between Bulu and Rembang), during a transgression following the Tawun Formation deposition, and the subsequent highstand of sea level.

c. **Wonocolo Formation**

The Wonocolo Formation was deposited during late Middle Miocene to early Late Miocene. The type locality of this formation is in the Wonocolo area, some 15 km northeast of Cepu (Fig. 3.1; Pringgoprawiro, 1983). It is characterised primarily by unstratified *Globigerina*-rich marls and clayey marls with rare intercalations of calcarenite. The Wonocolo Formation is characterised by a 20 to 80% planktonic/benthonic foraminiferal ratio (Pringgoprawiro, 1983), which indicates an inner to outer shelf depositional environment (Murray, 1991). Most of this formation developed during a highstand of sea level following sea level fall in the late Middle Miocene (see Chapter 4). The Wonocolo Formation attains a thickness exceeding 300 m and shows a coarsening-upward sequence into the overlying Ledok Formation.

In the areas between Cepu and Rembang this formation conformably overlies or is laterally equivalent to the Bulu Formation. The Wonocolo Formation in northeastern Java (Tuban and Paciran areas, Fig. 3.1 & 3.2) does not occur and Pliocene reefal limestones (Paciran Formation) rest directly on the Middle Miocene Tawun Formation (north of Bojonegoro) and the Early Miocene Kujung, Prupuh and Tuban Formations (in the Tuban and Paciran areas). This hiatus indicates the presence of a topographic high during the Late Miocene in these areas.
d. Ledok Formation

The Ledok Formation conformably overlies the Wonocolo Formation. In the area northeast of Cepu, the transition between the formations is characterised by an increasing frequency of calcarenite intercalations within the marl. This upward trend is interpreted to indicate a shallowing environment of deposition, closer to a high energy environment. In the area between Bojonegoro and Blora, the upper part of the formation developed as a tidally-influenced deposit, as indicated by abundant large-scale cross-beds of probably tidal sand waves. The planktonic foraminiferal species *Globorotalia plesiotumida* is common in the Ledok Formation and indicates Late Miocene age (Pringgoprawiro, 1983).

In the core of an anticline north of Ngimbang (Fig. 3.1) the equivalent Ledok Formation is characterised by unbedded marl and marly clay. In the Java Sea, the equivalent unit to the Ledok Formation is characterised by reefal limestone (Najoan, 1972). In contrast, the Ledok Formation as well as the Wonocolo Formation was not developed in the Tuban and Paciran areas (Duyfjes, 1938a) suggesting that topography remained high during the deposition of both formations.

Pringgoprawiro (1983) found that the planktonic/benthonic foraminiferal ratio decreases upwards through the formation (from 47% to 30%) which also suggests a shallowing environment of deposition. The lower part of the formation is typified by benthonic foraminifers: *Eponides praecintus*, *Nodogenerina scalaris*, *Uvigerina peregrina*, *Uvigerina flintii*, *Bulimina striata*, *Laticarinina pauverta*, *Robulus orbicularis*, *Epistoma bradyi*, *Dentalina neugobarina*, *Nodosaria* spp., *Elphidium* spp., *Bulimina marginata*. This faunal association indicates an inner shelf environment of deposition (Murray, 1991).

e. Mundu and Paciran Formations

The type locality of the Mundu Formation is in the Mundu area, 10 km west of Cepu (Fig. 3.1). It was deposited conformably on the Ledok Formation during the Early to Middle Pliocene (van Bemmelen, 1949; Lemigas/BEICIP, 1969; Pringgoprawiro, 1983). This formation is composed of unstratified yellowish-white to greenish-grey marl. In some intervals the Mundu Formation shows a sandy character due to a high content of foraminiferal tests. The formation is also typified by an association of outer neritic benthonic foraminifers. The original thickness of the Mundu Formation is uncertain due to post-Mundu erosion but 10 km north of Cepu, it attains a thickness of 340 m (Lemigas/BEICIP, 1969).
The Mundu Formation occurs widely in the East Java Basin. Major facies changes to reefal and calcarenitic limestone occur mainly along topographically high areas such as the Tuban area and on Madura Island. In the Tuban area (Figs 3.1 & 3.2) the limestone is termed the Paciran Formation (Pringgoprawiro, 1983) or the Karren Limestone (van Bemmelen, 1949), and on Madura Island is termed the Madura Formation (Situmorang et al, 1992; Aziz et al., 1993). Along the southern coast of Madura Island the equivalent unit to the Mundu Formation ranges from marls to calcarenite with common bioclasts of foraminiferal tests and algal fragments.

The Mundu Formation in the area between Bojonegoro and Pati exhibits a regressive sequence upward from outer neritic to inner neritic. The upper part of the Mundu Formation is composed of coquina intercalated with marl (Aminuddin et al., 1981). The bioclasts are commonly micritised smaller foraminifers and algal fragments. The unit is also characterised by the presence of interstitial glauconite embedded within foraminiferal tests. Pringgoprawiro (1983) assigned this limestone to the Selorejo Member of the Mundu Formation. The Selorejo Member thickens gradually from only few metres in the Bojonegoro area, to approximately 300 m in the southern Rembang area.

The Selorejo Member is also characterised by typical inner neritic benthonic foraminifers including the genera Nonion, Quinqueloculina and Ammonia/Rotalia (Phleger, 1960; Murray, 1991). Aminuddin et al. (1981) concluded that the Selorejo Formation was deposited during the Late Pliocene based on the presence of the planktonic foraminifers Globorotalia tosaensis, Globorotalia acostaensis, Globigerinoides trilobus, Globigerinoides sacculifer, Globigerinoides immaturus, Pulleniatina praecursor, Pulleniatina obliquiloculata and Orbulina universa.

f. Lidah Formation

The type locality of the Late Pliocene-Pleistocene Lidah Formation is in the Lidah Anticline, 10 km southwest of Surabaya (Fig. 3.1; Pringgoprawiro, 1983). In the Kendeng Zone the Lidah Formation was known as the clay facies of the Pucangan Formation (Duyfjes, 1938a,b,c,d). In the Rembang Zone it was called the Mergel Ton by the Bataafsche Petroleum Maatschappij (Table 3.1.) or the Blue Clays by van Bemmelen (1949; Table 3.2.).
The Lidah Formation is widely developed in the eastern part of the Kendeng Zone and the southern and western parts of the Rembang Zone (Fig. 3.1). It consists primarily of unstratified bluish-grey mudstone of lacustrine/tidal flat origin. The formation thickness varies between 300 and 550 m. In the eastern part of Kendeng Zone, this variation is particularly marked because of a lateral facies change between the upper part of the Lidah Formation and the volcanic facies of the lower part of the Pucangan Formation.

The field study indicated that the lower contact with the underlying Atasangin Formation, as observed near Kabuh and Sumberringin north of Jombang (Kendeng Zone), is conformable as shown by a gradual change from marl to calcareous mudstone and bluish grey mudstone. In contrast, in the Rembang Zone the Lidah Formation was deposited on the eroded Mundu Formation (Lemigas/BEICIP, 1969), which indicates a later development of the Lidah Formation than in the Kendeng Zone. A similar erosional contact is observed along the south coast of Madura Island, where the mudstone Lidah Formation rests directly on the tidal sand wave deposit Mundu Formation equivalent (Pasean Formation).

In areas southeast of Cepu and north of Bojonegoro the lower part of the formation is carbonaceous and pyritic. The middle part consists of glauconitic calcarenite with thin interbeds of marl which were called the Malo Member by Trooster (1937) and Lemigas/BEICIP (1969). The upper part is dominated by calcareous mudstone and marl with interbeds of coquina up to 2 m thick. The Lidah Formation can also be found in some places along the south coast of Madura Island (east of Kamal and south of Pamekasan, Fig. 3.2) where it lies on calcareous muddy sandstone of the Mundu Formation equivalent.

Field observations in the Sumberringin (Kedungwaru and Pucangan Anticlines) and Kedamean (Guyangan Anticline) areas reveal two prominent sandy intervals of up to 27 m in thickness. These intervals resemble a tidally influenced unit and contain marine molluscs (Gastropods, Echinoides and Arthropods [Balanus]; Duyfjes, 1938b,d), as well as benthonic foraminifers which are diagnostic of an inner neritic environment (offshore facies). These indicate that there were two marine incursions during the deposition of the Lidah Formation in this area.
3.2.2 Kendeng Zone

a. Kerek Formation

The Kerek Formation was deposited during the Middle to early Late Miocene. The type locality is in Kerek Village (8 km north of Ngawi) along the Solo River (de Genevraye & Samuel, 1972). This formation crops out only in the cores of anticlines in the Kendeng Zone (Fig. 3.1) with a thickness attaining 1000 m. The main components are medium-grained calcareous sandstone alternating with marl. Benthonic foraminifers such as *Eggerella bradyi*, *Melonis soldanii* and *Globobulimina pupoides* found within this formation (Van Gorsel & Troelstra, 1981) suggest an outer neritic environment.

Texturally the Kerek Formation shows a turbiditic character and van Bemmelen (1949) referred to it as a flysch deposit, whereas Muin (1985) regarded the formation as a submarine fan deposit. The abundance of volcanogenic coarse clastic fragments in this formation suggests a close association with the Middle Miocene volcanism [Banyak Volcano, van Bemmelen (1949)], which was present along the place presently occupied by Quaternary volcanoes. A sedimentological reinterpretation during the present study suggests that the Kerek Formation may not be associated with a submarine fan system. The palaeogeographic setting indicates that the formation was probably deposited on the apron of volcanoes. In addition, benthonic faunal analysis by Lemigas/BEICIP (1974) indicated deposition in an inner-middle neritic environment which is seldom characteristic of a submarine fan system (Walker, 1992). It is suspected that the turbiditic character of this formation may have resulted from storm-induced currents or been directly related to the volcanic activity. Similar settings to the latter have been reported by Busby-Spera (1988) in Baja, California (Mesozoic deposit) and Hathway (1994) in southwestern Viti Levu, Fiji (Oligocene-Miocene deposits). Towards the top of the Kerek Formation, the environment of deposition was more commonly inner neritic, as indicated by the occurrence of storm deposits.

b. Kalibeng Formation

The Late Miocene to Early Pliocene Kalibeng Formation is characterised by greenish or greyish white unstratified foraminiferal-rich marl. Bedding is always difficult to observe except where intercalations of calcarenitic sandstone are present. The type locality of the Kalibeng Formation is along the Kalibeng River, 15 km northwest of Jombang, East Java (Fig. 3.1). The Kalibeng Formation was formerly termed the Kalibeng Beds by Duyfjes (1938b) which were subdivided into the Lower (unstratified marl) and Upper (limestone and marl)
Kalibeng Beds. De Genevraye and Samuel (1972) renamed them the Lower and Upper Kalibeng Formations (note the use of term "Formation"). Based on the age and lithofacies differences, Pringgoprawiro (1983) proposed that the term Kalibeng Formation be retained for the Lower Kalibeng Formation and the new Sonde Formation be used for the Upper Kalibeng Formation.

Along the Solo river to the north of Ngawi, some 500 m of Kalibeng Formation occurs which, as in the type locality of this formation, is mainly composed of marl. A few thin intercalations of calcarenite and calcareous fine sandstone are present. In this section the Kalibeng Formation is also very rich in well preserved foraminifers. Benthonic foraminifers are all open marine species, such as *Oridorsalis umbonatus*, *Gyroidina neosoldanii*, *Planulina wuellerstorfi*, *Anomalina globulosa* and *Uvigerina auberiana* (van Gorsel & Troelstra, 1981).

In the Atasangin area (some 40 km east of Ngawi; Fig. 3.1), up to 200 m of volcaniclastic deposits occur in the upper part of the Kalibeng Formation. Pringgoprawiro (1983) suggested that this represented a new member of the Kalibeng Formation, and he called it the Atasangin Member.

The stratigraphic relationship between the Kalibeng Formation and the adjacent formations is difficult to define due to lack of bedding structures within the Kalibeng Formation. De Genevraye and Samuel (1972) and Pringgoprawiro (1983) considered that the Kalibeng Formation has a conformable relationship with both the Kerek and Sonde Formations. Their suggestion was based on the similarity in bedding orientation and faunal content in the Kalibeng, Kerek and Sonde Formations.

c. Sonde and Atasangin Formations

The term Sonde Formation was proposed by Pringgoprawiro (1983) to replace the term Upper Kalibeng Beds of Duyfjes (1938a,b) and van Bemmelen (1949), and the Upper Kalibeng Formation of de Genevraye and Samuel (1972). This formation was deposited in the Kendeng Zone during the Late Pliocene. The type locality of the Sonde Formation is in the Sonde area, along the Solo River, 15 km west of Ngawi (Fig. 3.1.). It is characterised mainly by coarse bioclastic limestone in the lower part and marl in the upper part. The bioclastic limestone has been called the Klitik Member of the Sonde Formation (Pringgoprawiro, 1983).
About 65 m of limestone in the Sonde Formation occurs in the Solo River section north of Ngawi. It shows a bedded calcarenite at the bottom and reefal limestone at the top. Van Gorsel and Troelstra (1981) considered the Sonde Formation as a regressive unit. Their benthonic foraminiferal analysis of the formation revealed an assemblage of middle shelf faunas at the base, and inner shelf faunas in the middle part of the formation. The upper part of the formation consists of about 80 m of sandy marl. The Sonde Formation in the Solo River section is unconformably overlain by 120 m of debris flow deposits of the Pucangan Formation.

In the eastern part of the Kendeng Zone this formation was developed in a completely different setting than in the type locality. The present study by the author (Chapter 3.3, and Chapter 5.4) suggests that in this area the Sonde Formation was developed in a relatively isolated basin. It can be divided into two main lithofacies, a sandstone facies (about 100 m thick) in the lower part and a marly facies (about 300 m thick) in the upper part. The sandstone facies consists primarily of prograding deltaic strata which rest conformably on the Kalibeng Formation. The marly facies consists of finely laminated diatomaceous marl with a few thin interbeds (15 cm) of fine sandstone. Based on the occurrence of benthonic foraminifers and facies characters, the marly facies is thought to have been deposited in an marginal marine environment (Chapter 5.4). The planktonic foraminiferal evidence suggests that the depositional timing of these facies is slightly earlier (Middle Pliocene, Chapter 3.3) than in the type locality (Late Pliocene).

The stratigraphic position of the sandstone facies is similar to the volcaniclastic unit of the Kalibeng Formation (Atasangin Member) found by Pringgoprawiro (1983) in the Atasangin area (some 40 km east of Ngawi; Fig. 3.1). Petrographical evidence suggests that the sandstone facies are volcanogenically derived (Chapter 5.4), and probably of similar origin to the Atasangin Member of the Kalibeng Formation. The marl facies has no equivalent in the type locality of the Sonde Formation. These results necessitate a reexamination of the validity of the Sonde Formation in the eastern part of the Kendeng Zone. In the present study the author proposes the term Atasangin Formation due to the differences in texture and age from the Sonde Formation in its type locality, and partly due to the facies similarity with the Atasangin Member of the Kalibeng Formation. The Atasangin Formation includes the Atasangin Member of Pringgoprawiro (1983).
d. **Pucangan Formation**

The type locality of the Pucangan Formation is on the south flank of the Pucangan Anticline (18 km north of Jombang; Fig. 3.1). It was proposed by Duyfjes (1938a,b,c,d) to consist of two main facies: bluish-grey mudstones of paludal origin in the lower part and volcanic facies in the upper part. As discussed earlier (Chapter 3.2.1f), the lower part is considered to be the Lidah Formation and the upper part remains as the Pucangan Formation (Pringgoprawiro, 1983).

The Pucangan Formation developed mainly in the Kendeng Zone during the Early Pleistocene. This formation consists of volcanic deposits and volcanic-derived sediment deposited in two different settings: tidally-influenced and alluvial fan/fluvial environments. According to Duyfjes (1938b,d) all of these volcanic materials were derived from the Wilis Volcano which was the only active volcano during the Early Pleistocene. A hominid skull fossil and a number of vertebrate fossils have been found within this formation, particularly in the conglomeratic facies related to the alluvial fan and fluvial systems (Duyfjes, 1938d).

e. **Kabuh Formation**

The term Kabuh Beds was used by Duyfjes (1938b,d) for the lithologic unit that conformably overlies the Pucangan Formation. Pringgoprawiro (1983) termed this unit the Kabuh Formation. It is only developed along the Kendeng Zone and was deposited during the Middle Pleistocene. It comprises two main lithofacies: marine mudstone and sandstone facies in the lower part and volcanic facies in the upper part.

The base of the widespread mudstone facies marks the boundary between this formation and the underlying Pucangan Formation (Duyfjes, 1938d). The present study (Chapter 5.8) indicates that this facies conformably overlies the fluvial facies of the Pucangan Formation and represents a shelf mud facies which marked a significant relative sea level rise in East Java during the Middle Pleistocene. The marine facies of the Kabuh Formation is well exposed on the flanks of anticlines north of Mojokerto with a thickness of about 15 m. A similar facies is also well developed in an anticline west of Pasuruan (see Fig. 3.1) where the marine facies consists of green mudstone in the lower part and mollusc-rich muddy sandstone in the upper part.
In some areas (i.e. near Kabuh and Sumberringin, 18 km north of Kabuh) the marine mudstone is absent and the volcanic facies of the Kabuh Formation lies directly on the Pucangan Formation. The volcanic facies was deposited along the southern edge of the Kendeng Zone. It consists of conglomeratic cross-bedded sandstone in which the fragments are mostly andesitic. The volcanic facies is also characterised by the presence of fresh-water molluscs (Melania and Unio) and abundant vertebrate fossils (Duyfjes, 1938b,d).

f. Notopuro Formation

The Notopuro Formation (Pringgoprawiro, 1983; formerly the Jombang Beds of Duyfjes, 1938b) lies unconformably on the Kabuh Formation. It consists mainly of volcanic breccia, tuff and tuffaceous sandstone and was deposited in the proximity of Late Pleistocene volcanoes (i.e. Lawu, Wilis and Arjuno volcanoes). The unconformable relationship with the underlying formation resulted from Late Pleistocene folding which also produced most of the anticlinorium in the Kendeng Zone (Duyfjes, 1938b; van Bemmelen, 1949).

3.3 BIOSTRATIGRAPHY

Biostratigraphic studies have been undertaken by several investigators on Miocene to Pleistocene strata in various places in East Java (e.g. Bolli, 1966; Pringgoprawiro & Baharuddin, 1979; van Gorsel & Troelstra, 1981; Pringgoprawiro, 1983; Hasjim, 1987). Most of these studies were based on planktonic foraminifers which are abundant in these strata.

Two commonly used planktonic foraminiferal zonations are the one by Bolli (1970) and Bolli and Premoli-Silva (1973), and the other by Banner and Blow (1967) and Blow (1969). Both zonations are similar, based on the first and/or last appearances of species. However, the zonation of Blow (1969) places greater emphasis on changes within evolutionary lineages, and zonal boundaries are assigned on the critical level within lineages. Bolli (1970) and Bolli and Premoli-Silva (1973) divided the Neogene into 19 major zones, with further subdivision carried out in the Early and Middle Pliocene, and Pleistocene (Fig. 3.3a). Banner and Blow (1967) and Blow (1969) subdivided the Neogene into 20 zones (Fig. 3.3b), and numerically labelled from N3 (Late Oligocene) to N23 (Late Pleistocene). The use of this numerical identification, in some respects, is more convenient than taxonomical identification as practiced by Bolli (1970) and Bolli and Premoli-Silva (1973) since stratigraphical order is more easily remembered. This is probably the reason for the wide use of the Blow (1969) zonation in the
Indonesian region. Until recently, no significant revisions have been made to these zonations, but attempts to numerically date the zonal boundaries have been carried out by Berggren (1972; 1973) and Berggren et al. (1985), particularly using integrated data of deep sea cores.

The Miocene/Pliocene boundary has been variously placed at the base (Berggren et al., 1985), middle (Banner & Blow, 1967) or top (Blow, 1969) of zone N18. The numeric age for the base of zone N18 is 5.2 Ma (Berggren et al., 1985; Harland et al., 1989). The top of zone N18 is the evolutionary level of Sphaeroidinellopsis subdehiscens to Sphaeroidinella dehiscens dehiscens, dated 5.0 Ma (Berggren et al., 1985). The Pliocene/Pleistocene boundary is coincident with the boundary of zones N21/N22, which is at the evolutionary level of species Globorotalia tosaensis to Globorotalia truncatulinoides (Banner & Blow, 1967). Berggren et al. (1985) and Harland et al. (1989) provided an age of 1.64 Ma for this boundary.

3.3.1 Miocene and Middle Pliocene strata

The Miocene and Pliocene strata in the East Java Basin are rich in foraminifers. The first biostratigraphical study carried out in the East Java area was by Bolli (1966) for the Bojonegoro-1 well (Fig. 3.2). It was followed by a study by Pringgoprawiro and Baharuddin (1979) for the TO-5 and TO-8 wells in the Tobo area 15 km south of Bojonegoro. Van Gorsel and Troelstra (1981) carried out detailed biostratigraphic and climatostratigraphic studies in the Solo River section north of Ngawi (Fig. 3.2). Biostratigraphic studies on some surface sections in the Rembang Zone have also been carried out by Pringgoprawiro (1983) and Hasjim (1987). The discussion below summarises the results of their investigations and estimates the timing of lithostratigraphic development. The discussion below also refers to the planktonic foraminiferal zonation developed by Blow (1969).

a. Solo River section

Biostratigraphy of the Solo River section, north of Ngawi, has been studied by Lemigas/BEICIP (1969) and van Gorsel and Troelstra (1981). These studies were particularly concerned with the Kalibeng Formation and the underlying Kerek Formation. Figure 3.4 summarises the biostratigraphic and climatostratigraphic studies in this section by van Gorsel and Troelstra (1981).

The zonal boundaries erected in both studies have been defined by the first appearance of planktonic foraminiferal species and by the first appearance of the succeeding species. Van
Gorsel and Troelstra (1981) have not indicated whether these appearances are evolutionary or migratory. However, they suggested that the species represent a reliable record of the changes in planktonic faunas at the time of deposition. This suggestion was based on the supposition that the Kerek and Kalibeng Formations were deposited in an open marine environment with no indication of transported or reworked faunas.

The Miocene-Pliocene boundary (N18/N19 boundary) in the section is about 150 m above the base of the Kalibeng Formation (Fig. 3.4). This boundary was determined by van Gorsel and Troelstra (1981) using the first appearance of *Sphaeroidinella dehiscens immatura*. The Pliocene-Pleistocene boundary (N21/N22 boundary) is determined by the first appearance of *Globorotalia truncatulinoides* and falls in the lower part of the Sonde Formation.

b. Puncakwangi and Kali Sampurna sections

Figures 3.5 and 3.6 summarise the biostratigraphic studies from the Puncakwangi and Kali Sampurna sections respectively by Pringgoprawiro (1983). He used the same species as van Gorsel and Troelstra (1981) to determine the Miocene-Pliocene and Pliocene-Pleistocene boundaries.

On these sections the Miocene-Pliocene boundary has been defined at the boundary between the Ledok and Mundu Formations which coincides with the first appearance of the planktonic foraminiferal species *Sphaeroidinella dehiscens immatura*. The first appearance of *Globorotalia truncatulinoides*, which marks the Pliocene-Pleistocene boundary, occurs at the boundary between the Mundu and Lidah Formations.

c. North Kabuh section

A biostratigraphical analysis of the north Kabuh section was also undertaken in the present study (Fig. 3.2). This section consists of 400 m of sediment represented mostly by the Atasangin Formation. The lower part of the section consists of the deltaic deposits of the Atasangin Formation underlain by the Kalibeng Formation. Figure 3.7 summarises the biostratigraphic analysis of this formation based on planktonic foraminifers in the 12 samples analysed. The results of the analysis should not be regarded as very accurate because of two major problems. These are: (1) that the analysis was based on a relatively wide sampling interval, and (2) the lithofacies, which was deposited in a marginal marine environment (see discussion in Chapter 5.4), is characterised by a low faunal diversity.
The analysis shows that the whole section is Pliocene in age. Deposition of the Atasangin Formation apparently commenced in the early N20 zone as indicated by foraminiferal species contained in the uppermost Kalibeng Formation in this area. The N20/N21 boundary in this section is determined by lack of *Pulleniatina primalis* and *Pulleniatina praecursor* in the samples from the upper part of the formation. This boundary is about 50 m below the suspected boundary between the Atasangin and Lidah Formations.

### 3.3.2 Late Pliocene and Pleistocene strata (Lidah Formation)

Since the discovery of vertebrate and hominid fossils in the Pucangan, Kabuh and Notopuro Formations in Central and East Java by Dubois in 1891 and von Koenigswald in 1936, as reported by Buijst (1936), and subsequent workers (Sartono, 1961; Jacob, 1978), the Quaternary sediments in the Kendeng Zone have been the subject of various stratigraphic studies.

In the eastern Kendeng Zone Ninkovich and Burckle (1978) analysed and compared the diatom assemblages studied by Reinhold (1937) with the diatom assemblages from the palaeomagnetically-dated deep sea core V28-179 located in the central Pacific (Shackleton & Opdyke, 1977). The Reinhold (1937) samples were taken from the upper part of the Atasangin Formation, the marine sandstone intercalated within the mudstone of the Lidah Formation and from the transitional strata between these formations (Fig. 3.8). Ninkovich and Burckle (1978) have arrived at an age of around 2.1 Ma for the base of the Lidah Formation. This result is in close agreement with the present biostratigraphic study on the north Kabuh section (Fig. 3.7). The latter study found that the boundary between the Atasangin and Lidah Formations lies about 50 m above the N20/N21 boundary. According to Berggren *et al.* (1985) the age of the N20/N21 boundary is about 3 Ma. Sartono *et al.* (1981) also arrived at a similar result. They carried out micropalaeontological analysis on the Lidah Formation in Perning area (10 km north of Mojokerto, Fig. 3.1). The zone N21 age for the Lidah Formation was based on the planktonic foraminiferal species *Globigerinoides extremus, Globorotalia cultrata exilis* and *Globorotalia tosaensis*.

### 3.4 CLIMATOSTRATIGRAPHY

A climatostratigraphic study was carried out by van Gorsel and Troelstra (1981) on the Kalibeng Formation in the Solo River section. Their study was based on the morphological
characters, coiling behaviour and abundance of some planktonic foraminiferal species which are sensitive to climatic change.

A morphological character change in response to varying climatic conditions has been shown by species *Orbulina*. Based on samples from surficial sediments and Indian Ocean waters, Be et al. (1973) has demonstrated that the diameter of the globular form of *Orbulina* is inversely correlated with latitudinal occurrence. A faunal test diameter between 600 to 800 \( \mu \text{m} \) is common in the tropical and subtropical areas, and less than 450 \( \mu \text{m} \) diameter occurs in the subtropical-subpolar transition zone. The species *Globorotalia* (*Neogloboquadrina* *) pachyderma* is known to be a typical species in the polar and subtropical regions (Kennett & Srinivasan, 1983). This species according to Jenkins (1967), Kennett (1968) and Bandy (1972) is sinistrally coiled in polar and subpolar regions, and dextrally coiled towards lower latitude. These indices have been used by van Gorsel and Troelstra (1981) to reconstruct the climatic events in the Solo River section. Cold events are indicated by a significant drop in *Orbulina* diameter and abundance increase of sinistrally coiled *Globorotalia pachyderma*. Additionally, van Gorsel and Troelstra (1981) also used the relative abundance of some other species which are typical in tropical regions (*Globorotalia menardii, Globigerinoides trilobus, Globorotalia altisspira, Pulleniatina* spp.) and subpolar regions (*Globigerina bulloides*) regions.

Van Gorsel and Troelstra (1981) revealed the presence of five climatic zones (Fig. 3.4). Zones I, III and V are cold periods and have been interpreted to be associated with sea level lowering during the early Late Miocene, end of Late Miocene and Pliocene. The base of zone V coincides with the transition from the Kalibeng Formation to the Sonde Formation. Based on the correlation with planktonic foraminiferal datums determined from several deep sea cores from the equatorial Pacific, the numerical ages for the tops of zones I and III are 5.8-6.1 Ma and 5.0-4.8 Ma respectively, and for the base of zone V is 2.6-2.8 Ma, according to van Gorsel and Troelstra (1981).

### 3.5 MAGNETOSTRATIGRAPHY

In the Perning area (10 km north of Mojokerto, Fig. 3.1), palaeomagnetic investigations were carried out in 1986-1988 by a Japanese and Indonesian joint research team (Hyodo et al., 1992). This investigation focussed on the Late Pliocene and Pleistocene strata, with the primary aim of dating the hominid fossil bearing strata.
Magnetostratigraphy is based on variations in remanent magnetisation within a sequence of strata, in response to polarity reversals of the geomagnetic field. The present normal polarity of the Earth’s geomagnetic field (Brunhes Epoch) has persisted since 0.78 Ma (Spell & McDougall, 1992), a revision of the value of 0.73 Ma of Berggren et al. (1985). A reverse epoch (Matuyama) occurred between 2.47 and 0.78 Ma (Berggren et al., 1985). There are three short geomagnetic events of normal polarity within this Matuyama Epoch, which are known as the Reunion (2.04 Ma), Olduvai (1.88-1.72 Ma) and Jaramillo (0.94-0.88 Ma) events (Berggren et al., 1985).

The investigations by Hyodo et al. (1992) in the Perning area, obtained reversed or intermediate magnetisations in most horizons of the Pucangan and Kabuh Formations. Two zones dominated by easterly magnetisation directions have been recognised and correlated to the Olduvai and Jaramillo events for the lower and upper zones respectively. The upper zone coincides with the sediment layer that contains hominid fossils. The Brunhes/Matuyama boundary is situated near the boundary between the Pucangan and Kabuh Formations. The interpretation of Hyodo et al. (1992) was based on the similarity of the geomagnetic declination swing pattern to that in the previously defined Quaternary succession in Sangiran, Central Java (some 20 km north of Solo; Fig. 3.2).

The results of the palaeomagnetic study differ from the results of K-Ar dating by Jacob and Curtis (1971) in the same area (Perning, Mojokerto). The radiometric dating was carried out on a tuff which lies just below the hominid layer and gave a date of 1.9 ± 0.4 Ma. Hyodo et al. (1992) found that the tuff deposit near the hominid layer is secondary and this was confirmed by the section measured during this study. Consequently, the radiometric dating appears to indicate the age of the primary tuff which is older than the Jaramillo event identified by palaeomagnetic measurements.

3.6 SUMMARY

Seven main lithostratigraphic units occur in the Rembang Zone and range in age from Middle Miocene to Late Quaternary. These include: the Tawun, Bulu, Wonocolo, Ledok, Mundu, Paciran and Lidah Formations. Nine lithostratigraphic units have been recognised in the Kendeng Zone during the same and younger time intervals, these are: the Kerek, Banyak, Kalibeng, Atasangin, Sonde, Lidah, Pucangan, Kabuh and Notopuro Formations. The
biostratigraphic studies have allowed an accurate lithofacies correlation between the Rembang and Kendeng Zones. Figures 3.9 and 3.10 provide the north-south lateral correlation between these zones.

The sedimentary origin of the Rembang and Kendeng Zones is different. This is particularly obvious in the Middle Miocene deposits of both zones. The paralic and shallow marine Tawun and Bulu Formations contain continentally derived clastics, such as quartz sandstone, while the Kerek and Banyak Formations were sourced from volcanism along the south of the Kendeng Zone. The Late Miocene and Pliocene deposits in both the Rembang and Kendeng Zones (Wonocolo, Ledok, Mundu, Paciran, Kalibeng, Atasangin and Sonde Formations) are highly calcareous (marly) which suggests deposition mainly in an open marine environment during a period of reduced terrigenous supply.

The widespread lacustrine/tidal flat mudstone of the Lidah Formation marked the end of Pliocene deposition in the Kendeng Zone, but the deposition of this mudstone continued during the Pleistocene in the Rembang Zone, while coarse clastic deposits (Pucangan, Kabuh and Notopuro Formations) associated with the Quaternary volcanism were developing in the Kendeng Zone.
CHAPTER FOUR
SEISMIC STRATIGRAPHY OF THE REMBANG ZONE

4.1 INTRODUCTION

The Rembang Zone is known as the petroleum-rich province of East Java. Intensive geological studies since the end of the 19th century have been undertaken principally by petroleum companies. Published geological reviews were mostly concerned with the general geology of the area. Detailed micropalaeontological studies, however, have been carried out by Bolli (1966), Pringgoprawiro and Baharuddin (1979), Pringgoprawiro (1983) and Hasjim (1987). These studies were concerned with the construction of biostratigraphic zonations for the Middle Miocene to Pliocene strata. Other detailed sedimentological studies which concentrated on the Middle Miocene strata were carried out by Muin (1985) and Ardhana (1993).

The objectives of this chapter are: (1) to delineate the chronostratigraphic units on seismic profiles; (2) to infer the palaeoenvironment and relative sea level changes under which the units were deposited; and (3) to compare the seismic stratigraphy and facies with the stratigraphy and facies defined by previous authors. This study is based primarily on the recognition of sedimentary facies defined on the basis of a seismic stratigraphic study. Unfortunately, only two regional seismic lines were made available for the study, which practically eliminates possible inferences about the lateral facies distribution. These two seismic lines have been provided by PPT Migas (an Indonesian petroleum and gas educational institute) and Pertamina (the state oil and gas company); both lines are located on East Java (Fig. 4.1).

The seismic stratigraphic analysis was carried out by applying the seismic stratigraphic method proposed by Vail et al. (1977), Posamentier et al. (1988) and Posamentier and Vail (1988). The interpretation of environmental facies from seismic facies units follows the criteria of Sangree and Widmier (1977), supported by petrographic analysis on selected rock samples obtained during field observations, and by the benthonic faunal analysis of Pringgoprawiro (1983) and Aminuddin et al. (1981).
4.2 SEISMIC SEQUENCE AND FACIES DEFINITION

Vail et al. (1977) developed the concept of seismic stratigraphy as a method of analysing depositional systems which involves a genetic approach for interpreting seismic reflection profiles. Seismic stratigraphy assumes that seismic reflections derived from sedimentary layers are essentially isochronous, which permits a grouping of seismic reflections into packages of sequences based on their stratal pattern. Mitchum et al. (1977) defined a depositional sequence as "a stratigraphic unit composed of a relatively conformable succession of genetically related strata bounded by unconformities or their correlative conformities". Vail et al. (1984) recognised two major types of unconformities which they called type I and type II unconformities. A type I unconformity is formed during subaerial exposure of the shelf produced by a rate of eustatic fall that exceeds the subsidence rate. In contrast a type II unconformity is formed when the rate of eustatic fall is less than the rate of subsidence and no subaerial exposure occurs. Sequences are named according to their type of underlying unconformity.

Mitchum and Vail (1977) outlined three generalised steps in studying stratigraphy through seismic data. These are:

(1) recognition, correlation, and age determination of seismic sequences
(2) recognition, mapping, and interpretation of seismic facies, and
(3) regional analysis of relative changes of sea level.

Seismic sequences are defined by recognising surfaces of discontinuity from reflection terminations. Surfaces of discontinuity are recognised by interpreting systematic patterns of reflection termination along the surfaces (Fig. 4.2; Vail, 1987). Mitchum et al. (1977, p. 57-59) have categorised a set of relationships between seismic reflectors (summarised in Figure 4.3) and the following are their explanations for the terms used.

**Lapout** is the lateral termination of a stratum at its original depositional limit. This is a general term for lateral termination of strata at their depositional pinchout.

**Baselap** is lapout at the lower boundary of a depositional sequence. This term is used when the onlap pattern cannot be distinguished from downlap.

**Onlap** is a baselap in which an initially horizontal stratum laps out against an initially inclined surface, or in which an initially inclined stratum laps out updip against a surface of greater initial inclination.
**Downlap** is a baselap in which an initially inclined stratum terminates downdip against an initially horizontal or inclined surface. **Toplap** is lapout at the upper boundary of a depositional sequence. **Erosional truncation** is the lateral termination of a stratum by erosion. **Structural truncation** is the lateral termination of a stratum by structural disruption. **Hiatus** is the total interval of geologic time that is not represented by strata at a specific position along a stratigraphic surface. **Offlap** is the progressive offshore shingling of sedimentary units within a conformable sequence, in which each successively younger unit leaves exposed a portion of the older unit on which it lies.

In seismic sequence stratigraphy, onlap, downlap and toplap are evidence of non-depositional hiatus. However, this hiatus may not necessarily be identical with a true sedimentary hiatus because of the limited resolving power of the seismic method. Sheriff (1977; 1985) demonstrated that this resolving power decreases for beds thinner than about 1/4 wavelength. In the case where 25 Hz seismic frequency is used, a bed thinner than 20 m (v = 2000 m/sec) will not be represented. In multichannel seismic data, deeper sedimentary strata will suffer more from this limitation, because during processing, frequency filtering parameters are commonly depth dependent and lower frequencies are passed. These filtering parameters are chosen on the basis that higher frequencies are attenuated more rapidly with distance (and time) travelled. Thus the terms: onlap, downlap and toplap may also be interpreted as a thinning of strata below the seismic resolution.

In recent developments, the seismic stratigraphic method has been adapted to the sequence stratigraphic concept outlined by Posamentier *et al.* (1988) and Posamentier and Vail (1988). This development appeared be an attempt to overcome the criticisms of the proprietary nature of seismic data on which the seismic stratigraphic concept relies, and to apply the principles previously used on seismic sections to outcrop and well log sections.

In the new development, Posamentier *et al.* (1988) and Posamentier and Vail (1988) have introduced some new terms and the concept of a systems tract. Two commonly used terms are accommodation space and parasequence. Accommodation space is defined as "the space made available for potential sediment accumulation" (Jervey, 1988). The term parasequence was introduced by van Wagoner (1985), and is defined as "a relatively conformable succession of genetically related beds or bedsets bounded by marine flooding surfaces and their correlative surfaces". Although it was stressed as a descriptive term (Posamentier & James, 1993), the use of this term is rather ambiguous. There is a tendency to use the term as a part of the
sequence hierarchy (e.g. Swift et al., 1987; Boyd, 1994) similar to the hierarchy of supercycle, cycle and paracycle of Vail et al. (1977) which reflects different orders of duration of relative sea level change.

Analysis of seismic facies within a seismic sequence is carried out primarily to define depositional environments. The descriptions of seismic facies are based on their reflection patterns (Fig. 4.3), as classified by Mitchum et al. (1977), and combined with other reflection attributes such as continuity, amplitude, frequency and geometry. Table 4.1 from Sangree and Widmier (1977) summarises the seismic characteristics and environmental facies interpretation of clastic seismic facies units.

The recognition of coastal onlap and toplap is fundamental to the delineation of seismic sequences. Vail et al. (1977) have proposed that eustacy is the driving mechanism for sequence evolution and he assumed that coastal onlaps which define sequence boundaries are globally synchronous. The elementary assumptions given by Vail et al. (1977) are that a relative sea level rise is indicated by a landward shift of coastal onlap, while downlap in the reverse direction indicates a relative sea level fall, and coastal toplap indicates a relative stillstand of sea level. Many arguments have been put forward regarding these proposals. Miall (1986) argued that seismic features such as coastal onlap are not necessarily related to sea level change, but could be due to an autocyclic switching of a distributary channel, as in the case of a submarine fan system, or indicate flexural subsidence rather than sea level change. Moreover, Miall (1986) pointed out that quantifying coastal onlap could face a serious problem when dealing with a divergent continental margin where a gradual thinning out of strata is more common than coastal onlap.

Schlager (1993) investigated the role of sediment supply within a supply-dominated system. In the case of constant eustatic sea level, a similar stratal pattern to that produced by sea level change can be generated through the varying of sediment supply. Allen and Posamentier (1993) have documented that the highstand systems tract on the Gironde (France) estuary has developed a seaward prograding pattern since about 4000 ka, inconsistent with the adjacent shoreline which shows sediment starvation. These studies indicate that in the absence of chronological data, stratigraphic correlation between parts of the basin based on stratal pattern may face errors. Posamentier and Allen (1993) suggested that relative sea level change determines the timing of sequence bounding surfaces rather than stratal pattern. Thus, a
stratigraphic correlation can only rely on chronostratigraphic surfaces caused by relative sea level change, such as unconformities and flooding surfaces.

The sequence stratigraphic concept has been applied in various basins, but many of these examples are from passive continental margins on which the subsidence rates are greater basinward. This chapter presents two examples of seismic sequence stratigraphy from two sub-basins in the Rembang Zone which have different tectonic behaviour. The first example (line PWD-23) is from a basin which resembles a half graben basin in character, and the other (line DNR-12) is from a basin which has similarities with a passive continental margin setting. A seismic facies interpretation was carried out along the seismic lines, but, it was not possible to generate maps showing lateral facies variations because of the limited extent of the available seismic data.

The inferences of sea level change in the study area were made using the method of Vail et al. (1977). By referring to the limitations pointed out by Miall (1986), the constructed curve has to be regarded strictly as a qualitative indicator of the relative position of assumed coastal onlap or toplap. The Vail et al. (1977) and Vail (1987) method also allows an estimation of palaeowater depth of the basin floor based on the profile of a prograding clinoform, a feature which has been used for the present study. However, since this method ignores post-depositional effects, such as compaction, the results have to be regarded as minimum depths.

4.3 SEISMIC SEQUENCE STRATIGRAPHY OF THE REMBANG ZONE

Two unmigrated 12-fold seismic lines, DNR-12 and PWD-23, were used for this study (Fig. 4.1). These lines lie mostly in the Rembang Zone and run north-south across the major east-west structural trend. The data for line PWD-23 was acquired in 1974 and reprocessed in 1988, whereas line DNR-12 was shot and processed in 1979. The source was dynamite, and 24 groups of 18 geophones were used (group interval 75 m for PWD-23 and 60 m for DNR-12). Processing included relative statics (sea level datum), pre-stack deconvolution, velocity analysis, residual statics, CDP stack and time variant filtering and scaling.

The interpreted lines for PWD-23 and DNR-12 are provided as Sections 4.1 and 4.2, respectively. Figures 4.6 and 4.7 show the chronostratigraphic correlation sections for lines PWD-23 and DNR-12, respectively. Appendix A contains velocity information for some shot
points along line PWD-23. Unfortunately, similar velocity information was not provided with seismic line DNR-12. The left portion of both seismic lines lies in the northern part of the Kendeng Zone, with extensive deep-seated faults on line DNR-12 (Sect. 4.2) representing the zonal boundary. In contrast to the Rembang Zone, the reflection quality is extremely poor in the Kendeng Zone, which definitely prevents a seismic facies study in this region, and moreover, makes it impossible to correlate units between these zones. The cause of these poor reflections is not clearly known, but it is probably due to the geological conditions in the Kendeng Zone where tight folding and faulting occur.

Strata deposited during the Middle Miocene to Pliocene can be sub-divided and assigned to seven seismic sequences EM1, MM1, MM2, LM1, LM2, PL1 and PL2. During the discussion, systems tracts of these sequences are referred to as the lowstand (LST), transgressive (TST) and highstand (HST) systems tracts, respectively, to indicate a rock unit deposited within a certain interval with respect to relative sea level position. Ages of these sequences were based on correlations with the biostratigraphic studies carried out in East Java by previous investigators (see Chapter 3). In addition, biostratigraphic studies by Bolli (1966) and Blow (1969) for the Bojonegoro-1 well (Figs 4.1 & 4.4), and by Pringgoprawiro and Baharuddin (1979) for the Tobo-08 well (Fig. 4.1) are extremely useful in defining the sequence ages on line DNR-12. The study by Kadar (1992) for the BNG-1 well (Banyubang area, Figs 4.1 & 4.5) is useful for line PWD-23. The reliability of these data relies on the presence of sandy shelf deposits that are widespread and relatively constant in thickness. This facies, the Ledok Formation, displays a very strong and laterally continuous reflection pattern, and is an easily identifiable seismic unit. This facies has been used as a seismic marker on both seismic lines PWD-23 and DNR-12. A biostratigraphic study by Pringgoprawiro (1983) of samples from the Ledok Formation from various places in the Rembang Zone suggested correlation with planktonic foraminiferal zones N17 and N18.

This study found that although the seismic lines PWD-23 and DNR-12 are only 25 km apart, the sequence characters and boundaries are expressed differently on the two sections. The two seismic lines may be regarded as representing the two main basinal areas in the Rembang Zone: the southwest extension of the East Bawean Trough (PWD-23) and the Central Deep (DNR-12, Fig. 4.10). The two elongate basins are separated by the JS-1 Ridge. The differences in seismic features may reflect differences in tectonic behaviour in these basins.
On line PWD-23 (Sect. 4.1) which stretches nearly 50 km from south of Rembang to south of Cepu, downlap directions and general facies trends suggest that all Middle Miocene to Pliocene sediments were deposited on a relatively gentle southward slope. The rate of basin subsidence during the Early and Middle Miocene was high and appears to have been more pronounced in the landward (margin) area which produced relatively thick sequences (EM1, MM1 and MM2) on the northern portion of the seismic line (PWD-23). This basin subsidence is suspected to be associated with the downward movement of a half graben system on the northern boundary of the basin, and the area beneath the Kedinding Anticline acted as a pivot point. During the Late Miocene and possibly Pliocene the rate of subsidence was reduced and the sequences deposited during this interval (LM1, LM2 and PL1) are relatively constant in thickness along the line.

Line DNR-12 (Sect. 4.2) spans about 30 km and was recorded across the Tertiary topographic high of the Tobo High and the basin to the north of it. The Tobo High stretches from east of Ngimbang to south of Cepu and is very well expressed by the gravity map (Fig. 2.4). Compared with the previous line (PWD-23), all Middle Miocene to Pliocene strata were deposited toward the north on relatively steeper topography (?shelf slope). Excessive lateral thickness anomalies of sequences are not indicated which suggests a low sediment supply and a relatively uniform subsidence rate throughout the Middle Miocene to Pliocene.

4.3.1 Sequence boundaries

The sequence boundaries were mostly defined by recognising surfaces on which seismic reflections systematically terminate. Criteria for recognising sequence boundaries were given by Vail et al. (1977) and outlined earlier. The boundaries of the recognised sequences on the studied seismic lines are characterised by onlaps and truncation of reflections. Ages of these boundaries were defined through correlation with the biostratigraphic studies carried out by Bolli (1966), Blow (1969), Pringgoprawiro and Baharuddin (1979) and Pringgoprawiro (1983) in several places in East Java.

Most of the sequence boundaries on line PWD-23 are type II. On this line (Sect. 4.1) the base of the sequence EM1 cannot be defined due to the poor seismic record. The base of its regressive unit (HST) was deduced from inconspicuous reflection appearances which are interpreted as a downlapping reflection pattern. The seismic velocity analysis at some shotpoints has indicated an abrupt decrease of interval velocity at the suspected positions of
this regressive base. The estimated age is Early Miocene, based on the palaeontological data of the BNG-1 well (Banyubang).

On line DNR-12 (Sect. 4.2) the lower boundary of sequence MM1 is difficult to define due to the character of the seismic data which are almost free from reflections. The base of sequence MM2 on this line is expressed by a very strong continuous reflection which is interpreted as an unconformity surface (type I unconformity). This boundary has not been reached by wells drilled in the area of line DNR-12. A rather strong and continuous reflection occurs on top of sequence MM2, which is also interpreted as an unconformity surface. This surface may represent the late Middle Miocene unconformity at the top of the Tawun Formation. Biostratigraphic studies on the Bojonegoro-1 well (Figs 4.1 & 4.4) by Bolli (1966) and Blow (1969) reveal that this hiatus occurs between 855 and 914 m in depth, which is in foraminiferal zone N14. The other type I sequence boundaries on line DNR-12 occur between sequences MM2 and LM1, LM1 and LM2, and probably between sequences LM2 and PL1.

In contrast with line DNR-12, erosional surfaces on line PWD-23 (Sect. 4.1) are mostly local and occur on the northern end of the section between sequences LM1/LM2 and probably between sequences LM2/PL1. The boundaries of sequences MM1/MM2 and MM2/LM1 on this line (Sect. 4.1) are not expressed by pronounced surfaces due to continuous deposition from sequence EM1 to MM1 and from MM1 to MM2. The prominent surfaces shown on this line between MM1 (TST) and MM1 (HST), and between MM2 (TST) and MM2 (HST) are in fact maximum flooding surfaces (surfaces of non-deposition) which occurred during the maximum elevation of sea level. On line PWD-23 these surfaces are marked by basinward terminations of reflectors beneath the surfaces, which are interpreted as apparent truncations, following a geometrical relationship suggested by Vail (1987) (Fig 4.2). By definition (Mitchum et al., 1977), these surfaces cannot be regarded as sequence boundaries. The sequence boundaries, thus, have to be found half way between the HST unit of the earlier sequence and the TST unit of the subsequent sequence.

A maximum flooding surface is also recognised within the sequence LM1 and lies on top of its transgressive unit (TST). Compared with the position of the transgressive units of sequences MM1 and MM2, the transgressive unit of sequence LM1 was shifted more landward, suggesting a relatively more rapid sediment starvation which may have resulted from an insufficiency of terrigenous clastic materials to keep pace with a rapid sea level rise. Due
to this, in the area south of Banyubang Anticline, the regressive unit (HST) of MM2 is separated by a surface of non deposition from the regressive unit (HST) of sequence LM1.

Correlation between these seismic sections, outcrop and well data (BNG-1, NGBU-1, Tobo-04 and Tobo-08; Fig. 4.1) confirms that the sequence boundaries LM1/LM2 and LM2/PL1 correspond to the lithostratigraphic boundaries of the Wonocolo/Ledok and Ledok/Mundu Formations respectively. The base of the Wonocolo Formation, according to the data from BNG-1 and NGBU-001 wells, was picked at the boundary between sequences MM2 (HST) and LM1 (HSTa) for the area south of Banyubang Anticline. The stratigraphically equivalent Bulu Formation (developed on the north flank of the Banyubang Anticline in the area between Blora and Rembang; Figs 3.1, 4.1 & 4.5) represents the transgressive unit of sequence LM1. Its reefal facies may have developed during the high sea level stillstand (HSTa). Stratigraphically, the sequence just below the Wonocolo and Bulu Formations should represent the Tawun Formation. The boundary of the Tawun and underlying Tuban Formations according to the NGBU-001 well data was picked at the boundary between the transgressive and regressive units of sequence MM1. Ages of the formation boundaries have been discussed in Chapter 3 and summarised in Table 3.4. Here, these ages are used to infer the ages of the seismic sequences.

4.3.2 Seismic facies

a. Line PWD-23

Discussion presented in this section refers to the interpreted seismic section, Section 4.1, and the chronostratigraphic correlation section Figure 4.6. The sequence terminology EM, MM, LM and PL correspond to their periods of formation: Early Miocene, Middle Miocene, Late Miocene and Pliocene respectively.

a.1 Sequences EM1 (regressive unit), MM1 and MM2

These sequences are composed of nearly symmetrical regressive-transgressive systems (Sects 4.1 & Fig. 4.6). As discussed earlier the boundaries between these sequences are type II unconformities. The base of sequence EM1 is unidentified due to poor data quality in the lower part of the seismic section, and only the regressive unit (HST) of sequence EM1 is recognised.
All of these sequences thicken landward (northward) suggesting greater subsidence and sedimentation rates landward. To the south, the deposition of the sequences was limited by a topographic high beneath the Kedinding Anticline. Early and Middle Miocene sediments above this high are relatively thin. The absence of erosional features suggests that this high was syndepositional, and the development ceased from the deposition of sequence LM1 (Wonocolo Formation) onward. This high is interpreted to be the southwest extension of the JS-1 ridge in the Java Sea (Fig. 4.10).

The transgressive and regressive units of these sequences are distinctive. The regressive units are characterised by a subtle sigmoid progradational pattern, and due to thinning southward the pattern tends to be shingled at the toes. The transgressive units are characterised by a progressive landward shifting (northward) of the apparent truncation features on which the subsequent highstand units downlap in a basinward direction. These regressive and transgressive patterns suggest that the combination of subsidence rate, sediment supply and sea level fluctuations might have successfully maintained a constant accommodation space for the terrigenous clastic to be deposited. The Banyubang well, BNG-1 (Fig. 4.5), indicates that these sequences are not characterised by pronounced vertical changes in sedimentary facies. They are represented by thick paralic deposits, consisting mostly of carbonaceous mudstone with thick interbeds of bioclastic limestone and glauconitic quartz sandstone. The bioclastic limestone and quartz sandstone dominate the upper part of sequence MM2. It is interpreted that at the end of this sequence the relative sea level slowly increased which is also evident from the significant landward shift of the transgressive unit (TST) of sequence LM1.

The sequence EM1 and the lower part of sequence MM1 are equivalent to the Tuban Formation (Ardhana, 1993). The boundary between this formation and the overlying Tawun Formation in the BNG-1 well is in fact difficult to define due to the similarity in sedimentary facies. However, the Bojonegoro-1 well (Fig. 4.4) has indicated the presence of a hiatus (early-Middle Miocene hiatus, at depth of 1842 m) between these formations. On line PWD-23 (Sect. 4.1) this boundary is tentatively placed on the prominent surface (maximum flooding surface) between the transgressive (TST) and the regressive (HST) units of sequence MM1.

The regressive unit (HST) of sequence MM2, according to BNG-1 well, corresponds with the Ngrayong Member of the Tawun Formation. Seismically, this formation is characterised by a parallel clinoform pattern with variable amplitude and continuity. The upper part of the
Tawun Formation (corresponding to sequence MM2), as observed in the Bulu, Plantungan and Jepon areas (Fig. 4.1), consists of paralic deposits which transgress into shallow marine facies. In the Jepon area, the paralic deposits consist of well-bedded, organic-rich mudstone (Fig. 4.11) with some thin interbedded coal seams. A quartz sandstone of littoral facies (Ngrayong Member) is very well exposed in the Plantungan area. Here, this sandstone is often intercalated with beds of orbitoidal (larger foraminiferal) limestone of up to 1 m thick (Fig. 4.12). Some of the foraminifer genera have been attributed to the Tawun Formation (van Bemmelen, 1949; Pringgoprawiro, 1983). The upper part of this formation is typified by the genera *Lepidocyclina*, *Miogypsinoidea* and *Cycloclypeus*. The living genera of larger foraminifers occur in shallow marine waters of the tropics or subtropics, and are common constituents in rocks containing coral reef deposits and calcareous algae (Moore *et al.*, 1952; Serra-Kiel & Reguant, 1984).

As indicated previously, the basin in which sequences EM1, MM1 and MM2 were deposited is characterised by a greater landward subsidence rate. This style is commonly associated with a tilt-block of a half graben basin (Leeder and Gawthorpe, 1987; Schlische, 1991), and the sediments are derived from the uplifted footwall. It should be noted that such basins commonly occur in an extensional tectonic system. Hamilton (1979) suggested that the present northward movement of the Indian-Australian Plate is about 10 cm/year. A similar northward compressional stress would occur if the tectonic setting during the Middle Miocene were similar to the present day, as indicated by Baumann (1982). Dewey (1980) indicated that an extensional tectonic system could occur in a retro-arc setting along an old line of weakness obliquely oriented to the subducting plate. This has been proven to occur along the Sumatran Fault system (Katili, 1970), and may also occur in the study area where lines of weakness (Fig. 4.10) resulting from a previous pre-Tertiary subduction complex are suspected to be present.

A greater landward subsidence rate has allowed a landward thickening of the Middle Miocene deposits. This style of sedimentation, in some cases, resembles foreland basin sedimentation where the greater basin subsidence is adjacent to the uplifted area where the sediments are sourced. Examples of the application of sequence stratigraphy to foreland basins are few. Swift *et al.* (1987) studied the Late Cretaceous Mesaverde Group, a foreland basin deposit in the Book Cliffs (Utah). They found that the anatomy of sedimentary sequences differs significantly from those found on passive margins. In the Mesaverde Group,
the transgressive deposits are offshore sand bodies derived from the high area in front of the basin, intertongued with the deltaic regressive deposits derived from the uplifted mountainous region. Figure 4.8 illustrates the sequence architecture models from passive margin and foreland basin settings. In the foreland basin setting, a sea level fall-induced sequence boundary has never been well represented. The sea level fall has never effectively exposed the landward margin because of more rapid subsidence than the seaward margin. This situation appears to be similar to the study area (line PWD-23), but both the transgressive and regressive deposits were derived from the uplifted area in the north. However, the influence of eustatic sea level change on the Middle Miocene sediment deposition is strongly displayed in the study area, as indicated by the almost sinusoidal advance and retreat of these deposits (Fig. 4.6).

a.2 Sequence LM1

Three units (TST, HSTa and HSTb) can be recognised within this sequence. The transgressive unit (TST) was deposited during a relatively rapid sea level rise. The deposition was restricted to landward (north) of Banyubang Anticline, while a surface of non deposition occurred in the basinal area (south of the Banyubang Anticline) between this sequence and the underlying sequence. The unit HSTa is the early regressive unit, and occurred during the highstand of sea level.

Unit TST is characterised by a medium to high amplitude and high continuity parallel reflection pattern. The spacing of reflections is somewhat irregular, probably indicating an irregular spacing of sedimentary interbeds. Data from the BNG-1 well confirms that this transgressive unit corresponds with the Bulu Formation which consists of calcarenite interbedded with quartz sandstone and marly mudstone. The total thickness of the formation in this well is about 80 m. A relatively thick (>10 m) reefal limestone consisting of larger foraminifers (Orbitoididae), bryozoans, echinoids and coral fragments occurs in the area near Bulu (Fig. 4.1). In this area the thickness of the formation reaches 240 m.

Unit HSTa is typified by a low to medium amplitude subparallel reflection pattern. Most of this unit was deposited in a low energy depositional setting. The maximum thickness is about 270 msec TWT (about 270 m) decreasing landward to about 150 msec TWT (about 150 m for a velocity of 2000 m/sec). Its stratigraphic position, which lies between the transgressive (TST, Bulu Formation) and prograding (HSTb) units implies deposition during
a highstand of sea level. With regard to its seismic character, most of unit HSTa may be a hemipelagic deposit. Well data (NGBU-001) confirm that it consists of foraminiferal-rich marl and corresponds with the lower part of the Wonocolo Formation. The planktonic/benthonic ratio in this formation ranges from 20 to 50% in the Bulu area and 60 to 80% in the Kawengan area (Pringgoprawiro, 1983), which indicate a facies variation from inner to outer shelf (Murray, 1991). Toward the north (Bulu area), the facies might have changed into reefal which is regarded as the upper part of the Bulu Formation.

The late regressive unit (HSTb) has sigmoidal prograding character which changes into a shingled reflection pattern due to basinward thinning. It is characterised by medium to high amplitude and good continuity. It thickens landward from about 70 to 110 msec TWT (about 70 to 110 m). In the area between the Banyubang and Plantungan Anticlines the upper part is locally truncated, indicating the presence of sedimentary bypassing. However, the presence of basinward thinning combined with a shingled progradation pattern instead of outbuilding clinoforms suggests that sediment supply was relatively low. This unit corresponds with the upper part of the Wonocolo Formation.

a.3 Sequence LM2

The sequence LM2 was deposited at the end of the Late Miocene. Recognition of reflection terminations and system tracts was difficult as the apparent depositional slope is relatively flat on section PWD-23. Seismically, this sequence is characterised by high amplitude continuous parallel reflection patterns. A relatively constant lateral thickness suggests a sheet-like or tabular external form. These features indicate a uniform subsidence rate. A weakening of reflection amplitude occurs in the area south of the Kedinding Anticline. This is interpreted as a facies change into marl-prone facies which possibly represents a deeper marine environment.

Sequence LM2 is correlated with the Ledok Formation which, elsewhere in the Rembang Zone, stratigraphically overlies the Wonocolo Formation. Field study in the areas near Ledok and Wonocolo found that the Ledok Formation consists of interbedded calcarenite and marl/marly mudstone. In the field, the transition from the Wonocolo to Ledok Formations is recognised by an increase in the number of calcarenite interbeds (Fig. 4.13). The upper part of the Ledok Formation is very sandy and glauconitic with an average grain size of about 1 mm. The benthonic foraminifers *Uvigerina, Bulimina, Laticarinina* and *Nodosaria* are
common in the Ledok Formation (Pringgoprawiro, 1983), suggesting a middle to outer neritic environment. But, the abundance of large crossbeds (> 1 m) in this formation suggests the deposition was influenced by tidal currents, which implies a shallower environment. Samuel and Johannes (1986) measured the orientation of these structures in an area north of Cepu and obtained a general southward trend of the dominant tidal currents. To the south, the Ledok Formation is correlated with the pelagic marl of the Kalibeng Formation in the Kendeng Zone.

These sedimentary features suggest that the Ledok Formation was deposited in a tide-dominated shelf environment. Dalrymple (1992) indicated that in this setting, tidal currents commonly range from 0.5 to 1.5 m/sec, which are strong enough to generate tidal sand waves. A modern example of this environment has been reported by Berne et al. (1991) to occur in the Bay of Bourgneuf (France). Here, sand waves range in amplitude from 0.7 to 9.4 m, and exhibit marked asymmetry, which has been attributed to the combined effect of wave and tidal currents. Dip measurements by Berne et al. (1991) on the internal laminae of sand waves suggest the occurrence of sandflow similar to those found in eolian dunes.

a.4 Sequence PL1 and PL2

Sequence PL1 was deposited during the Pliocene. Most parts of the sequence observed on the seismic sections show transgressive features, since to the south of the Kedinding Anticline it progressively onlaps landward (northward) on top of sequence LM2. The seismic character is a low amplitude and poor continuity subparallel reflection pattern, indicating a low energy depositional environment. The lithofacies interpretation was based particularly on correlation with outcrops along the seismic line, and suggests a correspondence with the marl forming the Mundu Formation. This marl is an outer neritic hemipelagic deposit characterised by abundant planktonic foraminifers.

A study by Aminuddin et al. (1981) in the area between Cepu and Rembang indicated that the Mundu Formation shoaled upward into an alternating coquina (Fig. 4.14), sandy limestone and marl beds [Selorejo Member, Fig. 4.14, Pringgoprawiro (1983)]. This unit thickens northward from few metres in the area south of Cepu to more than 200 m in the area south of Rembang. Aminuddin et al. (1981) found that many foraminifers contained in this unit represent reworked faunas. This was confirmed by the petrographic study (chapter 4.4.5) that showed most foraminifers in this unit are micritised. The micritisation may have occurred during a reworking of older units or during diagenesis. The Selorejo Member may be regarded
as the regressive unit of sequence PL1 where the sea level was nearly or at the high stillstand position.

Sequence PL2 is the uppermost sequence identified. South of the Kedinding Anticline (Sect. 4.1) the base of this sequence is characterised by an erosional surface. The internal character shows a high amplitude wavy seismic reflection pattern. It appears to lie horizontally which implies that deposition occurred after or contemporaneous with the Kedinding Anticline formation. This sequence is correlated with the tidal flat deposits of the Lidah Formation which is widely exposed in the Rembang Zone and the eastern part of the Kendeng Zone. Some authors agree that the development of this formation has led to the Late Pliocene base levelling through most of East Java (van Bemmelen, 1949; Lemigas/BEICIP, 1969). The real thickness is not known, but the seismic sections suggest up to 400 msec TWT (about 400 m).

b. Line DNR-12

Since Tertiary sedimentary outcrops along and near this seismic line are absent, the lithofacies interpretation of this seismic line was carried out through correlation with well data from Tobo-04, Tobo-08 and Bojonegoro-1 wells and additional reference to the criteria given by Sangree and Widmier (1977). The Tobo-04 and Tobo-08 wells are shallow wells, with total depths of 410 m and 432 m, respectively. They are only effective for the correlation of the upper sequences: LM2, PL1 and PL2. The Bojonegoro-1 well has a total depth of 2034 m. However, due to its position which is about 17 km away from line DNR-12, a good correlation with this line has not been achieved. Section 4.1 displays the interpreted section for line DNR-12 and Figure 4.7 summarises the chronostratigraphic correlation of this line. The sequence terminology used in these figures is similar to line PWD-23, and implies chronostratigraphic equivalence.

b.1 Sequences EM1, MM1 and MM2

A detailed discussion on sequences EM1 and MM1 is hampered by their seismic character which is nearly reflection free. The sequences probably consist of mudstone facies with some sandstone lenses as indicated by the presence of a few discontinuous and variable amplitude reflections. The recognition of systems tracts, on the basis of this seismic character, would be speculative and is not attempted.
Sequence MM2 on line DNR-12 is a type I sequence. Three systems tracts can be recognised in this segment of the line: the lowstand, transgressive and highstand systems tracts. The lowstand systems tract is represented by a relatively thin [< 110 msec TWT (110 m)] shingled prograding unit. The internal seismic character of this unit shows a high amplitude and continuous parallel reflection pattern. The lowstand unit is thought to have been deposited in a shallow marine environment and consists of terrigenous clastic strata. The water depth of basin floor was probably equal to the maximum thickness of the unit, 110 msec TWT (about 110 m).

The subsequent sea level rise stopped deposition of the lowstand unit and the transgressive systems tract developed on the topographic high. The reflections along the upper boundary of this tract backstep landward (southward), indicating a further retreat of sedimentation (note the difference in sedimentation direction compared with line PWD-23). The seismic reflection pattern which shows a mounded form and lack of internal continuity, suggests that it is composed of carbonate buildups.

The highstand systems tract is characterised by an oblique progradational pattern with reflectors toplapping against the upper sequence boundary. The internal reflection pattern shows a high continuity of medium amplitude parallel reflectors. Mitchum et al. (1977) attributed this type of seismic pattern to rapid basin infill and sedimentary bypassing resulting from a combination of high sediment supply and sea level stillstand. On the seismic line studied, a rapid basinward shift of the toplapping pattern, particularly above the northern flank of the topographic high, may have resulted from a relative sea level fall. In the distal part (basinward), the presence of northward toplapping suggests that this unit merged with a northerly derived depositional system.

b.2 Sequence LM1

This a type I sequence in which two main systems tracts can be differentiated from the seismic section: the lowstand and highstand systems tracts.

The lowstand systems tract consists of three distinct features which are interpreted to represent the sheet-drape, basin-floor fan and lowstand wedge. This interpretation refers to the criteria outlined by Mitchum (1985), Posamentier and Erskine (1991) and Posamentier et al. (1991). The basin-floor fan is characterised by a mounded external form, and the internal
reflections are somewhat variable in both amplitude and continuity indicating a variable current during deposition. Basinward (northward), it grades into a sheet-drape of probable hemipelagic deposits which is characterised by a medium amplitude parallel reflection pattern. Shelfward (southward), the basin-floor fan grades into a thin oblique progradational clinoform of a lowstand wedge.

Deposition of the highstand systems tract occurred as a result of the subsequent relative sea level rise. It can be separated into three main parts: proximal, middle and distal. The proximal part downlaps on the lower sequence boundary. The middle part consists of an oblique prograding wedge which grades into a tangential-oblique reflection pattern in the distal part. Toplapping occurring on top of the tract suggests sedimentary bypassing combined with a very high sediment supply. Similar to the regressive unit of sequence MM2, the most distal (prodelta) facies may have merged with a northerly derived depositional system as indicated by the presence of southward prograding clinoforms. The Tobo-04 well has penetrated about 40 m of the highstand deposit and indicated a lithofacies of greenish grey, glauconitic sandy marl. The well report assigned this facies to the Wonocolo Formation.

The highstand systems tract may have been deposited as a highstand prograding delta complex. However, a further detailed seismic study would be necessary to support the interpretation, with the specific aims of defining the lateral distribution, external geometry and associated seismic features (such as fluvial channelling). Highstand prograding clinoforms associated with a prograding delta complex have been reported by some authors. Suter and Berryhill (1985) recognised the occurrence of five Late Quaternary deltas on the northwest shelf margin of the Gulf of Mexico. They used only geophysical data to identify these delta systems, but the extent and coverage of the data allowed a recognition of mappable geomorphic patterns which are associated with high-angle clinoform reflection patterns of delta lobes and buried fluvial systems. A similar method combined with core drilling was used by Sydow and Roberts (1994) on the recognition of a shelf edge delta in the northeast Gulf of Mexico.

b.3 Sequence LM2

Sequence LM2 on line DNR-12 is a type I sequence. The seismic character of this sequence is very similar to that on line PWD-23. Its depositional setting, which was on a
relatively steeper slope, has allowed recognition of two major systems tracts, the lowstand (LST) and highstand (HST) tracts.

The lowstand systems tract is characterised by relatively low amplitude and high continuity tangential-oblique reflections which may be associated with a lowstand wedge mudstone facies. The highstand systems tract was deposited following this facies. The landward onlapping of the lower part of the highstand systems tract (HST) probably represents transgressive conditions and may be regarded as the transgressive systems tract. The maximum relative sea level rise is difficult to estimate because of poorly defined onlapping features in the landward direction.

The highstand deposits pinch out basinward, and are characterised by a high amplitude reflection pattern with frequent internal discontinuity suggesting a fairly shallow marine environment. These deposits downlap on top of sequence LM1 and the lowstand deposits of LM2. Two shallow wells, Tobo-04 and Tobo-08, penetrated the highstand deposits and indicate similar facies to the sequence LM2 on line PWD-23, which was assigned to the Ledok Formation in the earlier discussion. In these two wells, this formation is also characterised by interbedded fossiliferous, glauconitic sandstone and marl; the main terrigenous fragments are fine to medium-grained quartz.

**b.4 Sequences PL1 and PL2**

Sequence PL1 is identified by its progressive landward onlap on top of sequence LM2. This part may represent a transgressive systems tract. Seismically, this unit is characterised by a low amplitude and high continuity parallel reflection pattern. Data from the Tobo-04 and Tobo-08 wells indicate that the sequence consists mostly of foraminiferal-rich marl similar to the Mundu Formation recognised on line PWD-23. In the Tobo-08 well this formation is 70 m thick, thinning southward to about 30 m thickness in the Tobo-04 well. Thinning of the formation appears to be related to the depositional topography which was higher toward the south. The coquina and sandy limestone of the Selorejo Member, which characterise the upper part of the Mundu Formation, are still recognisable in the Tobo-08 well as a 30 m thick unit.

Sequence PL2 lies horizontally and onlaps on top of sequence PL1. Seismically, it is characterised by a relatively stronger amplitude reflection pattern than PL1, suggesting a more pronounced alternation of fine and coarse clastic strata. This uppermost sequence is widely
exposed along the seismic line and has a similar appearance to the tidal deposits of the Lidah Formation in Cepu area.

4.3.3 Relative sea level change

The inferred relative sea level changes for the Rembang Zone are presented in Figure 4.9. The curves were based only on the Middle Miocene to Pliocene coastal aggradations in line DNR-12 (Sect. 4.2 & Fig. 4.7) and are not normalised to remove local tectonic effects. Recognition of coastal onlaps on line PWD-23 is difficult, due to limited extent and poor data quality in the landward areas, and the interpretation of its relative sea level change is not attempted.

The inferred Pliocene sea level magnitude in Figure 4.9 was not based entirely on the seismic features on line DNR-12 due to the difficulty in defining the topmost position of onlap. The seismic study in Chapter Seven suggests that the relative sea level rise in the Early Pliocene was about 115 m. The sea level magnitude was also based on the foraminiferal examination by Pringgoprawiro and Baharuddin (1979) in the Tobo-08 well. The Pliocene sequence, PL1, consists of hemipelagic marl and was deposited in an outer neritic environment.

Figure 4.9 indicates that the Middle Miocene and Pliocene maximum sea level magnitudes were relatively higher than in the Late Miocene. These are consistent with the Late Tertiary global relative change of coastal onlap proposed by Haq et al. (1988). However, the number of Middle Miocene and Pliocene sequences in the study area is fewer than the number of sequences recognised by Haq et al. (1988) for the same period, and appears to be a generalisation of Haq's curve. Posamentier and James (1993) indicated that the number of sequences observed in a certain area will depend on many local factors. The four major variables include tectonic subsidence, eustatic change of sea level, sediment supply and climate (Vail, 1987). Because of these, each area may have its own relative sea level curve (Suter et al., 1987).

4.4 SEDIMENTARY PETROGRAPHY

This section briefly discusses the petrographic character of the Middle Miocene to Pliocene lithofacies in the area between Cepu and Rembang. All rock samples for the petrographic analysis came from outcrop. The analysis mostly used a transmitted light microscope. The
main objective of the petrographic study was to document the overall character of the lithofacies, particularly composition and provenance. Appendix B.1 summarises the results of the petrographic analyses.

4.4.1 Tawun Formation

Four samples from the Tawun Formation were analysed for this study. All of these samples were from exposures in the Bulu, Plantungan and Jepon (Banyubang) areas (Fig. 4.1) which represent the upper part of the Tawun Formation. Two main facies are developed: organic-rich mudstone with thin beds of fine quartz sandstone, and quartz sandstone (the Ngrayong Member). The mudstone is commonly parallel laminated. In the samples studied, the amount of fine-grained (average grain size of 0.2 mm) quartz is up to 20% (Fig. 4.15).

Quartz sandstones in the Tawun Formation are mostly free from mud, and commonly contain organic spots of possibly dried oil in their pore spaces. Quartz (up to 86%) is commonly characterised by undulose extinction and cracking (Fig. 4.16). Some polycrystalline quartz grains are sutured and sheared (Fig. 4.16). The grain size of quartz varies from 0.1 to 1.5 mm with mostly subangular shapes.

Folk (1974) used the quartz undulatory extinction and polycrystallinity as parameters to indicate its origin. He proposed that highly undulose quartz is diagnostic of metamorphic origin. A study of the significance of quartz undulatory extinction and polycrystallinity by Basu et al. (1975) concluded that these features can be used to distinguish plutonic quartz from low-rank metamorphic quartz in first-cycle sandstone. In the Tawun Formation, in all four sandstone samples studied, undulose extinction and polycrystalline quartz are present. Their metamorphic origin is also suggested by the presence of metamorphic rock fragments such as slate.

The seismic interpretation has indicated that all of the Tawun Formation in the area between Cepu and Rembang was derived from a northerly source. This suggests the importance of the Bawean Arch as the source of sediment during the Middle Miocene (Fig. 4.10). The paralic environment which characterised the Tawun Formation would have developed along the margin of the subaerially exposed Bawean Arch.
4.4.2 Bulu Formation

A study of an exposure about 50 m long in an area near Bulu (Fig. 4.1) indicates the formation is characterised by a transgressive sequence. The sequence begins with an organic-rich dark brown mudstone associated with a channel facies containing quartz-rich sandstone and conglomeratic-lag deposits. Above these deposits is about 30 m of interbedded fossiliferous mudstone and fine-grained quartz sandstone which are topped by a thick orbitoidal bioclastic limestone (Fig. 4.18).

The petrographic study was carried out on four limestone samples and one sandstone sample from the above section. In character, the channel sandstone resembles the quartz sandstone of the Ngrayong Member, but most of the pores contains organic streaks (?dried oil). The quartz grains (Fig. 4.17) have a similar appearance to those in the Ngrayong Member and are suspected to be derived from a similar source. The bioclastic limestone is bedded and typified by an abundance of larger foraminifers (Orbitoididae) and algal fragments (up to 6 mm). It also contains quartz grains, commonly less than 16%, with the grain size varying from 0.1 to 2 mm (Fig. 4.18). In most of the limestone samples studied, finely crystalline sparry calcite cement is well developed (as in Fig. 4.18).

4.4.3 Wonocolo Formation

Samples for the petrographic study of this formation were from the Kawengan area (some 20 km northeast of Cepu) and the upper part of the formation. Here, this formation is characterised mainly by foraminiferal-rich sandy marl. Figure 4.19 shows a typical specimen with carbonate mud constituting about 50%, mainly as matrix, pore fillings and micritisation of bioclasts. The bioclasts are mainly planktonic foraminiferal tests (< 1 mm size), mollusc and algal fragments (up to 6 mm size) and make up 40%. The remainder consists of terrigenous clastic detritus, composed mainly of angular to subangular monocrystalline quartz with an average grain size of 0.15 mm. Sparry calcite cement appears to be poorly developed, and is found only as a pore filling within some foraminiferal tests (e.g. Fig. 4.19).

X-ray diffraction analysis has been carried out on a sample from this formation, and the results are shown in Figure 4.23. The analysis has confirmed the microscopical examination and shows that the main constituents of this formation are carbonates. The most abundant carbonate is calcite which displays a very pronounced response at 29.43° 2θ, along with most of its subordinate peaks. Aragonite is also present and is suspected to represent calcareous
faunal fragments. The analysis also indicates the presence of very small amounts of smectite and kaolinite.

The presence of substantial carbonate mud is commonly interpreted to indicate deposition under fairly low energy conditions (Matthews, 1966). But the notable presence of quartz and algal fragments suggest that periodic bottom currents were active. Quartz, in this formation, is interpreted to have been derived from the Tawun Formation since in the Late Miocene, the topographic high (JS-1 Ridge) was still developing near Tuban (van Bemmelen, 1949; Lemigas/BEICIP, 1969) and exposing the Middle Miocene strata subaerially.

4.4.4 Ledok Formation

Four samples for petrographic analysis were obtained from the middle and top of the formation in the Ledok area (some 15 km north of Cepu) and the Wonocolo area (some 20 km northeast of Cepu). In the samples from the middle part of the formation, carbonate mud, bioclasts of foraminiferal tests and algal fragments constitute up to 69-88%. The terrigenous fragments, mostly monocrystalline quartz, form less than 12-37%. These fragments display subangular form with an average grain size of 0.3 mm.

In the upper part of the formation, the average grain size (0.6 mm) is significantly coarser than in the middle part. This is associated with shoaling of the sequence. The upper part is characterised by the appearance of up to 47% glauconite (Fig. 4.20). Almost all of the glauconite occurs as rounded peleoid grains, accompanied by some interstitial filling within foraminiferal tests. The rim of some glauconite grains appears to have been altered into iron oxides, probably during the present weathering cycle. Carbonate content is up to 69%, consisting mainly of foraminiferal tests and algal fragments. Undulose monocrystalline quartz constitutes up to 17%, with the grain size varying between 0.1 and 0.5 mm. Compared with the samples from the middle part of the formation, carbonate mud is essentially minor indicating deposition under a strong bottom current. This interpretation is supported by the fact that large crossbeds are common in the upper part.

The occurrence of glauconite in the upper Ledok Formation has never been discussed in detail. The mode of deposition, which was under strong bottom currents led Samuel and Johannes (1986) to the opinion that the glauconite formation was not in situ. The processes which lead to the formation of glauconite are still uncertain. Burst (1958) and Hower (1961)
believed that glauconite pellets are derived from a three-layer silicate lattice (clay minerals) through a transformation process involving supplies of potassium and iron, and a suitable oxidation potential. Odin and Matter (1981) argued that glauconite formation is independent of parent materials. They suggested that glauconite authigenically grows in substrate pores favoured by the microenvironment in the substrate, and is accompanied by progressive alteration and replacement of the substrate. The glauconite occurrence in the Ledok Formation appears to favour the Odin and Matter (1981) argument. There are three modes of glauconite precipitation in the Ledok Formation, as rounded peleoids, interstitial fillings and casts of bioclasts. Among these, the rounded peleoid type is the most common. The usual microscopic appearance is that it often contains well preserved very fine-grained quartz fragments, which imply that the glauconite may have been altered from quartz-bearing pellets.

There is a common agreement that glauconite formation is associated with a low sedimentation rate (Hower, 1961; Odin & Matter, 1981; van Houten & Purucker, 1984), and most commonly occurs in the open marine environment beyond the zone of fluvial influence (Odin & Matter, 1981). In contrast, abundance glauconite in the Ledok Formation occurs at the upper part of the formation, and was associated with strong tidal bottom currents. These suggest that glauconitisation might not have occurred simultaneously with substrate deposition, and possibly occurred later during the subsequent transgression.

4.4.5 Mundu Formation

The Mundu Formation consists of marly facies and limestone facies (Selorejo Member). The marly facies consists of hemipelagic marl and is very rich in planktonic and benthonic foraminiferal tests (up to 40% of the total carbonate; Fig. 4.21). Terrigenous clastic detritus is minor and mostly consists of very fine-grained quartz and feldspars. In the area near Cepu, the limestone facies of the Mundu Formation is known as the Selorejo Member (Pringgoprawiro, 1983). A sample from this unit shows that subangular plagioclase grains with an average size of 0.05 mm make up about 1% and glauconite amounts to less than 1%. The field study indicates that this member is typified by a high content of bioclastic material, including foraminiferal tests and molluscan shell fragments. All these clastics are cemented by sparry calcite, resulting in a total carbonate content of more than 90%. Microscopic study found that these bioclasts have been intensely micritised (Fig. 4.22).
4.5 DISCUSSION

a. Middle Miocene strata

The Middle Miocene strata economically are the most important strata in the Rembang Zone. The primary petroleum reservoir rock lies in the regressive quartz sandstone of the Middle Miocene Tawun Formation (Ngrayong Member). Little is known about organic source rocks, but the source potential of organic-rich mudstones in this formation has been suggested.

Several suggestions have been put forward regarding the setting and mode of deposition of the Middle Miocene strata. On a regional scale, Van Bemmelen (1949) postulated that the Rembang Zone was the marginal part of Sundaland (Sunda Shelf). The Early and Middle Miocene basin subsidence and sedimentation that occurred in this zone were results of diastrophism in southeast Sundaland. Regional studies by Lemigas/BEICIP (1969), Sudiro et al. (1973) and Ardhana (1993) indicated that during these periods the sedimentary basin was fragmented by northeast trending ridges which can be traced into the Java Sea (Fig. 4.10). As shown by the seismic interpretation, sedimentation occurred essentially within locally developed sub-basins which have different rates of subsidence and sediment supply. Particularly high subsidence and sedimentation rates occurred in the sub-basin forming the southwestward extension of the East Bawean Trough [Fig. 4.10 & Sect. 4.1 (seismic line PWD-23)], and most petroleum accumulations in the Rembang Zone occur in this sub-basin.

Some inferences about the depositional environment and palaeogeography for Middle Miocene sediments in this sub-basin have been made by various workers. Ardhana (1993) re-examined the depositional setting of the Tawun Formation based on outcrops and well data mostly from the area between Surabaya and Rembang (Fig. 4.1). His study relied heavily on the bentonic faunal association in the sediments and the turbidite model developed by Mutti and Ricci-Lucchi (1972). He assumed that most of the Tawun Formation was derived from the Bawean Arch (Fig. 4.10). In the south, Ardhana (1993) interpreted that the Tawun Formation developed as a slope facies and bathyal submarine fan system which are characterised by deep marine faunas such as Globocassidulina subglobosa, Hoeglundina elegans and Uvigerina crassicostata. In the area near the depositional high (Bawean Arch) Ardhana concluded that the formation developed as a tidally influenced deposit. This confirms a study by Muin (1985) in the area between Blora and Rembang (Fig. 4.1) in which he concluded that most of the facies developed as coastal sands and tidally influenced deposits.
Posamentier et al. (1988) and Posamentier and Erskine (1991) pointed out that a submarine fan system is most likely to occur during the development of a type-I sequence, which is characterised by the occurrence of a lowstand systems tract. The seismic study on line PWD-23 (Sect. 4.1) indicates that the Middle Miocene sequences are all type-II, and are not associated with lowstand systems tracts in which submarine fans commonly occur. Furthermore, the basin in which the Tawun Formation was deposited never reached a bathyal environment. A balance between high subsidence and sedimentation rates provided a constant amount of accommodation space throughout the Middle Miocene, and the sedimentation simply advanced and retreated depending on the eustatic sea level change. Additionally, faunas such as Globocassidulina, Hoeglundina and Uvigerina are proven to be poor palaeowater depth indicators since Phleger (1960) and Murray (1991) documented their occurrence ranging from middle neritic to bathyal environments.

Less attention has been given to the Middle Miocene strata developed in the basin between the JS-1 Ridge and the Tobo High (Fig. 4.10). In the past, the clastic sediments were assumed to be derived from a northerly source. In contrast to the neighbouring basin, the seismic interpretation (DNR-12, Sect. 4.2) has indicated that the sediment supply was relatively low and appeared to be derived from bordering highs both to the north and south. However, whether a similar pattern applies for the whole basin (Central Deep, Fig. 4.10) still has to be proved through a comprehensive seismic interpretation.

b. Late Miocene and Pliocene strata

In the Rembang Zone the Late Miocene and Pliocene are known as periods of widespread development of hemipelagic marl (Lemigas/BEICIP, 1969). Terrigenous deposition was significantly reduced, apart from some rather widespread deposits in the Ledok Formation. This marly depositional period was contemporaneous with the deposition of similar facies in the Kendeng Zone - the Kalibeng Formation. The underlying formation, the Kerek Formation, which is stratigraphically equivalent to the Wonocolo Formation, is characterised by volcanogenic detritus, indicating a close association with Late Miocene volcanic activity.

The mechanism which led to the widespread deposition of these marly facies is not fully understood. An explanation was given by Van Bemmelen (1949) whereby he associated them with the Middle Miocene "Banyak volcanism", which occurred along the area where the Quaternary volcanoes lie. An uplift which accompanied this Miocene volcanism caused some
subsidence on the adjacent Kendeng Zone and further tilting on the unstable area in the Rembang Zone. The present study indicates that the marly deposition was closely related to a prolonged high stillstand of sea level. A similar relationship is also observed in the Java Sea (Chapter 7), where the marl deposition during the Pliocene occurred following a relative high stillstand or slow fall of sea level.

4.6 SUMMARY AND CONCLUSIONS

The Rembang Zone consists of two main sub-basins which were the southwestward extensions of the East Bawean Trough and the Central Deep. During the Middle Miocene to Pliocene the number of sequences developed in these sub-basins is the same, but they have different seismic character as the result of different rates of subsidence and deposition in the sub-basins. Relatively low sedimentation and subsidence rates, particularly during the Middle Miocene, occurred in the southwestern part of the Central Deep.

The relative sea level fluctuations and timing during the Middle Miocene to Pliocene in general appear to be in agreement with the global relative coastal onlap charts established by Haq et al. (1988). However, the study failed to recognise some of the smaller fluctuations indicated by Haq et al. (1988).

The seismic and petrographic studies have indicated that sediments deposited in the sub-basins were derived from the adjacent highs. Thick Middle Miocene paralic deposits in the basin north of Cepu were derived from the Bawean Arch, whereas in the basin south of Cepu they were derived from the JS-1 Ridge and Tobo High. Most of these sediments are organic-rich mudstones and quartz sandstones which are good source and reservoir rocks in these areas.

The Late Miocene and Pliocene periods were dominated by hemipelagic marly facies deposition (Wonocolo and Mundu Formations), interrupted in the late Late Miocene by a period of relatively coarse clastic deposition forming the Ledok Formation. Relatively thick marls of the Wonocolo Formation and probably the Mundu Formation were deposited during a prolonged highstand of sea level during which terrigenous clastic input and deposition was reduced. Near the topographic highs the Wonocolo marl changed into deltaic (on the Tobo High) and into limestone dominated sequences (Bulu Formation in the area between Rembang and Blora).
CHAPTER FIVE
SEDIMENTOLOGY OF THE KENDENG ZONE

5.1 INTRODUCTION

This chapter examines the Late Miocene to Pleistocene sedimentology of the Kendeng Zone. The main objective of this chapter is to provide a parallel discussion with Chapter 4 which provided details about the Rembang Zone. Further, this chapter aims to review the palaeoenvironments and depositional settings under which the units were deposited. Due to the paucity of geophysical data the whole discussion presented in this chapter will be based on the lithofacies analysis of a number of measured vertical sections. Petrographic and benthonic microfaunal analyses are used to support the interpretation, particularly when the lithofacies analysis is ineffective.

The sedimentological discussion includes seven lithofacies units, these are the Kerek, Kalibeng, Sonde, Atasangin, Lidah, Pucangan and Kabuh Formations. The discussion emphasises the Atasangin, Lidah and Pucangan Formations, since the sedimentological settings of these units were not discussed in detail by previous authors (e.g. de Genevraye & Samuel, 1972; Pringgoprawiro, 1983). A detailed discussion on the Kerek, Kalibeng and Sonde Formations will not be given in this chapter. These formations have been widely studied by previous authors (Lemigas/BEICIP, 1969; van Gorsel & Troelstra, 1981; Muin, 1985). However, the previously defined depositional environments are discussed in the light of new data and environmental models. The upper part of the Kabuh Formation and the Notopuro Formation mainly consist of volcanic deposits (Duyfjes, 1938b,d), and due to the paucity of the data, will not be discussed.

The facies analysis is based on 14 measured vertical sections, with the eleven most complete sections presented in the discussion. The localities of these sections are given in Figure 5.1. This limited number of measured sections is primarily due to the scarcity of good outcrops which further limits the discussion about lateral facies relationships of some lithostratigraphic units. The analytical methods for the petrographic study are similar to those used in Chapter Four (i.e. transmitted light microscope and X-ray diffraction methods), and provide documentation about the general composition and provenance of the succession. The
benthonic faunal analysis aims to identify faunal assemblages which are characteristic of certain environments. This study focuses on the benthonic foraminifers and follows the criteria given by Phleger (1960) and Murray (1973; 1991).

5.2 KEREK FORMATION

5.2.1 Lithofacies

The lithofacies analysis of the Kerek Formation is based on a vertical section measured by Muin (1985) in the Solo River section north of Ngawi (Fig. 5.1). The generalised section is presented in Figure 5.2. The uppermost 26 m was measured during the study (Fig. 5.3), and was particularly aimed at examining the upper boundary of the formation.

The exposed Kerek Formation in the Solo River section attains a thickness of about 800 m (Fig. 5.2). The section consists of a monotonous repetition of sandstone and mudstone in which Muin (1985) found various sedimentary structures typical of Bouma-type sequences. Complete Bouma sequences (Ta-e) are rare, but some complete Bouma sequences are characterised by a 10 to 75 cm conglomeratic sandy unit with erosional base (Bouma facies Ta), grading into parallel-laminated sand (facies Tb; 30 to 110 cm thick), ripple-laminated and convolute-bedded sandstone of about 20 cm thick (facies Tc), parallel-laminated fine-grained sandstone of about 5 cm thick (facies Td), and pelagic marl (facies Te). Most of the coarse clastic units are characterised by abundant plagioclase, lithic fragments and amphibole, suggesting a very close association with volcanism. Two thick tuff beds (1 and 8 m) occur, indicating deposition during an active period of volcanism.

The uppermost 26 m of the Kerek Formation in the Solo River section is characterised by interbedded parallel-laminated calcareous mudstone and cross-bedded, ripple and parallel bedded medium- to coarse-grained sandstone (10 m thick; Figs 5.3 & 5.4). Bed thickness varies from 5 to 20 cm. The mudstone and sandstone are calcareous due to the abundant occurrence of planktonic foraminifers. These underlie a 10 m thick calcareous mudstone which is topped by about 6 m of calcirudite. The calcirudite is composed of angular to subrounded clasts of medium- to coarse-grained limestone with dioritic and andesitic rock fragments. Larger benthonic foraminifers are abundant in the calcirudite, but many of these specimens show indications of mechanical destruction and are thought to have been reworked from previous limestone.
5.2.2 Petrography

The petrographic study was carried out on four samples from the uppermost units in the formation (Appendix B.1); the stratigraphic position of the samples is indicated in the lower part of Figure 5.3. Two samples (KE1A & KE3) are characterised by medium-grained sandstone and sandy mudstone. They are composed of a mixture of planktonic foraminifers and terrigenous fragments which are often differentiated into contrasting laminae (1 to 3 mm thick; Fig. 5.6). The terrigenous clastics make up 60% of the sequence, with an average grain size of about 0.6 mm. They consist mainly of angular to subrounded plagioclase, lithic fragments (volcanic) and mafic minerals.

The texture of the calcirudite in the uppermost part of the formation indicates that it was deposited in a high energy environment; mud or carbonate mud is nearly absent. It is characterised by an abundance of coral fragments and larger benthonic foraminifers; genus Lepidocyclina is the most common (Fig. 5.7). Algal fragments are minor and planktonic foraminifers are absent. This unit is also characterised by the presence of about 35% terrigenous clastics. These are rounded, coarse-grained to pebble size diorite and andesite, plagioclase and very minor amounts of mafic minerals.

5.2.3 Depositional setting and discussion

Turbiditic deposits equated with the Kerek Formation are widely developed in the Kendeng Zone (van Bemmelen, 1949; Pringgoprawiro, 1983; de Genevraye & Samuel, 1972; Muin, 1985) where it attains a thickness of more than 1000 m. The main terrigenous clastic components in this deposit have been reworked from the Middle to Late Miocene volcaniclastic strata referred to as the Banyak Volcanism (van Bemmelen, 1949). The volcanism occurred along the area which is presently occupied by the Quaternary volcanic centres. Figure 5.10 shows a tentative palaeogeographic map of East Java in the Late Miocene.

The uppermost part of the formation has attracted some attention (e.g. van Bemmelen, 1949; van Gorsel & Troelstra, 1981; Pringgoprawiro, 1983). Based on sedimentary structures and vertical facies sequence, this part appears to be regressive, and was capped by the deposition of a relatively thick (6 m) calcirudite, which van Gorsel and Troelstra (1981) suggested was deposited in an outer neritic environment as indicated by a consistent appearance of benthonic foraminifers Eggerella bradyi, Melonis soldanii and Globobulimina pupoides. They suggested that the calcirudite represents an olistostromal deposit. This implies
that the calcirudite has a local extent. In contrast, a regional study by van Bemmelen (1949) found that the calcirudite is stratigraphically equivalent with a conglomeratic unit containing andesitic boulders deposited in some areas near Miyana (some 30 km east of Ngawi; Fig. 5.1). He suggested that these conglomeratic units represent a basal conglomerate which marks an unconformable relationship with the overlying Kalibeng Formation. The present study in the Solo River section indicates that the lower part of the Kalibeng Formation, just above the calcirudite, is characterised by a repetitive graded sequence typical of storm deposits (see Chapter 5.3). This implies that the depositional environment of the unit was shallow marine, which further confirms the van Bemmelen (1949) suggestion that calcirudite may represent a wave-dominated environment.

In the past, the Kerek Formation has been regarded as a submarine fan deposit. It has been referred to by van Bemmelen (1949) as a flysch deposit. This opinion was confirmed by van Gorsel and Troelstra (1981) who studied the benthonic faunal content, and by Muin (1985) who studied the lithofacies of this formation in the Solo River section. But the palaeogeography, depositional setting and timing of deposition suggest that the Kerek Formation should not be regarded as a normal submarine fan deposit.

Posamentier et al. (1991) indicated that the presence of sufficient slope length is essential to allow significant gravity flows and turbidity currents to develop. In support, Walker (1992) pointed out that in order to be preserved as turbidite, the deposits must occur below storm wave base, effectively at a minimum depth of 250-300 m. The extensive Plio-Pleistocene Mississippi fan system, for example, occurred in a bathyal setting in the Gulf of Mexico (Weimer, 1991). Textures and sediment composition of the Kerek Formation suggest that the deposition was near or probably on the apron of the Banyak volcanoes. A benthonic foraminiferal study in the Solo River section by Lemigas/BEICIP (1974) indicated that most of the formation in this area was deposited in an inner to middle neritic setting, a region which is effectively influenced by tidal currents (Dalrymple, 1992), fair-weather currents and more importantly storm-induced currents (Johnson & Baldwin, 1986; Walker & Plint, 1992).

The biostratigraphic studies by Lemigas/BEICIP (1974) and Pringgoprawiro (1983) indicated that most of the Kerek Formation in the Solo River section interfingered northward with the hemipelagic marly deposit of the Wonocolo Formation (Fig. 5.10). The seismic stratigraphic study in the Rembang Zone (Chapter 4) reveals that the latter was mostly
deposited during a sea level highstand. The Kerek Formation submarine fan deposition, if it occurred, contradicts the sequence stratigraphic concepts which suggest fans are most commonly deposited during a sea level lowstand (Posamentier et al., 1988; 1991). However, further validation may be required as the sequence stratigraphic concepts were tested mainly on passive margins, a different tectonic setting from the study area.

The present study suggests that at least two possible depositional modes occurred in the Kerek Formation. These are a storm-dominated mode and turbiditic current deposition related to the collapse of volcanic edifices. Because of the tectonic setting of the study area, both may occur simultaneously. This suggestion relies on the fact that (1) most of the formation is dominated by Bouma-type sequences, (2) deposition occurred in a shallow marine setting (Lemigas/BEICIP, 1974), (3) sediment composition and the presence of tuff layers suggest a close association with active volcanism, (4) high depositional slope was probably present, and (5) the regional tectonic setting suggests that seismicity was high (Hamilton, 1979).

Bouma-type sequences regarded as graded storm deposits have been recognised by Nelson (1982) who studied the occurrence of graded sand beds on the Bering Shelf. He estimated that during the major storms a coastal set up may occur and promote liquefaction of the upper 2 to 3 m of sediment. Storm-associated bottom currents can transport liquefied sediments more than 100 km offshore. Johnson and Baldwin (1986) documented that the thickness of a single graded storm unit can be up to 0.5 m, and is characterised by an erosive base, a basal lag deposit, horizontal low angle lamination which is possibly related to hummocky cross-bedding, wave ripple cross-lamination and a burrowed upper interval. With regard to the tectonic setting of the study area, similar deposition may have occurred through seismically-induced waves, such as tsunamis, which often characterise shelves near seismically active region (Coleman, 1969, cited in Johnson & Baldwin, 1986).

Turbiditic sequences related to volcanic activity have been reported by some authors. Busby-Spera (1988) documented three lithofacies developed during the deposition of the Mesozoic volcanioclastics in Baja, California; these are (i) tuff, (ii) lapilli-tuff and tuff breccia, and (iii) primary volcanic lithofacies. Based on the fact that lateral thickness variation is high, the tuff lithofacies was interpreted to have been deposited from subaqueous turbidity flows which followed a palaeo-low. The lapilli-tuff and tuff breccia lithofacies are thick bedded, are inversely graded in the proximal area (debris flow) and channelised, graded and stratified in
the distal parts, and resulted from high-density turbidity currents. The primary lithofacies consist of dacitic pyroclastic rocks and pillow lavas.

Similar facies and modes of deposition as the lapilli-tuff and tuff breccia lithofacies of Busby-Spera (1988) have been reported by Hathway (1994) in the Oligo-Miocene volcaniclastic arc assemblage in the intra-oceanic/fore-arc area of southwestern Viti Levu, Fiji. He recognised that in the proximal area beds are typically massive and may show inverse grading. In the distal part graded sandstone and mudstone units occur and show incomplete Bouma sequences. These units are characterised by sharp-bases, and topped by lime mudstone which may be bioturbated. He interpreted that the deposition was primarily from high-density mass flows which were probably initiated by the upslope collapse of unstable volcaniclastic debris. Based on the sedimentological character, most of the Kerek Formation may be categorised into the distal facies of the lapilli-tuff and tuff breccia lithofacies of Busby-Spera (1988) or the graded sandstone and mudstone of Hathway (1994). Figure 5.11 shows the inferred depositional setting of the Kerek Formation, based on Hathway’s (1994) model.

5.3 KALIBENG FORMATION
5.3.1 Lithofacies

A typical succession through the Kalibeng Formation is represented by the Solo River section (Fig. 5.3) where the total thickness is about 425 m. The section was measured on the southern flank of an anticline north of Ngawi (Fig. 5.1). Texturally, this formation is nearly homogeneous and massive, and consists mostly of foraminiferal-rich pelagic/hemipelagic marl. Biostratigraphic studies by Lemigas/BEICIP (1974) and van Gorsel and Troelstra (1981) indicated that deposition occurred from the Late Miocene to Pliocene (planktonic foraminiferal zones N17 to N21; Chapter 3.2.2 & 3.3.3). Pringgoprawiro (1983) suggested the formation is contemporaneous with the Ledok and Mundu Formations of the Rembang Zone.

The base of the formation is characterised by about 10 m of bioturbated (Cruziana ichnofacies), carbonaceous sandy marl which lies directly on the calcirudite forming the top of the Kerek Formation. Some thin (15-20 cm) interbeds of sandy calcarenite contain coarse-grained to pebble size mudclasts, and show planar crossbedding and ripple laminations. Some of these beds are lensoidal and show an erosional base. Upward, the number and thickness of the sandstone beds diminish and the unit is dominated by unstratified, foraminiferal-rich marl.
The deposition of the middle and upper parts of the formation appeared to have been influenced periodically by strong bottom currents which are evidenced by the presence of some relatively thick crossbedded sandy units. The 7 m thick unit in the middle part of the formation is dominated by calcareous fine-grained sandstone and is characterised by westward-trending trough crossbedding (Fig. 5.5). The uppermost part of the formation is characterised by planar crossbedded glauconitic calcarenite. Here, the sandy character is due to an abundant occurrence of planktonic foraminiferal tests.

5.3.2 Petrography

Six rock samples have been studied microscopically, all of them from calcarenite and calcareous sandstone interbeds within the marl. The stratigraphic positions of the samples are indicated by prefix KB in Figure 5.3, and Appendix B.1 summarises the results of the analysis. Calcarenite samples from the lower part are characterised by bioclasts, mostly consisting of foraminiferal tests which range in abundance from 20 to 75%, surrounded by carbonate mud matrix, giving a total carbonate content of up to 95%. Fine-grained terrigenous clastic grains are minor, less than 6%.

Calcareous sandstone interbeds occur in the middle of the formation. They are characterised by fine- to medium-grained terrigenous fragments (plagioclase and hornblende) which form about 32% of the rock (Fig. 5.8). Rounded limestone fragments (about two thirds of total carbonate) are mostly muddy limestone (reworked marl). Bioclasts (mostly foraminiferal tests) are minor (< 10% of total carbonate).

In the upper part of the formation an abundance of foraminiferal tests characterises the calcarenite. They reach 70% of the total carbonate and are cemented by micrite. Glauconite reaches 21% (sample no. KB15), and mostly occurs interstitially within foraminiferal tests (Fig. 5.9). Terrigenous clastic grains, which are commonly plagioclase fragments, are minor (less than 2%).

An X-ray diffraction analysis was carried out on two marl samples, primarily to define the type of mineral in the formation. All diffractograms (Fig. 5.12) reveal that calcite is the major component. Clay minerals (kaolinite and smectite), quartz, plagioclase and mafic minerals are minor, indicating a low terrestrial influence and that most of the material has been reworked from calcareous faunas.
5.3.3 Depositional environment and discussion

The lithofacies and petrographic studies indicate that the Kalibeng Formation shows little facies variation and only minor terrigenous clastic input. This implies that the formation is composed primarily of marine derived materials which have been deposited by suspension sedimentation throughout the accumulation of the formation. The dominant composition, which is mainly an association of carbonate mud and foraminiferal tests, is typical of a pelagic deposit. This deposit accumulated during a period without major bottom current activity by particle by particle settling from the overlying water (cf. Reineck & Singh, 1980).

Because of these sedimentary features, in the past, interpretation of the depositional environment of this formation appeared to rely only on the results of benthonic faunal analysis. Lemigas/BEICIP (1974) estimated that the environment of deposition was outer neritic to upper bathyal, based on the association of benthonic foraminifers Bathysiphon, Dentalina, Planulina, Stilostomella, Nodosaria, Lenticulina, Gyroidina, Pyrgo, Oridorsalis and Bulimina. An extreme result has been obtained by van Gorsel and Troelstra (1981) using a similar analysis. They estimated a palaeowater depth of about 1000 m, based on the benthonic foraminiferal assemblage of Oridorsalis umbonatus, Gyroidina neosoldanii, Planulina wuellerstorfi, Anomalina globulosa and Uvigerina auberiana, whereas a slightly shallower condition (about 400 m) occurred near the upper boundary. Their conclusion, in fact, appears to disagree with Murray (1991) who documented that these species are commonly present in a wide range of environments, from shelf to bathyal.

The present study indicates that beside benthonic faunal evidence, there are at least three parameters which have to be considered before a reliable depositional environment can be determined for the Kalibeng Formation. These are: (1) detailed sedimentary features, (2) palaeogeography during deposition, and (3) eustatic sea level changes during the Mio-Pliocene.

A climatostratigraphic study in the Solo River section by van Gorsel and Troelstra (1981; Chapter 3.4) recognised the presence of three cold periods during the deposition of the Kalibeng Formation, which occurred at the beginning (cold period I), middle (cold period III) and end (cold period IV) of the deposition (Fig. 3.4). Biostratigraphic studies by van Gorsel and Troelstra (1981) and Pringgoprawiro (1983) indicated that the segment of the Kalibeng Formation between the cold periods I and III is laterally equivalent with the Ledok Formation in the Rembang Zone; the segment between the cold periods III and V corresponds with the
Mundu Formation (Chapter 3.3 & 3.6, Table 3.4). The cold periods I and III probably correlate with the sea level falls recognised on seismic line DNR-12 (Chapter 4.3.2), which occurred before and after the Ledok Formation deposition (seismic unit LM2).

Discussion presented earlier indicated that the sandy facies in the Ledok Formation, on both seismic lines PWD-23 and DNR-12, was deposited on a shallow shelf influenced by tidal currents (seismic unit LM2). Thus, from the facies difference, the lower part of the Kalibeng Formation probably represents a deeper facies of the Ledok Formation. Although van Gorsel and Troelstra (1981) did not provide any sedimentological description for the Kalibeng Formation related to their defined cooling periods, a field study carried out by the author recognised the occurrence of some sedimentary features which possibly relate to these periods.

The lowermost part of the Kalibeng Formation in the Solo River section, where the cold period I occurs, is characterised by 6 m thick calcirudite, which from the texture and composition was probably deposited in a high energy environment (see discussion in chapter 5.2). This, in turn, is overlain by about 10 m of bioturbated (*Cruziana* ichnofacies), carbonaceous sandy marl with some thin interbeds of sandy calcarenite, containing pebble-size mudclasts, and showing planar crossbedding and ripple laminations. Some of these beds are lensoidal and show an erosional base. These features suggest that the unit was probably deposited during alternation between storm and fair-weather bottom currents. The *Cruziana* burrows in the marl suggest that the unit was deposited in a sublittoral environment which is commonly typified by moderate to low energy levels (Frey & Pemberton, 1984), and was beyond the reach of normal wave base (Walker & Plint, 1992). The presence of some lensoidal sandy calcarenites with an erosional base suggests that the unit may have been subjected to storm-generated waves (cf. Reinson, 1984; Johnson & Baldwin, 1986). During a storm event, a sandy unit was deposited by a storm-induced current, followed by a bioturbated marl unit during the next fair-weather period.

The middle part of the Kalibeng Formation in the Solo River section, suspected to be the stratigraphic position of cold period III, is characterised by trough cross-bedded sandstone. This 7 m thick unit is indicated in Figures 5.3 (position 230 m) and 5.5, and petrographically, is characterised by up to 50% terrigenous clastic detritus. This unit appears to have been deposited as sand waves, which are common in an environment dominated by tidal currents (Dalrymple, 1992). This implies that during the cold period the depositional environment
became a shallower tide-dominated area. Haq et al. (1988) indicated that the magnitude of eustatic sea level fluctuations during the Late Miocene was about 95 m.

With regard to these environments, the bathyal depositional environment for the Kalibeng Formation, as suggested by van Gorsel and Troelstra (1981), appears to be incorrect, although there was a combined significant sea level rise and basin subsidence.

The upper segment of the Kalibeng Formation between cold periods III and IV has been correlated with the Pliocene marl Mundu Formation in the Rembang Zone. The deposition of this marl was widespread (Fig. 5.13), and appears to be in response to the significant eustatic sea level rise in the Early Pliocene (cf. Haq et al., 1988). The seismic sequence stratigraphic study of the Java Sea (Chapter 7) indicates a relative sea level rise of about 115 m. It must have occurred relatively quickly, as evidenced by the lack of recognisable transgressive systems tracts (note that the seismic system used has about 2 m resolution), and was followed by a slow relative sea level fall. Due to these circumstances, in the Java Sea, most Pliocene deposits appear to have been deposited under regressive conditions, and as sea level slowly fell, most of the deposits would be in middle and outer neritic environments. This type of deposition is believed to have occurred in the Kendeng Zone, where due to its tectonic position, the subsidence rate was probably higher than in the Java Sea. Most of the marl in the Kalibeng Formation may be the result of prolonged regressive conditions because of the slow relative sea level fall, following a rapid sea level rise in the Early Pliocene. The low content of siliciclastic detritus in this formation, as evidenced by the petrographic data, appears to support this mode of deposition as such detritus would be deposited closer to the basin margin.

5.4 ATASANGIN FORMATION

The Atasangin Formation was developed in the eastern part of the Kendeng Zone, and was deposited in the Middle Pliocene (Chapter 3.2.2 & 3.3). The succession was first mapped and described by Duyfjes (1938a,b,d) and subsequently by de Genevraye and Samuel (1972), but interpretation of its depositional environment and setting has never been carried out. Two facies typically occur: a deltaic sandstone facies in the lower part and a finely-laminated siliceous marl facies in the upper part. The total thickness in the eastern Kendeng Zone is about 400 m.
Because of the scarcity of good outcrops, study of this formation could only be carried out on the north Kabuh section, which is located some 1.5 km north of Kabuh (Fig. 5.1). The Atasangin Formation in this section is underlain by the unstratified marly (prodelta) Kalibeng Formation, which here, was probably deposited in a shelf environment, based on the content of benthonic foraminifers *Bolivina, Bulimina, Cibicides, Laticarinina, Planulina* and *Siphonina* (Appendix C.2, sample no. NKabuh3; cf. Murray, 1973; 1991). Representative sections of the Atasangin Formation are given in Figures 5.14 and 5.15.

### 5.4.1 Lithofacies

#### a. Sandstone facies

The sandstone facies occupies the lowermost 90 m of the Atasangin Formation; Figure 5.15 shows the representative vertical section of this facies. Overall facies show a regressive upward sequence which based on sedimentary textures and structures, can be subdivided into two main subfacies: coarsening-upward sandstone subfacies, and fine- and coarse-grained sandstone facies.

##### a.1 Coarsening-upward sandstone subfacies

**Facies:**

This facies occupies the lowermost 35 m of the north Kabuh section, and is directly underlain by the Kalibeng Formation (Fig. 5.15).

The lowermost 17 m is characterised by alternating foraminiferal-rich mudstone and calcareous sandstone on a centimetre to decimetre scale, which is overlain by massive and parallel laminated sandstone consisting of fine-grained bioclastic (foraminiferal tests), plagioclase and lithic fragments (andesite).

The uppermost 18 m consists of a coarsening-upward sequence of medium- to coarse-grained sandstone. Thickness of successive sandstone units progressively increases from 20 cm to 3 m, and they are separated by 5 to 20 cm sandy mudstone beds which commonly contain burrow structures. Ripple lamination is the most common structure in the sandstone units. Mudclasts, 2 to 5 cm diameter, are common in the upper parts.
Interpretation:

A coarsening upward sequence is commonly associated with a prograding wave- or storm-dominated shoreface (Reinson, 1984; Bhattacharya & Walker, 1992) but it may also represent a delta front facies in a wave-dominated environment (Bhattacharya & Walker, 1992). The differentiation between these two environments appears to rely on the availability of three-dimensional control. As such control is absent in this study, a reliable interpretation may be difficult to reach. However, overall sequence tendency, which regresses toward an association of possible channel levee complexes in a delta plain environment (sub-section a.2) suggests that the coarsening-upward sequence may have been deposited in a delta front environment. In this environment, a coarsening-upward sequence occurs due to the amalgamation of a series of coalesced distributary mouth bar crests and their associated distal bars (Cotter, 1975).

The lower part, which is characterised by interbedded fine-grained sandstone and mudstone, is interpreted to represent a transition from the underlying prodelta facies (the marly Kalibeng Formation). The abundant foraminiferal tests in the observed facies indicate a relatively strong open marine influence.

a.2 Fine- and coarse-grained sandstone subfacies

Facies:

This subfacies overlies the upward-coarsening sandstone subfacies and attains about 54 m in thickness.

The fine-grained sandstone subfacies is characterised by interbedded sandy marl/mudstone and fine- to coarse-grained sandstone. In the lower facies (interval between 36 and 50 m; Fig. 5.15), the sandstone beds gradually thicken upward from 10 to 25 cm, are separated by 5 to 20 cm thick mudstone beds, and are characterised by common parallel and ripple laminations. Deformation structures (such as convolute bedding, Fig. 5.16) and mudclasts are also common. Some of these sandstone horizons are normally graded into parallel laminated mudstone (Fig. 5.16). In the upper facies (interval between 60 and 68 m, Fig. 5.15), the facies is characterised by interbedded parallel laminated mudstone and parallel bedded, fine- to coarse-grained sandstone. Some of the sandstone units are considerably thicker, up to 1.5 m, and contain very small amounts of planktonic foraminifers.
The coarse-grained sandstone subfacies consists commonly of fining-upward units with basal scours and lag deposits. The sandstone is moderately sorted, without muddy matrix and consists mainly of rounded andesite fragments. Crude bedding and parallel bedding are common in the coarser-grained sandstone. In the lower part (interval between 50 and 55 m; Fig. 5.15), the facies occurs as small scale channel-fills, which have an average width and depth of about 1.5 m and 60 cm respectively (Fig. 5.17). In the uppermost part, (interval between 83 and 85 m; Fig. 5.15), the facies is characterised by coarse-grained sandstone and breccia, which tend to be crudely bedded, and by consolidated blocks of mud (up to 10 cm). The main fragment composition is andesitic rock and mud blocks.

**Interpretation**

The fine-grained sandstone subfacies in the lower part (interval 36-50 m; Fig. 5.15) is characterised by some thin fining-upward beds and deformation structures. A fining-upward sequence commonly develops in fluvial deposits and is related to lateral migration of point bars (Allen, 1970). However, the absence of basal scours between this facies and the underlying facies (upward-coarsening sandstone facies) suggests that these fining-upward beds are not related to a point bar system. Moreover, Coleman and Prior (1982) indicated that fining-upward units could occur in sandy crevasse splays of a delta system.

The occurrence of deformed structures in this fine-grained sandstone facies (interval 36-50 m; Figure 5.15) indicates that the sediment was partially liquefied during deposition of the sandy facies (cf. Allen, 1977; Reineck & Singh, 1980). These sedimentary structures are not unique to a single environment and could develop in at least three different environments: channel bar deposits (Coleman, 1969), subaqueous levees of a delta front environment (Reineck & Singh, 1980), and turbidite systems (Middleton & Hampton, 1976). The close association of this fine-grained sandstone facies with fining-upward beds and distinct channel structures (interval 50-55 m; Fig. 5.15) suggests an occurrence in a channel-levee complex of a delta plain environment. The occurrence of foraminifera-bearing sandstone beds (interval 55-68 m; Fig. 5.15) indicates a rather strong marine influence (an interdistributary bay).

The coarse-grained sandstone facies in the upper part (interval 68-90 m; Fig. 5.15) was probably deposited as channel fill deposits, as evidenced by the occurrence of basal scours and lag deposits. The fining-upward sequences in this facies may be the result of lateral migration of point bars within a meandering river or distributary system, as characterised by Allen (1970)
and Walker and Cant (1984). The occurrence of coarse-grained sandstone, breccia and mud blocks in the uppermost part suggests possible temporary flood events. The coarse-grained sandstone facies, thus, may represent a mid to upper delta plain environment on which riverine processes are dominant (Wright, 1982).

b. Diatomaceous marly facies

The diatomaceous marly facies developed during a transgression following the deposition of the previous deltaic sandstone facies. The facies consists mostly of light grey, finely-laminated marls with some thin interbeds (15-20 cm) of very fine-grained calcareous sandstone. Diatom frustules, as well as radiolaria tests and spongidiscs, are abundant, indicating the study area was sufficiently nutrient-rich to allow a high productivity of diatoms. This diatomaceous marly facies always shows a consistent parallel-bedding enhanced by the diatom-rich beds. Massive beds occur locally particularly in the middle part of the succession (Fig. 5.14).

The depositional environment as defined from the benthonic faunal analysis (foraminifers) during this study (Chapter 5.4.3) is inconclusive, particularly because of the occurrence of mixtures of inner (brackish/hypersaline) and middle/outer neritic assemblages. Duyfjes (1938b) reported the occurrence of some fresh water molluscs (Unio) in the upper part of the siliceous marl in the Atasangin Formation on an area some 6 km northeast Kabuh (Fig. 5.30). This suggests that the diatomaceous marly facies was probably deposited in a marginal marine environment.

5.4.2 Petrography

The petrographic study was based on three representative samples; one from the deltaic sandstone facies was studied by using a transmitted light microscope whereas the other two (from the siliceous marly facies) were analysed by X-ray diffraction.

The sandstone sample consists mainly of andesitic (and some basalt) terrigenous clastic detritus (Appendix B.1). Rock fragments (30%) are the most abundant clastic component along with plagioclase (17%) and mafic minerals (13%) (Fig. 5.18). Fragments are cemented by sparry calcite (34%). The rock fragments are commonly oxidised and glauconitised with the grain size varying from 0.05 to 0.7 mm. They show a rounded form.
The X-ray diffractograms of the siliceous marly facies (Fig. 5.19) show minor occurrence of two clay minerals (smectite and kaolinite) which probably reflects a minor terrigenous influence or minimal weathering in the source area (Hardy & Tucker, 1988). The high siliceous content in the Atasangin Formation is reflected by the high content of quartz in the samples analysed. A high calcite content is displayed by sample no.NKabuh 13. This sample is from the middle part of the succession which is characterised by abundant calcareous fauna.

5.4.3 Benthonic fauna

Benthonic foraminiferal analysis was carried out on eight marly samples, mostly from the diatomaceous marly facies. The results of the analysis are summarised in Appendix C.2, and the stratigraphic positions of the samples are indicated on the vertical section Figure 5.14.

All samples analysed are characterised by low species diversity, averaged at 9 species per sample. This causes the maximum species diversity index of Murray (1973) to fall below 3. Murray (1991) indicated that a diversity index of less than 5 is typical of brackish or hypersaline marginal marine environments. A rather high species diversity (21 species) occurs in sample no. NKabuh16, which is stratigraphically located in the middle part of the succession (Fig. 5.14). This part corresponds with the occurrence of marl beds with massive structure, and is interpreted to represent normal salinity marine water.

The faunal association in most of the samples studied appears to be a mixture of inner and middle/outer neritic faunas (Appendix C.2). This is apparent in samples NKabuh12, NKabuh16 and JKB10, which represent the lower, middle and upper parts of the section (Fig. 5.14), respectively. The genera *Ammonia* and *Elphidium*, which commonly characterise brackish and hypersaline lagoons, occur associated with middle/outer neritic genera *Planulina*, *Laticarinina*, *Lenticulina*, *Bolivinita* and *Gyroidina* (cf. Phleger, 1960; 1965; Boucot, 1981; Murray, 1973; 1991).

Such unusual assemblages have been reported by some other authors, and could result by two possible means, upwelling or tidal currents. Sen Gupta et al. (1981) documented the unusually high abundance of *Bolivina subaenariensis* on the continental slope off Daytona Beach (Florida), whereas this species is generally very rare along the western North Atlantic margin. They suggested that because of the varying influence of the Gulf Stream and seasonal upwelling, the environment undergoes abrupt physical and chemical changes which favour the
species *Bolivina subaenariensis*. Sloan (1992) reported the occurrence of open marine-type ostracods in the Late Pleistocene Yerba Buena mud, San Francisco Bay (California), which elsewhere is commonly characterised by estuarine taxa. The Yerba Buena mud represents the deposits of a single transgressive event. The deposition commenced during the last glacial maximum with brackish and low salinity environments continuing to the present day bay (25-70 m below mean sea level). He interpreted that the open marine ostracods were introduced into the bay by tidal currents which became more prominent as sea level increased.

5.4.4 Discussion

The sedimentary texture and composition of the deltaic sandstone facies of this formation suggest that volcanic detritus was the main source. Lemigas/BEICIP (1974), de Genevraye and Samuel (1972) and Pringgoprawiro (1983) reported the occurrence of volcanic breccia at the same stratigraphic position as the deltaic sandstone facies in areas near the Pandan Volcano (Fig. 5.1). The absence of similar facies in some areas, such as Ngawi and Ngimbang (Fig. 5.1), indicate that the breccia and deltaic deposits are probably very localised around aprons of some volcanoes. Figure 5.20 displays the inferred palaeogeography during the deposition of the deltaic sandstone facies in the Middle Pliocene.

The facies difference between the deltaic sandstone facies and the underlying marly Kalibeng Formation suggests that the deltaic deposition accompanied a relative sea level fall. The occurrence of deltaic progradation during such times has been indicated experimentally by Koss *et al.* (1994). Biostratigraphic studies by Ninkovich and Burckle (1978) and the author during the present study (Chapter 3.3.1c and 3.3.2) indicated that the Atasangin Formation was deposited during the Middle Pliocene (planktonic foraminiferal zones N19/N20. Tentative correlation with the Haq *et al.* (1988) eustatic sea level curve suggests that the deltaic deposition corresponds with the eustatic sea level fall 3.8 Ma ago.

The siliceous marly facies was probably deposited during the transgression which followed deltaic deposition. Abundant diatoms, as well as other siliceous taxa, suggest that the environment of deposition was rich in nutrients. Jenkyns (1986) indicated that nutrient-rich water is introduced when deep-water inflow and shallow-water outflow prevails. Calvert (1966) documented the occurrence of large accumulations of diatomaceous siliceous sediments in the Gulf of California, and suggested that circulation of upwelling currents supplied dissolved silica from the Pacific Ocean and was responsible for the formation of these deposits.
A biostratigraphic study in the Solo River section (Figs 3.4 & 5.1) by van Gorsel and Troelstra (1981) indicated that during the Middle Pliocene, deposition of the marly Kalibeng Formation occurred in a middle/outer neritic environment (chapter 5.3). In the eastern Kendeng Zone, which is represented by the north Kabuh section, a shallower marginal marine environment was probably present as indicated by the low species diversity (?lagoonal/estuarine; Fig. 5.21). The unusual faunal assemblage in the siliceous marly facies in this section may reflect the occurrence of water circulation between the shallow marine environment in the eastern Kendeng Zone and the deeper marine facies in the western Kendeng Zone (represented by the Solo River section), which also led to the introduction of nutrient-rich water to allow the development of siliceous taxa.

5.5 SONDE FORMATION

5.5.1 Lithofacies

A vertical section of Sonde Formation was measured on an exposure along the east bank of the Solo River, north of Ngawi (Fig. 5.22). The formation was deposited during Middle to Late Pliocene mostly in the central and west Kendeng Zone (Chapter 3.2.2). It is conformable with the Late Miocene-Early Pliocene Kalibeng Formation.

The Sonde Formation consists of two main facies, the limestone and mudstone facies. The limestone facies is approximately 65 m thick, and consists of bedded calcarenite at the base and a coarse clastic limestone unit containing coral fragments and algae in the upper part (Fig. 5.23 and Fig. 5.24).

Van Gorsel and Troelstra (1981) identified the presence of benthonic foraminifers Rectobolivina, Bolivina, Uvigerina, Trifarina, Globocassidulina and Hoeglundina near the base, and species Amphistegina papillosa and Planorbulinella elatensis in the middle part of the limestone facies, from which they suggested a progressive shoaling of the sequence from an outer to inner neritic environment of deposition.

The calcareous mudstone facies of the Sonde Formation is about 80 m thick, and conformably overlies the limestone facies. This mudstone facies is poorly exposed but, from a few outcrops, it consists of massive calcareous sandy mudstone with abundant pebble-size mudclasts (intraclasts). This facies also contains planktonic and benthonic foraminifers, but according to van Gorsel and Troelstra (1981) most foraminifers were reworked from the Early
and Middle Pliocene sediments. These features suggest that the mudstone facies was deposited in a wave-dominated region.

5.5.2 Discussion

The vertical facies changes from the marl of the Kalibeng Formation and within the Sonde Formation clearly indicate an upward regressive sequence. In the central Kendeng Zone (including the Ngawi area), the stratigraphic position of the Sonde Formation, which is overlain directly by the debris flow deposit of the Pucangan Formation, suggests that the marine stage has ended in this area. This is in marked contrast to the eastern Kendeng Zone (including the Jombang and Mojokerto areas, and Madura Strait) where a high rate of basin subsidence resulted in a thick succession of siliceous marly facies of the Atasangin Formation and lacustrine mudstone of the Lidah Formation.

5.6 LIDAH FORMATION

5.6.1 Lithofacies

a. Kabuh and Mojokerto areas

The Lidah Formation in these areas was deposited in the Late Pliocene (Chapters 3.2.2 & 3.3.2), and conformably overlies the Atasangin Formation. The first detailed mapping in these areas was carried out by the Geological Survey of the Netherlands Indies in the 1930s (Duyfjes, 1938a,b,c,d). Despite a few subsequent studies (e.g. de Genevraye & Samuel, 1972; Sartono et al, 1981; Pringgoprawiro, 1983), the sedimentological characteristics of the Lidah Formation have never been examined in detail. Previous studies have mainly examined the stratigraphic context of the formation.

In the Kabuh and Mojokerto areas the Lidah Formation attains a thickness of 400 m. A representative vertical section in these areas is provided in Figure 5.25. The section was measured in the Sumberringin area, some 3 km northeast of Kabuh (Fig. 5.1). Here, the formation is characterised mainly by unstratified, homogenous dark blue mudstone. Sedimentary structures are absent and the bedding orientation can be measured only when a thin interbed of sandstone is present. This facies is also characterised by streaks of organic material and some fine-grained pyrite crystals indicating anoxic conditions at the time of deposition.
In these areas, there are at least two sandstone horizons which indicate temporary exposure to a marine environment. The lower sandstone horizon is about 10 m thick. A detailed study was not carried out on the lower sandstone facies due to poor exposure. A faunal study on rock samples from this unit indicates the presence of benthonic foraminifers including the species *Uvigerina peregrina*, *Bulimina marginata* and *Bolivina* (Appendix C.2) which suggests an inner to middle neritic environment (Phleger, 1960; Murray, 1991).

More detailed observations were made on the upper sandstone facies of the Lidah Formation. Figure 5.26 shows the generalised section of this facies, measured in the Sumberringin area, some 3 km northeast of Kabuh (Fig. 5.1). This facies is characterised by a regressive sequence, with a total thickness of 27 m, and can be subdivided into lower, middle and upper parts, which represent three different environments of deposition. The lower part (between 2.5 and 6.5 m in Fig. 5.26), consists of massive, bioturbated, muddy sandstone intercalated with ripple laminated, moderately sorted, medium-grained sandstone. Inner neritic benthonic foraminifers such as those found in the lower sandstone facies are abundant. This lower part is interpreted as a subtidal deposit, but the absence of slack-water mud drapes [common in this environment, Terwindt (1988)] and abundant bioturbation suggest an environment closer to offshore facies. The middle part (between 6.5 and 21 m in Fig. 5.26) is characterised by muddy sandstone which fines upward into massively bedded carbonaceous mudstone. Clay and organic pellets are abundant. This part probably represents a tidal flat environment. The upper part (between 21 and 29 m in Fig. 5.26) is dominated by fining upward sandstone facies. The base of the facies is erosional and contains coarse-grained to pebble-size, rounded fragments of andesite as a lag deposit. Planar crossbedding is common and ripple stratification is found in some thin intercalations of fine sandstone. This part is interpreted to represent a fluvial or tidal channel facies.

b. Nglampin area

The Lidah Formation is widely developed in the Rembang Zone. The Lidah Formation in the Nglampin area (some 25 km southeast of Cepu, Fig. 5.1) has been studied by Nahrowi *et al.* (1981). Here, only the uppermost 400 m is exposed. Their biostratigraphic study suggested that the sequence was developed in the Pleistocene (zones N22 and N23) based on the present of planktonic foraminifers *Globorotalia hirsuta* and *Globigerina calida calida*, and laterally corresponds with the Pucangan Formation in the Kendeng Zone. Figure 5.27
represents the vertical section of the Lidah Formation in the Nglampin area, measured by Nahrowi et al. (1981).

The exposed section can be divided into three main facies: a lower mudstone, an intermediate limestone and an upper mudstone facies. The lower mudstone facies is characterised by massive carbonaceous bluish-grey mudstone, and contains streaks of fine-grained quartz sandstone. Fine-grained pyrite crystals are occasionally found. Some thin interbeds (< 10 cm) of fine-grained calcareous sandstone are present and commonly show parallel and ripple lamination. This lower mudstone facies may have been deposited mostly in anoxic conditions, as were similar facies in the Kabuh and Mojokerto areas.

The limestone facies is characterised by sandy mudstone interbedded with calcarenite. The sandy mudstone interval varies from 40 cm in the lower part to 7 m in the upper part. The calcarenite is glauconitic and very rich in mollusc fragments which suggests the influence of strong waves during the deposition. The thickness of calcarenite varies from 15 cm to 1.5 m and shows a tendency of thickening upward. This facies may have been deposited in a fairly shallow marine (?shoreface/foreshore environment).

The upper mudstone facies consists of light grey calcareous mudstone and in some places is carbonaceous and bioturbated, suggesting a tidal flat environment of deposition. This facies is also characterised by thin interbeds (< 60 cm) of calcarenite, particularly in the upper part, which may represent a temporary occurrence of strong waves. Faunal evidence in this facies supports a tidal flat environment of deposition, as indicated by the occurrence of Ammonia (Nahrowi et al., 1981).

5.6.2 Petrography

a. Microscope analysis

A petrographic study was carried out on three sandstone samples from the upper sandstone facies of the Lidah Formation in the Sumberringin area (the results are summarised in Appendix B.1). The subtidal sandstone is characterised by the presence of fine-grained quartz fragments (5%), and mollusc fragments and foraminiferal tests (16%). These fragments are supported by a calcareous clay matrix. Feldspars are minor (< 4%) indicating a low volcanic influence. The quartz is mainly monocrystalline and shows undulose extinction indicating strain during its crystallisation. It was possibly reworked from the Middle Miocene sediments
The sandstone in the tidal channel deposit is characterised by a different mineralogical content to the subtidal deposit, and suggests a volcanic origin. This sandstone consists of highly altered, rounded to subrounded fragments of basaltic and andesitic rocks representing up to 72% of the total sediment. Plagioclase and mafic minerals represent up to 30% and 10% respectively of the total sediment. At the base of the channel the sandstone is glauconitic and cemented by carbonate mud.

b. X-ray diffraction analysis

X-ray diffraction analysis was carried out on three mudstone samples from the Kedamean and Nglampin areas (Fig. 5.1; Appendix B.2).

The diffractograms of the samples from the Nglampin area (Fig. 5.28) show a wide background spectrum which is known to be typical of organic-rich samples (Mandile & Hutton, 1994). This is in agreement with the general appearance of the mudstone facies which is carbonaceous. The diffractograms also indicate the presence of the evaporite mineral gypsum. The precipitation of evaporite minerals from brines is controlled by their relative solubility (Kendall, 1992). Thus the occurrence of gypsum indicates that the environment was saltier than normal sea water. With the assumption that sea water was the main source of brine (Kendall, 1992), during the deposition of the mudstone facies the area may have been subaerially exposed and only temporarily been flooded with sea water. This suggestion is also supported by the occurrence of siderite in some of the samples (Nglampin1, Nglampin7 & Kedamean4). Siderite is commonly associated with pedogenic concretions (Collinson, 1986).

The diffractogram of the sample from the Kedamean area indicates that smectite and kaolinite are the most common clay minerals in the Lidah Formation in this area. Smectite is often related to the alteration product of volcanic ash and kaolinite is related to the weathering of Al-rich silicates (Deer et al., 1992; Hardy & Tucker, 1988). With regard to the greater X-ray response of kaolinite than smectite (2.5 times; Weir et al., 1975), the diffractograms suggest that smectite is more abundant than kaolinite, indicating the source was mostly basic or intermediate volcanism where Al-rich silicate minerals are less common.
5.6.3 Discussion

The Lidah Formation is up to 500 m thick in the Kedamean area (north of Mojokerto). It is also extensively developed in East Java. Field observations during the present study along the south coast of Madura and Gili Raja Islands, and the seismic study on Madura Strait (Chapter 6.5.2) confirm the presence of similar facies in these areas. On the south coast of Madura and Gili Raja Islands, as well as in the Rembang Zone (Lemigas/BEICIP, 1969), this formation is underlain unconformably by Pliocene deposits (Mundu Formation and its equivalent). The lateral extension of the Lidah Formation, thus, appears to be confined within an east-west elongated basin, bordered in the north by the Rembang Zone anticlines which extend from the Blora area to Madura Island, and bordered in the south by the Plio-Pleistocene volcanic centres (Fig. 5.29). Petrographic evidence suggests that the Rembang Zone was exposed subaerially and was the source of sediments in the lower part of the formation, as well as the Plio-Pleistocene volcanism in the south which became dominant in the uppermost tidal channel facies.

The evidence presented here suggests that a large lacustrine basin was the site of deposition of the Lidah Formation in the Plio-Pleistocene, which was characterised by organic-rich and stagnant water (anoxic conditions as indicated by pyrite content). This stagnant water condition was probably responsible for the minor facies variations within this formation. The thickness, which attains 500 m in the Kendeng Zone during a relatively short time (Late Pliocene-Early Pleistocene), suggests that the subsidence and sedimentation rates were high. The occurrence of siderite and gypsum minerals, and interbeds of marine-influenced deposits in the Lidah Formation, further suggest that the lacustrine environment had been subjected to drying and flooding events which are suspected to be related to eustatic sea level changes. The Haq et al. (1988) eustatic sea level curve indicated that there are at least four recognisable fluctuations during the Late Pliocene-Pleistocene. A more detailed study in the Java Sea by the author (Chapter 7) indicates that during just the Late Pliocene-Early Pleistocene there are five sea level fluctuations (represented by seismic units JP1, JQ1a,b and JQ2a,b; Chapters 7.4.2 & 7.4.3).

The development of this stagnant lacustrine basin ceased when there was a significant relative sea level rise, as indicated by greater marine influence on the upper mudstone facies in the Nglampin section (Fig. 5.27), and the development of the Pucangan Formation in the Kabuh and Mojokerto areas, which is characterised by intercalations between tidally influenced
and volcaniclastic deposits. The transition from the Lidah to Pucangan Formations in Madura Strait is further discussed in Chapter Six.

5.7 PUCANGAN FORMATION

The Late Pliocene/Early Pleistocene Pucangan Formation (Chapters 3.2.2 & 3.5) of the Kendeng Zone conformably overlies the Lidah Formation. Its sedimentation mostly relates to the Quaternary volcanism along the south of this zone. In the eastern Kendeng Zone, where the present study is concerned, it is related to the Wilis Volcanism (Duyfjes, 1938b; van Bemmelen, 1949).

The first study of the Pucangan Formation was carried out by Duyfjes (1938b) and was followed by de Genevraye and Samuel (1972). Duyfjes’s (1938a,b,c,d) studies were very detailed, and his comprehensive geological maps are still very useful for contemporary researchers. However Duyfjes (1938b), as well as de Genevraye and Samuel (1972), did not examine the sedimentological aspects of the Pucangan Formation. The present study re-evaluates the environments of the depositional units of the Pucangan Formation, particularly in the light of contemporary facies models (e.g. Reineck & Singh, 1980; Walker, 1984, 1992; Reading, 1986). Poor outcrop is the main constraint to the study, and discussion presented in this section is based on widely spaced vertical sections.

5.7.1 Lithofacies

Examination of the Pucangan Formation was carried out at seven locations, that were considered to provide representative exposures of the general character of this formation: the Solo River, Kabuh, Kemlagi, Perning, Banyuurip, Kedamean and Raci areas (Figs 5.1. & 5.30). Because detailed mapping was not carried out during the study, lithostratigraphic correlation between the sections was facilitated by a detailed geologic map produced by Duyfjes (1938b,d) (Fig. 5.30).

The study reveals that the facies associations developed in the Pucangan Formation mostly result from the introduction of volcaniclastics into shallow marine settings, as shown by intercalations between marine-influenced and coarse volcaniclastic deposits. Based on textures, sedimentary structures, faunal content and facies association, most marine-influenced sedimentary units in the Pucangan Formation can be explained by using the tidal flat facies
model of Reineck and Singh (1980), Terwindt (1988) and Dalrymple (1992) (Fig. 5.31); and the barrier island facies model of Reinson (1984) (Fig. 5.32).

Textures, sedimentary structures and faunal content within units in the Pucangan Formation suggest that this formation can be divided into three major regressive sequences, where each sequence commences with a marine-influenced deposit and terminates with a vertebrate-bearing conglomerate (volcaniclastics). The marine-influenced deposits in all three sequences are always associated with abundant marine molluscs (*Volva, Erosaria, Erronea* and *Palmadusta*; Duyfjes, 1938b,d). Duyfjes (1938b,d) has used these three intervals of marine molluscs for biostratigraphic correlation and named them Molluscan Horizons 1, 2 and 3. The present study found that conglomeratic facies (volcanic facies) also represent useful lithostratigraphic markers. This latter facies occurs widely in the study area and is easily identified.

a. Lower sequence

Three measured vertical sections provide the basis of the discussion; these are located in the Solo River (north of Ngawi, Fig. 5.22), Kabuh (Fig. 5.33) and Kemlagi (Fig. 5.34) areas. Stratigraphically, this sequence directly overlies the Lidah Formation (in the eastern Kendeng Zone) or the Sonde Formation (in the central Kendeng Zone). Relatively good exposures occur in the Solo River section and on a road cut north of Kabuh village, but the sequence is poorly exposed in the Kemlagi area. The sequence subdivision refers the Kabuh section in which six units can be identified, referred to as units L1 to L6.

a.1 Muddy and matrix-supported conglomerate

**Facies:**

In the Solo River section this deposit attains 120 m thickness and comprises at least six sequences dominated by massive, unsorted, matrix-supported conglomerate (Figs 5.22, 5.35 & 5.36). Fragments vary from pebble-sized to about 1.5 m and include igneous lithologies and older pyroclastic fragments. Indications of inverse and normal grading are not obvious which suggests that the sediment gravity flows were highly viscous. In the uppermost part, conglomerate clasts are imbricated and planar cross-bedded, and the unit is overlain by interbedded planar cross-bedded sandstone and non-fossiliferous mudstone.

In the Kabuh section this facies occurs in two main horizons, denoted by L2 (about 32 m) and L4 (about 20 m, Fig. 5.33). Texturally these units are similar, characterised by matrix-
supported conglomerate which normally grades into coarse-grained sandstone. The lower portion of unit L2 displays inverse grading. The conglomerate fragments include pebble-sized andesite and boulder-sized tuffaceous mudstone clasts. A microscopical examination on the sandstone matrix of unit L4 revealed the presence of foraminiferal tests, suggesting subaqueous deposition.

**Interpretation:**

This facies consists mostly of debris flow deposits. This type of deposit is commonly characterised by muddy-matrix supported clasts, is unstratified and poorly sorted and shows no preferred orientation of clasts (Reineck & Singh, 1980; Walker, 1984). However, the strength of the muddy matrix may force larger clasts to concentrate above the base of the bed which results in a basal inverse grading or preferred clast alignment (Walker, 1984) as shown in unit L2. Reineck and Singh (1980) indicated that in more fluid debris-flow deposits, graded bedding is common and flat pebbles are oriented more horizontally. Bull (1977) and Walker (1984) suggested that debris flows are commonly associated with a steep depositional slope, and enhanced by lack of vegetation and short periods abundant water supply (Bull, 1977). The abundance of volcanic materials in this deposit suggests an association with volcanism. Duyfjes (1938b) and van Bemmelen (1949) related these deposits to the Wilis Volcanism.

Sediments in the upper portion of the Pucangan Formation in the Solo River section, display traction current structures associated with mudstone. These deposits can be interpreted as the result of reworking by fluvial processes. Van Bemmelen (1949), who studied this formation regionally, indicated that these deposits are widespread in the western part of the Kendeng Zone and are dominated by black mudstone. He reasoned that the deposits represent a lacustrine environment, as the result of damming drainage systems by debris flow deposits. A recent study by Stollhofen and Stanistreet (1994) on the interaction between volcanism and fluvial sedimentation in the Permo-Carboniferous Saar-Nahe Basin (southwest Germany) found a similar case. They demonstrated that sedimentary sections show a transformation from a pre-eruptive meandering fluvial system to a lacustrine system following lava eruption.

a.2 Fossiliferous muddy sandstone

**Facies:**

The units included in this facies are units L1, L3a and L5 of the Kabuh section (Fig. 5.33) and almost the whole lower sequence of the Kemlagi section (Fig. 5.34), except the
conglomeratic sandstone unit L6. This facies is essentially highly bioturbated muddy sandstone and sandy mudstone, but parallel bedding is often observed in units in the Kemlagi section, and cross-bedding occurs in unit L1 of the Kabuh section. A notable feature of these units is the common occurrence of planktonic and benthonic foraminifers. Benthonic foraminiferal analysis in unit L3a (Appendix C.2, sample no. JKB7) indicates the presence of an inner neritic faunal association such as the foraminiferal genera Ammonia, Cibicides, Eponides, Pseudorotalia and Quinqueloculina (Murray, 1973; 1991).

Interpretation:

The inferences of depositional environment based on the sedimentary structures are difficult due to strong bioturbation. But faunal evidence suggests that this facies may have been deposited in a shallow marine offshore facies. Unit L1, which is characterised by cross-bedding, probably was deposited in an environment closer to subtidal facies. This facies is commonly characterised by sandbars (Reineck & Singh, 1980). For the Kemlagi section, the interpretation is also supported by the fact that in this section, the conglomeratic units L2 and L4, and tidal flat unit L3b of the Kabuh section are missing. This is interpreted as the wedging out of these facies into deeper marine facies.

a.3 Cross- and ripple-bedded sandstone

Facies:

This facies includes units L3b and L5 of the Kabuh section, with a thickness of 40 m and 2 m respectively. Various types of large scale cross-beds bundled by mud and shell fragments are common in unit L3b, and bioturbation is common in unit L5. Cross-bedding cosets in unit L3b vary in thickness from 20 cm to 1.5 m, and often feature plant remains and mudclasts. Scouring and reactivation surfaces on cross-stratification are also observed. Interbeds of fine to medium sandstone with parallel to ripple laminations, combined with mud draping (mud flasers), are often developed between the cross-bedded cosets with the interbed thickness ranging from 5 to 75 cm (Figs 5.38 & 5.39).

Interpretation:

The depositional environments were defined particularly based on the sedimentary structures. The determination based on the vertical sequence of facies is practically impossible due to the occurrence of conglomeratic facies (debris flow) which replaced the normal trend of vertical facies change.
The associated sedimentary structures suggest that units L3b and L5 were deposited in a tidal flat environment. This environment occurs on a flat to gently sloping area and geographically extends from the supratidal zone through the intertidal zone, into the upper portion of the subtidal zone (Dalrymple, 1992). The facies distribution and sedimentary structures in these environments largely reflect the strength of tidal currents generated during flood and ebb relative to the geographic position. Tidal channels commonly occur on the outer edge of a tidal flat environment. A relatively strong tidal current occurs in such channels and produces a sandy character adjacent to the channel which gradually passes into mud near the high tide line (Dalrymple, 1992). As a common practice, the tidal flat has been divided into sand flat, mixed flat and mud flat subenvironments (cf. Reineck & Singh, 1980; Dalrymple, 1992; Fig. 5.31).

Unit L3b is characterised by various types of large scale cross-bedding bundled by mud and shell fragments. Terwindt (1988) and Dalrymple (1992) pointed out that bundle mudstone occurs during slack water periods, and reactivation surfaces represent erosion by subordinate currents of to the preceding structures (dunes) formed during flow of the dominant current. The dominance of cross-bedded sandy units further suggests a subtidal and outer portion of intertidal (sand flat) environment. The tidal current speeds during both flood and ebb tides are high in these environments (Dalrymple, 1992). Unit L5 is bioturbated, and may represent a mixed flat environment. Bioturbation is commonly weakest in the sand flat and becomes stronger toward the mudflat environment (Reineck & Singh, 1980).

### a.4 Normally graded conglomeratic unit

**Facies:**

This facies includes unit L6, and is observed in the Kabuh and Kemlagi sections (Figs 5.33 & 5.34). It is the most widespread unit and is an important stratigraphic marker. The facies is about 50 m thick in the Kabuh section, thinning to about 25 m in the Kemlagi section, and completely wedges out in the western part of the Guyangan Anticline (Fig. 5.30). In the Perning section (Fig. 5.37), the facies is not entirely exposed but forms a broad plain of conglomeratic sandstone in the core of the Kedungwaru Anticline.

The main character of this facies is conglomeratic, but the overall facies is dominated by medium- to coarse-grained sandstone. The conglomeratic facies, as observed in the Kabuh and Kemlagi sections, commonly represents the base of the facies. It is composed mostly of
massive to crudely bedded cobble-size fragments of rounded andesite, supported by a matrix of muddy coarse-grained sandstone. In the Kabuh section, the thickness of the conglomeratic unit attains 3 m. The conglomeratic facies tends normally to grade upward into medium- to coarse-grained sandstone. Here, the sandstone is commonly associated with planar and trough cross-beds. But it also includes pebble-size mudclasts and marine molluscs in the foreset laminae which indicate the presence of a marine reworking process.

**Interpretation:**
This facies probably represents an alluvial fan deposit as indicated by the dominance of coarse clastics. However, the occurrence of marine molluscs associated with crossbedding suggests a strong influence of tidal currents. These currents might also be responsible for its widespread occurrence.

b. Middle sequence

The middle sequence is a regressive sequence, about 50 m thick which lies on the conglomeratic unit L6 at the top of the lower sequence. The sequence is very well exposed in the Perning section (Fig. 5.30 & 5.37), but poorly exposed in the Kemlagi section (Fig. 5.34).

b.1 Fossiliferous muddy sandstone

**Facies:**
This facies includes unit M1, and lies on the conglomeratic sandstone of unit L6 (lower sequence). In the Perning section it is characterised by glauconitic sandstone with interbeds of carbonaceous mudstone. Comminuted mollusc shells as well as benthonic foraminifers (*Operculina*) are abundant, which is why Duyfjes (1938a,b,c,d) called this facies Molluscan Horizon 2.

**Interpretation:**
This facies is interpreted to have been deposited in a subtidal environment. *Operculina* commonly occurs in a slightly hypersaline inner shelf environment (Murray, 1991). The common occurrence of comminuted shells in this unit suggests a strong wave influence.
b.2 Interbedded mudstone and parallel, ripple-laminated sandstone

Facies:

This facies includes unit M2 in the Perning section (Fig. 5.37) and the unit above unit L6 in the Kemlagi section (Fig. 5.34). Its exposure in the Perning section is relatively good, and three main subfacies can be recognised: sandstone; mixed mudstone and sandstone; and conglomeratic subfacies.

The sandstone subfacies is relatively thick (2 to 6 m). There are three prominent sandstone beds. The lowermost sandstone is characterised by 2 m high epsilon cross-beds with well developed mud flasers between the foreset laminae (Fig. 5.40). The middle and upper sandstone horizons are similar, characterised by clean, planar and trough cross-bedded sandstone.

The mixed mudstone and sandstone subfacies is characterised by alternating parallel and ripple laminated sandstone and mudstone beds (Fig. 5.41). In the lowermost 12 m, mudstone is dominant with a few interbeds (5 to 10 cm thick) of parallel and ripple laminated fine-grained sandstone. This part features an abundance of plant remains and leaf fossils. In the middle and upper parts, mudstone and sandstone finely alternate (Fig. 5.43). Some of the sandstone beds reach 10 cm and are bundled by thin wavy and flaser bedded mudstone (Fig. 5.42).

The conglomeratic subfacies occurs as units less than 40 cm thick. Some of the beds in the upper part (just below unit M3) are normally graded and accompanied by basal scouring (Fig. 5.42). In the middle part, some beds (about 30 cm thick) show inverse grading from coarse sand to pebble size.

Interpretation:

The inferred depositional environment was based on sedimentary structures and facies association. This facies is interpreted to have been deposited in a tidal flat environment. The diagnostic criteria for tidal deposits have been discussed by Ginsburg (1975), Reineck and Singh (1980), Terwindt (1988) and Dalrymple (1992).

The most diagnostic features in this facies are the occurrence of mud flasers, epsilon cross-bedding and plant remains. Mud flasers occur due to small scale alternations of slack and
strong water current (Reineck and Wunderlich, 1968). Epsilon cross-bedding, which occurs in the lower sandstone subfacies, is commonly associated with a lateral migration of tidal channels, and the thickness of the crossbeds represents the channel height (Reineck & Singh, 1980); in this subfacies the channel height was about 2 m. The clean, planar and trough cross-bedded sandstone subfacies are believed to result from winnowing sand which should occur in a relatively strong tidal current region such as the sand flat and upper subtidal environments (Dalrymple, 1992).

Temporary strong wave (storm) conditions probably occurred in this tidal flat environment, and resulted in some conglomeratic subfacies. Basal scouring in these subfacies may represent storm erosion (Johnson & Baldwin, 1986), and inverse grading may represent temporary turbiditic currents which occurred during the storm.

b.3 Normally graded conglomerate

Facies:

This facies is well exposed in the Perning section, where it overlies the tidally-influenced deposit (Fig. 5.37). In the Kemlagi and Banyuurip sections this facies is absent, suggesting a limited lateral extent.

In the Perning section the facies is labelled M3, and consists of at least seven fining upward units. Each unit, which is separated by sandy mudstone, begins with a crudely bedded conglomerate fining upward into planar cross-bedded, medium- to coarse-grained sandstone. The evidence for subaerial deposition in this facies is the presence of some terrestrial vertebrate and hominin fossils [Pithecanthropus modjokertensis, Duyfjes (1938d), Fig. 5.45].

Interpretation:

Although the geomorphic evidence (associated fan-shaped geometry) is not obvious, this facies is interpreted as an alluvial fan deposit as indicated by the occurrence of conglomerate as the main facies (Rust & Koster, 1984) and vertebrate/hominid fossils. Most alluvial fans are characterised by water-laid deposits, dominated by horizontally stratified gravel (Rust & Koster, 1984). The occurrence of fining upward sequences in each unit, which shows a transition from coarse gravel to sand, is an indication of a water-laid deposit which reflects a gradual decrease in the particle/water depth ratio due to decrease of stream competence (Reineck & Singh, 1980; Rust & Koster, 1984).
The westward (Kemlagi section) and northeastward (Banyuurip section; Fig. 5.30) wedging out of this facies, with regard to the geographic position of the Wilis Volcano, suggest that this alluvial fan system was not directly related to the Wilis volcanism. It may have been developed locally due to a local topographical anomaly.

c. Upper sequence

The upper sequence is observed in the Perning, Banyuurip, Kedamean and Raci sections. The localities of these sections are shown in Figures 5.1 and 5.30, and the generalised sections are shown in Figures 5.37, 5.44, 5.48 and 5.50. Figure 5.49 shows the typical exposure in the Kedamean section. The thickness varies locally due to fluvial erosion in the uppermost sequence; in the Banyuurip section the sequence attains 50 m.

c.1 Interbedded mudstone and parallel-ripple laminated sandstone

Facies:

This facies includes unit U1, and stratigraphically lies directly on the normally graded conglomerate facies of the alluvial fan system (unit M3) in the Perning section. The lower part, as observed in the Banyuurip section, is characterised by intercalations of parallel-laminated, tuffaceous sandy mudstone and parallel to ripple laminated, fine- to medium-grained sandstone. The bed thickness varies from 2 to 10 cm. The facies is also characterised by the abundant occurrence of plant remains, including leaf imprints.

The middle part of the facies is characterised by poorly-sorted tuffaceous sandy mudstone. Bioturbation is strong which almost destroyed the parallel laminated bedding. Large *Gastrochaenolites* burrow structures (3 to 5 cm diameter; *Glossifungites* ichnofacies) are abundant (Fig. 5.46) in both the Perning and Banyuurip sections. This burrow shows the protrusive appearance of a tear-like boring structure. In the Perning section this part is also characterised by the abundance of marine mollusc fragments which Duyfjes (1938d) termed the Molluscan Horizon 3. Because the latter is widespread, he used it as a stratigraphic marker.

The upper part is characterised by finely laminated tuffaceous mudstone (Fig. 5.47), with interbeds of parallel and ripple laminated, medium-grained sandstone up to 70 cm thick. This subfacies is only observed in the Kedamean section.
**Interpretation:**

The abundance of plant remains in the lower part of the facies suggests that the environment was relatively close to or on a vegetated area. But textures and sedimentary structures suggest that the facies may not represent a supratidal environment, as commonly characterised by muddy facies (Terwindt, 1988). This lower part of the facies may represent a foreshore or shoreface region, where wave energy is present. In the upper part of the facies, the sedimentary texture and structures are similar to those in the lower part, and suggest a similar depositional environment.

In the middle part, bioturbation is strong which suggests a deeper setting than the lower and upper parts of the facies. Bioturbation has been regarded as a sensitive indicator of the depositional energy, and commonly diminishes toward the shore as wave energy increases (Reinson, 1984; Elliott, 1986). In addition, the *Gastrochaenolites* burrows, according to Frey and Pemberton (1984) and Pemberton *et al.* (1992), are included in the *Glossifungites* ichnofacies which commonly develops in firm un lithified substrates (dewatered muddy substrates) of a littoral or sublittoral environment.

The overall facies (unit U1), thus, appears to represent a transgressive-regressive sequence, as suggested by the transition of depositional environment from foreshore/shoreface to lower shoreface and back to foreshore/shoreface.

c.2 **Cross-bedded sandstone**

**Facies:**

This facies is very well exposed in the Banyuurip, Kedamean (Figs 5.44, 5.48 & 5.49) and Raci (Figs 5.50, 5.51 & 5.52) sections. There are two units included in this facies, units U2a and U2c, which are separated by interbedded sandstone and mudstone unit U2b.

The thickness of the cross-bedded sandstone facies varies from 1.5 to 7 m. They consist of coarse- and medium-grained sandstone, associated with erosional base and conglomeratic lag deposits. Planar and trough crossbedding are the common sedimentary structures. In the Kedamean section palaeocurrent direction, as indicated by foreset orientation, is toward the southwest.
**Interpretation:**

This facies is interpreted as fluvial channel facies based on the occurrence of basal erosion associated with channel lag deposits. Planar and trough cross-beds may represent linguoid or transverse bars and dunes in the channel (Miall, 1992). The patchy nature of the outcrops, unfortunately, does not allowed a further examination of the fluvial type.

This fluvial facies is well recognised on the seismic sections from Madura Strait and assigned to seismic unit C1a (Chapter 6.5.5). Its widespread occurrence is interpreted to correspond with a major sea level fall in the study area (particularly in the Kendeng Zone and Madura Strait).

c.3 Interbedded mudstone, sandstone and conglomeratic sandstone

**Facies:**

This facies includes subunits U2b and U2d, which are characterised by interbedded mudstone and sandstone; and conglomeratic sandstone. The interbedded mudstone and sandstone is typified by parallel and wavy laminated sedimentary structures. Subunit U2b in the Kedamean section, contains plant remains and leaf imprints, and subunit U2d in the Banyuurip section is bioturbated and rich in marine mollusc fragments and benthonic foraminifers (*Cibicides* & *Elphidium*). The conglomeratic sandstone is typified by massive structure, and consists dominantly of medium- to coarse-grained sandstone with abundant andesitic pebbles and mudclasts. The conglomeratic sandstone occurs in the Raci and Kedamean sections where the thickness varies from 0.5 to 1.5 m.

**Interpretation:**

The interbedded mudstone and sandstone facies is interpreted as a tidally-influenced deposit, based on the sedimentary textures and structures. The occurrence of benthonic foraminifers and bioturbation in subunit U2d suggests a lagoonal environment. The occurrence of plant remains in subunit U2b suggest a close relationship with a supratidal or estuarine environment. The conglomeratic sandstone in subunit U2b is massive and is composed of abundant intraclasts, suggesting a very quick depositional mode which may have occurred during a storm event.
5.7.2 Petrography

Fifteen samples from sandy units were examined, and most of these samples are poorly cemented and show varying degrees of weathering. The study found that there is no common classification of rocks in the Pucangan Formation due to the large variety of lithologies. However, the study has documented the occurrence of terrigenous clastic deposits of two different compositions which are strongly suspected to represent two different sources. The first clastic source is dominated by quartz, and the other contains volcanic-derived fragments such as andesite, plagioclase, pyroxene and hornblende. The occurrence of quartz is always affiliated with the marine facies at the beginning of sequences, while the second clastic is more widespread. Appendix B.1 summarises the results of the petrographic examination.

a. Lower sequence

The tidal sandstone unit (unit L3b; Fig. 5.53) is characterised by angular to subangular grains of plagioclase (21%), pyroxene (17%) and andesitic rock fragments (17%); the grain size varies from 0.4 to 0.75 mm. The sandstone is very porous, uncemented, with practically no muddy matrix, indicating a high energy depositional current.

The sandstone of the debris flow unit is characterised by angular to subangular grains of plagioclase (up to 60%), rock fragments (up to 22%) and pyroxene (up to 16%); the grain size varies from 0.3 to 1 mm. This sandstone is also characterised by the presence of foraminiferal tests (less than 2%) with no indication of reworking. This indicates that the debris flow unit was deposited in a marine environment.

Five petrographic samples have been studied from the subtidal facies in the lower sequence (from the Kemlagi section). The lower part of this offshore facies is characterised by muddy sandstone which is almost free of volcanogenic fragments (< 1%). The main components are angular to subangular quartz grains (up to 34%), and bioclasts of foraminiferal tests, algae and mollusc fragments with the average grain size 0.2 mm. All these fragments are cemented by carbonate mud which makes the total carbonate component up to 59% of the rock. A similar texture occurs in the middle sequence (sample no. KG7), but around 30% of the terrigenous clastic material is volcanogenic fragments (plagioclase and andesitic rock fragments). This part also contains streaks of organic materials. The uppermost part, which can be included in unit L6 (sample no KG1), consists of subangular to subrounded andesitic rock fragments (17%),
angular plagioclase (29%) and pyroxene (< 4%), with the grain size varying from 0.2 to 1.5 mm (Fig. 5.54).

The upward decreasing bioclastic content in the samples from the Kemlagi section suggests that the lower sequence is regressive. A fully marine environment may have occurred in the lower part which gradually became shallower upward where the deposition became influenced by volcanic rock fragments and plant debris. The quartz in the lower part has never been associated with Quaternary andesitic volcanism. The reason for the quartz abundance is probably the same as for the sandstone facies in the Lidah Formation in the Sumberringin section (some 3 km east of Kabuh). The quartz was derived from the exposed quartz-bearing formations in the Rembang Zone and transported southward to the Kendeng Zone through marine processes. The same case occurs in the marine fossiliferous sandstone facies in the middle (unit M1) and upper (unit U1) sequences.

b. Middle and upper sequences

The middle and upper sequences are represented by seven petrographic samples. Units M1 and U1 are quite similar, and are characterised mainly by carbonate mud matrix. As indicated earlier, these marine derived facies contain angular to subangular fine-grained quartz which is interpreted to represent a northerly derived terrigenous clastic component (Fig. 5.55). In unit M1, quartz is a minor constituent, but in unit U1 it is up to 7%. It is associated with volcanogenic fragments such as andesite (some are glauconitised) and plagioclase (about 14%), and bioclastic foraminifers (Operculina) and marine molluscs (Lamellibranch).

Three sandy samples from the channel facies (unit U2c) contain relatively little mud matrix; most of the fragments are lithic (up to 56%). These fragments (andesite) are rounded, highly altered, with an average grain size of 0.4 mm. Plagioclase and pyroxene attain 17% and 2%, respectively, and commonly are less altered and show angular form.

Unit U2b is characterised by the presence of mudstone beds which consist mostly of carbonate mud. Fine-grained (less than 0.15 mm) plagioclase and quartz are up to 12% and 7% of the rock respectively. It also contains less than 1% bioclasts. The sandstone beds in this deposit are moderately sorted, consisting of altered volcanic rock fragments (up to 56%), angular plagioclase (up to 17%) and pyroxene (< 1%).
5.7.3 Discussion

The Pucangan Formation in the eastern Kendeng Zone developed on the Lidah Formation which was, in turn, developed as an extensive lacustrine deposit. Thus the Pucangan Formation was deposited on a relatively flat-lying surface predefined during the deposition of the Lidah Formation. The lithofacies study indicates that the flat depositional slope persisted during the deposition of the Pucangan Formation in the eastern Kendeng Zone, as evidenced by the common occurrence of tidal flat deposits in this formation. There are three major conglomeratic and sandy facies (units L6, M3 and U2) in the Pucangan Formation which have been extensively developed across this relatively flat-lying area. They are believed to be the product of the Wilis Volcano through an alluviation process. Thus, the gross depositional setting of this formation can be regarded as a fan delta, similar to the ones described by Nemec and Steel (1988).

The first depositional model for the Pucangan Formation was proposed by Duyfjes (1938b). This model regarded the vertical development of the old Wilis Volcano in the surrounding marine facies. He described the volcanic facies laterally wedging out and changing to the muddy Lidah Formation (Fig. 5.56). The lateral development of the Pucangan Formation was assumed to correspond with the vertical development of the Wilis Volcano (Fig. 5.57). This model appears to have satisfied many authors working in the eastern Kendeng Zone (e.g. van Bemmelen, 1949; de Genevraye & Samuel, 1972; Pringgoprawiro, 1983) as there has been no attempt at further examination by the later investigators.

The development of coarse clastic deposits associated with alluviation in the Pucangan Formation appears to be intermittent. The subdivision into three major sequences in fact represents this. Marine processes occurred during the time intervals in between the deposition of these coarse clastic sequences.

The lower sequence is characterised by debris flow deposits. They may characterise a steep depositional slope and rapid deposition (Walker, 1975). Observations of the debris flow deposits in the Solo River section reveals a substantial amount of fine debris, which is apparently strong enough to support clasts of more than boulder size. The occurrence of these deposits is probably closely associated with the Wilis volcanism.
The conglomeratic facies in unit L6 (lower sequence) and unit M3 (middle sequence) lack a debris flow character. A close association with vertebrate fossils, particularly in unit M3, is rather suggestive of a period of volcaniclastic alluviation. Figure 5.58 displays the depositional model for unit M3.

5.8 KABUH FORMATION

5.8.1 Lithofacies

The Kabuh Formation formed during the Middle Pleistocene (Chapter 3.3.2 and 3.5). A sedimentological discussion of this formation can only be carried out on its lowermost part where it is well exposed in the Banyuurip, Kedamean and Raci areas (Figs 5.30 and Fig. 5.1). The vertical sections are shown in Figures 5.44, 5.48 and 5.50. The upper part of the Kabuh Formation consists of volcanic and fluvial facies, and presently lies in the synclinal area. In the easternmost Kendeng Zone it is rarely exposed and is often covered by thick recent alluvium.

In the Kedamean and Banyuurip sections, the lower part of the Kabuh Formation developed as a regressive barrier island setting (Fig. 5.32). Detailed observations in these sections (Figs 5.44; 5.48 & 5.49) identified four different facies which are commonly present in this setting: a shelf mud, lower shoreface, upper shoreface and foreshore facies. In addition, a non-marine facies overlies the foreshore facies. As shown in Figures 5.48 and 5.49, these marine facies are relatively thin (about 20 m). Duyfjes (1938b,d) reported that these facies gradually wedge out westward. In the synclinal area north of Perning (Fig. 5.30) the marine facies have been replaced by a non-marine facies characterised by the presence of non-marine molluscs and vertebrate fossils.

a. Green mudstone

This facies is interpreted to represent a shelf mud environment. It consists of massive, greenish grey, sandy mudstone, and was deposited on the fossiliferous, molluscan-rich tidal facies unit U2d (Pucangan Formation; Fig. 5.49). The present faunal study of rock samples from this facies in these sections indicates that benthonic and planktonic foraminifers are rarely present. Duyfjes (1938d) reported the occurrence of some foraminiferal species (e.g. *Ammonia* and *Globigerina*) in this facies from the Raci area.
The change from fluvial facies (unit U2c), passing through tidal facies (unit U2d) to the shelf mud facies appears to represent a very rapid transgression which only deposited thin units. The shelf mud facies was widely developed, probably due to a very low depositional gradient whereby a few metres increase of sea level will lead to the flooding of an extensive area. A similar facies can be recognised in the Raci area (Fig. 5.50). Here, the facies is unconformably overlain by the fluvial channel deposit forming the upper part of the Kabuh Formation.

b. Bioturbated muddy sandstone and parallel laminated mudstone

The first facies represents a lower shoreface environment; it consists of poorly sorted, muddy sandstone. In the Kedamean section this unit was channelised in the upper part (Figs 5.48 & 5.49) and filled with muddy sandstone which is rich in marine molluscs and bentonic foraminifers. Genus *Ammonia* is the most common fauna and typifies an inner neritic environment (Murray, 1973; 1991). Sedimentary structures are absent, possibly due to intensive bioturbation. Reinson (1984) suggested that the lower shoreface region occurs seaward of the break in the shoreface slope, at the toe of the barrier-island sedimentary prism (Fig. 5.32). The normal condition of the lower shoreface region is relatively low energy.

Upward, the facies changes to parallel-laminated mudstone with thin interbeds (< 10 cm) of ripple laminated fine-grained sandstone. Bioturbation is much less than in the lower shoreface deposit. This part is interpreted to represent a low energy upper shoreface environment.

d. Cross-bedded sandstone

Facies:

This facies may represent foreshore and non-marine environments. The lower portion represents the foreshore region, and is characterised by a series of coarsening and thickening upward sandstone units (Figs 5.48 & 5.49), separated by thin (< 5 cm) mudstone interbeds. The sandstone units contain prominent low-angle, large scale (about 1 m) planar cross-beds but no basal erosion is associated with these structures. Reinson (1984) indicated that the swash and backwash mechanism is often related to the formation of distinct subparallel to low angle, seaward-dipping, planar sedimentary structure in the foreshore region.
The non-marine facies was defined particularly on the basis of the vertical sequence trends and a mapping report by Duyfjes (1938d). Some outcrops in the Banyuurip area indicate the presence of conglomeratic sandstone and planar cross-bedded sandstone with mudstone interbeds.

5.8.2 Discussion

The Kabuh Formation is known as the last lithostratigraphic unit in East Java which was deposited under the influence of marine processes (Duyfjes, 1938b,d). Deposition commenced with a widespread development of shelf mud [green mudstone according to Duyfjes (1938d)] which, from its textural character, appears to have been deposited as a deeper facies than the marine-influenced deposits in the Pucangan Formation. This green mudstone facies is recognisable on the seismic sections in Madura Strait (seismic Section 6.1) and is referred to as the seismic unit C2a. It marked the beginning of a series of deposits (seismic unit C2 to C8) which are interpreted to represent frequent sea level cyclicity and rather large sea level rises and falls. A magnetostratigraphic study by Hyodo et al. (1992) in the Perning area (Fig. 5.37, Perning section) indicated that this green mudstone is younger than the Jaramillo event. This led to the speculation that the marine facies of the Kabuh Formation should represent the beginning of periodic and rather extreme sea level fluctuations over the last 900 ka which are very well recorded by the oxygen isotope record in the Pacific core V28-239 (Shackleton & Opdyke, 1976). Unfortunately, many of the cyclic marine facies in the Kabuh Formation in East Java are missing due to widespread erosion during the combined Late Pleistocene folding and sea level fall during the oxygen isotope stage 6. However, some of these events are recognised on the seismic sections from Madura Strait.

5.9 SUMMARY

The sedimentation history of the Kendeng Zone in the study area is complex. Here, the oldest outcropping lithofacies is the Middle-Late Miocene Kerek Formation. This formation consists of a thick (> 800 m) turbiditic deposit which was derived from the Banyak volcanic arc. A marl facies, the Kalibeng Formation, dominated deposition in the Early Pliocene. A short regression occurred in the Kendeng Zone in the Middle Pliocene but deltaic deposits (sandstone facies of the Atasangin Formation) only occurred in the eastern part of the zone.
Since the Middle/Late Pliocene two different tectonic characters have occurred. The central Kendeng Zone (including the Ngawi area) has undergone tectonic uplift, as indicated by regressive vertical facies sequences in this area which ended with the deposition of the wave-dominated unit of the Late Pliocene Sonde Formation. In contrast, in the eastern Kendeng Zone, including Madura Strait, very rapid basin subsidence occurred, combined with a high sedimentation rate. This resulted in the deposition of thick diatomaceous marl (marl facies of the Atasangin Formation) and lacustrine deposits (Lidah Formation). In the Plio-Pleistocene, activity of the Wilis Volcano commenced (Duyfjes, 1938b,d; van Bemmelen, 1949). Its clastic detritus was deposited as the marine Pucangan Formation followed by the Kabuh Formation where only the easternmost part of the zone was in a marine environment.
6.1 INTRODUCTION

Madura Strait is the eastward extension of the Kendeng and Randublatung Zones. It is a small east-west elongated basin, 130 x 65 km in size. The basin is synclinal, bordered in the north by the anticlinal zone of the Rembang Zone (Madura Island), and by Quaternary volcanoes in the south (see Fig. 2.8). Discussion given in Chapter 5 indicated that the eastern Kendeng Zone, including Madura Strait, has undergone very rapid subsidence since the Pliocene, which may be due to the northward compressional stress within the larger plate tectonic system in western Indonesia. Thick Quaternary deposits are indicated by some well data, and seismic records provide evidence of very high sedimentation rates.

This chapter discusses the distribution and chronology of the sedimentary facies developed in Madura Strait during the Quaternary, and briefly reviews its underlying Late Tertiary strata. The discussion relies mostly on the results of seismic interpretation, combined with lithofacies analysis, including petrographic and faunal analyses, of the Late Tertiary strata. Furthermore, this chapter can only provide a preliminary Quaternary sedimentological study, because of the absence of reliable well data and dating. The seismic stratigraphy of Madura Strait is interpreted by applying the sequence stratigraphic concepts developed by Vail et al. (1977), Posamentier et al. (1988) and Posamentier and Vail (1988). Sequence definition and terminologies used in this concept have been discussed in Chapter 4.2. Interpretation of clastic depositional facies from seismic facies is carried out based on the criteria outlined by Sangree and Widmier (1977, Table 4.1).

The sequence stratigraphic concept was originally applied to low resolution multifold seismic data associated with petroleum exploration and involved third order sequences or lower (Vail et al., 1977). The present study concerns Quaternary strata, for which the sequences are of higher order (fourth and fifth order). The thickness of such sequences is often beyond the resolution of multifold seismic systems. However, the applicability of this concept to such sequences is endorsed by Posamentier et al. (1988), Mitchum and van Wagoner (1991), and
Posamentier and James (1993) who indicated that the concept of sequence stratigraphy is scale independent.

In the last few years, a number of authors have applied sequence stratigraphic analysis to fourth and fifth order sequences [0.1-0.2 Ma and 0.01-0.02 Ma respectively, Mitchum & van Wagoner (1991)] using high resolution seismic data. On continental shelves, successful applications have been demonstrated by Suter et al. (1987) on the southwest Louisiana continental shelf; Tesson et al. (1993) on the outer Rhone continental shelf (France); and Hernandez-Molina et al. (1994) on the Spanish continental shelf. Okamura and Blum (1993) successfully applied sequence stratigraphic analysis to high resolution seismic data from the fore-arc region, southwest of Japan. The present study adopts the same principles as these studies, using high resolution seismic data (from a sparker system) which have about 2 m vertical resolution. Although this study is not intended to review the applicability of the sequence stratigraphic concept in general, its use in Madura Strait may be regarded as a test of its application to a small basin which has high subsidence and depositional rates. Furthermore, because this basin is relatively small in lateral dimension, the trans-basinal depositional style is able to be studied.

Five major sedimentary units can be identified on the seismic sections studied. Each of these units is constructed of two or more subunits which have distinct areal distributions, thicknesses and seismic characters. The upper four major units are bounded at their base by regional unconformities. They can be regarded as fourth order type I sequences, according to the criteria outlined by Vail et al. (1977) and Posamentier and Vail (1988), due to their duration of less than 1 Ma. The basal unconformities are associated with widespread and relatively thin fluvial deposits which possibly represent major sea level falls during the Early to Middle Pleistocene. The five major seismic units are referred to (from bottom to top) as units Pre-A, A, B, C, and D. Numerical and lower case characters affixed on these letters indicate the subunits. The seismic unit Pre-A lacks internal reflections, which does not allow further subdivision of this unit although it is relatively thick. This unit is underlain by Pliocene strata which in this discussion are referred to as the Mundu Formation.

The determination of Pleistocene eustatic sea level change in Madura Strait will not be attempted, because the basin subsidence rate is not known. In addition, recognition of coastal onlaps in many cases is difficult due to the very low depositional dip, and the formation of the
Late Pleistocene fold system which has disturbed the position of the sequences. The Pleistocene sea level can only be deduced tentatively based on the position of incisions and coastal onlaps relative to present sea level.

6.2 SEISMIC DATA

The data base for this study is drawn from seismic profiles totalling some 1500 line km. The seismic profiling was carried out by the Marine Geological Institute of Indonesia in 1989 and 1990, using a 600 joule high frequency single channel sparker system. Figure 6.1 shows the ship tracks during profiling in Madura Strait. Spacing of profiles was relatively regular and the north-south profiles were run perpendicular to the general east-west structural trend.

During profiling the sparker output power was released through a three tip array. Low/high filter settings for most of the sparker survey were 200-2000 Hz. The data were graphically recorded in analog format. Firing and recording sweep rates were 0.5 seconds, and the vertical recording scale was 0.5 sec, resulting in an average vertical exaggeration of 8 times on the records. Ship positioning during the seismic profiling was accomplished with the Magnavox MX 1157 satellite navigation system with point accuracy of 36 m.

Although the seismic system provides high resolution records, it has a maximum depth penetration of about 350 m. In some areas where the seismic wave attenuation is high, the penetration of the seismic system is significantly reduced. The penetration is also largely inhibited where medium to coarse-grained sediments are interpreted to be present. With such limitations and because the thickness of Quaternary sediments in East Java could reach 500 m, the discussion will mostly concentrate on Middle and Late Quaternary sediments in Madura Strait.

Figure 6.2 shows the location of seismic sections used for the discussion, but isochron (unit/sequence thickness in time units) and time structure maps presented in this discussion were derived from all the seismic lines indicated in Figure 6.1. Due to the absence of measured seismic velocity data, during interpretation the seismic velocity is assumed to be 1500 m/sec which is really only valid for the water column. On the seismic profiles presented in the discussion, the depth was calculated based on this velocity and should only be regarded as a rough estimate.
Analysis of seismic data used for this study includes four main steps: seismic interpretation, seismic horizon digitising, gridding and contouring. In the seismic interpretation, seismic horizons are recognised based on the criteria outlined by Vail et al. (1977) which was discussed in Chapter 4. However, because these criteria were derived for use in a petroleum exploration setting, mostly on shelf margins where depositional dips are usually obvious, a different approach will be attempted for this study. Sedimentary sequences in Madura Strait are commonly very thin and widespread. Due to rapid sea level oscillations and due to the very low depositional dip, recognition of seismic features such as coastal onlap is very difficult. Here, sequence boundaries have been identified using features such as fluvial erosion and anticline truncation.

The interpreted seismic lines were digitised by using a digitising tablet. The results from this step were the coordinates for the picked seismic horizons, depth (z value) and horizontal position along a line. The last value was used to obtain the absolute positions (x and y coordinates) of picked seismic horizons by comparing it with the satellite fixed positions. The discrete data obtained from this step are irregularly spaced along the seismic lines. Prior to contouring, these data have to be gridded with a predefined spacing. In this study, the inverse distance gridding method was applied, to obtain faster and reliable results. The digitising and gridding steps were carried out on an IBM PC, using software (see Appendix F for source codes) specially written by the author in BASIC and C++ computer languages. The gridded data obtained from this step were translated into the "Surfer" computer program data format and contouring was done using this computer program. The "Surfer" computer program is in fact able to do the gridding step; however, dealing with such a large amount of data, this program would consume a lot of time.

6.3 WELL DATA

To date, there have been no published studies of the Tertiary and Quaternary geology of Madura Strait. The results from two petroleum exploration wells drilled in this area, MS1-1 and MDB-1 (Fig. 6.1), indicated a low economic potential. The current study is probably the first attempt to discuss the Quaternary geology of this area.

Faunal analyses have been carried out on cuttings and sidewall core samples from these petroleum wells by the Indonesian Cities Service Inc. Company. Due to the aims of the
drilling, which were targeted on the Tertiary sequences, these wells do not provide good data control for the present seismic study. Three disadvantages of these wells for this Quaternary study are that: (1) they were drilled on or close to an anticlinal axis on which Quaternary sediments are very thin due to non-deposition or Quaternary erosion; (2) faunal and lithofacies analyses were carried out mostly on cuttings sampled at long intervals (every 30 ft or 9.1 m penetration); and (3) no geophysical logs are available for study. With such limitations, these wells only provide a rough estimation on the environment of deposition and age of stratigraphic boundaries. Detailed lithofacies correlation with the seismic data is practically impossible, and most of the lithofacies interpretation has to rely on the onshore geological data, such as from the Kendeng Zone and Madura Island.

**Well MS1-1**

This well was drilled on the southern flank of a Late Pleistocene anticline the crest of which forms Kambing Island (Fig. 6.1). The projected well position is shown on seismic line MD-MD' (Section 6.4). Accurate projection and stratigraphic correlation are difficult to achieve, due to the position of the well on the ramp side of an anticline. This well was drilled in a position which has a water depth of 27.4 m and a 24° SE average dip for the strata. The description below summarises the well report and concerns only the uppermost 700 m of sediment in the well (measured from the rotary stage, which was 22 m above sea level). Figure 6.3 summarises the planktonic foraminiferal analysis.

Between 49.4 and 76 m depth, the lithology is characterised by fossiliferous muddy sandstone, with abundant benthonic foraminifers and gastropods (up to 65%). A fairly massive calcareous mudstone is present between 76 and 277 m depth. The Pliocene-Pleistocene boundary (planktonic foraminifer zones N21/N22 boundary) has been picked at the base of this unit. It was defined by extinction of the planktonic foraminifer species *Globigerinoides obliquus* and *Globorotalia acostaensis*. However, it should be noted that some species such as *Globigerinoides ruber*, *Globigerinoides quadrilobatus trilobus*, *Pulleniatina obliquiloculata*, *Globorotalia cultrata*, *Orbulina universa*, *Hastigerina aequilateralis* and *Sphaeroidinella dehiscens*, which should remain until N23 (Late Pleistocene), have also disappeared. This indicates that changes in the environment of deposition may have contributed to the extinction of these species at this level in this area.
The occurrence of *Operculina*, *Amphistegina* and *Asterorotalia* in this calcareous mudstone suggests a fairly shallow water depth (Phleger, 1960; Murray, 1991), and a possible tidal flat environment as *Operculina* commonly occurs in a slightly hypersaline region (Murray, 1991). Stratigraphically, the calcareous mudstone is equivalent to the Pucangan Formation in the north of Mojokerto area, which is characterised by interbeds of shallow shelf, tidal flat and non marine deposits.

The calcareous mudstone is underlain by a sandy mudstone to a depth of 451 m. This sandy mudstone is typified by the appearance of lignite, plant remains, quartz, pyrite and glauconite. Planktonic and benthonic foraminifers occur sporadically. Between 451-585 m depth, the lithology consists mostly of massive mudstone with trace amounts of glauconite, limestone and plant remains. The consistent appearance of *Ammonia beccarii* indicates that the environment of deposition was lagoonal. This mudstone is stratigraphically equivalent to the Lidah Formation lacustrine deposit in the area north of Mojokerto which is mainly characterised by an organic-rich, dark blue mudstone (Chapter 5.6).

Below 585 m depth, the mudstone is calcareous and fossiliferous. The upward decrease in abundance of planktonic foraminifers in this Pliocene mudstone suggests that the environment of deposition was gradually shoaling from open marine-influenced (below 585 m depth) to lagoonal environments. Stratigraphically, this fossiliferous mudstone should correlate with the Mundu Formation marly facies (equivalent with the Kalibeng and Atasangin Formations).

**Well MDB-1**

Well MDB-1 is located 12 km east of the easternmost seismic line. In the uppermost 488 m, the lithology is dominated by a non-calcareous mudstone, with minor interbeds of silt and loose glauconitic sand. Below this, the facies becomes gradually more fossiliferous with depth until 690 m, where a clastic limestone unit underlies. The Plio-Pleistocene boundary was defined within the fossiliferous mudstone, at 673 m depth, based on analyses of planktonic foraminifers and nanoplankton. The age of the limestone, according to the well report, is Early to Middle Pliocene. Unfortunately, the biostratigraphic data are not available for this study.
The fossiliferous mudstone in this well is correlated with the similar facies and age in the MS1-1 well, below 585 m depth. The non-calcareous mudstone in the MDB-1 well probably correlates with the mudstone facies above 585 m depth in the MS1-1 well. Correlation between the position of the Pliocene-Pleistocene boundary and facies in both the MS1-1 and MDB-1 wells suggests that the lacustrine facies was developed earlier in the MS1-1 well (western part).

6.4 PHYSIOGRAPHY OF MADURA STRAIT

The bathymetry of Madura Strait is shown in Figure 6.4. In general, the sea floor depth increases towards the middle of the strait which has the form of a broad submarine valley. Some local highs in the strait correspond to buried small anticlinal or diapirc structures (Fig. 6.1). The deepest portion (more than 80 m) occurs in the easternmost part of Madura Strait. The average east-west sea floor morphological gradient is very low (0.5 m/km) and, in the north-south direction, it is relatively similar (2.2 m/km) on both northern and southern sides of Madura Strait.

In Madura Strait the orientation of geological structures is similar to the onshore Kendeng Zone. Seismic data indicate that anticlines, which are also developed in the onshore part, become progressively simpler eastward and are individually separated from one another. In this strait, the crests of anticlines are represented by two small islands (Kambing and Ketapang Islands, Fig. 6.1). Figure 6.29a shows a three dimensional view of the sea floor morphology. It is suspected that mud diapirism, as found in several places in the onshore area (north of Mojokerto area), has played an important role in the structural development in the Kendeng Zone and Madura Strait. In this strait, it is often accompanied by normal faulting.

6.5 SEISMIC FACIES ANALYSIS AND INTERPRETATION

The discussion presented in this section refers to some representative seismic lines, and their geographic locations are provided in Figure 6.2. However, all the surveyed lines were used to construct the isochron, seismic facies and palaeogeographic maps. The representative seismic sections are numbered as Sections 6.1 to 6.12.
6.5.1 Late Tertiary units

a. Seismic facies

Because of the limited penetration of the seismic system used, the Late Tertiary units can only be studied from the seismic sections recorded close to some anticlinal islands. The seismic line ML-ML’ (Sect. 6.12) is representative of these lines. It was recorded above an anticline which forms the Gili Genting and Gili Raja Islands (Figs 6.1 & 6.2). This line displays two contrasting geological features. On the southern synclinal side, the section is mostly occupied by late Late Pliocene (unit Pre-A) and Quaternary (units A to D) strata. A continuous and rapid subsidence, combined with rapid sedimentation, has allowed an accumulation of thick sequences that may attain more than 500 m in the synclinal axis. On this seismic section these sequences are not well displayed due to multiple reflection interference, but detailed discussion, based on other seismic profiles, will be presented in the subsequent sections.

On the anticlinal portion, three distinct Late Tertiary seismic sequences can be recognised and are referred to as units LT1, LT2 and LT3. The description below concerns these units. The lithostratigraphic correlations are mostly tentative, based on some geological mapping reports from the Madura, Gili Raja and Gili Genting Islands, and some direct observations during this study on these islands.

The sedimentary facies interpretation is curtailed by the absence of exposure and well data near seismic line ML-ML’ (Sect. 6.12), but a tentative correlation is possible through a stratigraphic correlation with the inland geological outcrops. On Madura Island geological mapping was carried out by Duyfjes (1938c), Situmorang et al. (1992) and Aziz et al. (1993). There is common agreement among these authors that four major Early/Middle Miocene to Pliocene stratigraphic units are exposed on this island, along the flanks of some east-west trending anticlines. These are the Early/Middle Miocene paralic and clastic limestone deposits equivalent to the Tawun and Bulu Formations respectively, and Late Miocene/Pliocene marly facies, and reefal and clastic limestone of the locally termed Pasean and Madura Formations respectively. The marly facies, Pasean Formation, is stratigraphically equivalent to the Mundu and Kalibeng/Atasangin Formations in the Rembang and Kendeng Zones (of East Java) respectively.
a.1 Units LT1 and LT2

The lowermost unit, LT1, is topped by an angular unconformity. Seismically, this unit is characterised by a relatively strong amplitude and a low continuity of reflectors, but some strong and continuous reflections occur on the northern side. Unit LT2 thickens northward, and is typified by medium amplitude and low continuity reflectors. Some internal baselapping patterns occur and are characterised by relatively thin, strong and continuous parallel reflection patterns.

Units LT1 and LT2 are interpreted to correspond with the Miocene paralic deposit and limestone respectively. This paralic deposit is dominated by organic-rich mudstone with interbeds of quartz sandstone and clastic limestone. The occurrence of benthonic foraminifers Ammonia sp., Brizalina sp., Triloculina sp. Quinqueloculina sp., Eponides sp. (Situmorang et al., 1992) suggests an inner neritic environment of deposition (Phleger, 1960; Murray, 1991). The limestone, according to Situmorang et al. (1992), is characterised by the occurrence of larger foraminifers Lepidocyclina sp., Cycloclipeus sp., Operculina sp., which are commonly associated with reefal deposits and calcareous algae (Moore et al., 1952; Serra-Kiel & Reguant, 1984).

The MS1-1 well report indicated that a similar paralic facies and age as the Tawun Formation occurred at 2627 m depth, at which the drilling was terminated. This indicates that intensive fold development had not commenced in the Miocene, which implies the widespread occurrence of paralic environments in the study area.

a.2 Unit LT3

Unit LT3 is characterised by a low amplitude subparallel reflection pattern with some strong and continuous parallel reflections. The true thickness of this unit cannot be determined on seismic line ML-ML’ (Section 6.12) due to the extensive Late Pleistocene erosion, but it may reach more than 300 m. The uppermost part tends to show a stronger amplitude reflection which is interpreted to represent remnants of reefal and clastic limestones. These limestones are widely exposed along the south coast of Madura Island and in the core of Gili Raja and Gili Genting Islands. These have been referred to as the Madura Formation by Situmorang et al. (1992), and are equivalent to the Paciran Formation. On these islands, the limestone facies is underlain by a marly mudstone which is interpreted to represent the lower amplitude subparallel reflection pattern of unit LT3. Situmorang et al. (1992) assigned the
marly facies beneath the Madura Formation limestone to the Pasean Formation which is a lateral equivalent of the Mundu Formation. This marly facies wedges out toward the axial line of Madura Island.

In the Kamal area, the limestone facies appears to have been deposited under the influence of tidal currents, which resulted in southward-dipping very large cross-beds of sand waves (Dalrymple, 1992; Fig. 6.5), and further suggests a fairly shallow shelf environment of deposition. Duyfjes (1938c) described a limestone succession developed on the western end of Madura Island and assigned it to the limestone facies of the Upper Kalibeng Beds. Stratigraphically, this limestone is equivalent to the shallow water reefal facies of the Mundu Formation. Such limestone facies is absent in MS1-1 well, suggesting that shallow marine facies only developed along Madura Island (southwestern part) and deeper marine facies occurred along Madura Strait.

In some areas on Madura Island (east of Kamal and south of Pamekasan) and Gili Raja Island, the limestone facies is unconformably overlain by a dark, carbonaceous mudstone similar to the Lidah Formation found in the Mojokerto area and MS1-1 well (above the Pliocene fossiliferous mudstone). This mudstone, based on its stratigraphic position and direct correlation with the seismic section, is assigned to seismic unit Pre-A (lacustrine deposit).

A petrographic study carried out on four clastic limestone samples (Pasean Formation) from the south coast of Madura, Gili Raja and Gili Genting Islands indicates the dominance of sand-size micritised bioclasts (20 - 90%) of foraminifers, algae, brachiopods, bryozoans and some rounded reworked limestone clasts (Appendix B.1). Rounded quartz grains are present (up to 30%) with the grain size commonly less than 1 mm. All these clastic components are cemented commonly by micrite. Figures 6.6 and 6.7 display the typical limestone under a polarised microscope.

The texture and composition of all these analysed samples appear to support a shallow shelf environment, where reef and carbonate sand bodies are commonly associated (Tucker & Wright, 1990). Quartz grains are interpreted to have been reworked from the Tawun Formation, which may have been exposed along the core of anticlines in Madura Island.
b. Microfaunal study

Planktonic and benthonic foraminiferal analyses have been carried out on a number of marly samples of the Pasean Formation from some areas on the south coast of Madura and Gili Genting Islands. The aim was to determine the age and depositional environment of the formation. Unfortunately, all samples represent the upper part of the formation, due to the rare exposure of the middle and lower parts.

Planktonic foraminiferal analysis was carried out on four samples and Figures 6.8, 6.9 and 6.10 summarise the result of this analysis. The presence of *Pulleniatina obliqueloculata* in the samples from Gili Genting Island (samples GG3 & GG7) suggests that the age of the upper part of the formation is not older than Late Pliocene (planktonic foraminiferal zone N19). The assemblages are equivalent to the Pliocene section in the MS1-1 well (Fig. 6.3).

Benthonic faunal analysis was carried out on nine samples from the Pasean Formation. As for the petrographic study, these samples were from exposures along the south coast of Madura Island (east of Kamal and southeast of Pamekasan), and from Gili Raja and Gili Genting Islands (Fig. 6.1). Appendix C.2 summarises the results of the benthonic faunal analysis.

A sample from east of Kamal indicates an inner neritic environment as shown by the appearance of the genera *Quinqueloculina* and *Elphidium*. Six samples from the Pamekasan area and the Gili Raja Island also suggest a similar environment, with all samples showing a consistent appearance of the genus *Elphidium* with the addition of the genera *Nonion* and *Quinqueloculina*. The high content of benthonic foraminifers in two calcareous mudstone samples from Gili Genting Island is evidence of a littoral environment. The typical fauna for this environment are the species *Ammonia gaimardii* and the genus *Pseudorotalia* (Phleger, 1960; Murray, 1973; 1991).

The fact that a limestone facies has developed in the upper part of the Pasean Formation, and is in turn unconformably overlain by a lacustrine unit (Pre-A), suggests that there was a shallowing trend in the environment of deposition. Since this faunal evidence is from the upper part of the Pasean Formation, a deeper marine facies than littoral or inner neritic may be present in the lower part. Thus, a transgression may have occurred after unit LT2
deposition. The occurrence of this Pliocene sea level rise has been indicated during a sequence stratigraphic study in the Java Sea (see Chapter 7.4.2).

### 6.5.2 Unit Pre-A

The deepest seismic reflections obtained by the seismic survey come from the Pre-A unit. On most seismic lines studied it is characterised by a lack of reflections and tends to be chaotic. Parallel to subparallel reflection patterns, combined with medium relative amplitude, high frequency and high continuity, do occur particularly in the upper part of the unit near its upper boundary. Isochron maps of the unit cannot be constructed, as its lower limit mostly lies beyond the depth of penetration of the seismic system. The inferred thickness on some anticlinal sites is 300 msec TWT (about 225 m).

Correlation with well MS1-1 suggests that this unit consists of non calcareous mudstone featuring plant remains, lignite and pyrite. A similar facies crops out on the south coast of Madura Island (east of Kamal and south of Pamekasan) and Gili Raja Island, and lies unconformably on the limy facies of the Pasean and Madura Formations. Based on the stratigraphic position and similarity in facies, unit Pre-A is interpreted as the eastern extension of the Lidah Formation. This interpretation is also supported by an X-ray diffraction analysis carried out on the sample from the Kamal area (Fig. 6.11b), which shows a similar character to the diffractogram of the Lidah Formation from the Nlampin area (Fig. 5.28). The organic hump which characterises the richness of organic materials is well displayed as well as calcite and siderite. However, the X-ray diffractogram of a sample from Gili Raja Island (Fig. 6.11a) shows the dominance of quartz which indicates a strong terrigenous clastic influence.

The Lidah Formation has been interpreted mostly as a lacustrine deposit (Chapter 5.6), which was widely developed in the Rembang Zone and eastern Kendeng Zone. The extent of this facies in Madura Strait appears to be confined to the synclinal basin of this strait. The fact that it has an unconformable relationship with its underlying units in the Kamal area, suggests that the deposition was later on the basin margin, and possibly followed a relative rising of base level (?sea level) position.

### 6.5.3 Unit A

The beginning of unit A was marked by a depositional hiatal surface due to a relative sea level rise which terminated unit Pre-A deposition. This sea level rise shifted the depocentre
landward, and led to sediment starvation of the basinal area. Three subunits can be identified within unit A, these are: subunits A1, A2 and A3, which are very apparent on the seismic records from the easternmost area. Westward extension of these sequences is hard to trace due to high seismic wave attenuation by the overlying sediments in the western part of Madura Strait.

a. Subunit A1

Subunit A1 is well displayed on the seismic lines from the eastern part of Madura Strait (Sects 6.8 & 6.11). The lower boundary of this unit is prominent (bottom boundary of major sequence A) indicating a high contrast in physical properties with the underlying unit (Pre-A) which has continuous and parallel reflection patterns. No erosional features are associated with this boundary suggesting an abrupt relative sea level rise at the end of unit Pre-A deposition. The internal character is typified by an upward loss in acoustic transparency - almost reflection free at the lower part which upward gradually becomes subparallel to hummocky patterns with variable amplitude and low to high continuity. This indicates an upward coarsening of lithofacies which may relate to the shoaling of the subunit (regressive). This subunit attains 35 msec TWT (about 26 m, Fig. 6.12) in thickness and occurs in the middle of the eastern part of the strait. The thinning trend toward the basin flanks suggests that this subunit regressed from both the north and south margins of the basin.

b. Subunit A2

At the end of subunit A1 deposition, sea level rose again. The lower boundary of this unit represents a depositional hiatus as the result of a rapid relative sea level rise which led to sediment starvation. Almost the whole of sequence A2 is characterised by an upward loss in acoustic transparency (Sects 6.5, 6.8 & 6.11), indicating regressive (shoaling) deposits, according to the criteria of Chiocci (1994). Southeast dipping sigmoid-progradational reflections occur in the southeast area (Sect. 6.5) which indicate the presence of sedimentary bypassing in the northern part of the region at the end of subunit A2 deposition. Toplapping which truncates the prograding clinoforms can be regarded as a type I unconformity. The estimated depositional basin floor depth as indicated by the height of the prograding profile is about 35 m.

The maximum thickness of this unit of about 55 msec TWT (about 41 m, Fig. 6.13) occurs only locally, probably related to local subsidence. The unit thins toward the north and
may onlap onto unit A1, but no clear onlapping features are seen due to poor data. An anomalous thickness associated with the progradational clinoforms on the southeastern area is absent, indicating that the clinoforms may relate to lateral southeastward movement of the 'shelf' slope rather than a prograding delta system. The clinoform direction suggests that the sediments were northerly derived.

c. Subunit A3

Sea level rose again after the deposition of the regressive sequence in subunit A2. Subunit A3 is almost completely eroded by the subsequent relative sea level fall. The presence of this sequence can only be observed on Sections 6.5, 6.8, 6.9 and 6.11. The seismic facies shown by the remainder of the sequence is a high amplitude, high continuity parallel configuration pattern which suggests shallow marine clastic deposits.

6.5.4 Unit B

A major sea level fall occurred at the end of unit A deposition which led to the subaerial exposure of Madura Strait, creating a type I unconformity. Unit B consists of a number of subunits which can be observed on Sections 6.5, 6.8, 6.9 and 6.11.

a. Subunit B1a

Subunit B1a displays complex channel cut and fill features. Seismically it is characterised by parallel to hummocky reflection patterns combined with low continuity and variable amplitude. The maximum determined thickness is 50 msec TWT (about 37 m, Fig. 6.14). It appears that thick deposits occur along the axial line of the syncline (Madura Strait), pinching out northward (Sect. 6.8) and southward. These suggest that a river system was developed along the median line of the syncline.

b. Subunit BV

A sea level rise led to the deposition of the volcanic-derived subunit BV which downlaps onto subunit B1a in the northward direction. Seismically, subunit BV is characterised by a nearly reflection free pattern indicating a nearly homogeneous deposit. Strong and relatively continuous reflections occur at the top of the subunit where they are also associated with channelling (Sects 6.8 & 6.9). The isochron map Figure 6.15 displays the limited areal extend of this subunit. The conical form of the isochron map which thickens southeastward has
suggested that the deposition was related to the volcaniclastic detritus derived from the Lurus-Ringgit-Beser Volcanic complex.

The Lurus-Ringgit-Beser Volcanic complex comprises three leucite bearing volcanoes (Lurus, Ringgit and Beser Volcanoes). The volcanism occurred in the Early Pleistocene and their deposits conformably overlie the neritic to littoral deposits of Plio-Pleistocene age (van Bemmelen, 1949). These ages are consistent with the well data (MS1-1) which assigns unit Pre-A to the Late Pliocene. A relatively steep depositional setting is shown in the seismic sections (Sects 6.5, 6.8 & 6.9) suggesting that this subunit was deposited in a fan-delta setting (Nemec & Steel, 1988). Stratigraphically, subunit BV is comparable to the fan-delta deposits of the Pucangan Formation in the eastern part of the Kendeng Zone (Chapter 5.7) which were mostly derived from the Wilis Volcano.

A similar setting, in which sedimentation was directly correlated with volcanic activity, has been reported by Ballance (1988) who studied the Late Jurassic Huriwai delta of New Zealand. He documented that volcaniclastics were fed to a marine environment through a braided plain delta system, and that there was evidence that the system received periodic influx related to volcanic eruptions.

c. Subunits B1b, B2, B3 and B4

These subunits can be recognised only in the eastern part of Madura Strait (Sects 6.5, 6.8 & 6.11). They downlap on subunits B1a and BV in a southeasterly direction. In the western part, the recognition is hindered particularly because of high sound attenuation, multiples and only subtle facies differences between the subunits. Due to the lack of erosional features on top of each unit, these units may be regarded as parasequences (not sequences) in the sense of van Wagoner (1985) and Posamentier and James (1993) (see Chapter 4.2 for definition).

In the eastern part of Madura Strait, although the downlap patterns of subunits B1b, B2 and B3 are recognised, no significant form of prograding clinoforms has been developed indicating no sedimentary bypassing. However, the subunits overall show a southeastward progradation as indicated by the relative position of downlaps. Also apparent is that the depositional slopes were very low and the subunits tend to be characterised by an upward loss in acoustic transparency which indicates a shoaling upward (Sect. 6.11). The thickness variations in these subunits is shown in Figures 6.16, 6.17 and 6.18, which all display a
southeasterly thinning. Some localised thickenings on subunit B3 are related to local synclinal subsidence. These isochron maps seem to underline the importance of the straits between the Madura, Gili Raja and Gili Genting Islands which have acted as feeder channels from the Madura Island to Madura Strait.

6.5.5 Unit C

A relative sea level fall occurred at the end of unit B deposition and Madura Strait was exposed subaerially (type I unconformity). Unit C consists of at least eight northwesterly derived subunits (subunit C1 to C8) and two southerly derived subunits (included within subunit CV). As for the deposition of the marine facies unit B, northwesterly derived sedimentation appears to be dominant and shows a consistent southeastward 'shelf' progradation. Isochron maps, Figures 6.19 to 6.24, also suggest the important of the straits between Madura, Gili Raja and Gili Genting Islands as feeder channels, in addition to sediments derived from the west. The uppermost subunits (C6, C7 and C8) mostly have been eroded during the sea level fall at the end of unit C deposition. The presence of these subunits is noted on seismic lines from the southeastern part of Madura Strait by their clinoform patterns (Sects 6.5 & 6.8). The other subunits are mappable throughout the strait.

a. Subunits C1a and C1b

Subunit C1a was deposited unconformably on top of unit B. This subunit is characterised by subparallel to hummocky reflection patterns with variable relative amplitude and low continuity. Association with channelling is common. Subunit C1a is interpreted as fluvial deposits. The isochron map (Figure 6.19) shows that this subunit is widely distributed, with an average thickness of about 30 msec TWT (about 22.5 m). Locally thick deposits occur along an east-west trending synclinal zone in the northern half of Madura Strait which probably indicates an early development of Pleistocene folding that restricted the development of fluvial subsystems to these synclines.

A relative sea level rise occurred after the deposition of subunit C1a. Subunit C1b downlaps on subunit C1a (Sects 6.1, 6.2 & 6.10); however, the downlapping appears to be very subtle due to the low depositional gradient. Seismically it is characterised by parallel to subparallel reflection patterns with variable amplitude and continuity. A relatively high continuity with strong reflections can be found particularly in the western part of Madura Strait (Sects 6.1 & 6.2) which is interpreted as very shallow marine deposits. The isochron map
Figure 6.20 indicates that this subunit was only thin and was restricted to the western and northern part of Madura Strait. This subunit has a thickness of about 15 msec TWT (about 11 m), and is totally absent in the southeastern part of the area.

Based on the relative position and facies difference between subunits C1a and C1b, it is suggested that they may represent lowstand and highstand systems tracts respectively. The absence of subunit C1b on the southern part can be regarded as due to sediment starvation, since the sediments, which mostly derived from western and northern parts, were mostly deposited close to the western and northern basin margins because of a sea level rise.

b. Subunit C2

Subunit C2 has been derived from the north and western part of the basin. It was deposited contemporaneously with subunit CV. In the eastern part of Madura Strait subunit C2 shows a southward downlap. Its internal reflection pattern is characterised by an upward loss in acoustic transparency which indicates a shallowing upward regressive sequence (Sects 6.8 & 6.11). The upper part shows a subparallel reflection pattern with low amplitude, possibly representing shallow marine clastics. The thickness attains 20 msec TWT (about 15 m), and thins southward where the subunit is probably amalgamated with the southerly derived subunit CV (Fig. 6.21).

In the western part of Madura Strait this subunit also shows a regressive sequence. The lower part (C2a, Sect. 6.1) is characterised by a reflection free pattern suggesting a homogeneous sediment, probably homogeneous mudstone. In the upper part, the subunit (C2b) is channellised and associated with a low continuity, variable amplitude parallel reflection pattern. This part is interpreted as a non-marine fluvial facies.

c. Subunit CV

Subunit CV was deposited northward, contemporaneous with the northerly derived subunit C2. The distribution of this subunit is mostly confined to the southeastern part of Madura Strait. The distal part is shown in Sections 6.5, 6.8 and 6.9. The proximal part has mostly been eroded during the sea level fall at the end of unit C deposition. On these sections this subunit is characterised by relatively strong and high continuity reflections of oblique (Sects 6.5 & 6.8) to parallel patterns (Sect. 6.9). The isochron map (Fig. 6.21) shows that this subunit very rapidly thickens southward (landward). As for subunit BV, the sediments are
suspected to have been derived from volcanic deposits of the Lurus, Ringgit and Beser Volcanoes. Both subunits BV and CV were not developed as extensively as the northerly derived subunits. This is shown in Sections 6.5 and 6.8 where subunit BV has been overlain by subunits B1, B2 and B3, and subunit CV by subunits C4, C5, C6, C7 and C8. Apparently the sediment supply from the south rapidly diminished soon after the deposition of subunits BV and CV, indicating periodic as opposed to continuous volcanic activity.

d. Subunits C3 and C4

Subunits C3 and C4 are regressive deposits. On the seismic lines from the eastern part of Madura Strait these subunits are characterised by an upward loss in acoustic transparency (Sects 6.8, 6.9 & 6.11). The lower parts are characterised by a reflection free pattern indicating homogeneous deposits of possibly mudstone. The upper parts are characterised by parallel reflection patterns with low to high amplitudes indicating shallow marine facies with frequent intercalations of mudstone and sandstone. Prograding clinoforms within the slope facies are not well developed, probably due to low sediment supply. A thin oblique progradational pattern is only shown in subunit C4.

In the western part of the strait, the upper parts of the subunits are characterised by parallel reflection patterns with variable amplitude and low continuity. They are also associated with channelling (Sects 6.2 & 6.10). Eastward these patterns change to a parallel pattern with low amplitude (Sect. 6.10). These features suggest a lateral facies change from non-marine in the west to shallow marine in the east. Section 6.10 also shows the presence of an eastward parallel-oblique progradation pattern in subunit C4. This is interpreted as an eastward prograding coastal facies (Sect. 6.10).

The isochron maps (Figs 6.22 & 6.23) indicate that the thickness of subunits C3 and C4 is relatively similar, averaged at 20 msec TWT (about 15 m). The subunits thin and downlap southeastward, suggesting a general progradation in this direction.

e. Subunit C5

Subunit C5 represents a type I sequence, composed of two subunits - C5a and C5b. Figure 6.24 is an isochron map of combined C5a and C5b. Subunit C5a is a lowstand deposit (wedge) which accumulated as sea level fell at the end of subunit C4 deposition. The areal distribution of this subunit is very limited and is confined to the basin floor in front of the
prograding slope subunit C4 which is located in the eastern part of Madura Strait (Sects 6.8 & 6.9). In Figure 6.24 subunit C5a is not represented, but can be found beneath the local thickening in the middle east of the area. A separate isochron map was not constructed due to the difficulty in defining the upper boundary of this subunit on most sections. Seismically it is characterised by a mounded form, and probably baselaps on both subunits CV and C4. Its internal reflection patterns, as shown on Section 6.8 are parallel and show a medium amplitude and high continuity. Its deposition may have contemporaneously occurred with the channelling and prograding coastal facies at the upper part of subunit C4 (Sects 6.2 & 6.10).

As the sea level rose, deposition of subunit C5a ceased. Subunit C5b developed as a highstand regressive deposit (Sects 6.8 & 6.10). The regressive situation apparently occurred as the result of a combined sea level stillstand and high sediment supply. This is obvious in the eastern part of the area where sedimentary bypassing occurred, combined with extensive southeastward sigmoid prograding clinoforms (Sects 6.5, 6.8 & 6.9). The estimated water depth of basin floor measured from the clinoform profile was about 70 msec TWT (about 52 m). Figure 6.24 shows that the thickness of the prograding lobe reaches 55 msec TWT (about 41 m). This southeastward prograding unit is interpreted as a prograding delta system. Unfortunately, the recognition of the complete delta system is hindered, as its proximal facies was eroded during the sea level fall at the end of unit C deposition. However, the direction of clinoform dip suggests that the delta system was sourced from the northwest, and again indicates that the strait between Madura and Gili Raja Islands acted as a feeder channel.

A much smaller delta system contemporaneous with the previous delta system probably developed in the western part of the area as displayed on Section 6.10. It is separated from the previous delta system by a shallow marine facies, indicated by a seismic facies with a high amplitude and high continuity parallel pattern on Figure 6.24. In the proximal direction (westward) the delta system is characterised by high amplitude hummocky clinoforms associated with channelling (Sect. 6.10 & Fig. 6.24). This part is interpreted as a fluvial facies. The delta front developed on a very low slope and profile. Seismically it is characterised by a thin complex sigmoid-oblique reflection pattern with strong amplitude indicating a low sediment supply and high energy depositional setting. Eastward these features rapidly change into a low to high amplitude parallel reflection pattern of probably mudstone-prone facies (Sect. 6.10).
f. Subunits C6, C7 and C8

These subunits can only be recognised by their strong and continuous reflections on top of prograding clinoform facies in the eastern part of the area (Sects 6.5, 6.8 & 6.9). Their characteristic 'shelf-type' facies have mostly been eroded during the sea level fall at the end of unit C deposition.

The stacking pattern of subunit C6 on Sections 6.8 and 6.9 suggests that the sea level rose after the deposition of subunit C5b. This increased the accommodation space and the sediments aggraded on top of subunit C5b. In contrast, the remnants of subunits C7 and C8 do not permit evaluation as to whether these subunits were deposited in transgressive or regressive settings.

Seismically subunit C6 is characterised by a parallel reflection pattern combined with low amplitude indicating a low energy depositional setting. The internal pattern of clinoform facies, subunits C7 and C8, is characterised by a lack of reflections which indicates homogeneous sediments of probably mudstone.

6.5.6 Unit D

The beginning of unit D is marked by pronounced erosion which is related to a significant sea level fall at the end of unit C deposition. This sea level fall exposed the whole Madura Strait subaerially, and part of unit C6 and most of units C7 and C8 were eroded. The prolonged exposure was also encouraged by the commencement of the major Pleistocene folding event. A number of subunits can be recognised in unit D, but most of them are too thin to be mapped. In addition, the much less pronounced subunit boundaries make their recognition difficult. Only the lowermost subunit (subunit D1a) was mapped seismically as a distinct unit, and a combined isochron map of the remaining units was produced. The present sea floor is the upper boundary of unit D.

a. Subunit D1a

On the seismic sections subunit D1a is distinctive and characterised by variable amplitude and poor continuity of reflections. It is interpreted as an extensive fluvial deposit. In the upper part parallel to hummocky reflection patterns are shown, which may be associated with ponds or floodplain deposits (Sect. 6.8). This part is associated with a river system (Sects 6.9 & 6.11) which was later abandoned during the subsequent sea level rise. Figure 6.26 displays
the positions of observable channel cuts which support the occurrence of an axial river system in the synclinal basin. The flanks of the basin contributed sediment to this river system. Toward the east, an eastward running main channel developed and exhibited a high sinuosity.

The isochron map Figure 6.25 indicates that the thickness of subunit D1a varies from 0 to 70 msec TWT (0 to 45 m). This map and the mesh diagrams in Figures 6.29b and 6.29c clearly show that the deposition of D1a was strongly controlled by the formation of Late Pleistocene folds. Thick deposits are confined to the synclinal areas. The westward relative thickening of this subunit is due to the fact that the downwarping was more intensive in two distinct centres in the western part of the basin.

Apart from the local control by the Late Pleistocene fold formation, the general depositional style of this fluvial system is similar to subunits B1a and C1a. Miall (1982) pointed out that river flows basically can be generalised into transverse, flowing directly from the uplifted area across structural grains, and longitudinal, flowing parallel to the strike of the basin axis. The river systems in Madura Strait (B1a, C1a and D1a) are included in the second category, where the basin controlling the flow has a relatively simple synclinal geometry.

A modern example of such a longitudinal river system is the Po River system of northern Italy, which has been studied by Ori (1993). This river system, which supplies most of the detritus into the Adriatic Sea, is confined within the asymmetrical subsiding Po Basin, which is bounded to the north by the Alps mountain chain and to the south by the active thrust-faulted belt of the Apennines. Alluvial fan and contributary river systems occur on both flanks of the basin, and flow into the main longitudinal Po River. A complex interplay between different subsidence rates within the basin, climatic conditions and different types of sediment sources, has led to the difference in fluvial facies and body geometries between the north and southern flanks of the basin. In the southern sector, where thrust faulting is active, fan and plain systems are thin and are formed by a few depositional processes. In contrast, the northern sector is characterised by laterally widespread fan systems formed by many sedimentation events. Although the study area has a different setting, this model of fluvial sedimentation may be similar to the more active alluvial fan sedimentation occurring in the vicinity of the active Quaternary volcanism in the south.
b. Subunit D1b

Rapid rise of relative sea level occurred after the deposition of subunit D1a and the depocentres shifted toward the basin margin. Drainage patterns created on top of subunit D1a were completely abandoned. Sections 6.1 and 6.7 suggest that the top of this subunit (top of D1) occurs very close to 66 msec TWT (-50 m from the present sea level). While the lowest observable previous sea level lowstand feature (fluvial channelling) on top of subunit D1a occurs at 140 msec TWT (about -105 m from the present sea level; Sect. 6.9). Assuming very minor structural deformation, the sea level rise from the previous lowstand was at least 74 msec or 55 m.

In the western part subunit D1b is relatively thin (Sect. 6.1), characterised by medium amplitude, relatively continuous reflection patterns, and is associated with small channels. It is suspected that here, this subunit was deposited in a very flat area in a very low energy environment (?tidal flat).

In the deeper setting (Sects 6.2 & 6.3), subunit D1b is typified by a reflection free character of possibly homogeneous mudstone, associated with mounded and irregular shapes of probably patch reefs (bioherms). A very strong reflector on Section 6.3 (line MC-MC'), which is possibly associated with an undulating sheet carbonate surface, has produced pronounced multiples and diffractions.

In the topographically steeper area (Sect. 6.7) subunit D1b is wedge shaped and shows a relatively continuous, medium amplitude reflection pattern. Figure 6.27 illustrates the palaeogeography during the deposition of subunit D1b. The maximum thickness is about 21 msec TWT (16 m).

c. Subunit D2

Subunit D2 comprises a number of units that resulted from repetitive small transgressions and regressions. Two depocentres were present along the southern and northern margins of Madura Strait. Deposition along the southern margin was less extensive compared to the northern margin, which was probably due to a lower sedimentation rate. On Section 6.3 (line MC-MC') where the topography was relatively steep, four wedges of transgressive/regressive facies can be identified and are referred as D2a to D2d. These wedges progressively shifted
basinward which may suggest a progradation or gradual relative sea level lowering while oscillating.

A similar mode of deposition, but on a larger scale, has been documented by Tesson et al. (1993) on the Rhone (France) continental shelf. They associated the wedges with alternating episodes of coastal progradation and transgression, which are believed to occur during the latest Pleistocene glacial lowstands.

Extensive development of subunit D2 occurred on the northern margin of Madura Strait (Sect. 6.6). It is interpreted as a prograding wedge of delta facies, fed through a channel system in between Madura and Gili Raja Islands. The thick deposit of the delta system forms the main component in the combined isochron map Figure 6.28, and appears as a pronounced bulge. The thickness reaches more than 30 m. This extensive development probably resulted from a high sedimentation rate and the presence of folds (Fig. 6.29b) on the sides of the delta which acted as a barrier to wave processes, and also concentrated sedimentation by funnelling through a narrow gap.

d. Subunit D3

Subunit D3 is the uppermost observable sequence on the seismic records. The lower boundary is probably an erosional surface which is very prominent in the northern part of the strait where the upper parts of deltaic deposits have been truncated (Sects 6.6 & 6.7). The thickness of subunit D3 may reach 15 m (Sect. 6.6). Details of this subunit are mostly obscured by peg-leg multiples at the top of the subunit.

The lower boundary of subunit D3 is associated with strong and irregular reflections which may have resulted in very poor seismic data quality underneath (Sects 6.6 & 6.7) and obscured details of the earlier strata. The lateral extent of such reflections appears to follow the extent of the previous deltaic deposit, subunit D2. A similar phenomenon has been reported by Chiocci (1994) in association with Late Pleistocene delta systems, interpreted as the result of interstitial gas within the delta system. However, a close examination on some sections in the study area has indicated that such a feature occurs because of a very thin layer in between the regressive deltaic deposit (subunit D2) and the overlying subunit (D3), which may be associated with transgressive lag deposits over the deltaic deposits.
6.6 DISCUSSION

6.6.1 Correlation with onshore data

Quaternary sediments are widely exposed in the onshore part of the East Java area (Kendeng Zone) along Late Pleistocene anticlinal flanks. Discussion regarding their depositional environments has been given in Chapter 4. Correlation between these deposits and Quaternary sediments developed in Madura Strait has been made possible due to the fact that during the Early and Middle Pleistocene East Java and Madura Strait formed a single elongated basin (van Bemmelen, 1949).

During the Middle Pleistocene relative sea level rose significantly in East Java. This was marked by a widespread deposition of shelf mud (green mudstone) which signifies the beginning of the Kabuh Formation deposition (Chapter 5.8; Figs 5.48, 5.49 & 5.50). The stratigraphic position of this mudstone, which overlies the fluvial and coastal deposits (unit U2) of the Pucangan Formation, has provided a good seismic marker due to the difference in physical properties between the two formations. In the closest seismic line to the onshore area (line MA-MA', Sect. 6.1), the mudstone facies is represented by seismic facies subunit C2a, and the fluvial and coastal facies are represented by subunits C1a and C1b respectively.

A lithofacies study on the outcrops on the anticlinal flanks north of Mojokerto (Chapter 5.7; Figs 5.30, 5.37, 5.44 & 5.48) indicates that sediments below the fluvial deposits (unit U2) of the Pucangan Formation are mostly tidally influenced deposits. Alluvial fan deposits (unit M3) occur in the section from the Perning area (Fig. 5.37) where they also contain abundant vertebrate fossils and hominid remains. These deposits according to Duyfjes (1938b) and van Bemmelen (1949) were sourced from the Wilis Volcano. The stratigraphic position of these deposits, which are below unit U2, is exactly similar to the fan system (subunit BV) developed in the southeastern part of Madura Strait (Sects 6.5, 6.8 & 6.9). This later fan system occurred as the result of activity of the Lurus, Ringgit and Beser Volcanoes. Figure 6.30 displays the correlation between outcrops in the area north of Mojokerto and seismic sections MB-MB' (representing the western part) and ME-ME' (representing the eastern part).

The lithofacies study in the northern part of the Mojokerto area indicates that the Pucangan Formation is underlain by thick lacustrine mudstone deposits of the Lidah Formation, which attains 500 m. Field observation in the Banyuurip area (Fig. 5.30) indicates that the facies is almost entirely massive. The thickness and lack of bedding in this mudstone facies would
produce a prominent, thick, reflection-free unit in the seismic sections studied. These features occur in unit Pre-A which is widespread in Madura Strait. Well data (MS1-1) indicate that this unit is characterised by non-calcareous mudstone occurring at a depth of 277 to 451 m, which was deposited during the Late Pliocene. Field studies in the Gili Raja Island (Fig. 6.1) found that this same facies lies on the Pliocene marly facies (Pasean Formation) which is equivalent to the Mundu Formation in East Java.

6.6.2 Depositional timing

a. Units A, B and C

Absolute and relative dating have never been carried out on Quaternary sequences in Madura Strait. In this study, the inferences on the depositional timing of Quaternary sequences in Madura Strait are based particularly on the correlation of Madura Strait sequences with the inland sequences (north of Mojokerto). The previous section has discussed the facies relationship between these areas.

A recent palaeomagnetic study was carried out in the Perning area, north of Mojokerto City (some 35 km west of Madura Strait) by a Japan-Indonesia joint research team in 1986-1988 (Hyodo et al., 1993; Chapter 3.5). The palaeomagnetic measurements that were carried out on mudstone and tuffaceous mudstone samples have revealed the presence of two Quaternary magnetic events, the Olduvai and Jaramillo. The results of these measurements are shown on the measured section in Figures 5.37 and 6.30. The Jaramillo magnetic event is coincident with the alluvial fan deposits (unit M3) which in the seismic study is correlated with the subunit BV.

Based on the previous correlations, unit C in the seismic sections has to be younger than the Jaramillo magnetic event. The two important features that occurred during the deposition of this unit were (1) the presence of two significant and widespread subaerial exposures which may indicate significant relative sea level falls, and (2) the presence of highly repetitive sequences. The first subaerial exposure occurred at the beginning of subunit C1a deposition and the second occurred after the deposition of subunit C4. These features suggest that the deposition of subunits in unit C corresponds to the relatively regular climatic cycles during the Brunhes Normal Epoch (Fig. 6.32), and the two significant relative sea level falls are speculatively correlated to the glacial maximum stages 16 and 10 where the intensities of glaciation were relatively higher.
b. Unit D

Inferences on depositional timing of subunit D were based particularly on the assumption that structural deformations were minor after the deposition of subunit D1a. The positions of erosional features with respect to the present position of sea level were then compared with the adjusted sea level curve of Huon Peninsula (New Guinea) by Chappell and Shackleton (1986, Fig. 6.31) and the oxygen-isotope record compiled by Harland et al. (1989, Fig. 6.32).

Subunit D1a is bounded at its base and top by prominent erosional surfaces. The deepest fluvial channelling at the top of this subunit occurs at depth of about 140 msec TWT (-105 m from the present sea level, see Sect. 6.9). Between this fluvial channelling and the erosional surface at the top of subunit D2 the sequence is characterised by a number of small marine units (D1b, D2a, D2b, D2c, & D2d; Sects 6.3 & 6.6). These suggest that unit D1a was deposited during the sea level lowering of oxygen isotope stage 6. Deposition was enhanced by the development of folds, and consequent increased erosion, during this time. Subunits D1b, D2a, D2b, D2c and D2d on Section 6.3, and subunits D1b and D2 on Sections 6.6 and 6.7, which were deposited after folding, can be interpreted to have resulted from minor sea level oscillation after the oxygen isotope stage 6 (Figs 6.31 & 6.32).

Subunit D3 is the uppermost observable unit in the study area. As shown on Sections 6.6 and 6.7, the base of this subunit is probably an erosional surface. It is interpreted that the erosion took place during the oxygen isotope stage 2 (the last glacial maximum). Sea level lowering during this time may have reached more than 100 m below the present sea level (Fig. 6.31; Chappell & Shackleton, 1986).

6.6.3 Sedimentary source

The study has indicated that Madura Strait has been fed from three different sources of sediments, in the south, west and north. Seismic features and isochron maps show that the south and westerly derived sediments were not as effective a source as northerly derived until the Middle Pleistocene. This probably relates to the fact that sediments from the south and western parts of the basin were volcanogenic. Their supply depended on the intermittent volcanic activity, and lateral deposition would occur when a sea level rise provided an accommodation space. In contrast, sediments from the north were the erosion products of Madura Island. The absence of Quaternary marine deposits in this island (Duyfjes, 1938c; Aziz et al., 1993; Situmorang et al., 1992) suggests that the island may have been subaerially
exposed since the end of the Pliocene. In addition, the Miocene and Pliocene sediments were then folded during the Plio-Pleistocene with at least two pairs of east-west trending anticline-synclines occurring along this island (Duyfjes, 1938c; Aziz et al., 1993; Situmorang et al., 1992). The folds have allowed the development of axial river systems which may have become effective erosion agents and maintained a continuous sediment supply to Madura Strait.

6.7 SUMMARY AND CONCLUSION

The seismic stratigraphic study in Madura Strait revealed the presence of four major Quaternary sedimentary units (units A, B, C and D). These units are bounded at their base, above the latest Pliocene unit Pre-A, by a surface of non-deposition or erosion. Due to their durations being less than 1 Ma, these Quaternary units can be classified as fourth order sequences. Detailed analysis of these units has further found the presence of smaller subunits. The lateral distribution of these subunits is strongly controlled by their depositional setting. The southerly derived subunits, although attaining a significant thickness, are commonly limited in their lateral extent. The direction of subunit thickening suggests that these units were derived from the volcanic products of the Lurus, Ringgit and Beser Volcanoes. The north and northwesterly derived subunits, in contrast, were mostly extensively developed and show a cyclical pattern of lowstand and highstand. The two laterally extensive delta units were related to these sources.

The depositional timing of depositional units in Madura Strait has been inferred by correlation with the micropalaeontologically dated well MS1-1, the palaeomagnetically dated section in the Perning area (north of Mojokerto), the marine oxygen isotope record and the adjusted sea level curve of the Huon Peninsula (New Guinea). A rough estimation on the depositional timing has been obtained. The deposition of unit A occurred in the Early Pleistocene, probably starting close to the Olduvai magnetic event. Unit B was deposited during the late Early Pleistocene to early Middle Pleistocene. Unit C was deposited in the middle Middle Pleistocene to late Middle Pleistocene, and unit D was deposited during the Late Pleistocene.
CHAPTER SEVEN
LATE TERTIARY AND QUATERNARY SEISMIC SEQUENCE STRATIGRAPHY OF THE SOUTHEAST JAVA SEA

7.1 INTRODUCTION

The southeast Java Sea is the submerged part of the Sunda Shelf (Chapter 2), and lies in a different tectonic and geological setting from Madura Strait. In Madura Strait, a north-south compressional tectonic stress is dominant, and the sedimentary basin is simply an east-west synclinal basin, which is bordered in the north by the anticlinal Rembang Zone (Madura Island), and by a zone of active Quaternary volcanism in the south. In contrast, the southeast Java Sea lies on a relatively stable continental shelf, particularly during the Quaternary, as indicated by the present study which found less structural deformation.

Marine geological investigations in the southeast Java Sea have mostly been carried out as part of regional studies on the Sunda Shelf. Based on regional geophysical data, Ben-Avraham and Emery (1973) noted that Tertiary sedimentation in the southeast Java Sea occurred in basins which were bounded mostly by northeastward trending faults. This finding was confirmed by Sudiro et al. (1973) and Bishop (1980), who further indicated that these basins are commonly half grabens. The present study confirms the finding by these investigators that during the Plio-Pleistocene, marine sedimentation continued over these basins. However, this study also finds that synclinal east trending basins occurred during the Pliocene and Early Pleistocene. Sedimentation in these basins gradually changed basin morphology to a relatively flat plain at the end of the Quaternary.

This chapter discusses the sedimentary facies distribution and chronology in the southeast Java Sea during the Late Tertiary (particularly the Pliocene) and Quaternary. This discussion is an extension of similar discussions in Chapters Four and Six which examined the northern part of East Java and Madura Strait respectively. As in these previous chapters, the discussion relies heavily on seismic data, which have been interpreted by applying the sequence stratigraphic concepts developed by Vail et al. (1977), Posamentier et al. (1988) and Posamentier and Vail (1988). In many aspects, this study faces the same problems as in
Madura Strait due to the lack of reliable dating, as well as published geological studies. Thus, this study represents a preliminary interpretation of the sequences in this area.

The seismic sequence stratigraphic analysis presented in this chapter is considered to be a general overview in terms of sequence delineation. Detailed analysis cannot be carried out on most sequences, because of three major problems: (1) the limited penetration of the seismic system used, often resulting in seismic signals too weak to be interpreted, (2) presence of strong multiple reflections, particularly in the area where strong reflectors, such as unconformity surfaces, are present, (3) occurrence of complex channel cuts and fills. The first two problems apply to the Miocene sequence and the lower part of the Pliocene sequence, and the last applies to the Quaternary sequences.

7.2 DATA

The data base for this study is drawn from seismic profiles totalling some 3750 line km in the Java Sea (Fig. 7.1). All geophysical data were obtained from the Marine Geological Institute of Indonesia which ran the survey in 1989/1990. The seismic system used is similar to that in the Madura Strait survey (Chapter 6), a single channel 600 Joule sparker system, fired every 0.5 seconds. In addition, a 300 Joule boomer system was run in the area near the mouth of the strait between East Java and Madura Island. The seismic signals were not tape recorded, but directly band pass filtered (200-2000 Hz) and graphically recorded in analog format during the survey. Due to this technique, no further data processing was carried out. The estimated vertical exaggeration on all seismic sections presented in this chapter is about 21 times for the sparker system and about 17 times for the boomer system. The ship positions during the survey relied mostly on the Magnavox MX 1157 satellite navigation system with point accuracy of 36 m. The profiles were oriented north-south and spaced 5 to 10 km apart.

The isochron (thickness in time units) and time structure maps presented in this chapter are the results of digitisation of all the seismic data following the seismic interpretation. The computer programs used are the same as those used in the Madura Strait study, and the source codes are provided in Appendix F. On the seismic profiles presented, the depth was calculated based on an assumed seismic velocity of 1500 m/sec, because of the absence of seismic velocity data. This depth is in fact valid for the water column only, and should be regarded as a rough estimation. Figure 7.2 shows the locations of seismic lines used during the discussion.
Stratigraphic control for the analysis of seismic data was provided by six petroleum exploratory wells, JS1-1, JS2-1, JS3-1, JS8-1, JS10-1 and JS16-1. A graphic summary of biostratigraphic analyses of these wells (except for the JS1-1 well) is provided in Appendix E. Although more well data are available for the study than in Madura Strait (Chapter Six), the study suffers from a similar problem. Biostratigraphic and lithofacies analyses done on these wells were based on well cuttings which were commonly sampled every 30 ft (9.1 m) penetration. These data may not be accurate, but have narrowed the age estimation of the stratigraphic time markers. A more serious problem is that the well data may not be applicable to the Quaternary sequences, because most of the wells were drilled on top of anticlines where Late Tertiary and Quaternary sediments are thin or absent due to erosion and sediments wedging out.

7.3 STRUCTURAL GEOLOGY AND BATHYMETRY

The discussion presented in Chapter 2 indicated that northeast-trending structures are prominent in the southeastern part of the Java Sea (Figs 2.2 & 7.3), and have been the major control for the Early Tertiary sedimentation. Many of these structures are half grabens that formed on the pre-Tertiary shelf (Bishop, 1980). These major features were interpreted by Ben-Avraham and Emery (1973) as resulting from past interaction between the Eurasian and Indian-Australian lithospheric plates, the principal ridges probably being part of an island arc system active during the Late Cretaceous-earliest Tertiary (Bishop, 1980). Such an island arc complex has been deduced from the occurrence of pre-Tertiary ophiolites indicating a subduction complex cropping out in Central Java, and in Southeast Kalimantan (van Bemmelen, 1949) possibly representing a previous subduction complex (Katili, 1989).

The Karimunjawa Arch is the dominant ridge in the eastern Java Sea, which extends into the offshore area of southern Kalimantan as a broad positive feature (Bishop, 1980). It is capped by the Karimunjawa Islands on which pre-Tertiary quartzite and phyllitic shale, cut by basic dykes, and probable Quaternary fissure-eruptive sheets crop out. This arch is separated by the narrow, northeast trending West Florence Deep from the Bawean Arch. The Bawean Arch is characterised by alkaline volcanism of the latest Neogene or Quaternary and the steeply dipping Miocene marine strata (van Bemmelen, 1949).
The present seismic study on the southeastern part of the Java Sea found that the Pliocene and Early Pleistocene sedimentation was still partly controlled by the structural features mentioned above. The sediments were derived primarily from the two positive features, and from the growing anticline of Madura Island. The structures became ineffective from about the Middle Pleistocene when the morphology became relatively flat, and the Middle and Late Pleistocene strata are characterised by extensive channel cut and fills that have resulted from frequent Quaternary sea level fluctuations. Intense faulting in the study area is absent, but active faulting, at least until the Late Pleistocene (as indicated by faulted Late Pleistocene sediments) occurs on the borders of the major structural elements in the study area (Fig. 7.3).

Figure 7.4 displays the three dimensional mesh diagrams which represent the present situation at the top of the Miocene, Pliocene and Pleistocene (bathymetry) strata respectively. The pronounced highs in the north represent the Karimunjawa and Bawean Arches. It is shown that besides sedimentation occurring in the West Florence Deep (a half graben system), the Pliocene sedimentation was strongly controlled by an east-west, asymmetrical synclinal structure between the Karimunjawa and Bawean Arches in the north, and the east trending ridge in the south which extends from the north of Java to Madura Island. This ridge is the anticlinal zone termed the Rembang Zone in the previous chapters. More complex Pliocene deposition occurred in the area north of Madura Island where a small east trending ridge was developing during the deposition which allowed the accumulation of a locally thick Pliocene deposit.

The present bathymetry of the study area (Figs 7.4a and 7.5) may reflect the geological processes in the Late Pleistocene. At present the deepest water setting in the study area is approximately 85 m. The gentle, gradual northeastward deepening may partly relate to the growing anticlines in the northern part of East Java and Madura Island. As revealed by the present seismic analysis, the average gradient is 0.4 m/km. In the area closer to the East Java and Madura coasts, the bathymetry rapidly shallows due to the Late Pleistocene coastal deposits derived from these areas. In the northwestern Rembang Zone, the deposits were sourced mainly from the Quaternary Muria Volcano on the peninsula northwest of Rembang. In the far north of the map (Fig. 7.5) the bathymetry shows pinch and swell structures which may reflect an underlying structural feature.
7.4 SEISMIC STRATIGRAPHY

Seismic analysis indicates that the Late Tertiary and Quaternary sediments in the study area can be subdivided into three major seismic units. These are referred to as units JM, JP and JQ, which correspond to the Miocene, Pliocene and Pleistocene ages respectively. Ages of these units are based on biostratigraphic data from petroleum exploratory wells in the Java Sea provided by the Cities Service Indonesia Inc. and Pertamina Company, and outcrop geological studies on Madura Island by Situmorang et al. (1992) and Aziz et al. (1993). The units identified are bounded at their bases by regional unconformities [type I unconformity according to the Posamentier & Vail (1988) criteria], and each of these units consist of two or more subunits which are characteristic in areal distribution, thickness, and seismic character.

7.4.1 Unit JM

This unit can only be observed on the structurally high areas, such as near the Bawean and Karimunjawa Arches, and Madura Island. Its internal reflection patterns and areal distribution are poorly defined, particularly because of the limited penetration of the seismic system used and strong multiple reflections. The lower boundary is unidentifiable, but the upper boundary is a regional unconformity as shown by a pronounced erosional surface on the structurally high areas (Sects 7.1, 7.3 & 7.8). The estimated age of the unit is Miocene, based on the correlation with the micropalaeontological data of several wells in the study area (Appendix E). The Miocene strata in the JS8-1 and JS3-1 wells are typified by the occurrence of larger foraminifers which are characteristic of this age, such as *Lepidocyclina*, *Miogypsina*, *Flosculinella* and *Alveolinella* (Rahaghi, 1984; Hamaoui, 1984; Butterlin, 1984). The top of the Miocene strata in most wells was also defined by an abrupt increase in planktonic foraminifers. This change indicates the occurrence of a significant sea level rise at the end of the Miocene, since larger foraminifers are commonly associated with shallow marine waters (Moore et al., 1952; Serra-Kiel & Reguant, 1984), in contrast with planktonic foraminifers (Murray, 1991).

The seismic lines Sections 7.1 and 7.7 suggest that unit JM can further be subdivided into subunits JM1 in the lower part and JM2 in the upper part. The subunits appear to be separated by an erosional surface. On the Karimunjawa and Bawean Arches units JM1 and JM2 are characterised by a medium amplitude, continuous parallel-subparallel reflection pattern (Sects 7.1 & 7.7) which is interpreted as interbedded sandstone and mudstone facies. These deposits
are overlain by a reefal limestone on both arches, and possibly on Madura Island as reported by Situmorang et al. (1992) and Aziz et al. (1993). The mounded external form of the carbonate buildups of unit JM2 is clearly shown on Sections 7.3 and 7.5, located on the southern flank of the Bawean Arch. On the southeastern flank of the Karimunjawa Arch, the tops of subunits JM1 and JM2 are characterised by a rough, undulating surface (Sect. 7.1), which is interpreted as karst topography formed during the subsequent sea level fall of each sequence. The lateral extent of this morphology is limited, and is not present in the deeper setting. This suggests that the limestone formed only during sea level highstands, which implies that the underlying siliciclastics are lowstand deposits.

Some seismic lines were run close to Bawean Island and suggest that unit JM (JM1) is probably equivalent to the Miocene strata exposed on this island. A geological study by van Bemmelen (1949) reported the occurrence of limonitic sandstone, interbedded with lignite, marl and crystalline limestone on Bawean Island. He indicated an age of Early to Middle Miocene based on the occurrence of larger benthonic foraminifers *Lepidocyclina, Cycloclypeus* and *Alveolina bontangensis* which is in agreement with the evidence from JS8-1 and JS3-1 wells (Appendix E). In the Rembang Zone, three major Middle to Late Miocene seismic units, EM2, LM1 and LM2, correspond to the Tawun, Wonocolo and Ledok Formations respectively (Chapter 4.3). The Middle Miocene Tawun Formation is a paralic marine facies, and shows a facies similarity with those developed in Bawean Island. The Late Miocene Wonocolo and Ledok Formations are shelfal facies, and are characterised by hemipelagic marl and tidally-influenced glauconitic sandstone respectively. The absence of the Late Miocene deposits (Wonocolo and Ledok Formations) in the topographically high areas, such as Bawean Island, suggests that the deposits were basinally restricted. Unfortunately the limited penetration of the seismic system used does not permit recognition of this Late Miocene deposit.

### 7.4.2 Unit JP

Unit JP is relatively thick and was deposited following unit JM2. It consists of two type-I sequences, subunits JP1 and JP2, with subunit JP1 forming the major sequence. Correlation between the seismic study and the micropalaeontological data from some petroleum exploratory wells indicates that these subunits developed during the Pliocene. This age was determined particularly by the appearance of the species *Sphaeroidinella dehiscens*. The first appearance of this species (5.05 Ma; Berggren et al., 1985) has been used to indicate the boundary of
planktonic foraminiferal zones N18/N19 (Banner & Blow, 1967; Blow, 1969). The top boundary of unit JP is an erosional surface marking extensive subaerial exposure in the study area at the end of the Pliocene. Subunits JP1 and JP2 may be classified as third order sequences as their duration was no more than 1 Ma (Vail et al., 1977). The sediment sources of these subunits were mainly the Karimunjawa and Bawean Arches in the western half of the study area (Sects 7.1 & 7.3). In the eastern half, the deposits were sourced from both the Bawean Arch and Madura Island, but the distribution was complicated by the development of folds (Sect. 7.8).

a. Subunit JP1

The deposition of this subunit is well described by the theoretical model of shelf margin sequence stratigraphy by Posamentier et al. (1988). Two main systems tracts can be defined, the lowstand and highstand systems tracts. The transgressive systems tract on most of the seismic lines studied is absent or unidentified, probably due to a combination of low depositional slope, low sedimentation rate and rapid sea level rise which did not permit formation of a seismically resolvable transgressive unit. The ideal cross-sectional view for the western half of the area is shown on Section 7.3. Figure 7.6 displays the time structure map for the top of the Miocene strata on which unit JP1 was deposited, and Figure 7.7 displays the isochron map of total unit JP (subunits JP1 and JP2). These maps indicate that the Pliocene basin in the western half of the study area was still influenced by the normal fault movement of the half graben system in the West Florence Deep. The occurrence of the deepest basin and accumulation of the thickest Pliocene sediments in this trough (particularly along the normal faults) has further suggested probable faster subsidence and sedimentation rates. In the eastern half of the area, the influence of the previous structural configuration (Figs 7.3 & 7.4c) is not really obvious. The Pliocene structural development (east-west trending folds) had more influence on the sedimentation, as indicated by the trends of basin morphology and the Pliocene sediment accumulation (Figs 7.6 & 7.7).

Lowstand systems tract

A sea level fall occurred after the deposition of unit JM2. This exposed most of the positive relief areas in the study area. Karstification occurred on the Karimunjawa Arch and possibly on the Bawean Arch (Sects 7.1 & 7.5). The estimated minimum relative sea level fall was about 130 msec TWT (about 100 m), measured from the profile between the highstand reefal facies of unit JM2 and the nick on Section 7.3. This nick point is regarded as the
position of sea level during the lowstand, based on the seismic features which show an erosional surface on the shelf above it. Consequently, the position of the 'shelf' edge before erosion should be put some distance higher than this point. Figure 7.6 displays the palaeogeographic map during the sea level lowstand, plotted on the time structure contours to top of the Miocene strata.

Pronounced valley and fluvial incisions occurred on the Bawean Arch (Sects 7.3 & 7.7). Most of these incisions are traceable on seismic sections and ended basinward in the lowstand systems tract. The channel and valley incisions occur during the interval of sea level fall, according to the sequence stratigraphic concept (Posamentier & Vail, 1988). During this active incision, minimum stream deposition occurs and sediments are fed directly to the stream mouths near the shelf edge and are ultimately deposited as turbidites where abrupt decreases in slope gradient occur (Posamentier et al., 1991). The lowstand fans (basin floor fans) are not observed in the study area due to the limited penetration of the seismic data. But some interpreted submarine canyons which may be associated with the feeder channels are observed beneath the lowstand wedge. The canyon on Section 7.3 in fact can be traced landward, and is associated with the subaerial valley incision shown on Section 7.7.

As sea level reached the lowest point, a seaward prograding lowstand wedge developed (Sect. 7.3). The bottom set on most seismic lines is not observed due to weak seismic signals. The topset is nearly horizontal (about 0.5° dip), but the parallelism and concordance of internal strata that commonly characterise a low depositional angle (Mitchum et al., 1977) are not well displayed, and the pattern tends to show an oblique progradation. It is suspected that the sediment supply was relatively high during the lowstand. Wedges of lowstand deposits are also observed on some sections (Sects 7.6 & 7.8), and are suspected to be at a similar stratigraphic position as the above prograding deposits. On Section 7.6, at about the same level as the lowstand prograding clinoforms on Section 7.3, some mounded forms are displayed. The flank of the lower lobe is rather steep (about 1°) compared with the overlying lobes (about 0.25°).

The origin of such mounded and prograding clinoform structures has been discussed by several authors (e.g. Mitchum, 1985; Posamentier & Erskine, 1991), and could be associated with submarine fan lobes. These mounded forms according to Berg (1982) can also be associated with a fluvial-dominated delta system. But the mounded forms on Section 7.5 show
well-developed channel levees and lack evidence of surface marine processes which are commonly shown by a flat surface (of the delta plain). Therefore, it is interpreted that these fan systems are parts of the lowstand wedge. Posamentier et al. (1991) indicated that a channel-levee complex (early lowstand wedge) may be formed prior to the development of the prograding complex (late lowstand wedge). The formation of the lowstand wedge is contemporaneous with the deposition of the channel and valley fills (Sects 7.3 & 7.7; Posamentier & Vail, 1988; Posamentier et al., 1991).

Although the timing and related seismic features have strongly suggested the occurrence of a shallow submarine fan system, modern analogues are lacking. Walker (1992) indicated that although turbidity currents can occur at any depth, turbidites are only preserved below storm wave base at a minimum depth of 250-300 m, which is not the case in the study area. The Late Pleistocene deep marine fan systems developed in the Gulf of Mexico (Weimer, 1990; 1991) and off the Amazon delta system (Flood et al., 1991) are good examples. In these areas there is also sufficient slope length to allow turbidity currents to develop. Such slope length and depth are absent in the study area, and thus the above interpretation has to be tested further through detailed mapping combined with well coring.

**Highstand systems tract**

A rapid rise of relative sea level occurred after the deposition of the lowstand systems tract. The estimated relative sea level rise is difficult to define due to the post-Pliocene erosion, but the profile of the highstand clinoform on Section 7.3 suggests that it may have reached 150 msec TWT (about 115 m). Since there is no transgressive systems tract developed (within seismic resolution), the highstand systems tract appears to be directly deposited on the lowstand systems tract. A very thin condensed section may occur between them, which may also coincide with the first flooding surface. Most of the Pliocene deposits in the study area in fact consist of highstand deposits.

On the Karimunjawa Arch, obvious highstand prograding clinoforms are not displayed, probably due to very low depositional dip (about 0.1°). Here, the early highstand systems tract is characterised by a medium to high amplitude and high continuity parallel reflection pattern (Sect. 7.1). Sangree and Widmier (1977) suggested this type of pattern represents interbedded strata deposited during high and low energy conditions, and commonly associated with shallow marine clastic units deposited mainly by wave transport processes. A mounded form, which
interfingers with surrounding marine clastics, is well displayed on Section 7.1 and interpreted as a highstand bioherm [shelf carbonate according to Tucker & Wright (1990)].

On the Bawean Arch, the early highstand deposit is characterised by relatively steep oblique progradational clinoforms, with the depositional dip reaching 0.75° (Sect. 7.3). It is also associated with some probable channel structures (Sect. 7.5). These features are restricted in extent to the inner shelf of the Bawean Arch, and are interpreted to indicate a highstand prograding delta complex. Chronis et al. (1991) documented that Late Quaternary deltaic progradation in the Gulf of Patras, Greece, was greatest during the slowdown of sea level change and the period of stable sea level. The fact that the delta system on the Bawean Arch occurred following a sea level rise suggests a similar mode of delta progradation as in the Gulf of Patras.

On and to the north of Madura Island, some east-trending anticlines developed during the Pliocene. During the sea level highstand, these structural highs were flooded allowing development of highstand carbonate facies, which shows a pronounced mounded form on Section 7.8. On the north flank of Madura Island, the highstand limestone is characterised by a medium continuity and medium amplitude parallel reflection pattern (Sect. 7.10). A landward projection to Madura Island suggests that it is equivalent to the locally known Madura Formation reported by Situmorang et al. (1992) and Aziz et al. (1993). This formation is equivalent to the Paciran Formation in the Rembang Zone, East Java, which is the limestone facies of the Mundu Formation (see Chapter 3.2.1). On Madura Island, the Madura Formation consists of up to 250 m of interbedded reefal and sandy limestone and marl (Situmorang et al., 1992; Aziz et al., 1993). Situmorang et al. (1992) reported the occurrence of benthonic foraminifers *Operculina* sp., *Amphistegina* sp. and *Gypsina* sp. in this formation, which indicate an inner shelf environment (Murray, 1973; 1991).

The late highstand tract was well developed in the western part of the study area. It formed in response to a relatively slow sea level fall, and the basin overall experienced a progressive shoaling. The late highstand systems tract is characterised by prograding clinoforms in which the clinoform angles gradually decrease to nearly horizontal at the end of the sequence. The clinoforms progressively downlap on the shelf, lowstand wedge and possibly basin-floor fan (?). These features are shown on Section 7.3 and in the smaller window of Section 7.2 where the pattern appears to be divergent. The change from the early
The highstand systems tract appears to be gradual with the early highstand delta being slowly abandoned.

During deposition of the slope and basin facies of the highstand tract, water depth may have reached 200 m (estimated from the clinoform profile), resulting in the accumulation of planktonic foraminifer-rich deposits. Such deposits occur in some of the petroleum exploratory wells drilled in the study area, e.g. JS10-1 and JS8-1. The mudstone facies in the lower Pliocene strata in both wells (between 280 m depth and the top of the Miocene) represent the basin and slope facies whereas the overlying sandy facies represents the shelf facies. Most of the planktonic foraminifers in fact occur in the mudstone facies.

b. Subunit JP2

Subunit JP2 appears to be the result of a smaller relative sea level fluctuation than the previous one. The area affected by this sea level oscillation is limited and appears to be concentrated in the deep portion (below contour 210 msec TWT) of the basin as indicated on the time structure map for the top of the Pliocene (Fig. 7.8). During the deposition of this unit a large part of the study area remained exposed. In the western part of the study area, subunit JP2 is recognised as a thin prograding complex occurring on the erosional surface on top of subunit JP1 (Sects. 7.2 & 7.3). This erosional surface should be correlated with a lowstand of sea level and the prograding complex with the highstand deposits.

A thick deposit of up to 80 msec TWT (about 60 m) occurs locally in the deep area north of Madura Island. Section 7.8 show the occurrence of a lensoidal lowstand unit, which is interpreted as a lowstand deposit. This deposit displays some diffraction features which at least indicate a rough surface. It is suspected that the deposit is a coarse clastic which has been derived from the adjacent highstand deposits (unit JP1, on Madura Island and on the anticline north of it).

A subsequent relative sea level rise led to the deposition of a highstand systems tract which is characterised by prograding clinoforms, onlapping onto the basin margins. This deposit has led to the shoaling of the basin prior to subaerial exposure of the study area at the end of the Pliocene.
7.4.3 Unit JQ

Unit JQ is a Quaternary sequence, deposited following the sea level fall which exposed the whole study area at the end of the Pliocene. Figures 7.4a, 7.4b and 7.4c are mesh diagrams of the present bathymetry and the present top of Pliocene and Miocene, respectively. Roughly, these figures suggest that the basin morphology where unit JQ was deposited appeared to be relatively flat compared with the Pliocene basin. Some local subsidence occurred, in the western part of the area, probably related to the normal fault movement bordering the West Florence Deep and Bawean Arch. On the eastern part, the local subsidence was related to the east trending folding which had been occurring since probably the Late Miocene. On the isochron map of unit JQ, shown in Figure 7.9, the trend of the deposition follows the basin topography and thick sediment accumulation occurred in locally deep areas.

The seismic characters and sedimentation patterns of unit JQ differ significantly from the preceding unit JP. They appear to be strongly influenced by extreme and rapid sea level fluctuations. Such fluctuation during the Quaternary has been demonstrated by Emiliani (1955, 1966) and Shackleton and Opdyke (1973, 1976 & 1977) through the oxygen isotope records of deep sea cores, which they related with the orbitally-induced fluctuations of global ice volume. These glacio-eustatic sea level fluctuations are particularly apparent since the Jaramillo magnetic event (0.97 Ma; Harland et al., 1989) with a relatively constant period. Donovan and Jones (1979) have calculated that during the period of maximum Pleistocene glacial advance the sea level was lowered by about 100 m. These studies are consistent with the results of Bloom et al. (1974) in their work on marine terraces in the Huon Peninsula, New Guinea, and the findings of Mesolella et al. (1969) on marine terraces in Barbados. Chappell and Shackleton (1986) indicated that in the last glacial maximum some 20 ka ago, sea level was up to 130 m lower than the present level. More recently, Fairbanks (1989) used cores drilled from offshore Barbados coral reefs to indicate that the sea level during the last glacial maximum was about 121 m below present level. He also found that deglaciation toward the present level was not monotonic, as indicated by the occurrence of a termination during 11-10 ka (Younger Dryas event).

Present water depth in the study area does not exceed 85 m and the bathymetry is relatively flat. With respect to basin subsidence, such conditions may have persisted in most of the study area since the end of the Pliocene, and subjected the area to two extremes in sea level during the Quaternary. Sea level fluctuations are indicated by the occurrence of a
repetitive sequence beginning with marine deposits and ending with extensive cut and fill structures, representing the interglacial and glacial periods respectively. In terms of duration, these sequences may be regarded as fourth order sequences.

Similar features have been reported by some authors in other shelf areas and have been related to Quaternary glacio-eustatic sea level changes. Piper and Perissoratis (1991) have demonstrated that the Late Quaternary sea level history inferred from seismic stratigraphy in the North Aegean continental margin (Greece) correlates with the global eustatic sea level record based on oxygen isotopic curves. In the southwest Louisiana continental shelf, Suter et al. (1987) recognised six depositional sequences within the Late Quaternary deposits, five of which are believed to relate to glacio-eustatic fluctuations.

The present seismic study identified nine seismic subunits in the study area, excluding Holocene deposits. These subunits are characterised mostly by parallel to subparallel reflection patterns or are reflection free. Each of them ended with channel cut and fill along the upper part (Sects 7.2 & 7.3) and is interpreted to represent marine deposition and fluvial channelling respectively. In some areas the thickness of these subunits appears to be similar (Sect. 7.2) which may indicate constant subsidence rates and periods of sea level fluctuation. Because these subunits have a similar seismic character and do not represent thick deposits, lateral correlation is difficult and tends to be speculative. But they can be grouped into five subunits, JQ1 to JQ5, based on the occurrence of widespread unconformities on top of each group. These unconformities are commonly associated with rather deep and wide fluvial channelling.

a. Subunits JQ1 and JQ2

These subunits were deposited during the Early Pleistocene, based on their stratigraphic position overlying Pliocene unit JP. A stratigraphic subdivision between these subunits in the western part of the study area is rather speculative, but clear differentiation can be made in the eastern part due to the occurrence of a relatively large sea level fall at the end of subunit JQ1 deposition (Sects 7.4 & 7.8).

Seismic features, such as rapid basinward thinning (Sect. 7.8) and pronounced anticlines (Sects 7.2, 7.3 & 7.4), and the isochron map (Fig. 7.10) indicate that their distribution was still influenced by local development of folds. In the western part, the subunits asymmetrically
thicken southward, while on the north of Madura Island local thickening occurred in the synclinal areas between this island and Bawean Island.

Subunit JQ1 can be further subdivided into subunits JQ1a and JQ1b based on the occurrence of an internal erosional surface (Sects 7.2, 7.3 & 7.8). In the western part, these subunits lie subhorizontally and onlap on the Pliocene unit JP. The thickness of subunit JQ1a reaches 100 msec TWT (about 75 m) in the deepest portion of the basin, and gradually thins toward the basin margin. The maximum thickness of subunit JQ1b is about 60 msec TWT (about 45 m). The thickness variation is mainly due to local subsidence, post depositional erosion and a gradual thinning because of the rising of the basin margin. Subunit JQ1b onlaps on subunit JQ1a on the southern margin, and on subunit JP when subunit JQ1a wedges out (Sect. 7.3). The seismic character of these subunits is similar, a subparallel reflection pattern with medium amplitude and medium continuity which suggests deposition in a shallow marine environment (Sangree & Widmier, 1977). Subunits JQ1a and JQ1b in the western part may be regarded as the units responsible for the flatness of this area. The later subunit, JQ2, was deposited on a flat surface and has an extensive coverage although its thickness is less than 35 msec TWT (26 m). Seismically, this subunit is characterised by a similar appearance to subunits JQ1a and JQ1b, and probably was deposited in a similar environment.

To the north of Madura Island a local deepening occurred (Fig. 7.10 & Sect. 7.8), and subunits JQ1a and JQ1b are characterised by northward prograding clinoform deposits, indicating the sediments were derived from Madura Island (Sect. 7.8). The clinoform profile provides a very rough water depth estimate for the basin centre which is about 135 msec TWT (100 m). Subunit JQ2, in this area, is characterised by an upward loss in acoustic transparency, almost reflection free at the lower part and gradually becoming subparallel to hummocky patterns with variable amplitude and low to high continuity at the upper part. This indicates a shoaling (regression) of the unit before finally being exposed subaerially.

b. Subunits JQ3 and JQ4

Subunits JQ3 and JQ4 can further be subdivided into three (JQ3a, JQ3b and JQ3c) and two (JQ4a and JQ4b) respectively. These subdivisions can only be carried out in a limited area where the subsidence and sedimentation rates were relatively high, as indicated by the thick deposit on the combined isochron map (Figure 7.12).
Subunits JQ3a and JQ3b are similar in seismic character, showing a medium amplitude, subparallel reflection pattern which probably represents a shallow marine environment. Subunit JQ3c is reflection free, indicating most probably homogeneous mudstone. In the eastern part, these subunits are rather subtly represented, but southward prograding clinoforms on Section 7.4 and complex channelling on Section 7.8 may be regarded as representative of these subunits.

Subunit JQ4 is extensively distributed and characterised by an almost reflection free character suggesting a nearly homogeneous deposit probably of mudstone. In the western part, subunits JQ4a and JQ4b are very thin to absent which may indicate a low depositional rate (Sects 7.2 & 7.3). A relatively thick deposit of up to 45 msec (about 34 m) occurs in the eastern part. Here, a further subdivision of the subunit is difficult. But parasequences, showing northward prograding clinoforms, might be identified with subunits JQ4a and JQ4b (Sects 7.4 & 7.8). Each of the parasequence appears to be associated with a channel notch which shifted as the parasequence prograded.

c. Subunit JQ5

Subunit JQ5 consists of a single reflection-free sequence of possibly homogeneous mudstone. The isochron map is shown in Figure 7.14. The maximum thickness is about 30 msec TWT (about 22 m) with a little variation on the western part of the map. On some parts to the north of Madura Island this subunit is too thin to be identified, but locally thick deposits of up to 25 msec TWT (about 19 m) occur in a limited area (Fig. 7.14). The seismic sections from the area near the coasts of East Java and Madura Island suggest that subunit JQ5 thickens landward (Sects 7.9 & 7.10), and the parasequence shows a northward prograding clinoform. A relatively thick deposit of up to 50 msec TWT (about 37 m) occurs at the mouth of a narrow strait between East Java and Madura Island. It is suspected that a delta system was present at this location and this strait served as a channel which linked the Java Sea with the sediment source area (East Java).

The base of subunit JQ5 is characterised by extensive fluvial channelling. Figure 7.13 shows the position of channels with one major channel up to 3 km wide and 20 msec TWT deep (about 15 m) occurring in the northern part of the area. The occurrence of this channelling can be related to the curve of the oxygen isotope record of deep sea cores (Fig. 7.15) and sea level position of the Huon Peninsula (Fig. 6.31). The base of subunit JQ5
should be correlated with the sea level minimum during oxygen isotope stage 6. This age is consistent with the occurrence of similar features found in Madura Strait occurring on top of subunit D1a at a depth of 140 msec TWT (about 105 m).

Subunit JQ5 appears to be the uppermost sequence on most of the seismic sections. Its upper boundary is rather subtle due to absence of non-marine reworking processes during the last glacial maximum which ended the deposition of unit JQ5, although the glaciation may have lowered the sea level to -130 m (Chappell & Shackleton, 1986). Some notches, however, appear on some seismic sections and are interpreted to represent small scale channels during the glacial maximum, but their lateral extent is difficult to trace. On a regional scale, however, Kuenen (1950) recognised drowned stream courses within the Java Sea, based on echo sounder data (Fig. 2.7). Two river systems with dendritic patterns were present, originating from a topographically high area southwest of Kalimantan, flowing northward to the China Sea and eastward to the Wallace Deep. With regard to the present bathymetry of the Java Sea, these courses should be the most recent and should be regarded as the result of the last glacial sea level lowering.

d. Holocene deposits

The Holocene deposits include sediments deposited after the last glacial maximum (oxygen isotope stage 2). On the shelf (Java Sea) these sediments are relatively thin (< 1 m), and in general are not well represented on the seismic sections. They are recognised mainly from the results of shallow sediment coring (Dewi, 1993). The thickness varies locally, and apparently depends on the geographical position and the palaeomorphology of previous lowstand surfaces. In the area between East Java and Bawean Island, the thickness is commonly less than 1 m, and the sediment is characterised mostly by homogeneous greenish grey silty-sand to sandy-silt. Near the base the sediment is commonly oolitic and contains foraminifers Ammonia beccarii and Operculina sp. (Dewi, 1993) which are typical of a lagoonal environment (Murray, 1973; 1991). Thus, this lower part of the sediment is thought to be due to the early stage of flooding in the study area after the last glacial maximum.

A rather thick Holocene deposit occurs near the mouth of the Solo River and can be regarded as a part of the Solo River delta deposit (locality shown in Fig. 7.2). Section 7.9 is the seismic section (boomer system) taken close to the Solo River Delta system. Here, the Holocene deposits may reach 5 m, but show a rapid thinning seaward. A detailed study of
Holocene sedimentation in the Solo River Delta has been carried out by Hoekstra (1993). His study indicates that the growth pattern of the delta exhibits very little lateral accretion, and the maximum rate of deposition occurs in front of the main channel which resulted in a prolongation of the delta. He also found that most of the Solo River sediments, which are mostly mud-dominated, are deposited in the deltaic environment and only small portion escape to the sea. Hoekstra (1993) postulated that if this condition is similar to all deltaic environments on the northern coast of Java, most of the recent sediments in the Java Sea are not fluvially derived from Java. This may explain the lack of Holocene vertical accretion in the Java Sea, because most of the sediments would be derived from previous deposits or marine organisms.

7.5 DISCUSSION

The Miocene basin configuration of the southeastern Java Sea is poorly known, and it is suspected that the basin was still strongly influenced by the northeast-trending structures. These structures are half grabens and have been the major control for the Early Tertiary sedimentation. On the northern part of East Java (Rembang Zone) these structures were still the dominant factor in controlling the Middle and Late Miocene sedimentation (Chapter 4). Although some elements of these structures were still active until the Pleistocene, their effectiveness in controlling the sedimentation during the post-Miocene was diminished. The Pliocene sedimentation, in general, occurred in east-trending synclinal basins which indicate the dominance of a northward tectonic compressional stress. This continued until the Early Pleistocene, as is indicated by some local thickening of the Early Pleistocene deposits. Since then, further basin development appears to have ceased, and a tectonically stable condition may have been reached.

Posamentier and Vail (1988) pointed out that depositional stratal patterns and distribution of lithofacies are controlled primarily by relative sea level change. The above discussion indicates that basin evolution in the study area, particularly since the Pliocene, is relatively well understood. The regional basin subsidence occurred at a relatively slow rate. This has led to the belief that eustatic sea level change was the only major factor in controlling the stratal pattern and lithofacies distribution in the study area.
7.6 SUMMARY

The seismic stratigraphic study revealed that the Late Tertiary and Quaternary sediments in the southeast Java Sea can be subdivided into three major depositional sequences: units JM, JP and JQ. These units, which represent the Miocene, Pliocene and Quaternary deposits respectively, are typified by different areal distributions, thicknesses, and seismic character.

Unit JM is poorly known due to the limited penetration of the system used. However, seismic sections from structurally high areas indicate the presence of two subunits, JM1 and JM2, which are characterised by subparallel reflection patterns of possibly paralic deposits in the lower part, and mounded and undulating forms of possibly reefal/karstified limestones in the upper part. The age and lithofacies of this unit resemble the Tawun (paralic deposit) and Bulu (limestone) Formations developed in the Rembang Zone.

Unit JP attains 220 msec TWT thick (about 165 m) and can be subdivided into subunits JP1 and JP2, with subunit JP1 being the most extensively developed. In the area where the clastic deposition was dominant, such as on the south flank of the Bawean Arch, subunit JP1 has developed systems tracts that resemble a shelf margin sequence, with sediments derived mostly from this arch. Highstand carbonates of this unit were developed on some anticlinal highs in the eastern part of the area and in the shelfal facies south of the Karimunjawa Arch. They may represent a low rate of clastic deposition. Subunit JP2 was developed in a limited area, particularly in the deeper part of the basin, which indicates that the sea level fluctuation was much lower than during the previous sequence. Based on this difference, a tentative correlation with the Haq et al. (1988) eustatic sea level curve suggests an age of 3.8 Ma for the boundary between subunits JP1 and JP2.

Unit JQ consists of nine small subunits, excluding the Holocene subunit, which represent at least nine Quaternary sea level fluctuations. Each subunit is relatively thin, and tends to be distributed widely because of deposition on a relatively flat lying area. The seismic characters are very similar, subparallel reflection or almost reflection free patterns representing marine deposits, topped by extensive fluvial channelling which is obvious in the western part of the study area. This repetitive succession is thought to represent highstand and lowstand of sea level respectively.
The nine subunits in unit JQ can be grouped into five subunits, JQ1 to JQ5, based on the occurrence of relatively larger sea level falls as indicated by deeper and wider fluvial incisions. A tentative correlation with the oxygen isotope records of deep sea cores (Fig. 7.15) can be made with an assumption that a fluvial incision reflects a significant relative sea level fall which exposed most of the Java Sea. Because the maximum water depth at present is about 85 m, the magnitude of sea level fall should be larger than this value. The oxygen isotope records (Harland et al., 1989; Fig. 7.15) suggests that relatively large sea level falls occurred after the Jaramillo Event, and correspond with oxygen isotope stages 22, 16, 12, 10, 6 and 2. Subunit JQ5, which underlies the Holocene deposits, is bounded at the top and base by fluvial erosional surfaces, which are interpreted to correspond with the oxygen isotope stages 2 and 6 respectively (Late Pleistocene deposition). Subunit JQ4 consists of two smaller subunits which are separated by a fluvial erosional surface. Subunit JQ4 is inferred to have been deposited during the interval between stages 10 and 6 (late Middle Pleistocene). The subunit JQ3, which consists of three smaller subunits, is interpreted to have been deposited during the interval between stages 16 and 10 (middle Middle Pleistocene). The boundaries between subunits JQ2 and JQ1, and subunits JQ1 and unit JP are difficult to infer due to the absence on the seismic sections of obvious features related to sea level fall.
8.1 INTRODUCTION

This study has involved a re-evaluation of the geological history of four aspects of the East Java Basin including: (1) a reassessment of the onshore sedimentological history of the Rembang and Kendeng Zones; (2) a reinterpretation of the Quaternary successions in the eastern Kendeng Zone; (3) a seismic stratigraphic interpretation of the Quaternary succession in Madura Strait; and (4) a seismic stratigraphic interpretation of the Late Tertiary and Quaternary successions in the southern Java Sea. The following discussion summarises briefly the major findings of this study.

8.2 BASIN DEVELOPMENT

During the pre-Tertiary to Early Tertiary, a series of arc and subduction complexes were systematically developed in the western Indonesian region (Fig. 8.1). The sutures between these complexes apparently have remained as lines of weakness which provide flexibility within the basement of the East Java Basin. These zones of pervasive weakness have influenced the development of the East Java Basin in the Late Tertiary and Quaternary. Differential basement block faulting and associated uplift led to the development of a fragmented basin with numerous small sub-basins, each characterised by contrasting sedimentation histories. Differential block uplifts were probably caused by the oblique orientation (northeast-trending) of the pre-Middle Tertiary sutures relative to the present subduction zone (east-trending), which produced wrench faulting accompanied by vertical movement (Situmorang et al., 1976). Half grabens related to this faulting influenced the pattern of sedimentation during the Middle Miocene (Tawun Formation) in the Rembang Zone (Chapter 4) and Pliocene (and probably earlier) sedimentation in the southern Java Sea (Chapter 6).

Since the Late Miocene, east-trending anticlinal zones developed and were superimposed on the previous northeast-trending structures. The anticlinal Rembang Zone, which extends from the Blora area to Madura Island, became a dominant structure which controlled
sedimentation during the Plio-Pleistocene and divided the East Java Basin into two major basinal areas (Fig. 8.2). In the north, a basinal area formed between the Karimunjawa and Bawean Arches and the Rembang Zone. In the south, a basinal area developed between the Rembang Zone and the chain of volcanism along the median line of Java. The east-trending orientation of these basins is parallel to the Neogene-Quaternary subduction zone in the south of Java and suggests the dominance of a northward tectonic compressional stress. The smaller sedimentary thickness and the wider dimension of the north basinal area suggest that the compressional stress here is less than in the south basinal area.

The peak development of these basins apparently occurred in the Late Miocene and Pliocene. Sedimentation patterns, such as local thickening in the north basinal area (southern Java Sea), suggest that basin development ceased in the Early/Middle Pleistocene. Since then, the basin has become flatter due to infilling, promoting thin and widespread deposition of Middle and Late Pleistocene sediments. In contrast, rapid basin subsidence and sedimentation occurred in the south basinal area at least until the late Middle Pleistocene, when extensive east-trending folding, accompanied by diapiric movement, gave rise to the anticlinal Kendeng Zone.

8.3 SEDIMENTATION HISTORY
8.3.1 The onshore area during the Middle Miocene and Pliocene

A re-assessment of the Middle Miocene to Pliocene sedimentary history of the Rembang Zone was carried out using seismic stratigraphic interpretation, while outcrop data were used for the Kendeng Zone. The three main stratigraphic markers used to achieve a reliable lateral correlation between these zones are: (1) the determined sequence boundaries of the Rembang Zone strata, (2) the results of biostratigraphic analysis of previous workers (e.g. Pringgoprawiro, 1983; van Gorsel & Troelstra, 1981), and (3) the results of the climatostratigraphic study of van Gorsel and Troelstra (1981) on the Late Miocene and Pliocene strata in the Kendeng Zone. These were backed up by sedimentological data such as vertical facies sequences.

Middle Miocene strata (seismic unit MM2) in the Rembang Zone, as recognised from seismic sections (Chapter 4), were deposited in two small basinal areas (Fig. 4.10). These basins are half grabens that resulted from basement fragmentation. Stratal patterns revealed that the direction of sediment transport was mainly from the north in the southern part of the
East Bawean Trough, and from both the south and north in the southern part of the Central Deep (Fig. 4.10). These patterns of sedimentation were not recognised by earlier investigators (e.g. Lemigas/BEICIP, 1969; Ardana, 1993) who assumed that most sediments were derived from the north (Bawean Arch).

Late Miocene sediments (Wonocolo and Ledok Formations) in the Rembang Zone were mostly deposited in a shelfal area during sea level highstands. The seismic unit LM1, which corresponds with the Wonocolo Formation, is mostly hemipelagic foraminiferal-marl, but deltaic deposition probably occurred near the Tobo high area (Fig. 4.10). The Ledok Formation (seismic unit LM2) was deposited on a shallow shelf as a tidally-influenced deposit.

A geological reinterpretation of the Kendeng Zone was carried out based on examination of outcrops of the Late Miocene and Pliocene strata. Here, sedimentation is the result of volcanic activity along the southern Kendeng Zone with marine deposition occurring to the north of the volcanic complex. The Kerek Formation, a turbiditic deposit, was formed in this setting. The influence of volcanism diminished by the late Late Miocene, and the resultant sediments are dominated by pelagic marls mapped as the Kalibeng Formation.

The deposition of pelagic marl became widespread in the Pliocene, and apparently followed a global eustatic sea level rise recognised by Haq et al. (1988). In the Rembang and Kendeng Zones, the marl is represented by the Mundu (seismic unit PL1) and Kalibeng Formations, respectively. The Atasangin Formation, which was deposited on top of the Kalibeng Formation, appears to result from a minor relative sea level fall in the Middle/Late Pliocene.

A relative sea level fall occurred in the Late Pliocene and led to the development of a stagnant lacustrine basin in a broad synclinal area south of the Rembang Zone (Fig. 8.2). This resulted in the widespread deposition of organic-rich mudstone mapped as the Lidah Formation.

8.3.2 The onshore area (eastern Kendeng Zone) and Madura Strait during the Quaternary

A reinterpretation of the onshore Quaternary succession has been carried out, particularly in the eastern Kendeng Zone, to facilitate correlation with the offshore (Madura Strait)
Quaternary succession interpreted from seismic data. The eastern Kendeng Zone - Madura Strait area is characterised by syndepositional folding which formed a small east-trending basin, located between the Rembang Zone (and Madura Island) in the north and Quaternary volcanoes in the south. The basin is characterised by rapid subsidence and sedimentation, which resulted in more than 200 m of Quaternary deposits.

In the eastern Kendeng Zone - Madura Strait area, deposition resulted from a complex interplay between Quaternary volcanism along the south of the Kendeng Zone, marine sediments derived from the exposed Rembang Zone in the north, and Quaternary relative sea level changes. Field observations indicate that the volcanioclastic deposits are coarse-grained and, in the seismic section from Madura Strait, they are associated with relatively steep depositional slopes (of alluvial fan/fan delta systems). In contrast, the northerly-derived deposits are associated with marine fossiliferous mudstone containing a small number of quartz grains. The present detailed seismic stratigraphic study of Madura Strait indicates that the timing of deposition of these Quaternary deposits was strongly controlled by relative sea level changes. During a major relative sea level fall, the eastern Kendeng Zone and Madura Strait were subaerially exposed, as indicated by thin widespread fluvial and alluvial fan deposits.

One of the major considerations of previous Quaternary geological investigations in the eastern Kendeng Zone was the dating of hominid bearing strata (Jacob & Curtis, 1971; Hyodo et al., 1992). The fact that vertebrate and hominid fossils occur associated with fluvial/alluvial fan deposits (as reported by Duyfjes, 1938a,b,c,d) suggests that Quaternary sea level fluctuations were an important factor in determining the habitat of human progenitors. This emphasises the importance of delineating the timing of sea level fluctuation in this area. The present study used the published Quaternary oxygen isotope curve of deep sea cores of Harland et al. (1989) to infer the depositional timing of these fluvial deposits. Two of the three main Quaternary fluvial horizons in Madura Strait have been correlated with two major eustatic sea falls during the oxygen isotope stages 16 (seismic unit C1a) and 6 (seismic unit D1a) respectively.

8.3.3 Pliocene and Quaternary sedimentation in the southern Java Sea

The Pliocene to Quaternary history of sedimentation in the southern part of the Java Sea has been assessed through seismic stratigraphic interpretation. The limited penetration of the seismic system used in this study did not permit a detailed evaluation of the Miocene
sedimentation in this area. Stratal patterns, as observed on structurally high areas, such as the Karimunjawa and Bawean Arches, indicate the occurrence of a parallel stratal pattern of probably shallow-marine deposits associated with carbonate buildups of mounded form. Outcrop data from Bawean Island suggest that the Miocene deposits are mainly paralic facies.

Basin development and sedimentation patterns, particularly during the Pliocene and Quaternary, are well understood. Here, a broad asymmetrical synclinal basin developed probably in the Late Miocene to Pliocene between the relatively stable Sunda Shelf in the north and the anticlinal Rembang Zone in the south. During the Pliocene, sediments were primarily derived from the stable Sunda Shelf (Karimunjawa and Bawean Arches) and the Rembang Zone (Madura Island). The stratal patterns are mainly controlled by relative sea level changes. For example, stratal patterns in the major Pliocene deposit (unit JP1) suggest that deposition occurred during a slow relative sea level fall.

Pleistocene sedimentation was highly influenced by frequent sea level fluctuations. In the southern Java Sea (north basinal area; Fig. 8.2), basin filling during the Pliocene had produced a flat basin topography by the end of the Pliocene. Nine thin seismic subunits are recognised in this area and are believed to represent at least nine Quaternary sea level fluctuations. Each subunit is relatively thin, and tends to be distributed widely because of deposition on a relatively flat-lying area. The seismic characters of these subunits are very similar; subparallel reflection or almost reflection free patterns representing marine deposits, topped by extensive fluvial channelling. This repetitive succession is thought to represent highstands and lowstands of sea level, respectively, which can be related to published sea level/oxygen isotope curves (e.g. Harland et al., 1989).

8.4 NOTES ON THE APPLICATION OF SEQUENCE STRATIGRAPHY IN THE EAST JAVA BASIN

Sequence stratigraphy was primarily developed for basins that undergo slow and steady differential subsidence which increases toward the basin centre. Such characteristics are commonly displayed by shelf margin settings and divergent continental margin settings (Jervey, 1988). The widespread application of sequence stratigraphic concepts is common in these settings (e.g. Suter et al., 1987; Piper & Perissoratis, 1991).
The present study of the East Java Basin indicates that at least three types of sub-basins occur in this area. They include (1) sub-basins showing the character of a half graben setting; (2) sub-basins showing the character of a shelf margin setting; and (3) sub-basins showing the character of a shelf setting. Sequence stratigraphic concepts have been applied to these basin settings, but some remarks need to be raised related to the applicability of the concepts in these basins.

A half graben basin setting formed during the Middle Miocene in the southern part of the East Bawean Trough (Chapter 4.3.2a). This basin is characterised by a greater subsidence rate toward the basin margin (Bawean Arch; Fig. 4.10). The sediments were derived mostly from the adjacent uplifted footwall. The absence of erosional features in this basin suggests that relative sea level falls have never exceeded the rate of subsidence. As a result, the sequence boundaries are difficult to place, which further emphasises the concerns of Swift et al. (1987) who suggested that sequence stratigraphic concepts are not applicable in basins characterised by a faster subsidence rate landward, compared to rates of eustatic sea level fall. The present study indicates that although sequence boundaries are subtle, maximum flooding surfaces resulting from retreating stratal patterns related to transgression are observable. These surfaces provide an important basis to correlate stratigraphic units between sub-basins.

Sub-basins showing the character of shelf margin settings include Madura Strait during the Quaternary and the Java Sea during the Pliocene. These basins are characterised by a greater subsidence rate toward the basin centre. The present study revealed that the development of systems tracts within this basin setting requires further research. In many cases, transgressive systems tracts are missing, and sequences are only represented by lowstand and highstand systems tracts. The incomplete development of systems tracts appears to depend on (1) palaeotopography and (2) order and magnitude of sea level change. Most parts of the basins in Madura Strait and the Java Sea are characterised by a relatively low morphological gradient. During a relative sea level rise, a rapid retreat (landward) of the depocentre results in sediment starvation in the basin centre and margin (slope). This retreat appears to increase if the order and magnitude of sea level change increases. More obviously, incomplete development of systems tracts occurs in the shelf setting, such as the Java Sea. Where the basin morphology is relatively flat and sea level fluctuations were rapid and of high magnitude as in the Quaternary, this effect is particularly pronounced. The sedimentary sequences are thin and extensive. The lowstand systems tract is only represented by fluvial channelling which was
later abandoned during a relative sea level rise, and the highstand systems tract is represented by a thin marine unit.

8.5 CONCLUSION

The existence of previously uninterpreted seismic data in Madura Strait and the Java Sea, combined with onshore seismic and outcrop data has provided a unique opportunity to study the development of the East Java Basin. The extent of the data and the application of seismic sequence stratigraphic concepts have allowed a comprehensive study of the timing of deposition and deformation of the basin fill; the mode of deposition and sedimentation patterns within the sub-basins; and the role of sea level changes on stratal patterns.


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Figure 1.1. Geographic location of study area.
Figure 2.1. Late Cainozoic tectonic elements of the western Indonesian region (after Hamilton, 1988).
Figure 2.2. Structural configuration of Java and surroundings (after Baumann, 1982).
Figure 2.3. Geographic locations of some back-arc basins in the western Indonesian region (after Ben-Avraham & Emery, 1973).
Figure 2.4. Three dimensional mesh diagram of the Bouguer anomaly in East Java, based on the measurements by the Bataafse Petroleum Maatschappij and the Geological Survey of Indonesia.
Figure 2.5. Geographic location of pre-Tertiary and Early Tertiary outcrops, indicated by black spots.
Figure 2.6. Paleogeographic evolution of the western Indonesian region (after Baumann, 1982).
Figure 2.7. Drainage pattern on the submerged Sunda Shelf interpreted from echosounder profiles (after Kuenen, 1950).
Figure 2.8. Physiographic map of central and east Java (after van Bemmelen, 1949).
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Figure 3.1. Geologic map of the East Java (modified from Pringgoprawiro, 1983).
### Figure 3.3: Tropical planktonic foraminiferal zonation:

Figure 3.4. Biostratigraphic zonation and palaeoclimate of the Solo River section, north of Ngawi, according to van Gorsel and Troelstra (1981, modified).
<table>
<thead>
<tr>
<th>SERIES</th>
<th>BLOW</th>
<th>ZONATION</th>
<th>FORMATION</th>
<th>PLANKTONIC FORAMINIFERA</th>
</tr>
</thead>
<tbody>
<tr>
<td>PLEISTOCENE</td>
<td>N23</td>
<td>LIDAH</td>
<td></td>
<td>Biota Datum</td>
</tr>
<tr>
<td></td>
<td>N22</td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td></td>
<td>N21</td>
<td></td>
<td></td>
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<tr>
<td></td>
<td>N20</td>
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<td></td>
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<tr>
<td></td>
<td>N19</td>
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<td></td>
<td>N18</td>
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<tr>
<td></td>
<td>N17</td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td></td>
<td>N16</td>
<td></td>
<td></td>
<td></td>
</tr>
</tbody>
</table>

<table>
<thead>
<tr>
<th></th>
<th>G. calida calida</th>
<th>G. truncatulinoides</th>
</tr>
</thead>
<tbody>
<tr>
<td></td>
<td>G. tosanensis</td>
<td>P. obliquiloculata</td>
</tr>
<tr>
<td></td>
<td></td>
<td>G. truncatulinoides</td>
</tr>
<tr>
<td></td>
<td>S. dehiscens immatura</td>
<td>P. obliquiloculata</td>
</tr>
<tr>
<td></td>
<td></td>
<td>S. dehiscens immatura</td>
</tr>
<tr>
<td></td>
<td>S. subdehiscens paeneadehiscens</td>
<td>G. plesirotumida</td>
</tr>
<tr>
<td></td>
<td></td>
<td>S. subdehiscens paeneadehiscens</td>
</tr>
</tbody>
</table>

Figure 3.5. Biostratigraphic zonation of the Puncakwangi section, south of Rembang (after Pringgoprawiro, 1983).
<table>
<thead>
<tr>
<th>SERIES</th>
<th>BLOW ZONATION</th>
<th>FORMATION</th>
<th>LITHOLOGY</th>
<th>PLANKTONIC FORAMINIFERA</th>
<th>BIODATUM</th>
<th>INTERVAL ZONE</th>
</tr>
</thead>
<tbody>
<tr>
<td>Pleistocene</td>
<td>N22</td>
<td>LIDAH</td>
<td></td>
<td>Globorotalia truncatulinoides</td>
<td></td>
<td></td>
</tr>
<tr>
<td>N21</td>
<td></td>
<td></td>
<td></td>
<td>Globorotalia tosaensis</td>
<td></td>
<td></td>
</tr>
<tr>
<td>N20</td>
<td></td>
<td>MUNDO</td>
<td></td>
<td>Pulleniatina obliquiloculata</td>
<td></td>
<td></td>
</tr>
<tr>
<td>N19</td>
<td></td>
<td></td>
<td></td>
<td>Globorotalia tosaensis</td>
<td></td>
<td></td>
</tr>
<tr>
<td>N18</td>
<td></td>
<td></td>
<td></td>
<td>Sphaeroidinella dehiscens immatura</td>
<td></td>
<td></td>
</tr>
<tr>
<td>N17</td>
<td></td>
<td>LEDOK</td>
<td></td>
<td>Sphaeroidinellopsis subdehiscens paenedehiscens</td>
<td></td>
<td></td>
</tr>
<tr>
<td>N16</td>
<td></td>
<td>WONOcolo</td>
<td></td>
<td>Globorotalia tumida plesiotumida</td>
<td></td>
<td></td>
</tr>
<tr>
<td></td>
<td></td>
<td></td>
<td></td>
<td>Globorotalia acostaensis</td>
<td></td>
<td></td>
</tr>
</tbody>
</table>

Figure 3.6. Biostratigraphic zonation of the Kali Sampurna section, Bojonegoro (after Pringgoprawiro, 1983).
<table>
<thead>
<tr>
<th>FORMATION</th>
<th>EPOCH</th>
<th>LITHOLOGY</th>
<th>PLANKTONIC FORAMINIFERA</th>
</tr>
</thead>
<tbody>
<tr>
<td></td>
<td></td>
<td></td>
<td>Globigerinoides aculeifer</td>
</tr>
<tr>
<td></td>
<td></td>
<td></td>
<td>Globigerinoides obesa</td>
</tr>
<tr>
<td></td>
<td></td>
<td></td>
<td>Globigerinoides tumidae</td>
</tr>
<tr>
<td></td>
<td></td>
<td></td>
<td>Globigerinoides papillosus</td>
</tr>
<tr>
<td></td>
<td></td>
<td></td>
<td>Globigerina marginata</td>
</tr>
<tr>
<td></td>
<td></td>
<td></td>
<td>Globigerina equilateralis</td>
</tr>
<tr>
<td></td>
<td></td>
<td></td>
<td>Globigerina conglobata</td>
</tr>
<tr>
<td></td>
<td></td>
<td></td>
<td>Globigerina dextroita</td>
</tr>
<tr>
<td></td>
<td></td>
<td></td>
<td>Globigerina acutangis</td>
</tr>
<tr>
<td></td>
<td></td>
<td></td>
<td>Globigerina crassa</td>
</tr>
<tr>
<td></td>
<td></td>
<td></td>
<td>Globigerina scrobiculata</td>
</tr>
<tr>
<td></td>
<td></td>
<td></td>
<td>Sphaeroidinella listidae</td>
</tr>
<tr>
<td></td>
<td></td>
<td></td>
<td>Pulvinatina obliquiloculata</td>
</tr>
<tr>
<td></td>
<td></td>
<td></td>
<td>Pulvinatina praelata</td>
</tr>
<tr>
<td></td>
<td></td>
<td></td>
<td>Pulvinatina miliacea</td>
</tr>
<tr>
<td></td>
<td></td>
<td></td>
<td>Planktonic foraminifera</td>
</tr>
</tbody>
</table>

Figure 3.7. Biostratigraphy of the Atasangin Formation on the north of Kabuh section, based on the planktonic foraminiferal analysis carried out during the present study. Dots indicate the stratigraphic position of the samples analysed.
<table>
<thead>
<tr>
<th>AGE MA</th>
<th>MAGNETIC STRAT.</th>
<th>DIATOM RANGES</th>
<th>FACIES</th>
<th>FORMATION</th>
<th>AGE MA</th>
</tr>
</thead>
<tbody>
<tr>
<td>1.5</td>
<td>Pleistocene</td>
<td></td>
<td>Marine</td>
<td>LIDAH</td>
<td>-1.5</td>
</tr>
<tr>
<td></td>
<td></td>
<td></td>
<td></td>
<td>(Dark blue clays)</td>
<td></td>
</tr>
<tr>
<td>2.0</td>
<td>Pliocene</td>
<td></td>
<td></td>
<td>Intermixed marine marls</td>
<td>-2.0</td>
</tr>
<tr>
<td>2.5</td>
<td>Pliocene</td>
<td></td>
<td></td>
<td>Transition series</td>
<td>-2.5</td>
</tr>
<tr>
<td>3.0</td>
<td>Pliocene</td>
<td></td>
<td></td>
<td>ATASANGIN</td>
<td>-3.0</td>
</tr>
<tr>
<td></td>
<td></td>
<td></td>
<td></td>
<td>(Well bedded marls)</td>
<td></td>
</tr>
</tbody>
</table>

Figure 3.8. Ranges of diatoms contained in the Sonde and Lidah Formations (modified from Ninkovich & Burckle, 1978).
Figure 3.9. Lithostratigraphic correlation between the Rembang and Kendeng Zones. The Puncakwangi and Kali Sampurna measured sections are from Hasjim (1987), and the Solo River section was measured during the present study.
Figure 3.10. Tentative north-south cross section showing the lateral and vertical relationships between lithostratigraphic units in the Rembang and Kendeng Zones.
### Table 3.1. Stratigraphic subdivision of the Rembang Zone according to Trooster (1937).

<table>
<thead>
<tr>
<th>AGE</th>
<th>FORMATION</th>
<th>STAGE</th>
<th>LOCAL FACIES</th>
</tr>
</thead>
<tbody>
<tr>
<td>PLEISTOCENE</td>
<td>Mergel Ton</td>
<td>Turn-Domas</td>
<td>Malo Limestone</td>
</tr>
<tr>
<td></td>
<td>Tambakromo</td>
<td></td>
<td>Selorejo Beds</td>
</tr>
<tr>
<td>MIocene</td>
<td>Late</td>
<td>Mundu</td>
<td>Kenang Horizon</td>
</tr>
<tr>
<td></td>
<td>Globigerina</td>
<td>Ledok</td>
<td>Grovergroene</td>
</tr>
<tr>
<td></td>
<td>Orbitoid Limestone (OK)</td>
<td>Ngrayong Sand (Upper OK)</td>
<td>Wonocolo</td>
</tr>
<tr>
<td></td>
<td></td>
<td>Lower OK</td>
<td></td>
</tr>
<tr>
<td></td>
<td>Upper Kujung</td>
<td>Upper Kujung</td>
<td>Prupuh Limestone</td>
</tr>
<tr>
<td></td>
<td>Kujung</td>
<td>Middle Kujung</td>
<td></td>
</tr>
<tr>
<td></td>
<td></td>
<td>Lower Kujung</td>
<td></td>
</tr>
</tbody>
</table>

### Table 3.2. Stratigraphic subdivision of the Kendeng and Rembang Zones according to van Bemmelen (1949).

<table>
<thead>
<tr>
<th>AGE</th>
<th>KENDENG ZONE</th>
<th>REMBANG ZONE</th>
</tr>
</thead>
<tbody>
<tr>
<td>Pleistocene</td>
<td>Late</td>
<td>Notopuro Breccias</td>
</tr>
<tr>
<td></td>
<td>Middle</td>
<td>Kabuh Beds</td>
</tr>
<tr>
<td></td>
<td>Early</td>
<td>Pucangan Beds</td>
</tr>
<tr>
<td></td>
<td>Late</td>
<td>Balanus Limestone</td>
</tr>
<tr>
<td></td>
<td></td>
<td>Sonde Marls</td>
</tr>
<tr>
<td>Pliocene</td>
<td>Early &amp;</td>
<td>Lower Kalibeng Beds</td>
</tr>
<tr>
<td></td>
<td>Middle</td>
<td>Klik Limestone</td>
</tr>
<tr>
<td></td>
<td>Late</td>
<td>Upper Kalibeng Beds</td>
</tr>
<tr>
<td>Miocene</td>
<td>Middle</td>
<td>Banyak Beds</td>
</tr>
<tr>
<td></td>
<td>Lower</td>
<td>Upper Kerek Beds</td>
</tr>
<tr>
<td></td>
<td></td>
<td>Lower Kerek Beds</td>
</tr>
<tr>
<td></td>
<td></td>
<td>Pelang Beds</td>
</tr>
</tbody>
</table>
Table 3.3. Stratigraphic subdivision of the Kendeng and Rembang Zones according to Pringgoprawiro (1983).
<table>
<thead>
<tr>
<th>AGE</th>
<th>Blow Zone</th>
<th>KENDENG ZONE</th>
<th>REMBANG ZONE</th>
</tr>
</thead>
<tbody>
<tr>
<td>PLEISTOCENE</td>
<td>N23</td>
<td>Notopuru Formation</td>
<td>Daster Member</td>
</tr>
<tr>
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<td>N22</td>
<td>Kubeh Formation</td>
<td>Malo Member</td>
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<tr>
<td>PLIOCENE</td>
<td>N21</td>
<td>Peegang Formation</td>
<td></td>
</tr>
<tr>
<td></td>
<td>N20</td>
<td>Liru Formation</td>
<td></td>
</tr>
<tr>
<td></td>
<td>N19</td>
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<td></td>
</tr>
<tr>
<td></td>
<td>N18</td>
<td>Sebes Formation</td>
<td>Paciran Formation</td>
</tr>
<tr>
<td></td>
<td>N17</td>
<td>Ariran Formation</td>
<td></td>
</tr>
<tr>
<td>LATE</td>
<td>N16</td>
<td>Atanung Formation</td>
<td></td>
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<tr>
<td></td>
<td>N15</td>
<td>Mekan Formation</td>
<td></td>
</tr>
<tr>
<td></td>
<td>N14</td>
<td>Taun Formation</td>
<td></td>
</tr>
<tr>
<td>MIDDLE</td>
<td>N13</td>
<td>Talu Formation</td>
<td></td>
</tr>
<tr>
<td></td>
<td>N12</td>
<td>Mejuk Formation</td>
<td></td>
</tr>
<tr>
<td></td>
<td>N11</td>
<td>Ngejej Formation</td>
<td></td>
</tr>
<tr>
<td></td>
<td>N10</td>
<td>Tuban Formation</td>
<td></td>
</tr>
<tr>
<td></td>
<td>N9</td>
<td>Prupah Formation</td>
<td></td>
</tr>
<tr>
<td>EARLY</td>
<td>N8</td>
<td>Kujung Formation</td>
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</tr>
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<td></td>
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<td></td>
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<td>N6</td>
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</tr>
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</tr>
<tr>
<td></td>
<td>N4</td>
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<td>N3</td>
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<td></td>
<td>P19</td>
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</tbody>
</table>

Table 3.4. Stratigraphic subdivision of the Kendeng and Rembang Zones used in the present study.
FIGURES AND TABLES TO CHAPTER FOUR
Figure 4.2. Upper section showing reflection termination pattern and types of discontinuities, lower section showing sequences and systems tracts (after Vail, 1987).
Figure 4.3. Typical seismic reflection patterns used to illustrate seismic facies (after Mitchum et al., 1977).
### BIOSTRATIGRAPHIC NOTES

<table>
<thead>
<tr>
<th>AGE</th>
<th>European Standard Stages</th>
<th>Local Stages</th>
<th>ZONE N.20 to N.21 (216 m)</th>
<th>Zones of Blow (1969)</th>
<th>Zones of Bolli and Bermudez</th>
<th>Depth in metres</th>
</tr>
</thead>
<tbody>
<tr>
<td>PLEISTOCENE</td>
<td></td>
<td></td>
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<td></td>
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<td></td>
<td></td>
<td></td>
<td></td>
</tr>
</tbody>
</table>

**BIOSTRATIGRAPHIC NOTES**

- The beds above 101 metres are possibly referable to Zone N.21 but no direct evidence. The beds are considered as being prior to the advent of *Globorotalia (G) truncatulinaidas*.
- Beds referred by Bolli (1966) to the '7 Equivalent of *Globorotalia truncatulinaidas*/*Globorotalia infaeta* Zone, but no *G. truncatulinaidas* seen in Bojonegoro Well No.1.
- HIATUS: *Globorotalia* *obliquus* extremus Zone.

- First appearance of *Globorotalia* (?), *acostaeis* (e.a)...
- Last occurrence of *Globorotalia* (?), *sinokensis*... Numerous small diatoms in Kawenqan Well No. 9...
- Last occurrence of *Cassigarina* *formosa*...

- *GA*... specimens of *Globorotalia* *tahiti forma typica* observed at 1,800 m...
- *HIATUS*: Zones N.9 (port), N.10 and N.11 (port) missing...
- Evolutionary first appearance of *Orbulina* op...
Figure 4.5. Stratigraphic column of the BNG-1 well (Banyubang Anticline), summarised from well log and palaeontological examination reports.
Figure 4.6. Schematic chronostratigraphic correlation section of seismic section PWD-23 (Sect. 4.1). The lithologies are based on the well and outcrop data.
Figure 4.7. Schematic chronostratigraphic correlation section of seismic section DNR-12 (Sect. 4.2). The lithologies are interpreted from seismic facies character and well data.
Figure 4.8. Comparison of sequence architecture between (A) passive margin and (B) foreland basin (after Swift et al., 1987).
Figure 4.9. Late Cainozoic global eustatic-cycle chart of Haq et al. (1988) compared with the inferred relative changes of sea level in the study area. The measurement was based on coastal aggradation on seismic section DNR-12 (Sect. 4.2).
Figure 4.10. Early and Middle Miocene basement architecture of the Rembang Zone and southeast of Java Sea, arrows indicate direction of deposition [compiled from Bishop (1980), Sudiro et al. (1973) & Ardhana (1993)].
Figure 4.11. Outcrop of the organic-rich mudstone facies of the Tawun Formation in the Jepon area, some 8 km east of Blora (Fig. 4.1.).

Figure 4.12. Typical exposure of the quartz sandstone (Ngrayong Member) of the Tawun Formation in the Plantungan Anticline. About 1 m thick bed of orbitoidal limestone is shown near the top and marked by an arrow.

Figure 4.13. Interbedded marl and calcarenite in the upper part of the Wonocolo Formation which typifies the boundary with the overlying Ledok Formation, exposed on the south flank of the Wonocolo Anticline, some 20 km northeast of Cepu (Fig. 4.1.).

Figure 4.14. Coquina beds (Selorejo Member) of the Mundu Formation crop out in the Selorejo area, some 10 km northwest of Cepu (Fig. 4.1.).
Figure 4.15. Photomicrograph of organic-rich mudstone of the Tawun Formation from the Jepon area. Sample no. BGJ X, crossed nicols.

Figure 4.16. Photomicrograph of quartz sandstone (Ngayong Member) of the Tawun Formation from the Plantungan Anticline. Sample no. N1, crossed nicols.

Figure 4.17. Photomicrograph of quartz sandstone of the Bulu Formation from the Bulu area, some 20 km north of Blora (Fig. 4.1). Fragment on the right is a quartzite fragment. Sample no. BL2, crossed nicols.

Figure 4.18. Orbitoidal limestones of the Bulu Formation from the Bulu area, some 20 km north of Blora (Fig. 4.1). Bioclasts are mostly foraminifer genus Lepidocyclina. Sample no. BL16, crossed nicols.
Figure 4.20. Photomicrograph of glauconitic sandstone of the upper part of the Ledok Formation from the Wonocolo area, some 20 km northeast of the Selojo Member from the Selojo Formation from the Selojo Member, some 10 km northeast of Cepu (Fig. 4.1). Sample no. L4, plane polarised light.

Figure 4.19. Typical photomicrograph of foraminifer-rich sand of the Wonocolo Antcline, some 20 km northeast of Cepu (Fig. 4.1). Sample no. WI, crossed nicons.

Figure 4.21. Foraminifer-rich marl of the Mundu Formation from the Ledok Antcline, some 15 km north of Cepu (Fig. 4.1). Sample no. MI, plane polarised light.

Figure 4.22. Photomicrograph of the Selojo Member of the Mundu Formation from the Selojo Member, some 20 km northeast of the Selojo Member from the Selojo Formation from the Selojo Member, some 10 km northeast of Cepu (Fig. 4.1). Sample no. S1, crossed nicons.
Figure 4.23. X-ray diffractogram of the Wonocolo Formation. The sample (sample no. W1) is from the Wonocolo area, some 20 km northeast Cepu (Fig.4.1). Ca=calcite, Ar=aragonite, Qz=quartz, Mo=smectite, Ko=kaolinite.
| Table 4.1. Seismic characteristics and environmental facies interpretation of clastic seismic facies units (after Sangree & Widmier, 1977). |
FIGURES TO CHAPTER FIVE
Figure 5.1. Geographic map of areas under discussion in Chapter 5. Triangles indicate presently active volcanoes.
Figure 5.2. Generalised vertical section of the Kerek Formation in the Solo River section. (summarised from Muin, 1985).
**SOLO RIVER SECTION**

<table>
<thead>
<tr>
<th>UNIT</th>
<th>LITHOFACIES</th>
<th>ENVIRONMENT</th>
</tr>
</thead>
<tbody>
<tr>
<td></td>
<td>Glaucocitic calcarenite, planar cross-bedded, composed mostly of foraminiferal tests.</td>
<td>Outer neritic</td>
</tr>
<tr>
<td></td>
<td>Marl, unstratified and homogenous, few interbeds of calcarenite (&lt; 40 cm).</td>
<td>Outer neritic</td>
</tr>
<tr>
<td></td>
<td>Calcareous fine to medium-grained sandstone, trough cross-bedded.</td>
<td>Inner neritic (?)</td>
</tr>
<tr>
<td></td>
<td>Marl, unstratified and homogenous, few interbeds of calcarenite (40 cm to 1 m).</td>
<td>Outer neritic</td>
</tr>
<tr>
<td></td>
<td>Sandy marl, bioturbated, with thin beds (15–20 cm) of planar cross-bedded sandy calcarenites with erosional base.</td>
<td>Inner neritic</td>
</tr>
<tr>
<td></td>
<td>Calcirudite, composed of benthonic larger foraminifers, coral and volcanic rock fragments.</td>
<td></td>
</tr>
<tr>
<td></td>
<td>Calcareous sandy mudstone, some beds of medium-grained, ripple-laminated sandstone.</td>
<td></td>
</tr>
<tr>
<td></td>
<td>Parallel-laminated calcareous mudstone interbedded with parallel- and ripple-bedded, medium-grained sandstones, 5–20 cm bed thickness.</td>
<td></td>
</tr>
</tbody>
</table>

Figure 5.3. Generalised vertical section of the uppermost Kerek Formation and the Kalibeng Formation, measured along the Solo River north of Ngawi.
Figure 5.4. An outcrop of the turbiditic deposits of the Kerek Formation (upper part) in the Solo River section; intercalations of calcareous sandstone and sandy mudstone.

Figure 5.5. Trough cross-bedded calcareous sandstone in the middle part of the Kalibeng Formation exposed in the Solo River section.
Figure 5.6. Photomicrograph of calcareous sandstone of the Kerek Formation from the Solo River section. Although not very obvious, the specimen is laminated between planktonic foraminiferal-rich (left part) and terrigenous-rich (right part) laminae. Sample no. KE1A, crossed nicols.

Figure 5.7. Photomicrograph of coarse calcarenite, part of calcirudite, from the uppermost part of the Kerek Formation in the Solo River section. Larger foraminifers shown are Lepidocyclina, the rock fragment on the right is diorite. Sample no. BU1, crossed nicols.

Figure 5.8. Photomicrograph of calcareous sandstone from the middle part of the Kalibeng Formation in the Solo River section. Sample no. KB13A, crossed nicols.

Figure 5.9. Photomicrograph of foraminiferal-rich glauconitic calcarenite from the upper part of the Kalibeng Formation in the Solo River section. Most of glauconite grains are interstitial, formed within foraminiferal tests. Sample no. KB15, plane polarised light.
Figure 5.10. Tentative palaeogeographic map of East Java in the Late Miocene. (arrows indicate direction of deposition)
Figure 5.11. Inferred depositional setting of the Kerek Formation, drawn based on the model of Hathway (1994) for the Oligo-Miocene volcaniclastic sedimentation in Viti Levu, Fiji.
Figure 5.12. X-ray diffractograms of the samples from A: the lower and B: upper parts of the Kalibeng Formation on the Solo River section (sample no. KB9 and KB14 respectively; Fig. 5.3). Mo=smectite, Ko=kaolinite, Qz=quartz, Ca=calcite, Ab=albite, Mi=mica, Ho=hornblende, Px=pyroxene.
Figure 5.13. Tentative palaeogeographic map of East Java in the Early Pliocene.
Figure 5.14. Generalised vertical section of the Atasangin Formation, measured on the area north of Kabuh (see Fig. 5.1 for the location).
Figure 5.15. Vertical section of the deltaic sandstone facies of the Atasangin Formation, measured in the area north of Kabuh (see Fig. 5.1 for the location).
Figure 5.16. Deformation structure underlain by a fining upward unit in the delta front facies of the deltaic sandstone facies, Atasangin Formation, in the area north of Kabuh.

Figure 5.17. Distributary channel fills in the deltaic sandstone facies, Atasangin Formation, in the area north of Kabuh.

Figure 5.18. Photomicrograph of calcareous sandstone from sandstone facies Atasangin Formation in the area north of Kabuh. Sample no. JKB14, crossed nicols.
Figure 5.19. X-ray diffractograms of the samples from A: lower (NKabuhl13) and B: upper (NKabuhl18) parts of the siliceous marl Atasangin Formation (see Fig. 5.14 for stratigraphic position). Mo=smectite, Ko=kaolinite, Cl=chlorite, Qz=quartz, Ca=calcite, Ab=albite, Si=siderite.
Figure 5.20. Tentative palaeogeographic map of East Java in the Middle Pliocene.
Figure 5.21. Tentative palaeogeographic map of East Java in the Late Pliocene.
Figure 5.22. Generalised vertical section of the Sonde Formation and the lower sequence of the Pucangan Formation in the Solo River, north of Ngawi (see Fig. 5.1 for the location).
Figure 5.23. An exposure of the bedded limestone Klitik Member of the Sonde Formation (Klitik Limestones, van Bemmelen, 1949) in the Solo River section.

Figure 5.24. A closer view of the bedded limestone Klitik Member in Figure 5.23.
<table>
<thead>
<tr>
<th>UNIT</th>
<th>LITHOFACIES</th>
<th>ENVIRONMENT</th>
</tr>
</thead>
<tbody>
<tr>
<td>PUCANGAN FORMATION</td>
<td>Parallel-laminated, poorly sorted sandstone interbedded with sandy mudstone, abundant comminuted marine molluscan shells.</td>
<td>Subtidal</td>
</tr>
<tr>
<td>400</td>
<td></td>
<td></td>
</tr>
<tr>
<td>300</td>
<td>Bluish grey sandy mudstone, unstratified, common organic streaks.</td>
<td>Lacustrine Anoxic</td>
</tr>
<tr>
<td>LIDAH FORMATION</td>
<td>See Figure 5.26</td>
<td></td>
</tr>
<tr>
<td>200</td>
<td>Bluish grey mudstone, unstratified, homogenous, common organic streaks.</td>
<td>Lacustrine Anoxic</td>
</tr>
<tr>
<td>100</td>
<td>Ripple-laminated calcareous fine sandstone (1 m thick).</td>
<td></td>
</tr>
<tr>
<td></td>
<td>Bluish grey mudstone, unstratified, homogenous, common organic streaks.</td>
<td></td>
</tr>
<tr>
<td>ATASANGIN FORMATION</td>
<td>Fossiliferous, sandy mudstone intercalated with muddy sandstone, structure unobserved.</td>
<td>Subtidal</td>
</tr>
<tr>
<td>m</td>
<td>Mudstone, dark grey, unstratified, homogenous.</td>
<td></td>
</tr>
<tr>
<td></td>
<td>Marl, unstratified, fossiliferous, thin interbeds of medium-grained calcarenite</td>
<td></td>
</tr>
</tbody>
</table>

Figure 5.25. Generalised vertical section of the Lidah Formation in Sumberringin area, north of Jombang (see Fig. 5.1 for the location).
### Sumberringin Section, Jombang

<table>
<thead>
<tr>
<th>UNIT</th>
<th>Lithofacies</th>
<th>Environment</th>
</tr>
</thead>
<tbody>
<tr>
<td></td>
<td>Bluish grey, mudstone, unstratified.</td>
<td>Lacustrine</td>
</tr>
<tr>
<td></td>
<td>Medium to coarse-grained sandstone, conglomeratic at base, fining upward into fine-grained sandstone, common planar cross-and ripple-stratifications.</td>
<td>Tidal/fluvial channel</td>
</tr>
<tr>
<td></td>
<td>Muddy sandstone fining upward into carbonaceous mudstone, unstratified, abundant clay and organic pellets.</td>
<td>Mud flat</td>
</tr>
<tr>
<td></td>
<td>Interbedded between massive, poorly sorted muddy sandstone and medium-sorted, ripple-laminated, medium-grained sandstone. Both contain abundant benthonic foraminifers. Bluish grey mudstone, unstratified, homogenous, common organic streaks.</td>
<td>Subtidal</td>
</tr>
<tr>
<td></td>
<td></td>
<td>Lacustrine Anoxic</td>
</tr>
</tbody>
</table>

Figure 5.26. Vertical section of the upper sandstone facies Lidah Formation in the Sumberringin area (see Fig. 5.1 for the location).
Light-grey calcareous mudstone, unstratified, carbonaceous and bioturbated in some parts, some interbeds of calcarenite (< 60 cm).

Tidal flat

Calcarenite interbedded with calcareous sandy mudstone, glauconitic, rich in comminuted shell fragments.

Shoreface/foreshore(?)

Bluish grey mudstone, unstratified, carbonaceous, streaks of fine-grained quartz and pyrite, some thin interbeds (< 10 cm) of fine-grained, ripple-laminated, calcareous sandstones.

Lacustrine Anoxic

Figure 5.27. Generalised vertical section of the Lidah Formation in Nglampin area, some 30 km southeast Cepu (after Nahrowi et al. 1981; see Fig. 5.1 for the location).
Figure 5.28. X-ray diffractograms of the samples from (a) lower (Nglampin1) and (b) middle (Nglampin7) parts of the Lidah Formation on the Nglampin area (see Fig. 5.27 for stratigraphic positions). Mo=smectite, It=illite, Ko=kaolinite, Qz=quartz, Si=siderite, Ca=calcite, Gp=gypsum.
Figure 5.29. Tentative palaeogeographic map of East Java in the late-Late Pliocene.
Figure 5.30. Geologic map of the Jombang-Mojokerto area (modified from Duyfjes, 1938b,d).
Figure 5.31. Depositional environments of a tidally-influenced area (after Dalrymple, 1992).

Figure 5.32. Various subenvironments in a barrier-island system (after Reinson, 1984).
**KABUH SECTION, JOMBANG**

<table>
<thead>
<tr>
<th>UNIT</th>
<th>LITHOFACIES</th>
<th>ENVIRONMENT</th>
</tr>
</thead>
<tbody>
<tr>
<td>150</td>
<td>Conglomerate, massive, cobble-size fragment of andesitic rocks, normally graded into planar cross-beded, coarse-grained sandstone, contains pebble size mud clasts, shell fragments of marine mollusc, and thin beds of calcareous mudstone.</td>
<td>Reworked alluvial(?) (L6)</td>
</tr>
<tr>
<td>100</td>
<td>Tuffaceous mudstone</td>
<td></td>
</tr>
<tr>
<td>50</td>
<td>Conglomerate, massive to crudely bedded, cobble-size fragment of andesitic rocks, normally graded into planar-crossbedded, coarse-grained sandstone, contains pebble-size mud clasts.</td>
<td>Tidal flat (L5)</td>
</tr>
<tr>
<td></td>
<td>Parallel-laminated mudstone, strongly bioturbated.</td>
<td>Debris flow (L4)</td>
</tr>
<tr>
<td>0</td>
<td>Matrix-supported conglomerate, cobble-size fragments of andesitic rocks and mud clasts, normally graded into tuffaceous, poorly-sorted sandstone, contains foraminifers.</td>
<td></td>
</tr>
<tr>
<td></td>
<td>Sandstone, medium- to coarse-grained, moderately sorted, contains plant remains, various primary structures (large scale planar, simple and trough cross- stratifications) with well developed mud flaser and scouring.</td>
<td>Tidal flat (L3b)</td>
</tr>
<tr>
<td></td>
<td>Sandy mudstone, greyish green, rich in benthonic foraminifers.</td>
<td>Offshore facies (L3a)</td>
</tr>
<tr>
<td></td>
<td>Matrix-supported conglomerate, inverse to normally graded, pebble to boulder-size fragment of volcanic rocks and tuffaceous mudstone clasts.</td>
<td>Debris flow (L2)</td>
</tr>
<tr>
<td></td>
<td>Sandstone, planar cross-stratification, interbedded with bioturbated muddy-sandstone. Both contain abundant comminuted shell of marine molluscs and foraminiferal tests.</td>
<td>Subtidal (L1)</td>
</tr>
</tbody>
</table>

Figure 5.33. Vertical section of the lower sequence of the Pucangan Formation to the north of Kabuh (see Fig. 5.30. for the location).
Coarse-grained sandstone, conglomeratic, structure unobserved.
Mudstone, grey, massive, bioturbated, rich in marine mollusc.

Fine-grained sandstone interbedded with mudstone, parallel-laminated, greyish blue, highly carbonaceous.

Coarse-grained sandstone, medium sorted, conglomeratic, planar and trough cross-bedding structures, contains shells of marine mollusc and thin interbeds of lapilli tuff.

Interbedded fine-grained sandstone and mudstone, parallel-laminated, greyish blue, highly carbonaceous.

Sandy mudstone, light grey, massive, strongly bioturbated, rich in marine mollusc fragments and foraminifers.

Mudstone, dark blue, massive and highly carbonaceous.

Interbedded fine-grained sandstone and mudstone, parallel-laminated, greyish blue, highly carbonaceous.

Alluvial fan ?
Upper sequence

Middle sequence
Tidal flat

Lower sequence
(L1-L5)

Offshore facies

Lacustrine
Anoxic

Figure 5.34. Generalised vertical section of the Pucangan Formation in the area north of Kemlagi (see Fig. 5.30. for the location).
Figure 5.35. Debris flow deposits of the Pucangan Formation exposed along the Solo River section.

Figure 5.36. Similar to Fig. 5.35. This deposit has a non-erosive contact with the underlying mudstone.
**PERNING SECTION, MOJOKERTO**

<table>
<thead>
<tr>
<th>UNIT</th>
<th>LITHOFACIES</th>
<th>ENVIRONMENT</th>
</tr>
</thead>
<tbody>
<tr>
<td>Kabuh Formation</td>
<td>Conglomeratic sandstone, parallel bedded and epsilon cross-bedded.</td>
<td>Fluvial channel ?</td>
</tr>
<tr>
<td></td>
<td></td>
<td>(U2)</td>
</tr>
<tr>
<td></td>
<td>Sandstone, poorly sorted, medium to coarse-grained, conglomeratic.</td>
<td>Lower shoreface (U1)</td>
</tr>
<tr>
<td>Upper sequence</td>
<td>Mudstone, grey, massive, bioturbated, contains marine molluscs fragments.</td>
<td></td>
</tr>
<tr>
<td></td>
<td>Sandy mudstone, very poorly sorted, highly bioturbated, rich in marine</td>
<td>Alluvial fan (M3)</td>
</tr>
<tr>
<td></td>
<td>mudstones, large-scale burrows of <em>Gastrochaenolites</em>.</td>
<td></td>
</tr>
<tr>
<td></td>
<td>Stacked fining upward units, crudely-bedded to planar cross-bedded</td>
<td>Tidal flat (M2)</td>
</tr>
<tr>
<td></td>
<td>conglomerate, medium to coarse-grained sandstones. Each sequence separated</td>
<td></td>
</tr>
<tr>
<td></td>
<td>by sandy mudstone, contains vertebrate and hominid fossils.</td>
<td></td>
</tr>
<tr>
<td></td>
<td>Tuffaceous mudstone, parallel-laminated, scoured by small-scale (15-30 cm)</td>
<td></td>
</tr>
<tr>
<td></td>
<td>channels and filled by marine-fossiliferous sandstone.</td>
<td></td>
</tr>
<tr>
<td></td>
<td>Sandstone, medium-grained, medium-sorted, trough cross-bedding.</td>
<td></td>
</tr>
<tr>
<td></td>
<td>Sandstones, planar cross-bedded, medium-sorted, medium- to coarse-grained.</td>
<td></td>
</tr>
<tr>
<td></td>
<td>Sandstone, medium-grained, ripple-laminated with well developed parallel and</td>
<td></td>
</tr>
<tr>
<td></td>
<td>wavy-bedded bundled mudstone.</td>
<td></td>
</tr>
<tr>
<td></td>
<td>Ripple-laminated fine-grained sandstone with interbeds of reverse and normally-graded conglomeratic sandstone.</td>
<td>Subtidal (M1)</td>
</tr>
<tr>
<td></td>
<td>Mudstone, parallel-laminated, interbedded with parallel to ripple-laminated</td>
<td></td>
</tr>
<tr>
<td></td>
<td>fine-grained sandstone, abundant plant remains and leaf imprints between</td>
<td></td>
</tr>
<tr>
<td></td>
<td>laminae.</td>
<td></td>
</tr>
<tr>
<td></td>
<td>Sandstone, medium-sorted, medium-grained, glauconitic, contains bentonic</td>
<td></td>
</tr>
<tr>
<td></td>
<td>foraminifers, abundant comminuted shell of marine mollusc, and thin</td>
<td></td>
</tr>
<tr>
<td></td>
<td>interbeds of carbonaceous mudstone.</td>
<td></td>
</tr>
<tr>
<td></td>
<td>Coarse-grained sandstone, conglomeratic, structure unobserved.</td>
<td></td>
</tr>
</tbody>
</table>

Figure 5.37. Vertical section of the middle sequence of the Pucangan Formation on the north of Perning (see Fig. 5.30 for the location). Palaeomagnetic dates are from Hyodo *et al.* (1992).
Figure 5.38. Parallel and ripple-laminated sandstone with well developed mud drapes occurring between two sets of planar cross-bedding. These characterise the tidal flat deposit unit L3b (Kabuh section).

Figure 5.39. Simple cross-bedding with mud flaser occurring in the tidal flat deposit unit L3b (Kabuh section).

Figure 5.40. Epsilon cross-bedding with mud drapes developed between laminae in the middle part of tidal deposit unit M2 (Perning section).

Figure 5.41. Parallel-laminated mudstone interbedded with ripple-laminated sandstone in the middle part of tidal deposit unit M2 (Perning section).
Figure 5.42. Ripple-laminated sandstone with well developed flaser bedding, associated with conglomeratic sandstone of small scale channel deposits in the uppermost part of unit M2 (Perning section).

Figure 5.43. Fine intercalation of sandstone and mudstone overlies a tidal channel deposit in the middle part of unit M2 (Perning section).
Figure 5.44. Vertical section of the upper sequence Pucangan Formation and the lower part of the Kabuh Formation in Banyuurip area (see Fig. 5.30 for the location).
Figure 5.45. A monument erected to commemorate the finding of a hominid skull of the *Pithecanthropus Modjokertensis*. The skull was found in the conglomeratic unit M3 of the Perning section.

Figure 5.46. Well bedded sandy mudstone of unit U1 (upper sequence), bioturbated at the upper part and characterised by *Glossifungites* burrow.

Figure 5.47. Finely laminated mudstone and fine sandstone in the upper most part of unit U1 (upper sequence).
Figure 5.48. Vertical section of the upper sequence Pucangan Formation and the lower part of the Kabuh Formation on the south of Kedamean (see Fig. 5.30 for the location).
Figure 5.49. A typical exposure on the Kedanean section. The grey mudstone unit of shelf mud facies marks the boundary between the Pucangan and Kabuh Formations.
**Figure 5.50.** Vertical section of the upper sequence Pucangan Formation and the lower part of the Kabuh Formation in Raci Anticline (see Fig. 5.1. for the location).
Figure 5.51. Typical exposure in the Raci Anticline. Prominent unit at the lower part is unit U2a, the upper sequence Pucangan Formation, which is underlain by calcareous mudstone. The uppermost unit marked by arrow is the Notopuro Formation; below this unit is the fluvial channel deposit Kabuh Formation which overlies a shelf mud facies of this formation.

Figure 5.52. Coarse sandstone of unit U2a (upper sequence, Pucangan Formation) on the Raci Anticline, characterised by large scale trough cross-bedding.
Figure 5.53. Photomicrograph of a volcanic derived sandstone from the lower sequence Pucangan Formation on the Kabuh area (see Fig. 5.33 for the stratigraphic position). Sample no. JKB6, crossed nicols.

Figure 5.54. Photomicrograph of a sandstone facies from the lower sequence Pucangan Formation on the Kemlagi area (see Fig. 5.34 for the stratigraphic position). Sample no. KG1, crossed nicols.

Figure 5.55. Photomicrograph of a calcareous sandy mudstone from the upper sequence Pucangan Formation on the Banyuurip area (see Fig. 5.44 for the stratigraphic position). Sample no. Banyuurip4, crossed nicols.
Figure 5.56. Tentative palaeogeographic map of East Java in the Early and Middle Pleistocene.
Figure 5.57. Schematic development of the Lidah and Pucangan Formation in the eastern Kendeng Zone (after Duyfjes, 1938b).
Figure 5.58. Depositional model of the Pucangan Formation at the end of alluvial fan unit M3 deposition (not to scale).
FIGURES TO CHAPTER SIX
Figure 6.1. Ship tracks during shallow seismic reflection profiling, and observed anticlines in Madura Strait.
Figure 6.2. Locations of seismic lines used for discussion in this chapter.
**Figure 6.3.** Biostratigraphy of the uppermost 700 m of the MS1-1 well (summarised from the well report).

<table>
<thead>
<tr>
<th>Epoch</th>
<th>Depth (m)</th>
<th>Foraminifers</th>
<th>Benthonic</th>
<th>Planktonic</th>
</tr>
</thead>
<tbody>
<tr>
<td>PLEISTOCENE</td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>Pliocene</td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td></td>
<td>100-</td>
<td><strong>Operculina ammonoidea</strong></td>
<td></td>
<td><strong>Globoquadratina americana</strong></td>
</tr>
<tr>
<td></td>
<td>200-</td>
<td><strong>Ammonia craticulata</strong></td>
<td></td>
<td></td>
</tr>
<tr>
<td></td>
<td>300-</td>
<td><strong>Ammonia obliqua</strong></td>
<td></td>
<td></td>
</tr>
<tr>
<td></td>
<td>400-</td>
<td><strong>Globorotalia obliqueuosclta</strong></td>
<td></td>
<td></td>
</tr>
<tr>
<td></td>
<td>500-</td>
<td><strong>Globorotalia virgulata</strong></td>
<td></td>
<td></td>
</tr>
<tr>
<td></td>
<td>600-</td>
<td><strong>Globorotalia arankesi</strong></td>
<td></td>
<td></td>
</tr>
</tbody>
</table>

**Est. Lithology**

- Operculina ammonoidea
- Ammonia craticulata
- Ammonia obliqua
- Globorotalia obliqueuosclta
- Globorotalia virgulata
- Globorotalia arankesi

**Foraminifers**

- Benthonic
- Planktonic
Figure 6.4. Bathymetry of Madura Strait (contours in metres). The data was digitised from seismic sections with an assumed seismic velocity = 1500 m/sec.
Figure 6.5. Very large scale planar crossbedding which characterises the upper part of the Pasean Formation in an area 15 km east of Kamal, Madura Island.

Figure 6.6. Photomicrograph of clastic limestone from the Pasean Formation from 15 km east of Kamal, Madura Island (Fig. 6.1). Sample no. Madura-18, crossed nicols.

Figure 6.7. Photomicrograph of clastic limestone from the Pasean Formation from 15 km south of Pamekasan, Madura Island (Fig. 6.1). Sample no. PK5, crossed nicols.
Figure 6.8. Ranges of planktonic foraminiferal species from the Pasean Formation in the Kamal area (sample no. Madura-8).

<table>
<thead>
<tr>
<th>Blow (1969) zone</th>
<th>7</th>
<th>8</th>
<th>9</th>
<th>10</th>
<th>11</th>
<th>12</th>
<th>13</th>
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Figure 6.9. Ranges of planktonic foraminiferal species from the Pasean Formation in the area south of Pamekasan (sample no. Pamekasan-8).
Figure 6.10. Ranges of planktonic foraminiferal species from the Pasean Formation in the Gili Genting Island (sample no. GG-3 & GG-7)

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|                  | 3. *Globigerinoides sacculifer* |     |     |     |     |     |     |     |     |     |     |     |     |     |     |     |     |     |
|                  | 4. *Orbulina universa*         |     |     |     |     |     |     |     |     |     |     |     |     |     |     |     |     |     |
|                  | 5. *Globigerinoides elongatus* |     |     |     |     |     |     |     |     |     |     |     |     |     |     |     |     |     |
|                  | 6. *Globorotalia dutertrei*     |     |     |     |     |     |     |     |     |     |     |     |     |     |     |     |     |     |
|                  | 7. *Globorotalia tumida*       |     |     |     |     |     |     |     |     |     |     |     |     |     |     |     |     |     |
|                  | 8. *Globigerinoides ruber*     |     |     |     |     |     |     |     |     |     |     |     |     |     |     |     |     |     |
|                  | 9. *Sphaeroidinella dehiscens* |     |     |     |     |     |     |     |     |     |     |     |     |     |     |     |     |     |
|                  | 10. *Pulleniatina obliquiloculata* |     |     |     |     |     |     |     |     |     |     |     |     |     |     |     |     |     |
Figure 6.11. X-ray diffractograms of samples from unit Pre-A exposed in (a) the Gili Raja Island, sample no. GR6; and (b) east of Kamal, sample no. BH5-17. Mo=smectite, It=illite, Ko=kaolinite, Qz=quartz, Ca=calcite, Si=siderite, Cl=chlorite, Ch=chamosite, Ar=aragonite.
Figure 6.12. Seismic facies map of subunit A1 superimposed on the isochron map. Contours in milliseconds TWT (1 msec = 0.75 m, seismic velocity = 1500 m/sec).

Figure 6.13. Seismic facies map of subunit A2 superimposed on the isochron map. Contours in milliseconds TWT (1 msec = 0.75 m, seismic velocity = 1500 m/sec).
Figure 6.14. Seismic facies map of subunit B1a superimposed on the isochron map. Contours in milliseconds TWT (1 msec = 0.75 m, seismic velocity = 1500 m/sec).

Figure 6.15. Seismic facies map of subunit BV superimposed on the isochron map. Contours in milliseconds TWT (1 msec = 0.75 m, seismic velocity = 1500 m/sec).
Figure 6.16. Isochron map of subunit B1b. Contours in milliseconds TWT (1 msec = 0.75 m, seismic velocity = 1500 m/sec).

Figure 6.17. Isochron map of subunit B2. Contours in milliseconds TWT (1 msec = 0.75 m, seismic velocity = 1500 m/sec).
Figure 6.18. Isochron map of subunit B3. Contours in milliseconds TWT (1 msec = 0.75 m, seismic velocity = 1500 m/sec).

Figure 6.19. Seismic facies map of subunit C1a superimposed on the isochron map. Contours in milliseconds TWT (1 msec = 0.75 m, seismic velocity = 1500 m/sec).
Figure 6.20. Seismic facies map of subunit C1b superimposed on the isochron map. Contours in milliseconds TWT (1 msec = 0.75 m, seismic velocity = 1500 m/sec). Patterned indicates partly or completely eroded.

Figure 6.21. Seismic facies map of subunits C2 (deposited southward) and CV (deposited northward) superimposed on the isochron map. Contours in milliseconds TWT (1 msec = 0.75 m, seismic velocity = 1500 m/sec). Arrows indicate the directions of downlaps, patterned indicates partly or completely eroded.
Figure 6.22. Seismic facies map of subunit C3 superimposed on the isochron map. Contours in milliseconds TWT (1 msec = 0.75 m, seismic velocity = 1500 m/sec). Arrows indicate the directions of downlaps, patterned indicates partly or completely eroded.

Figure 6.23. Seismic facies map of subunit C4 superimposed on the isochron map. Contours in milliseconds TWT (1 msec = 0.75 m, seismic velocity = 1500 m/sec). Arrows indicate the directions of downlaps, patterned indicates partly or completely eroded.
Figure 6.24. Seismic facies map of subunit C5 (C5a and C5b) superimposed on the isochron map. Contours in milliseconds TWT (1 msec = 0.75 m, seismic velocity = 1500 m/sec). Arrows indicate the directions of downlaps, patterned indicates partly or completely eroded.

Figure 6.25. Seismic facies map of subunit D1a superimposed on the isochron map. Contours in milliseconds TWT (1 msec = 0.75 m, seismic velocity = 1500 m/sec). Patterned indicates partly or completely eroded.
Figure 6.26. Stream courses occurring on top of unit D1a. The main channel, on the middle east, was meandering and flowed easterly.
Figure 6.27. Palaeogeographic map during the subunit D1b deposition superimposed on the contour time structure map of top subunit D1a.

Figure 6.28. Palaeogeographic map during the subunit D2 deposition superimposed on the isochron map of combined subunits D1b, D2 and D3. Arrows indicate the directions of downlaps.
Figure 6.29. Three dimensional mesh diagrams of the time structures: (a) top subunit D2 or bathymetry (D3 is very thin over the area and ignored); (b) top subunit D1a or bottom subunit D1b; and (c) bottom subunit D1a or top subunit C. Note the increasing fold complexity westward (c) and rapid filling of synclines by subunit D1a (c & b).
Figure 6.30. Correlation between geological outcrops in the Kedamean and Perning sections, and seismic sections MB-MB' and ME-ME'.
Figure 6.31. (A) Sea level curve in the last 240 ka for Huon Peninsula, New Guinea; (B) $^{18}$O record from east equatorial core V19-30; (C) Recalculated sea level curve after correlation with $^{18}$O record from core V19-30 (after Chappell & Shackleton, 1986).
Figure 6.32. Ages of seismic units correlated with magnetostratigraphy and oxygen isotope records of Harland et al. (1989).
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Figure 7.6. Palaeogeographic map during the deposition of lowstand systems tract subunit JP1 (Early Pliocene), plotted on the time structure contours to top of Miocene. Contours in milliseconds TWT (1 msec = 0.75 m, seismic velocity = 1500 m/sec).
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Figure 7.8. Time structure map to top of the Pliocene unit JP. Subunit JP2 covers the area lower than contour 210 msec TWT. Contours in milliseconds TWT (1 msec = 0.75 m, seismic velocity = 1500 m/sec).
Figure 7.9. Isochron map of the total Quaternary unit JQ. Contours in milliseconds TWT (1 msec = 0.75 m, seismic velocity = 1500 m/sec).
Figure 7.10. Isochron map of combined subunits JQ1 and JQ2 (Early Pleistocene). Contours in milliseconds TWT (1 msec = 0.75 m, seismic velocity = 1500 m/sec).
Figure 7.11. Time structure map to top of subunit JQ2. Contours in milliseconds TWT (1 msec = 0.75 m, seismic velocity = 1500 m/sec).
Figure 7.12. Isochron map of combined subunits JQ3 and JQ4 (Middle Pleistocene). Contours in milliseconds TWT (1 msec = 0.75 m, seismic velocity = 1500 m/sec).
Figure 7.13. Time structure map to top of subunit JQ4, stippled areas indicating the position of channel cuts. Contours in milliseconds TWT (1 msec = 0.75 m, seismic velocity = 1500 m/sec).
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Figure 7.15. Inferred ages of seismic units, based on the biostratigraphic analysis on petroleum exploratory wells in the Java Sea and the oxygen isotope records of Harland et al. (1989) from deep sea cores.
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Figure 8.2. Simplified palaeogeographic map of the East Java Basin during Pliocene to Middle Pleistocene.
APPENDICES
### APPENDIX A. LIST OF AVAILABLE VELOCITY DATA FOR LINE PWD-23

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APPENDIX B.1. TABULATION OF PETROGRAPHIC DESCRIPTIONS

EXPLANATION

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NOTE: THE RELATIVE ABUNDANCE SCALE USED HERE CONSIDERS THE INDIVIDUAL PETROGRAPHIC COMPONENT AS 100 % CORRESPONDING TO THE GROUP PERCENTAGE (IN BOLD).
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APPENDIX B.2. X-RAY DIFFRACTION DATA

Key to abbreviation

Ab : Albite
Ar : Aragonite
Bo : Boehmite
Ca : Calcite
Ch : Chamosite
Cl : Chlorite
Gp : Gypsum
Ho : Hornblende
It : Illite
Ko : Kaolinite
Mi : Mica
Mo : Smectite
Px : Pyroxene
Qz : Quartz
Si : Siderite
APPENDIX B.2. X-RAY DIFRACTOGRAMS.

Sample no. W1 (Wonocolo Formation)

Sample no. KB9 (Kalibeng Formation)
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**Note:**
- Relative abundance: ? = doubtful, 1 = rare, 2 = common, 3 = abundant
- The relative abundance scale considers total planktonic species in the individual sample as 100%
APPENDIX C.2. BENTHONIC FORAMINIFERAL DATA.

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Note:
- Relative abundance: 1 = rare; 2 = common; 3 = abundant
- The relative abundance scale considers total benthonic species in the individual sample as 100%
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Note:
- Relative abundance: 1 = rare, 2 = common, 3 = abundant
- The relative abundance scale considers total benthonic species in the individual sample as 100%
## APPENDIX D. LIST OF ROCK SAMPLES USED IN THIS STUDY

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APPENDIX E.1. GRAPHIC SUMMARY OF PETROLEUM EXPLORATORY WELL REPORT FOR JS2-1 WELL.

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<tr>
<td>Sphaeroidinellus subdehiscens</td>
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<td>Histioglena sp.</td>
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<td>Globigerinoides ruber</td>
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<td>Globigerinoides carinamensis</td>
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<tr>
<td>Globorotalia cf. subrotacea</td>
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<tr>
<td>Globorotalia cf. inflata</td>
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### APPENDIX E.2. GRAPHIC SUMMARY OF PETROLEUM EXPLORATORY WELL REPORT FOR JS-3-1 WELL

#### MIOCENE

<table>
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<th>Depth (m)</th>
<th>Est. Lithology</th>
<th>Planktonic Foraminifera zones</th>
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<td>100-300</td>
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<td>Recrystallized</td>
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<td>Globigerinoides trilobus</td>
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#### PLIOCENE

<table>
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<tr>
<td></td>
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</tr>
<tr>
<td></td>
<td></td>
<td>Globigerinoides trilobus</td>
</tr>
</tbody>
</table>

#### Epoch

- **Leptocyclina anguiosa**
- **Leptocyclina boreensis**
- **Miogypsinia boreensis**
- **Miogypsinia cuthmanni s.l.**
- **Flasculinella globulosa**
- **Pseudorotalia**
- **Asterorotalia**
- **Globigerina paebullioide**
- **Globigerinoides trilobus**
- **Globorotella humerosa**
- **Globorotella plesiotumida**
- **Globigerinoides obliquus extremus**
- **Pulvinatina precurror**
- **Sphaerolinella dehiscens**
- **Globorotella pseudopina**
APPENDIX E.3. GRAPHIC SUMMARY OF PETROLEUM EXPLORATORY WELL REPORT FOR JS8-1 WELL.
### APPENDIX E.4.

**Graphical Summary of Petroleum Exploratory Well Report for JS10-1 Well.**

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<table>
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<tr>
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<td></td>
<td>Globigerinoides obliquus extremus</td>
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<td>Globorotalia cf. flexuosa</td>
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<td>Globorotalia plesirotumida</td>
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**Planktonic Foraminifers**
### Graphic Summary of Petroleum Exploratory Well Report for JS16-1 Well

#### MIOCENE

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<th>Est. Lithology</th>
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<td>Benthonic Foraminifera zones</td>
</tr>
<tr>
<td>600</td>
<td>Pseudorotalia spp.</td>
</tr>
<tr>
<td>500</td>
<td>Asterorotalia puechella</td>
</tr>
<tr>
<td>400</td>
<td>Operculina spp.</td>
</tr>
<tr>
<td>300</td>
<td>Cyclolypeus carteri</td>
</tr>
<tr>
<td></td>
<td>Operculina spp.</td>
</tr>
<tr>
<td></td>
<td>Cyclolypeus gumbelianus</td>
</tr>
<tr>
<td></td>
<td>Globigerinoides ruber</td>
</tr>
<tr>
<td></td>
<td>Globigerinoides obliquus</td>
</tr>
<tr>
<td></td>
<td>Globigerinoides quadrilobatus</td>
</tr>
<tr>
<td></td>
<td>Orbulina spp.</td>
</tr>
<tr>
<td></td>
<td>Globigerinoides trilobus</td>
</tr>
<tr>
<td></td>
<td>Globigerinoides sacculifer</td>
</tr>
<tr>
<td></td>
<td>Sphaeroidinella dehiscens</td>
</tr>
<tr>
<td></td>
<td>Pullemnata obliqueloculata</td>
</tr>
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<td>Globorotalia cultrata</td>
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<td>Globorotalia menardii</td>
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<tr>
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</tr>
<tr>
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<td>Globorotalia cerasiformis</td>
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<tr>
<td></td>
<td>Hastigerina siphonfera</td>
</tr>
<tr>
<td></td>
<td>Globorotalia scitula</td>
</tr>
<tr>
<td></td>
<td>Biorbulina bilobata</td>
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#### PLIOCENE

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<td>400</td>
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<td>Globorotalia scitula</td>
</tr>
<tr>
<td></td>
<td>Biorbulina bilobata</td>
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</tbody>
</table>
APPENDIX F.1. COMPUTER PROGRAM TO DIGITISE SHIP POSITION ON MAP

1 REM Written by Susilohadi 1992, using BASIC
10 CLS : CLEAR
20 PRINT "PROGRAM TO DIGITISE UNKNOWN POINT ON MAP"
30 PRINT
40 PRINT "The program needs at least two reference points,"
50 PRINT "and maximum of five points"
60 PRINT
70 REM Digitise reference points
80 INPUT "File name of data : ", NF$
90 OPEN NF$ FOR APPEND AS #2
100 PRINT "Put map on digitiser tablet and begin to digitise the reference points"
110 DIM X(5), Y(5), P#(5), Q#(5)
120 FOR I = 1 TO 5
130 PRINT "Reference point no. : "; -
140 PRINT I
150 GOSUB 1160
155 IF C = 2 THEN 290
160 X(I) = A: Y(I) = B
170 BEEP
180 PRINT "Enter the coordinate :"
190 INPUT "Latitude (degree, minute, second) : ", DER, MIN, DET
195 Q#(I) = 3600000# - ((3600 * DER) + (60 * MIN) + DET)
200 IF I <> 2 THEN 290
210 PRINT "Longitude (degree, minute, second) : ", DER, MIN, DET
215 P#(I) = (3600 * DER) + (60 * MIN) + DET
220 NEXT I
230 IF C <= 1 THEN 330
240 BEEP: PRINT : PRINT " NOT ENOUGH REFERENCE POINT"
250 CLS : RUN
260 Z = I - 1: OZ = 0
270 REM Measure the distance ratio (pj) and angle (alpha)
280 FOR I = 1 TO Z
290 FOR J = (I + 1) TO Z
300 PJ# = PJ# / OZ
310 FOR I = 1 TO Z
320 FOR J = (I + 1) TO Z
330 OZ = OZ + 1
340 JPET# = ((P#(I) - P#(J)) ^ 2 + (Q#(I) - Q#(J)) ^ 2) ^ .5
350 JDIG# = ((X(I) - X(J)) ^ 2 + (Y(I) - Y(J)) ^ 2) ^ .5
360 IF (X(J) - X(I)) = 0 THEN SDTJ# =0
370 IF (Y(J) - Y(I)) = 0 THEN SDTJ# =0
380 GOTO 450
390 SDTJ# = ATN((Y(J) - Y(I)) / (X(J) - X(I)))
400 IF (Q#(J) - Q#(I)) = 0 THEN SETQP# =0
410 IF (P#(J) - P#(I)) = 0 THEN SETQP# =0
420 GOTO 490
430 ALPHA# = ALPHA# + (SDTJ# - SDT#)
440 NEXT J
450 NEXT I
460 PRINT " Coordinate reference points are :": A$; X(I); Y(I)
470 PRINT " longitude latitude"
480 PRINT FOR I = 1 TO Z
490 A$ = " "
500 B$ = ": xxx "
510 CS = " "
520 PRINT USING AS; X(I); Y(I);
530 OT# = P#(I): GOSUB 1220
540 PRINT USING BS; DJ; MT; DT;
550 PRINT USING CS; DJ; MT; DT;
560 NEXT I
570 PRINT : PRINT
580 REM Determination of arbitrary points
590 PRINT " Begin to digitise arbitrary points"
600 GOSUB 1160
610 IF C = 2 THEN 1130
620 K = K + 1
630 IF I = 1 TO Z
640 JTTK# = ((A - X(I)) ^ 2 + (B - Y(I)) ^ 2) ^ .5 * PJ#
650 IF B = Y(I) AND A = X(I) THEN SDTJK# = 0: GOTO 780
CONTINUED

770  SDTTK# = ATN((B - Y(I)) / (A - X(I)))
780  SPTTK# = SDTTK# - ALPHA#
790  DLAT# = ABS(JTTK# * SIN(SPTTK#))
800  DLON# = ABS(JTTK# * COS(SPTTK#))
810  IF A <= X(I) AND B > Y(I) THEN 910
820  IF A <= X(I) AND B <= Y(I) THEN 850
830  IF A > X(I) AND B > Y(I) THEN 940
840  IF A > X(I) AND B <= Y(I) THEN 880
850  OLAT# = Q#(I) - DLAT#
860  OLON# = P#(I) - DLON#
870  GOTO 960
880  OLAT# = Q#(I) - DLAT#
890  OLON# = P#(I) + DLON#
900  GOTO 960
910  OLAT# = Q#(I) + DLAT#
920  OLON# = P#(I) + DLON#
930  GOTO 960
940  OLAT# = Q#(I) + DLAT#
950  OLON# = P#(I) + DLON#
960  LAT# = LAT# + OLAT#
970  LON# = LON# + OLON#
980  NEXT I
990  LAT# = LAT# / Z
1000 LON# = LON# / Z
1010 S# = 324000# - LAT#
1020 R# = LON#
1030 PRINT "Coordinate point no. : ";
1040 PRINT K;
1050 PRINT " is : ";
1060 OT# = R#: GOSUB 1220
1070 PRINT USING BS$; DJ; MT; DT;
1080 OT# = S#: GOSUB 1220
1090 PRINT USING CS$; DJ; MT; DT;
1100 PRINT
1120 REM Digitiser control
1160 OPEN "COM1:9600,N,7,2,CS,DS" FOR RANDOM AS #1
1165 PRINT "Press red button when finish"
1170 PRINT #1, "P";
1180 INPUT #1, A, B, C
1200 CLOSE #1
1210 RETURN
1220 REM Change to degree format
1230 DJ = INT(OT# / 3600)
1240 MT = INT((OT# - (DJ * 3600)) / 60)
1250 DT = OT# - (3600 * DJ) - (60 * MT)
1260 RETURN
APPENDIX F.2. COMPUTER PROGRAM TO DIGITISE SEISMIC SECTION

CLS : CLEAR
DIM in(4)
PRINT "*** COMPUTER PROGRAM TO DIGITISE SEISMIC SECTION ***";
PRINT "Written by Susilohadi in 1992, using QBASIC"
PRINT "FUNCTION OF BUTTONS ON THE DIGITISING PUCK :"
PRINT > Button no. 1 (red) to digitise seismic horizon"
PRINT > Button no. 3 (blue) to change seismic sector to be digitised"
PRINT > Button no. 4 (white) to quit"
PRINT INPUT "File name of data : ",nf$
INPUT "Interval of digitising (hour,minute,second) : ",jtb,mtb,dtb
PRINT INPUT "Moving to the left (i) of right (a) : ",tb$
PRINT PRINT "PUT SEISMIC SECTION ON THE DIGITISING TABLET"
PRINT PRINT 1400 REM Subroutining data entry
REM variables: jsi=left horizontal distance
REM jsa=right horizontal distance
REM xas=abiss at zero point
REM yas=ordinat at zero point
REM js=average horizontal distance
REM s=ship time travel
PRINT " DIGIT SEISMIC BORDERS :"
1410 GOSUB 1300
IF z = 2 OR z = 3 THEN 1410
IF z = 4 THEN END
nv = y - (mv * x)
xas = (mv * xas) + n
yas = (mv * yas) + n
jsi = ((((xas - xai) ^ 2) + ((yas - yai) ^ 2)) ^ .5) * jh
jsa = ((((xaa - xas) ^ 2) + ((yaa - yas) ^ 2)) ^ .5) * jh
jsi = jsi + dw#
jsa = -jsa + dk#
js = INT((jsi + jsa) / 2)
PRINT " Point position (hour:minute:second) : ";
jj = INT(js / 3600)
mm = INT((js - (jj * 3600)) / 60)
dd = INT(js - (jj * 3600) - (mm * 60))
PRINT jj;
PRINT mm;
PRINT dd;
PRINT " Time travel (TWT) : ";
s = (((xas - x)^2 + (yas - y)^2)^.5) * jv
PRINT USING "##.##";s;
PRINT " msec."
OPEN nf$ FOR APPEND AS #2
PRINT #2, USING "####.####",";js,s
CLOSE #2
GOTO 1420
1100 REM subroutine to pick values from digitiser
REM variables: x=abiss; y=ordinat; z=button number
OPEN "com1:9600,n,8,1" AS #1
PRINT #1,chr$(0)
PRINT #1,"B"
FOR i=0 to 4
in(1)=asc(INPUT$(1,1))
next i
z=7 and in(0)
x=128*(127 and in(2))+(127 and in(1))
y=128*(127 and in(4))+(127 and in(3))
CLOSE #1
RETURN
1300 REM subroutine to change sector to be digitised
REM (actified by button 2 or 3 on the digitiser puck)
REM variables earlier border : jw=hour; mw=minute; dw#=second
CONTINUED

REM  xai=abscis minimum; yai=ordinat minimum
REM  xbi=abscis maximum; ybi=ordinat maximum
REM variables later border: jk=hour; mk=minute; dk#=second
REM  xaa=abscis minimum; yaa=ordinat minimum
REM  xba=abscis maximum; yba=ordinat maximum
REM other variables: jh=horizontal unit; jv=vertikal unit
REM  nk=minimum scale value; nb=maximum scale value
IF z = 2 and tb$ = "a" or z = 2 and tb$ = "A" THEN
  jw = jk:mw = mk:dw = dk
  jk = jk + jtb:mk = mk + mtb:dk = dk + dtb
  IF dk >= 60 THEN mk = mk + 1:dk = dk - 60
  IF mk >= 60 THEN jk = jk + 1:mk = mk - 60
  GOTO 1320
END IF
IF z = 2 and tb$ = "i" or z = 2 and tb$ = "I" THEN
  jk = jw:mk = mw:dk = dw
  jw = jw + jtb:mw = mw + mtb:dw = dw + dtb
  IF dw >= 60 THEN mw = mw + 1:dw = dw - 60
  IF mw >= 60 THEN jw = jw + 1:mw = mw - 60
  GOTO 1320
END IF
1310 INPUT " Type in hour,minute,second left side of the section : ", jw, mw, dw
1320 PRINT " Batas kiri : ";jw;":";mw;":";dw
PRINT " Batas kanan : ";jk;":";mk;":";dk
PRINT " Digit TWT minimum and maximum left side"
GOSUB 1100:BEEP
  xai = x: yai = y
GOSUB 1100:BEEP
  xbi = x: ybi = y
PRINT " Digit TWT minimum and maximum right side"
GOSUB 1100:BEEP
  xba = x: yaa = y
GOSUB 1100:BEEP
  xba = x: yba = y
mh = (yaa - yai) / (xaa - xai)
mh = mh + ((yba - ybi) / (xba - xbi))
mh = mh / 2
mv = -1 / mh
n = ((yai - mh * xai) + (yaa - mh * xaa)) / 2
INPUT " TWT minimum : ", nk
INPUT " TWT maximum : ", nb
jv = (nb - nk) / ((((xbi - xai) ^ 2) + ((ybi - yai) ^ 2))) ^ .5
jv = jv + ((nb - nk) / ((((xba - xaa) ^ 2) + ((yba - yaa) ^ 2))) ^ .5))
jv = jv / 2
jh = (dk# - dw#) / ((((xaa - xai) ^ 2) + ((yaa - yai) ^ 2))) ^ .5
jh = jh + ((dk# - dw#) / ((((xba - xbi) ^ 2) + ((yba - ybi) ^ 2))) ^ .5))
jh = jh / 2
RETURN
APPENDIX F.3. COMPUTER PROGRAM TO CONVERT GRIDDED DATA TO 'SURFER' COMPATIBLE FORMAT (ASCII)

```vbnet
DIM z(13000)
PRINT "PROGRAM TO CONVERT GRIDDED DATA"
PRINT "FROM XYZ FORMAT (ASCII) TO 'SURFER' (ASCII) COMPATIBLE FORMAT"
PRINT 'Written by Susilohadi, 1992, using QBASIC'
PRINT
'INPUT "Name of file to convert : ", nfu$
'INPUT "New file name : ", nfh$
OPEN nfu$ FOR INPUT AS #1
  INPUT #1, x
  INPUT #1, y
  INPUT #1, zz
CLOSE #1
xspc = x
yspc = y
OPEN nfu$ FOR INPUT AS #1
  spacel:
  INPUT #1, x
  INPUT #1, y
  INPUT #1, zz
  IF xspc = x AND (NOT EOF(1)) THEN GOTO spacel
  xspc = ABS(xspc - x)
CLOSE #1
OPEN nfu$ FOR INPUT AS #1
  space2:
  INPUT #1, x
  INPUT #1, y
  INPUT #1, zz
  IF yspc = y AND (NOT EOF(1)) THEN GOTO space2
  yspc = ABS(yspc - y)
CLOSE #1
OPEN nfu$ FOR INPUT AS #1
  INPUT #1, xmin
  INPUT #1, ymin
  INPUT #1, zmin
  zmin = -zmin
  zmax = zmin
CLOSE #1
i = 1
OPEN nfu$ FOR INPUT AS #1
  WHILE (NOT EOF(1))
    INPUT #1, x
    INPUT #1, y
    INPUT #1, z(i)
    'z(i) = -z(i)
    PRINT USING "#######.## #######.## ####.##"; x; y; z(i)
    IF xmin >= x THEN xmin = x ELSE xmin = xmin
    IF ymin >= y THEN ymin = y ELSE ymin = ymin
    IF zmin >= z(i) THEN zmin = z(i) ELSE zmin = zmin
    IF xmax <= x THEN xmax = x ELSE xmax = xmax
    IF ymax <= y THEN ymax = y ELSE ymax = ymax
    IF zmax <= z(i) THEN zmax = z(i) ELSE zmax = zmax
    i = i + 1
  WEND
CLOSE #1
PRINT xmin, xmax
PRINT ymin, ymax
PRINT zmin, zmax
xnum = ((xmax - xmin) / xspc) + 1
ynum = ((ymax - ymin) / yspc) + 1
OPEN nfh$ FOR OUTPUT AS #2
PRINT "DSAA"
PRINT USING "#### ####"; xnum; ynum
PRINT USING "#### ####"; xmin; xmax
PRINT USING "#### ####"; -ymax; -ymin
PRINT "DSAA"
PRINT #2, USING "#### ####"; xnum; ynum
PRINT #2, USING "#### ####"; xmin; xmax
PRINT #2, USING "#### ####"; -ymax; -ymin
PRINT #2, USING "#### ####"; xspc; yspc
```

CONTINUED

PRINT #2, USING "####.## ####.##"; zmin; zmax
'PRINTING AND SAVING
FOR k = ynum TO 1 STEP -1
  FOR 1 = 1 TO xnum
    PRINT USING "####.##"; z((xnum * (k - 1)) + 1);
    PRINT #2, USING "####.##"; z((xnum * (k - 1)) + 1);
    IF (1 / 10) - INT(1 / 10) = 0 THEN PRINT : PRINT #2, ;
  NEXT 1
  PRINT : PRINT
  PRINT #2,
  PRINT #2,
NEXT k
CLOSE #2
BEep
END
APPENDIX F.4. COMPUTER PROGRAM FOR GRIDDING SEISMIC DATA

//Written by Susilohadi 1992, using C++ language
#include <stdio.h>
#include <math.h>
#include <conio.h>
#include <iostream.h>
#include <ctype.h>
#include <stdlib.h>
#include <dos.h>

unsigned int lon[6000]; //longitude saved as integer of (xaxis-xmin)*10
int lat[6000]; //latitude saved as integer of (yaxis-ymin)*10
int depth[6000], //depth saved as integer of (zaxis-zmin)*10
i, //number of data point
a, b, c; //dummy variables
float xmin, //longitude minimum on file
xmax, //longitude maximum on file
ymin, //latitude minimum on file
ymax, //latitude maximum on file
zmin, //depth minimum on file
zmax, //depth maximum on file
x_spc, //grid spacing on x-axis
y_spc, //grid spacing on y-axis
power, //power, weight
xminest, yminest;
float distsort[64], zsort[64];

char
inpfile[50], //file name of data to be retrieved
outfile[50]; //file name of gridded data to be saved

FILE *file_in, //for ascii x,y,z file to be retrieved
*file_out; //for ascii x,y,z grided data to be saved

void heading(void);
void subheading1(void);
void subheading2(void);
void getfilename();
void readmaxmin();
void readdataO;
void prtmaxmin();
void grdspc();
float distance(float, float, float, float);
float xrange(float, float, float, float);
float yrange(float, float, float, float);
float truex(int);
float truey(int);
float truez(int);
void readmaxmin();
void saving(float, float, float);
void rectangular(float, float, float, float);
void ellipsoidal(float, float, float, float);
void area();
void closestpoint();
void freefind(int);
void quadrantfind(int);
void octantfind(int);
void notes(void);
void beep(void);

void main()
{
    int selectmenu;
    start:
    clrscr();
CONTINUED

heading();
menuopt:
selectmenu = getche();
switch(selectmenu)
{
    case '1': area();
        break;
    case '2': closestpoint();
        break;
    case '3': exit(1);
        default : beep();
        goto menuopt;
}

gotoxy(15, 25);
printf("CALULATION COMPLETED, PRESS ANY KEY");
getch();
goto start;

//estimate value based on points in the given area
void area()
{
    int search, azim;
    float mayor, minor;
    clrscr();
    subheadingl();
    getfilename();
    readmin();
    prtmmin();
    grdspc();
    gotoxy(53, 16);
    search = toupper(getche());
    // printf(search);
    gotoxy(44, 17)
    cin >> minor;
    gotoxy(44, 18)
    cin >> mayor;
    gotoxy(44, 19)
    cin >> azim;
    readdata();
    if (search == 'R')
    {
        rectangular(mayor, minor, azim);
    }
    else
    {
        ellipsoidal(mayor, minor, azim);
    }
    return;
}

//estimate z based on rectangular area of searching
void rectangular(float mayor, float minor, int azim)
{
    int pointh = 0;
    float x_est, //longitude of estimated point
         y_est, //latitude of estimated point
         x_point, //longitude of known point
         y_point, //latitude of known point
         dist_P, //distance between estimated and known point
         dist_R, //maximum distance of known point allowed
         x, y,
         truelon, truelat, truedepth, vdepth, wdepth, value,
         distance_P, distance_R[5], xe, ye,
         n_P, me[4], X_R, y_R,
         n_P, n_horiz, n_vert, ne[5];

    me[1] = me[2] = tan(azim * 3.14159/180);
    for(y = ymin; y <= ymax; y += y_spc)
    {
        for(x = xmin; x <= xmax; x += x_spc)
        {
            ++pointh;
CONTINUED

vdepth = 0;
wdepth = 0;
value = 0;
switch(azim)
{
    case 0: break;
    case 90: break;
    case -90: break;
    default:
        n_horiz = y - (me[1] * x);
        n_vert = y - (me[3] * x);
        ne[1] = n_horiz - (minor / cos(azim * 3.14159 / 180));
        ne[3] = n_vert - (mayor / sin(azim * 3.14159 / 180));
        ne[4] = n_vert + (mayor / sin(azim * 3.14159 / 180));
    }
for (a = 1; a != i; ++a)
{
    gotoxy(17, 21);
    printf("%d",a);
    truelon = truey(a);
    truelat = truey(a);
    truedepth = truez(a);
    distance_P = distance(x, truelon, y, truelat);
    if (distance_P == 0)
    {
        vdepth = truedepth;
        wdepth = 1;
        break;
    }
    if (truelon != x)
    {
        m_P = (truelat - y) / (truelon - x);
        n_P = y - (m_P * x);
        switch(azim)
        {
            case 0:
                if (truelon == x)
                {
                    distance_R[1] = minor;
                }
                else
                {
                    distance_R[1] = mayor / cos(atan(m_P));
                    distance_R[2] = minor / cos((90 * 3.14159 /180) - atan(m_P));
                    if (distance_R[1] >= distance_R[2])
                    {
                        distance_R[1] = distance_R[2];
                    }
                }
                if (distance_P <= distance_R[1])
                {
                    vdepth += truedepth/(pow(distance_P, power));
                    wdepth += 1/(pow(distance_P, power));
                }
                break;
            case 90:
                if (truelon == x)
                {
                    distance_R[1] = mayor;
                }
                else
                {
                    distance_R[1] = minor/ cos(atan(m_P));
                    distance_R[2] = mayor/ cos((90 * 3.14159 /180) - atan(m_P));
                    if (distance_R[1] >= distance_R[2])
                    {
                        distance_R[1] = distance_R[2];
                    }
                }
                if (distance_P <= distance_R[1])
                {
                    vdepth += truedepth/(pow(distance_P, power));
                }
        }
    }
}
CONTINUED

wdepth += 1/(pow(distance_P, power));
}
break;
case -90:
    if (truelon == x)
    {
        distance_R[1] = mayor;
    }
    else
    {
        distance_R[1] = minor/ cos(atan(m_P));
        distance_R[2] = mayor/ cos((90 * 3.14159 /180) - atan(m_P));
        if (distance_R[1] >= distance_R[2])
            distance_R[1] = distance_R[2];
    }
    if (distance_P <= distance_R[1])
    {
        vdepth += truedepth/(pow(distance_P, power));
        wdepth += 1/(pow(distance_P, power));
    }
    break;
default :
    if (truelon == x)
    {
        distance_R[1] = mayor / cos((90-azim) * 3.14159 /180);
        distance_R[2] = minor / cos(azim * 3.14159 / 180);
        if (distance_R[1] >= distance_R[2])
            distance_R[1] = distance_R[2];
    }
    else
    {
        for (b = 1; b <= 4; ++b)
        {
            x_R = xrange(m_P, n_P, me[b], ne[b]);
            y_R = yrange(m_P, n_P, me[b], ne[b]);
            distance_R[b] = distance(x, x_R, y, y_R);
            if (distance_R[1] >= distance_R[b])
                distance_R[1] = distance_R[b];
        }
    }
    if (distance_P <= distance_R[1])
    {
        vdepth += truedepth/(pow(distance_P, power));
        wdepth += 1/(pow(distance_P, power));
    }
}
if (vdepth == 0 & wdepth == 0)
{
    value = 0;
}
else
{
    value = vdepth / wdepth;
}
gotoxy(14, 23);
printf("%d", pointh);
gotoxy(26, 23);
printf("%7.2f", x);
gotoxy(42, 23);
printf("%7.2f", y);
gotoxy(58, 23);
printf("%5.2f", value);
saving(x, y, value);
}

//estimate z based on ellipsoidal area of searching
void ellipsoidal(float mayor, float minor, float azim)
{
    int pointh = 0;
    float dist_P, //distance between estimated and known point
    dist_R, //maximum distance of known point allowed
    x, y, truelon, truelat, truedepth, vdepth, wdepth, value,
    distance_P, distance_R, x_R, y_R, m_P, n_P;
    for(y = ymin; y <= ymax; y += y_spc)
    {
        for(x = xmin; x <= xmax; x += x_spc)
            ++pointh;
        vdepth = wdepth = value = 0;
        for (a = 1; a <= i; ++a)
        {
            truelon = truelat = truedepth = 0;
            gotoxy(17, 21);
            printf("%d", a);
            truelon = truex(a);
            truelat = truey(a);
            truedepth = truez(a);
            distance_P = distance(x, truelon, y, truelat);
            if (distance_P == 0)
            {
                vdepth = truedepth;
                wdepth = 1;
                break;
            }
            if (truelon == x)
            {
                y_R = minor * sin((90 - azim) * 3.14159 / 180);
                x_R = mayor * cos((90 - azim) * 3.14159 / 180);
            }
            else
            {
                m_P = (truelat - y) / (truelon - x);
                x_R = mayor * cos((azim * 3.14159 / 180) + atan(m_P)
                y_R = minor * sin((azim * 3.14159 / 180) + atan(m_P));
            }
            distance_R = distance(0, x_R, 0, y_R);
            if (distance_P <= distance_R )
            {
                vdepth += truedepth/(pow(distance_P, power));
                wdepth += 1/(pow(distance_P, powe));
            }
        }
        if (vdepth == 0 & wdepth == 0)
            value = 0;
        else
        {
            value = vdepth / wdepth;
        }
        gotoxy(14, 23);
        printf("%d", pointh);
        gotoxy(26, 23);
        printf("%7.2f", x);
        gotoxy(42, 23);
        printf("%7.2f", y);
        gotoxy(58, 23);
        printf("%5.2f", value);
        saving(x, y, value);
    }
}

//estmate z based on closest point values
void closestpoint()
{
    int numpoint, search;
    float dist_P, //distance between estimated and known point
    x, y, truelon, truelat, truedepth, vdepth, wdepth, value;
    clrscr();
CONTINUED

    subheading2();
    getfilename();
    readmaxmin();
    prtmaxmin();
    rdspc();
    repeatcall1;
    gotoxy(52, 16);

    search = toupper(getch());
    /* if(search != 'F' || search != 'Q' || search != 'O')
    
    beep();
    goto repeatcall1;
    */
    repeatcall2:
    gotoxy(37, 17);
    cin >> numpoint;
    if(numpoint >= 64 & search == 'F')
    {
        notes();
        beep();
        goto repeatcall2;
    }
    if(numpoint >= 16 & search == 'Q')
    {
        notes();
        notes();
        goto repeatcall2;
    }
    if(numpoint >= 8 & search == 'O')
    {
        notes();
        beep();
        goto repeatcall2;
    }
    readdata();
    switch(search)
    {
    case 'Q': quadrantfind(numpoint);
        break;
    case 'O': octantfind(numpoint);
        break;
    default : freefind(numpoint);
    }

    //~~~~~~~~~~~~~~~~~~~~~~~~~~~~~~~~~~~~~~~~~~~~~~~~~~~~~~~~~~~~~~~~~~~~~~~~~~~~~~~~~~~~~~~~~
    void freefind(int numpoint)
    {
        int pointh = 0;
        float truelon, truelat, truedepth,
            x, y, dist_P, vdepth, wdepth, value;
        --numpoint;
        for(a = 0; a <= numpoint; ++a)
        {
            truelon = truelat = truedepth = 0;
            truelon = truex(a);
            truelat = truey(a);
            zsort[a] = truez(a);
            distsortfa] = distance(x, truelon, y, truelat);
        }
        sorter(0, numpoint);
        for(y = ymin; y <= ymax; y += y_spc)
        {
            for(x = xmin; x <= xmax; x += x_spc)
            {
                ++pointh;
                vdepth = wdepth = value = 0;
                for(a = 1; a <= i; ++a)
                {
                    truelon = truelat = truedepth = 0;
                    gotoxy(17, 21);
                    printf("%d", a);
                    truelon = truex(a);
                    truelat = truey(a);
                    }
truedepth = truez(a);
dist_P = distance(x, truelon, y, truelat);
if(dist_P == 0)
{
    vdepth = trudepth;
    wdepth = 1;
    break;
}
if(dist_P <= distsort[numpoint])
{
    distsort[numpoint] = dist_P;
    zsort[numpoint] = truedepth;
}
sorter(0, numpoint);
if(dist_P != 0)
{
    for(a = 0; a <= numpoint; ++a)
    {
        vdepth += zsort[a]/(pow(distsort[a], power));
        wdepth += 1/(pow(distsort[a], power));
    }
    value = vdepth / wdepth;
    gotoxy(14, 23);
    printf("%d", pointh);
    gotoxy(26, 23);
    printf("%7.2f", x);
    gotoxy(42, 23);
    printf("%7.2f", y);
    gotoxy(58, 23);
    printf("%5.2f", value);
    saving(x, y, value);
}
return;

//estimate z based on quadrant method
void quadrantfind(int numpoint)
{
    int sect[4];
    int num, numr, numint;
    int pointh = 0;
    float phi4 = 90;
    float truelon, truelat, truedepth,
    x, y, dist_P, vdepth, wdepth, value;
    float deltax, deltay;
    float angle, numfloatl, numfloat2;
    --numpoint;
    for(y = ymin; y <= ymax; y += y_spc)
    {
        for(x = xmin; x <= xmax; x += x_spc)
        {
            ++pointh;
            vdepth = wdepth = value = 0;
            for(a = 1; a <= i; ++a)
            {
                truelon = truelat = truedepth = 0;
                gotoxy(17, 21);
                printf("%d", a);
                truelon = truex(a);
                truelat = truey(a);
                truedepth = truez(a);
                dist_P = distance(x, truelon, y, truelat);
                if(dist_P == 0)
                {
                    vdepth = trudepth;
                    wdepth = 1;
                    break;
                }
                deltax = (truelon - x);
                deltay = (truelat - y);
                if(deltay >= 0)
int sort1, sort2;
float opsortd, opsortz;

for(sort1 = begin; sort1 <= end; ++sort1)
{
    for(sort2 = sort1; sort2 <= end; ++sort2)
    {
        if(distsort[sort1] >= distsort[sort2])
        {
            opsortd = distsort[sort1];
            opsortz = zsort[sort1];
            distsort[sort1] = distsort[sort2];
            distsort[sort2] = opsortd;
            zsort[sort1] = zsort[sort2];
            zsort[sort2] = opsortz;
        }
    }
}
return;

//print heading
void heading(void)
{
    printf("GRIDDING PROGRAM USING INVERSE DISTANCE METHOD:

1. Area
2. Closest Points
3. Quit

Note: <> Data must be in ASCII form -> X Y Z.

Press the option key:");
}

//print subheading1 (area)
void subheading1(void)
{
    gotoxy(20, 1);
    printf("INVERSE DISTANCE METHOD (AREA SEARCH)\n\n");

    printf(" File name of data to be retrieved : \n\n");
    printf(" File name of gridded data : \n\n");
    printf(" The program found that maximum of: \n\n");
    printf(" X = and Y =\n\n");
    printf(" minimum of:\n\n");
    printf(" X = and Y =\n\n");
    printf(" Grid spacings on x-axis : , and on y-axis :\n\n");
    printf(" Top left coordinate: X = and Y =\n\n");
    printf(" Bottom right coordinate: X = and Y =\n\n");
    printf(" Number of point to be estimated =\n\n");
    printf(" Weighting, power of distance :\n\n");
    printf(" Searching pattern (R)ectangular or (E)llipsoidal :\n\n");
    printf(" Length of minor axis in unit :\n\n");
    printf(" Length of mayor axis in unit :\n\n");
    printf(" Azimuth of minor axis (± 90°, integer):\n\n");
    printf(" Calculation ->\n\n");
    printf(" Results:\n\n");
    printf(" point no. = ; X = ; Y = ; Z =\n\n");
}

//print subheading2 (closest point)
void subheading2(void)
{
    gotoxy(20, 1);
    printf("INVERSE DISTANCE METHOD (CLOSEST POINT)\n\n");
u = 0;

CONTINUED
v = 0;
for (q = ymin; q <= ymax; q += y_spc)
{
    ++v;
    u = 0;
    for (p = xmin; p <= xmax; p += x_spc)
    {
        ++u;
        ++pointnum;
    }
}
goxy(37, 14);
printf("%d", pointnum);
goxy(33, 15);
cin >> power;
return;
}

float xrange(float m_xrange, float n_xrange, float me_xrange, float ne_xrange)
return((n_xrange - ne_xrange)/(me_xrange - m_xrange));

float yrange(float m_yrange, float n_yrange, float me_yrange, float ne_yrange)
return(((n_yrange - ne_yrange)/(me_yrange - m_yrange)) * me_yrange) +
ne_yrange));

//read file names
void getfilename(void)
{
goxy(37, 3);
gets(inpfile);
goxy(37, 4);
gets(outfile);
}

//save results
void saving(float absis, float ordinate, float height)
{
    file_out = fopen(outfile, "a");
    fprintf(file_out,"%7.2f %7.2f %5.2f\n", absis, ordinate, height);
    fclose(file_out);
    return;
}

//calculate the real values of x
float truex(int idx)
{
    float valuex;
    valuex = (lon[idx] / 10) + xminest;
    return(valuex);
}

//calculate the real value of y
float truey(int idx)
{
    float valuey;
    valuey = (lat[idx] / 10) + yminest;
    return(valuey);
}

//calculate the real value of z
float truez(int idx)
{
    float valuez;
    valuez = (depth[idx] / 10) + zmin;
    return(valuez);
}

//sort array to ascending order
void sorter(int begin, int end)
CONTINUED

fscanf(file_in, "%f\n", &zaxis);
if (zmax <= zaxis)
    zmax = zaxis;
else
    zmax = zmax;
if (zmin >= zaxis)
    zmin = zmin;
else
    zmin = zmin;
}
xminest = xmin;
yminest = ymin;
fclose(file_in);
return;
}
// ==============================================================
void readdata()
{
    float xaxis, yaxis, zaxis, mayor;
gotoxy(4, 20);
printf("Extend area to ":");
cin >> mayor;
i = 1;
file_in = fopen(inpfile, "r");
while(!feof(file_in))
{
    fscanf(file_in, "%f\n", &xaxis);
    fscanf(file_in, "%f\n", &yaxis);
    fscanf(file_in, "%f\n", &zaxis);
    if (xaxis <= (xmax + mayor) & yaxis <= (ymax + mayor)
        & xaxis >= (xmin - mayor) & yaxis >= (ymin - mayor))
    {
        lon[i] = (xaxis - xminest) * 10;
        lat[i] = (yaxis - yminest) * 10;
        depth[i] = (zaxis - zmin) * 10;
        i++;
    }
}
fclose(file_in);
return;
}
// ==============================================================
void prtmxmn(void)
{
    gotoxy(8, 7);
    printf("%.2f", xmax);
    gotoxy(28, 7);
    printf("%.2f", ymax);
    gotoxy(8, 9);
    printf("%.2f", xmin);
    gotoxy(28, 9);
    printf("%.2f", ymin);
}
// ==============================================================
//read and calculate grid spacing
void grdspc()
{
    int pointnum;
    int u, v;
    float p, q;
gotoxy(28, 11);
cin >> x_spc;
gotoxy(31, 11);
cin >> y_spc;
gotoxy(31, 12);
cin >> xmin;
gotoxy(47, 12);
cin >> ymin;
gotoxy(31, 13);
cin >> xmax;
gotoxy(47, 13);
cin >> ymax;
pointnum = 0;
else

{  
  if (dist_P <= distsort[num*8+numpoint])
      
      distsort[num*8+numpoint] = dist_P;
      
      zsort[num*8+numpoint] = truedepth;
      
      sorter(num*8, num*8+numpoint);
      
      }

}

if (dist_P != 0)
{
  for (num = 0; num <= 7; ++num)
      
      for (a = 0; a <= sect[num]; ++a)
      
          vdepth += zsort[num*8+a]/(pow(distsort[num*8+a], power));

          wdepth += 1/(pow(distsort[num*8+a], power));

      }

  value = vdepth / wdepth;

  gotoxy(14, 23);
  printf("%d", pointh);
  gotoxy(26, 23);
  printf("%7.2f", x);
  gotoxy(42, 23);
  printf("%7.2f", y);
  gotoxy(58, 23);
  printf("%5.2f", value);

  saving(x, y, value);
}

// sub-routine for distance calculation
float distance(float x1, float x2, float y1, float y2)
{
  float x = x2-xl;
  float y = y2-yl;
  float xy = pow(x, 2) + pow(y, 2);
  return(sqrt(xy));
}

// find maximum and minimum values of data
void readmaxmin()
{
  float xaxis, yaxis, zaxis;

  file_in = fopen(infile, "r");
  xmin = 648000;
  ymin = 648000;
  while (!feof(file_in))
  {
      fscanf(file_in, "%f\n", &xaxis);
      if (xmax <= xaxis)
          xmax = xaxis;
      else
          xmax = xmax ;
      if (xmin >= xaxis)
          xmin = xaxis ;
      else
          xmin = xmin ;
      fscanf(file_in, "%f\n", &yaxis);
      if (ymax <= yaxis)
          ymax = yaxis ;
      else
          ymax = ymax ;
      if (ymin >= yaxis)
          ymin = yaxis ;
      else
          ymin = ymin ;
  }
}
CONTINUED

```c
vdepth += zsort[num*16+a]/(pow(distsort[num*16+a], power));
wdepth += 1/(pow(distsort[num*16+a], power));

value = vdepth / wdepth;
```

```c
// gotoxy(51,1);
// printf("vdepth=%f; wdepth=%f",vdepth,wdepth);

gotoxy(14, 23);
printf("%d",pointh);

gotoxy(26, 23);
printf("%7.2f", x);

gotoxy(42, 23);
printf("%7.2f", y);

gotoxy(58, 23);
printf("%5.2f", value);

saving(x, y, value);
```

```c
//estimate z based on octant method
void octantfind(int numpoint)
{
    int sect[8], num;
    int pointh = 0;
    float phi8 = 0.78539;
    float truelon, truelat, truedepth, x, y, dist_P, vdepth, wdepth, value;
    float tann;
    double angle;

    --numpoint;
    for(y = ymin; y <= ymax; y += y_sp)
    {
        for(x = xmin; x <= xmax; x += x_sp)
        {
            ++pointh;
            vdepth = wdepth = value = 0;
            for(a = 1; a <= i; ++a)
            {
                truelon = truelat = truedepth = 0;
                gotoxy(17, 21);
                printf("%d", a);
                truelon = truey(a);
                truelat = truey(a);
                truedepth = truez(a);
                dist_P = distance(x, truelon, y, truelat);
                if (dist_P == 0)
                {
                    vdepth = truedepth;
                    wdepth = 1;
                    break;
                }
                if(truelon == x)
                {
                    angle = phi8;
                }
                else
                {
                    tann = (truelon - x) / (truelat - y);
                    angle = atan(tann);
                }
                for(num = 0; num <= 7; ++num)
                {
                    if(angle >= (num * phi8) & angle < ((num+1) * phi8))
                    {
                        if(sect[num] < numpoint)
                        {
                            distsort[num*8+sect[num]] = dist_P;
                            zsort[num*8+sect[num]] = truedepth;
                            ++sect[num];
                        }
                    }
                }
            }
        }
    }
}
```

```
```
if(deltax >= 0)
{
    angle = fabs((float)atan(deltay/deltax));
} else
{
    angle = 180 - fabs((float)atan(deltay/deltax));
}
else
{
    if(deltax >= 0)
    {
        angle = 180 + fabs((float)atan(deltay/deltax));
    } else
    {
        angle = 360 - fabs((float)atan(deltay/deltax));
    }
// angle = angle * 180 / 3.14158;
// gotoxy(1,1);
// printf("angle=%f",angle);
for(num = 0; num <= 3; ++num)
{
    if(angle >= (num * phi4) & angle < ((num+1) * phi4))
    {
        printf("dist=%f; angle=%5.2f",dist_P,(float)angle);
        gotoxy(1,1);
        printf("sectnum: %d",sect[num]);
        if(dist_P < numpoint)
        {
            distsort[num*16+sect[num]] = dist_P;
            zsort[num*16+sect[num]] = truedepth;
            printf("dist=%f; z=%f",dist_P,truedepth);
            printf("dist=%f; z=%f",distsort[num*16+sect[num]],zsort[num*16+sect[num]]);
            ++sect[num];
        } else
        {
            sorter(num*16, num*16+numpoint);
            for(numr = 0; numr <= numpoint; ++numr)
            {
                numfloatl = fabs(distsort[num*16+numr] - dist_P);
                if(dist_P <= distsort[num*16+numpoint])
                {
                    distsort[num*16+numpoint] = dist_P;
                    zsort[num*16+numpoint] = truedepth;
                    gotoxy(1,1);
                    printf("dist=%f; z=%f",dist_P,zsort[num*16+numpoint]);
                    gotoxy(1,num*4+2);
                    printf("%3.1f-%3.1f: angle: %f; D: %f; Z: %f",(num* phi4),((num+1) * phi4),
                    (float)angle,dist_P,truedepth);
                }
                sorter(num*16, num*16+numpoint);
            }
        }
    }
}
if(dist_P != 0)
{
    for(num = 0; num <= 3; ++num)
    {
        for(a = 0; a <= sect[num]; ++a)
        {
            gotoxy(1,num+1);
            printf("D: %f; Z: %f",zsort[num*16+a],distsort[num*16+a]);
            printf("arr: %d",num*16+a);
            if(zsort[num*16+a] != 0 & distsort[num*16+a] != 0)
CONTINUED

printf(" File name of data to be retrieved : \n");
printf(" File name of gridded data : \n\n");
printf(" The program found that maximum of: \n");
printf(" X = and Y =\n");
printf(" minimum of: \n");
printf(" X = and Y =\n");
printf(" Grid spacings on x-axis : , and on y-axis :\n");
printf(" Top left coordinate: X = Y =\n");
printf(" Bottom right coordinate: X = Y =\n");
printf(" Number of point to be estimated: \n");
printf(" Weighting, power of distance :\n");
printf(" Searching pattern: (F)ree, (Q)uadrant, (O)ctant :\n");
printf(" Number of point in each sector :\n\n");
printf(" Calculation ->\n");
printf(" Results:\n");
printf(" point no. = ; X = ; Y = ; Z =\n");
}

//--print notes on error
void notes(void)
{
  gotoxy(5, 17);
  printf("Maximum number of point on:");
  gotoxy(5, 18);
  printf("Free search is 64, Quadrant search is 16 and Octant search is 8");
  delay(1500);
  gotoxy(5, 17);
  printf("\n");
  gotoxy(5, 18);
  printf("\n");
}

//--sound a beep on error
void beep(void)
{
  sound(880);
  delay(100);
  nosound();
  delay(100);
  sound(440);
  delay(100);
  nosound();
}
Volume 2 of this thesis contains seismic sections that are too large to be scanned. The seismic sections can be viewed in the Archives of the University of Wollongong Library.