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Late Cenozoic geohistory of NW Buru, Indonesia and plate tectonic implications

INDONESIAN-DUTCH SNELLIUS-II EXPEDITION
1984-1985

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Abstract

Mio-Pliocene deposits of Buru have been investigated in order to provide additional data concerning the timing, rate and magnitude of vertical movements in the northwestern end of the outer Banda Arc structure. Geohistory analysis of field and laboratory data of two sections recorded in NW Buru provide broad age – depth constraints only. Four episodes in the history of uplift and subsidence are distinguished and discussed in the light of present plate tectonic concepts.

Deposition of the Early Miocene Hotong Formation, bathyal sandy clays and marls with intercalated turbidites, followed after a mid Tertiary period of open folding, uplift, erosion and, finally, partial resubmergence. These changes may be related to plate tectonic interaction with southeast Asia and not necessarily with the Australian continent – from which Buru initially derived – which is now colliding with the Banda Arc. The Middle – Late Miocene (and possibly the Early Pliocene) left no sediments in Buru; it was a time of differential uplift (up to 20 cm/ka), related to reorganisation of the regional deformation pattern, i.e. evolution of the Banda Arc system. During a poorly dated Pliocene depositional episode shallow marine fan-delta clastics were laid down in giant prograding sets, up to 50 m high and dipping up to 20 degrees, thus witnessing continuing differential movements in NW Buru. The same pattern of deformation exists to the present day.

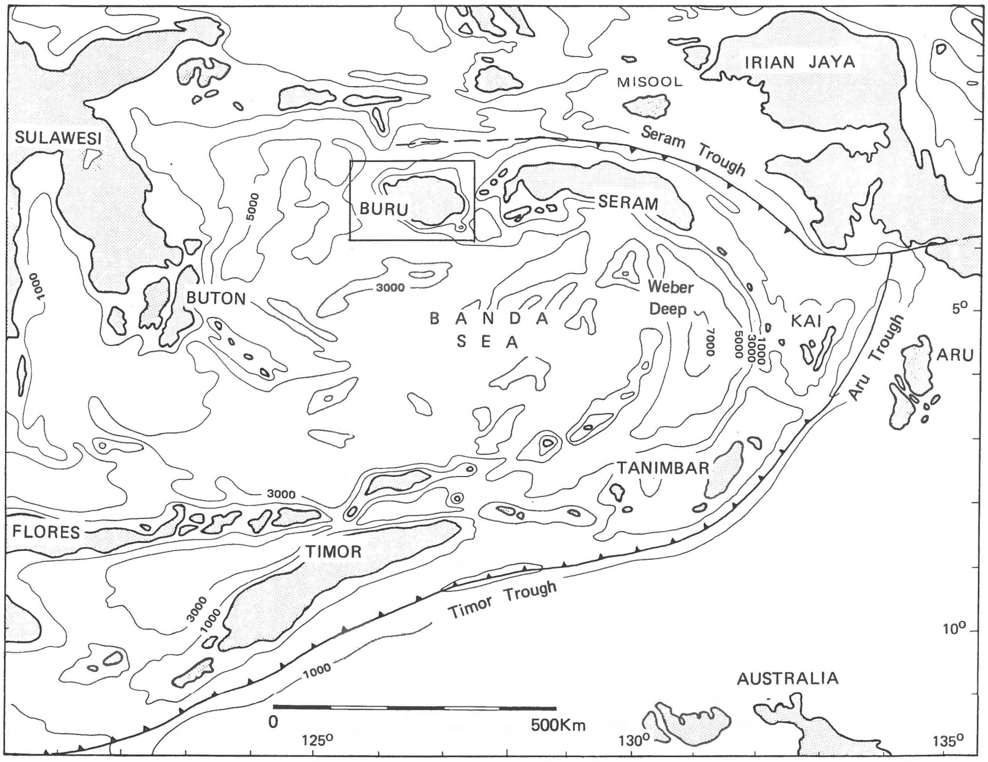

Fig. 1. Map showing location of Buru and the East Indonesian Banda Island Arc with surrounding areas. Depths in metres.

Introduction

Eastern Indonesia is a region of convergence between three major lithospheric plates, i.e. the Australian Plate, the Pacific Plate and the Eurasian Plate. Consequently its geology is complex and dominated by many active and completed collision zones. The largest active collision belt is the Banda Arc (Fig. 1). This paper discusses the tectonic and sedimentary history of Late Cenozoic sediments of NW Buru. Buru forms the northwestern end of the Banda Arc and with its maximum diameter of 150 km, the island is of intermediate size, compared to the other islands of the Banda Arc. Geologically, it is a fault-bounded structural high, lacking a complex thrust belt such as present in other islands of the outer arc. Elevations exceed 2000 m in the northwestern part (Mount Ghegan, 2736 m). Especially to the south and west the island is surrounded by deep, oceanic basins with depths exceeding 5000 m. Buru is separated from Seram by a strait more than 1000 m deep.

Buru is one of the least investigated islands of the Banda arc. It was visited in 1985 within the framework of the joint Indonesian–Dutch Snellius-II Expedition. One of the main objectives of the marine geological and geophysical research programme was to put constraints on existing theories concerning the Late Cenozoic evolution of the Banda Arc and Banda Sea basins. This paper is part of an onshore research campaign carried out on various islands of the outer arc, in order to determine the timing, rates and magnitudes of Late Cenozoic vertical movements in uplifted parts of that outer arc (De Smet et al., 1987).

Fieldwork on Buru concentrated on the Bara Bay area, in the northwest of the island (Fig. 2), where the most extensive outcrops of Neogene are located (Tjokrosapoetro et al., 1981). Since the island is densely forested and has no roads, access from the coast to outcrop in the forest was gained spotwise, along streambeds discharging considerable quantities of water. For locality details the reader is referred to the Progress Report of this campaign (Fortuin, 1986).

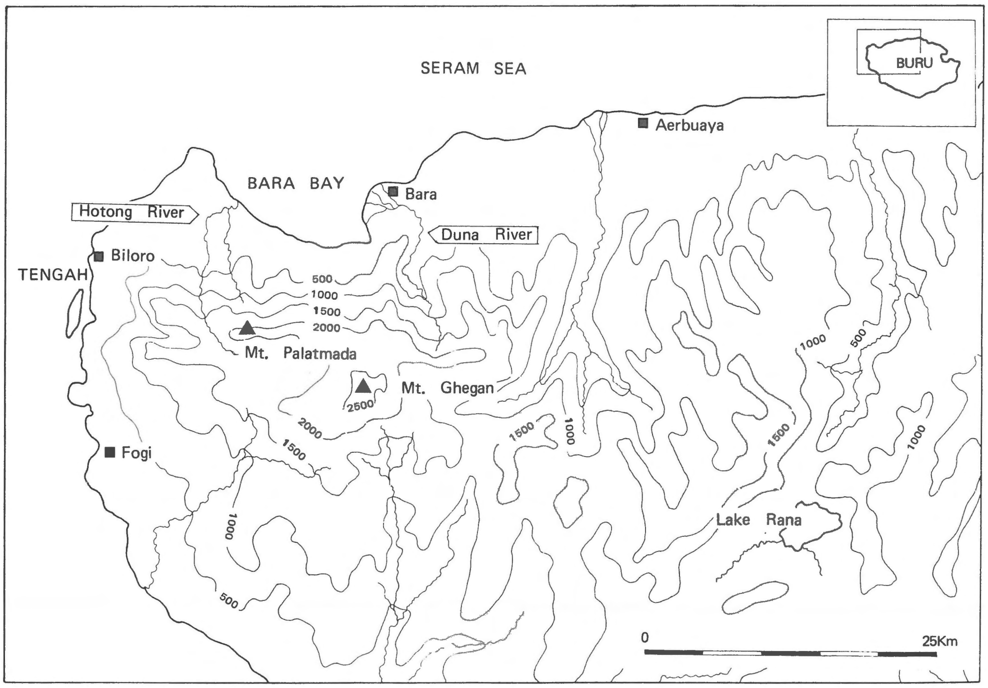

Fig. 2. Topographic map of NW Buru, showing the 500 m contour interval and location of the sections studied.

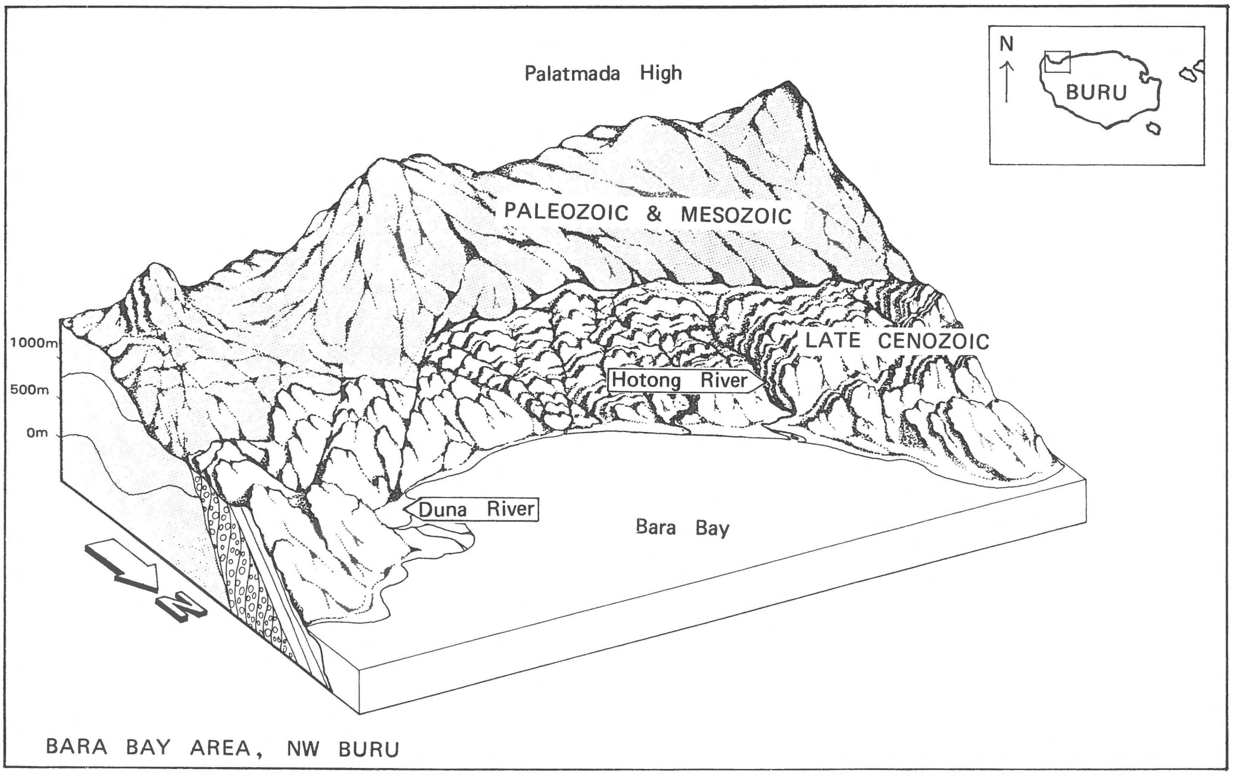

Fig. 3. Geomorphological block diagram of the Bara Bay area. Length of the block is 20 km. Vertical exaggeration × 2.

Northwest Buru is the most rugged part of the island. The relief is characterized by a central high, dominated by resistant Triassic dolomitic limestones, such as those forming Mount Ghegan. Immediately north of this highest peak, an approximately E–W oriented crest borders the Neogene of Bara Bay (Fig. 3). This area of elevated Mesozoic rocks is here called Palatmada High, after Mt. Palatmada (2050 m; Fig. 2).

Geological setting

The overall structure of Buru is relatively simple, a raised and faulted anticlinorium, with an approximately northwest–southeast oriented axis. In the denuded central parts Paleozic metamorphic rocks are widely exposed (Fig. 4). They are flanked by a thick Triassic cover, which is extensively exposed in W Buru. Only in this part of the island the overlying Mesozoic is exposed. Tertiary strata of W Buru are mainly confined to the coastal areas.

Buru is interpreted as a rifted piece of Paleozoic–Triassic Gondwana contiment, derived from the former northern margin of the Australian continent (Pigram & Panggabean, 1984). Stratigraphic similarities of the Paleozoic–Mesozoic deposits of Buru and Seram indicate the paleogeographic relationship of the islands (Rutten, 1927; Van Bemmelen, 1949; Tjokrosapoetro & Budhitrisna, 1982; Hartono & Tjokrosapoetro , 1984). These similarities led Pigram & Panggabean (1984) to distinguish Buru and Seram as a single microcontinent. Because the structure of Seram is more complex, assembled of elements from different plates (Hartono & Tjokrosapoetro, 1984; De Smet et al., in preparation), this paleogeographic relationship may be restricted to only the western parts of Seram. At present the islands are decoupled and Buru is moving southwestward relative to Seram (Tjokrosapoetro & Budhitrisna, 1982). Pliocene volcanism on the islands of Ambelau and Ambon,

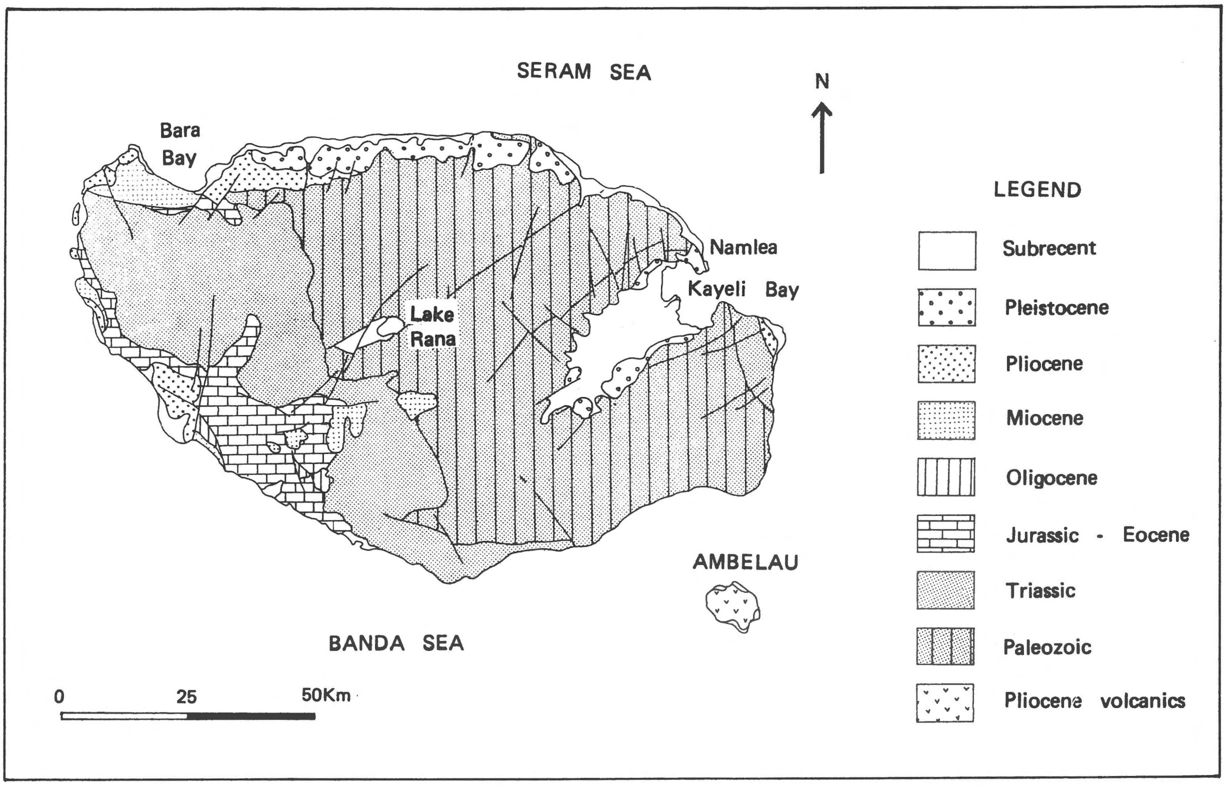

Fig. 4. Geological sketch map of Buru, simplified after Tjokrosapoetro et al., 1981.

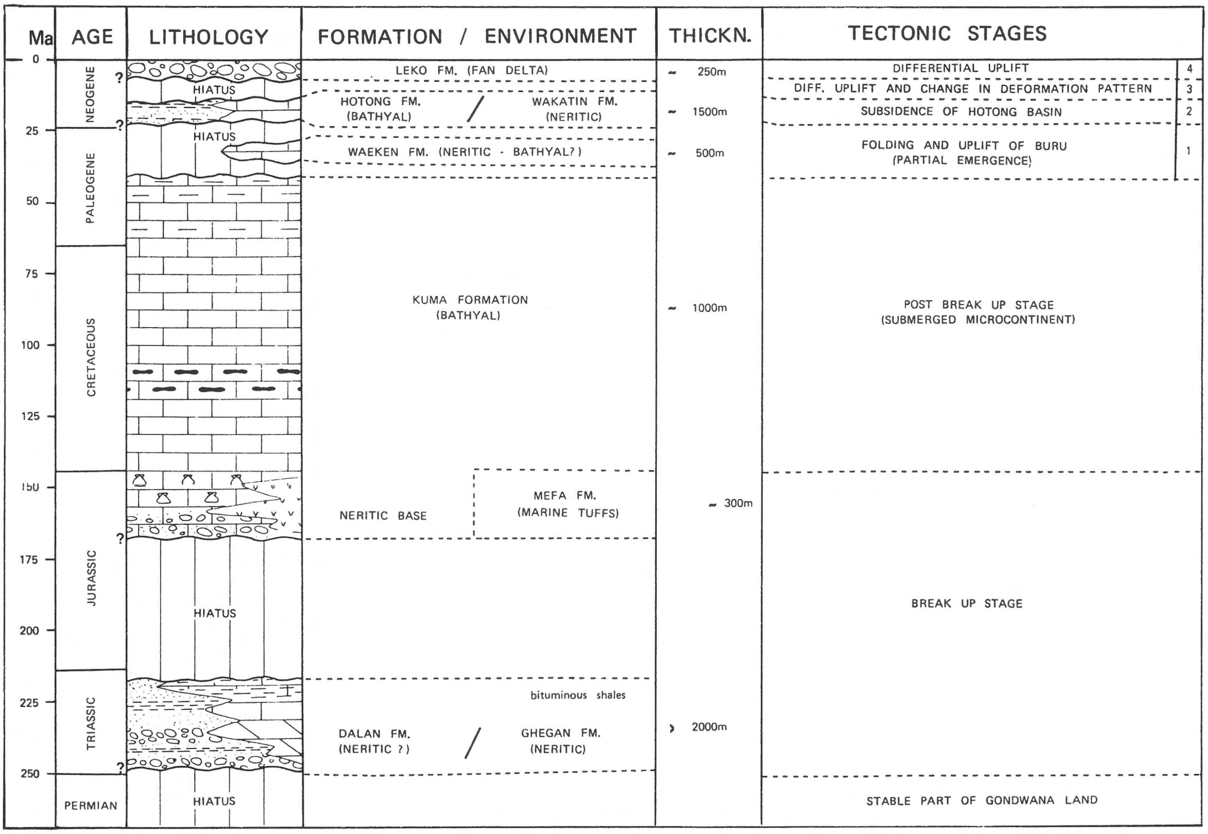

Fig. 5. Stratigraphic and tectonic framework of Buru, modified after Tjokrosapoetro et al. (1981) for the pre-Neogene. The numbers indicated for the Late Cenozoic tectonic stages refer to the episodes described in the text.

situated just to the south of respectively Buru and Seram, affirms close plate tectonic ties between the latter two islands.

Although the general stratigraphic framework of Buru has been firmly established by Tjokrosapoetro et al. (1981), details are lacking. A summary is given in Fig. 5 and below. The Paleozoic–Mesozoic fits into a rift-drift model (Pigram & Panggabean, 1984).

The oldest rocks of Buru are sediments metamorphosed in the greenschist to lower amphibolite facies that are overlain by lower grade metamorphics derived from a flysch-type facies. Unconformably overlying the Paleozoic basement are Triassic strata, attaining a total thickness of at least 2500 m. Of these, a dominantly clastic suite forms the Dalan Formation, whereas the mainly calcareous strata are named after Mt. Ghegan. Dark, bituminous shales are intercalated within the younger part of this latter unit.

The supposed break-up of Buru from Gondwanaland is shown by the presence of an angular unconformity between the Triassic strata and the Late Jurassic–Eocene Kuma Formation. Basaltic lavas and tuffs (Mefa Formation) interfinger with Jurassic highly fossiliferous strata and form the base of the Kuma Formation (Fig. 5). Most of this Late Jurassic–Eocene unit consists of pelagic marls, limestones and cherts, indicating continued submergence and bathyal conditions, which continued until the end of the Eocene.

Important tectonic activity resumed just before or during the Oligocene. Oligocene sediments consist of neritic – upper bathyal sandy marls and calcilutites (Waeken Formation), overlying either the previous deeper unit, or with an angular unconformity, the Triassic.

The Miocene is represented by the Early Miocene, bathyal, Hotong Formation and isolated occurrences of Early–Middle Miocene coralline limestones which make up the Wakatin Formation. These limestones unconformably overlie Triassic sediments and the Kuma Formation. The Hotong Formation is found in coastal areas of W and NW Buru. It is best exposed in the Hotong river, which was studied during our field campaign. The formation is characterized by repetitions of sand, silt and shale (or marly clays) which show a dull, dark grey colour. Andesitic lavas, volcanic breccias and tuffs are reported to interfinger locally with the base of the Hotong Formation, which shows that renewed diastrophism occurred in the Early Miocene.

The youngest Neogene deposits are the predominantly conglomeratic beds of the Leko Formation. This unit unconformably overlies the Hontong Formation, as well as older rock units. Fossiliferous intercalations in the type area suggest a Pliocene age. The formation is found in many coastal areas of W Buru and was studied in the vicinity of Bara village.

Pleistocene terrace deposits are either clastic (well developed east of Bara Bay), or are of reefal origin, such as at the western end of Bara Bay (Fig. 3). These terraces show a seaward tilt.

The structure of NW Buru is characterized by the seaward tilting of the Neogene deposits, which surround the Mesozoic rocks of the Palatmada High at the northern, western and southwestern side. The tilt demonstrates that the high is the result of young uplift of the area along a W–E oriented and W-dipping axis. Its development started during the Miocene and continued in the Pleistocene. Strata of the Hotong Formation are dipping 30–60 degrees away from the high and those of the Leko Formation 10–30 degrees. The dip of the Pleistocene terraces is less than 10 degrees.

Sections studied

Hotong Formation (Early Miocene, N4–N8)
The formation has been studied by recording the type sucession exposed along the Hotong river. Here, about 500 m of sediments are exposed in a monoclinal succession. The very base of the section, consisting of a supposed fault contact with the underlying Mesozoic, could not be reached. From aerial photographs it is estimated that the recorded section starts about 200–400 m above this contact. Because of large non-exposed intervals along the river, especially in the downstream parts, the total thickness of the entire succession can only be guessed at and may well exceed 1000 m. The lithological column (Fig. 6) has been compiled by extrapolating the non-exposed stratigraphic interval from the dip and distance measurements. Over forty samples were taken for micropaleontological analysis. The younger part of the Hotong Formation is somewhat better exposed in a tributary (Rumbia River).

Lithology: The basal 100 m thick interval consists of sandstones with thin pelitic intercalations and some prominent conglomeratic interbeds at the very base, giving rise to a waterfall. The conglomerate components are poorly rounded and sorted, consisting of quartzites, schists and other metamorphics. Also the sandstones (slightly calcareous) are poorly sorted. Horizontal lamination is a very common structure. Bedding is irregular at a dm scale. The sorting and regularity of the bedding increases distinctly upwards in this basal part. Plant debris is commonly intercalated; some *Zoophycos* burrows were observed.

Higher in the section dark grey, homogeneous pelitic strata dominate (mainly clayey marls and silts). Frequently intercalated sandy beds are somewhat calcareous. Bed thickness fluctuates rapidly (Fig. 7). Also, the degree of sorting and regularity of the bedding fluctuates. Conglomeratic intervals not only occur at the base, but also at about 450 m and towards the top of the section.

Turbidites form an essental element of the entire recorded succession. In general, turbidite beds are vaguely developed because of the muddy aspect of the sediment or the absence of distinctly graded a-intervals. At some levels calciclastic turbidites are intercalated. Reworked limestone fragments, larger foraminifera, such as Lepidocyclinids, algal and coral debris are very common in these beds. Downstream in the river channel, where no expo-

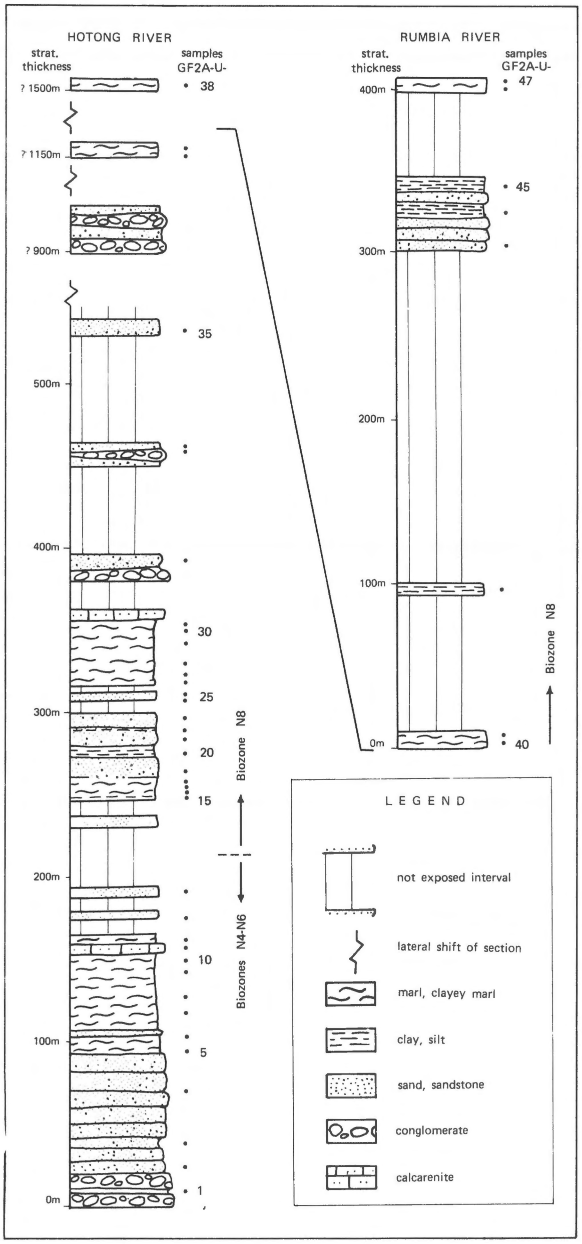

Fig. 6. Lithological column of the Hotong Formation, as recorded in the type section along the river Hotong and tributary (Rumbia river) for youngest strata.

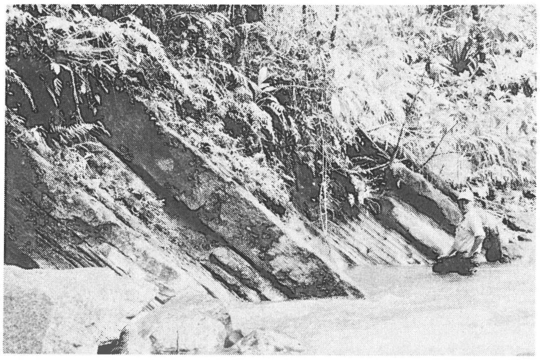

Fig. 7. Turbidites in a Hotong river outcrop, at the level of 290 m in the lithological column of Fig. 6. The basal thick calcareous sandstone beds amalgamate, showing poorly visible grading and horizontal lamination. Upward, the pelitic interbeds alternate with thin sandy strata, rich in cross-lamination, indicating NW directed paleocurrents.

sures were found, many blocks of the same lithology have been observed, suggesting that such beds are more frequently intercalated. The abundant larger foraminifera and limestone fragments in these blocks suggest that they originate from the partly coeval Wakatin Formation, although also fragments from the older Kuma Formation are included.

In the level between 250–300 m (Fig. 7) several paleocurrent measurements, based on sole marks and a Bouma-c interval, indicate transport in northwesterly directions.

Age: Biostratigraphic correlations and age indications presented here and in the following section follow the chronostratigraphic correlation schemes of Berggren et al. (1985). The planktonic foraminifera indicate that the Hotong Formation has been deposited within the Early Miocene: The basal 180 m of the recorded section yields species, such as *Globigerina venezuelana, Catapsydrax dissimilis* and *Globigerinoides primordius,* indicating deposition within plankton Zones N4–N6. The poor preservation of the fauna prevents designation to an age interval more specific than the rather long N4–N6 time interval. From 250 m onwards the faunas can be assigned to Zone N8. Characteristic species are *G. bisphericus, Globigerinatella insueta, Praeorbulina transitoria, P. sicana* and *P. glomerosa* *curva.* Because of the poorly preserved character of the section, with most of the sediments belonging to zone N8, it is assumed that only the upper part of the combined N4–N6 zones is represented in the base of the Hotong Formation.

Paleobathymetry: Most samples were collected from the finer clastic interbeds (clays-silty sands). Reworking of faunal elements of shallow water origin is evident in various samples. This must be due to frequent occurrence of turbidites. Preservation of the foraminifera is rather poor and specimens frequently are deformed, especially the arenaceous ones. The assemblages are characterized by a very high percentage of planktonic foraminifera, often exceeding 95%. Such faunas are most likely to be found in relatively deep, pelagic environments (Van Marle et al., 1987). The benthic foraminiferal content of the samples supports this idea of deposition in a bathyal depositional environment: forms as *Planulina wuellerstorfi, Laticarinina pauperata, Siphonina bradyana, Pullenia bulloides* and *Oridorsalis umbonatus* suggest, together with the high plankton percentages a paleodepth of 1000–1500 m. Amongst the benthic foraminifera the arenaceous forms are generally well represented, with *Adercotryma glomerata, Alveolophragmium subglobosum, A. scitulum* and *Vulvulina pennatula.* More details on the faunal aspects will be given by Van Marle (PhD thesis, in preparation).

Discussion: The depositional environment of the Hotong Formation evidently was deeper than littoral-neritic, as suggested by previous authors (Tjokrosapoetro et al., 1981). The abundance of metamorphic detritus in the formation indicates that the Paleozoic basement was subaerially exposed in the hinterland. In between the exposed land and the Hotong basin carbonate shoals existed, as is indicated by the presence of bioclasts of a type characteristic of the partly coeval Wakatin Formation. We therefore interprete that the Wakatin Formation formed carbonate shoals along the margin of the uplifted central parts of Buru during deposition of the Hotong sediments.

Leko Formation (? Pliocene)

This unit is named after the village of Leko in W Buru. It incorporates conglomerates, sandstones and limestone interbeds, unconformably underlying Quaternary terrace deposits in many coastal areas of Buru. The unit is widespread in NW Buru, where it unconformably overlies the Hotong Formation. The sediments investigated are exposed along the Duna river, south of the village of Bara (Fig. 2).

Lithology: The Duna river passes through a gorge-like valley, with up to 100 m high walls, entirely consisting of fairly cemented calcareous conglomerates of the Leko Formation. The outcrops were studied over a distance of 4 km.

The succession essentially consists of irregularly alternating coarser and finer intervals, with minor sandy or loamy intercalations. The conglomerates are polymict. Grey and white mudstones are dominant, but sandstones, quartzites, phyllites and other metamorphics are common as well. The average size of the fragments varies strongly from bed to bed (Fig. 8). Both gravel beds and boulder beds with blocks measuring up to 1 m in diameter have been observed. The strata are arranged in giant progading sets, up to 50 m high and dipping up to 20 degrees (after correction for the regional tectonic dip which amounts some 10 degrees). Individual foreset beds are up to 1.5 m thick. Although the internal geometry of the conglomerate bodies could not be studied in detail, the following characteristics apply: Individual prograding units seem to continue for many tens of metres in the downcurrent direction, which is roughly northeast. Individual beds are both inverse normally and non-graded (Fig. 8a–c), deposited as debris flows. Sorting, roundness and grainsize fluctuate rapidly. Evidently the supply of debris varied considerably during the outbuilding of the sets.

Within the sets curved truncations occur, followed by less inclined and less well organised conglomerates. Such intervals may show the characteristics of debris flow deposits, with outsized clasts embedded in a finer grained matrix (Fig. 8d). These features are interpreted as a result of slumping of coarse clastics due to unstability of the set

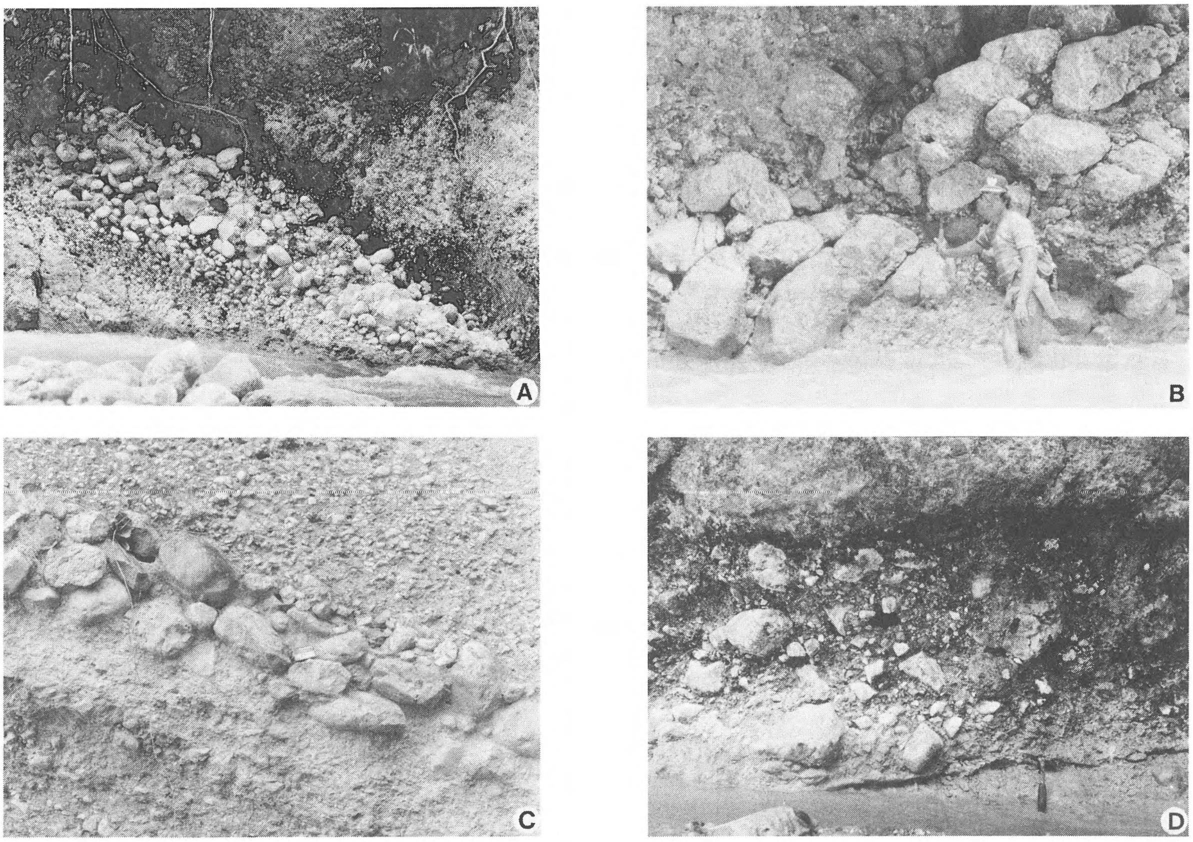

Fig. 8. Sedimentary structures of the coarse clastic Leko Formation. The beds form part of giant progading sets. The tectonic dip is under 10 degrees. A, Bed showing distinct reverse grading in well sorted clastics. B, Bed composed of large, flattered and subrounded blocks, exceeding 1.00 m length. C, Interval of well sorted clastics, showing positive grading. Film box (6 cm) for scale. D, Disorganised conglomerate bed with abundant angular fragments, probably deposited from viscous debris flow. Hammer for scale.

slopes. In addition, sets may be scoured and filled-up again.

Fine-grained beds are uncommon and could only be observed from a distance. We only could sample some loamy, perhaps reworked relics of bottom sets, now incorporated within the giant sets. In but one case a pocket of finer clastics yielded marine microfossils. Macrofossils have not been observed. Fossiliferous interbeds, rich in coral debris are, however, reported from the type area (Tjokrosapoetro et al., 1981).

Age: The limited fauna encountered in one of the samples does not permit an age assignment. We estimate a Pliocene age, following Tjokrosapoetro et al. (1981).

Paleobathymetry: The benthic foraminifera encountered mainly belong to shallow water genera as *Ammonia* and *Rotalia.* Because representatives of open marine environments are absent, deposition in a neritic environment is concluded.

Discussion: The progradational character of the conglomerates, expressed by mega sets up to 50 m thick and locally including marine fossils in the type area, points to deposition in a coarse clastic fan-delta. Rapid sedimentation from debris flows, regular sediment failure by overloading, and resedimentation are suggested as the main processes regulating fan-deltaic sedimentation (Postma, 1984; Postma & Roep, 1985). The steep inclination of the foreset beds indicates that they were deposited on the subaquaous upper delta slope of the Gilbert-type (cf. Postma & Roep 1985; Colella et al., 1987). The fossiliferous strata reported from the type area of the Leko Formation (Tjokrosapoetro et al., 1981) probably either belong to the somewhat deeper parts of the delta system, or were deposited in reefal environments laterally from the coarse clastic delta lobes.

The sometimes very coarse and poorly rounded fragments point to erosion of a nearby hinterland. Concerning the extraordinary size of the prograding sets and the large areal extent of the formation in NW Buru, a major rejuvenation of the relief must be concluded.

Previous investigators (cf. Rutten, 1927; Tjokrosapoetro et al., 1981) overestimated the thickness of the Leko Formation because they did not recognize the initial sedimentary dip of the strata. Because of its progradational character the real thickness of the formation is probably under 250 m in the Bara area.

Alluvial deposits

Coarse clastic alluvial deposits occur in large quantities at the mouth of river Duna in the eastern part of Bara Bay, where they form part of a fan delta system (Figs. 2 and 4). These deposits may be regarded as a modern analogue of the Leko depositional system. Upstream along the Duna river alluvial clastics have been reported by early investigators of this island (quoted by Verbeek, 1908) and have escaped attention ever since. They were of interest because of the presence of boulders of bituminous Mesozoic claystones in the river channel which geochemically resemble the Seefelder Schichten of Austria (Von John, 1906) that are presently known to be source rock of major hydrocarbon accumulations. These bituminous boulders indeed are present in notable quantities and must have been derived from the younger parts of the Triassic Ghegan Formation (S. Tjokrosapoetro, personal communication), which exposed in the hinterland. Because Late Triassic–Early Jurassic bituminous rocks also occur in Seram, which beds are the primary source of the Seram oil (O'Sullivan et al., 1985), we want to stress their possible importance as a hydrocarbon source.

Patterns of Cenozoic uplift and subsidence

From the estimated ages of the Hotong and Leko Formations we distinguish 4 episodes in the tectonic and sedimentary history of NW Buru.

1. Before the Miocene

Clastics of various types of metamorphics in the Hotong Formation indicate that the Paleozoic basement was already subaerially exposed. This implies considerable vertical movements and denudation of Buru prior to the onset of Hotong sedimentation. Since its separation from Gondwanaland Buru had become a firmly submerged part of a rifting microcontinent and did not emerge before the Late Eocene. Then uplift of Buru took place along a roughly NW–SE oriented axis, as indicated by the present distribution of the Triassic–Eocene rocks. The uplift and deformation terminated the post-breakup stage, which was characterized by marine, bathyal depositional environments since the Late Jurassic. Relicts of Oligocene neritic–bathyal sediments (Waeken Formation), which unconformably underlie the Hotong Formation in SW Buru, suggest that these represent an intermediate step towards the formation of the Neogene basin conditions. We therefore conclude that somewhere in mid Tertiary times a major tectonic reorganisation took place. When Hotong deposition started in the Early Miocene, denudation already had removed the Mesozoic cover at some places. Estimating a maximum duration of some 20 Ma from the first emergence of Buru until the onset of Hotong sedimentation and a minimum thickness of 3000 m for the removed Mesozoic cover, a rate of denudation of at least 15 cm/ka is suggested. Such a rate conforms with data for intermediate–high relief areas (Stow et al., 1985) and agrees well with drastic tectonic changes.

2. Early Miocene

The geohistory diagram for the Bara Bay area (Fig. 9) is based on the data collected by our team. The continuous bathyal character of the deposits investigated suggests that the depositional depth of the Hotong Formation remained more or less constant, which is confirmed by the fairly consistent

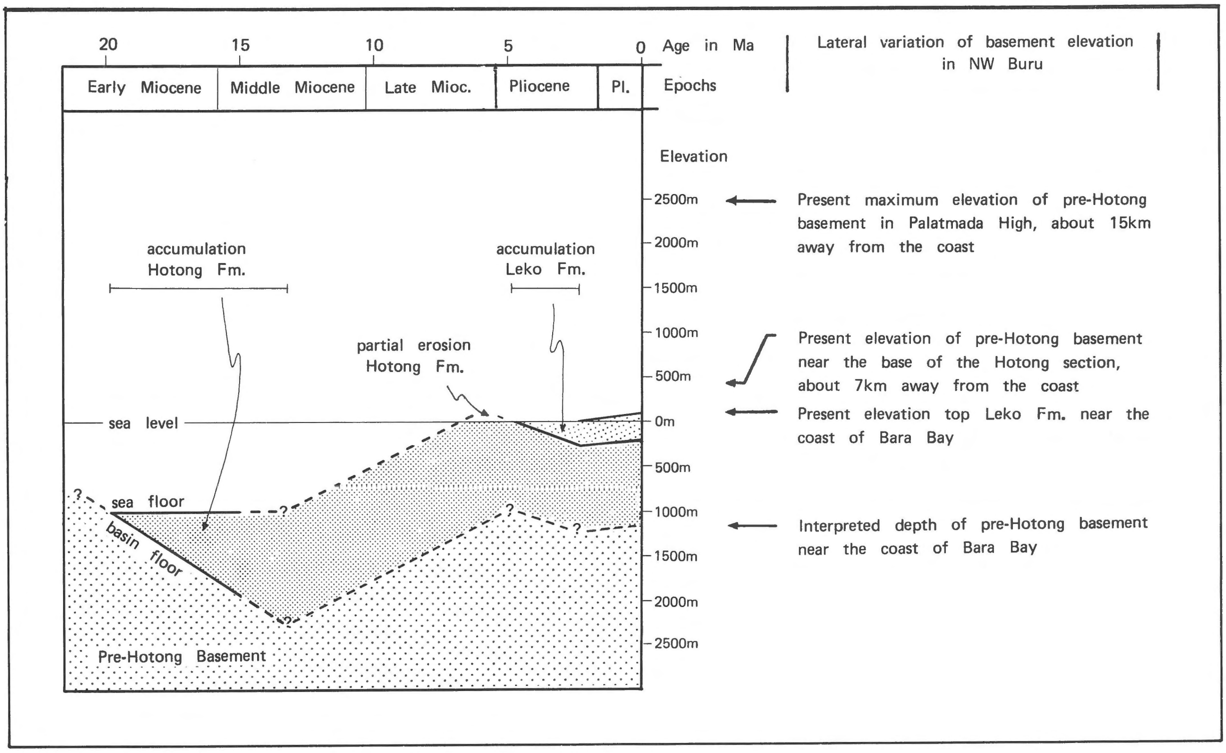

Fig. 9. Composite geohistory diagram for the Late Cenozoic deposits of NW Buru, Bara Bay area, with indication of the lateral change in elevation of the pre-Hotong basement towards the Palatmada High.

facies pattern. Thus an equilibrium between subsidence and sedimentation is probable. The Hotong Formation recovered is estimated to have had a maximum duration of 5 Ma, between 20–15 Ma. With a thickness of at least 1000 m, the rate of subsidence and sedimentation may have been in the order of 20 cm/ka and possibly more.

To calculate vertical movements from sedimentary sequences correctly, these sequences should be measured vertically. In the case of the Hotong section, where the outcrops are subhorizontal and extend over 3.5 km, the real vertical succession may be somewhat different. Moreover, because of the limited age constraints the calculated rate of subsidence is only an approximation.

3. Middle Miocene – (?) Early Pliocene

This episode comprises the interval between the deposition of the Hotong and Leko Formations. We estimate that deposition of the latter unit started less than 5 Ma BP, so that this interval roughly covers the period of 15–5 Ma BP.

The unconformable position of the shallow marine Leko deposits on top of the bathyal Hotong deposits and older basement rocks, proves that this episode was characterized by uplift, associated with tilting and erosion. Uplift of at least 1000 m took place near Bara Bay; it was even greater to the south, towards the Palatmada High (Fig. 3). Since the distribution of the post-Miocene sediments can not be related to the NW–SE structural trend of the basement rocks as in case of the older formations, a change in the regional deformation pattern must have occurred. Apparently, the center of uplift shifted from central East Buru to western Buru, as is also indicated by the spatial relations of formations along the northern coast of the island (geological map of Buru, Tjokrosapoetro et al., 1981; Fig. 4), where the oldest rocks underlying the post-Miocene deposits are found in the East, while the oldest post-Miocene sediments are cropping out in the west.

In the geohistory diagram (Fig. 9), the interpreted minimum uplift of the Hotong beds in the Bara Bay area is indicated. The uplift of the Palatmada High, which is separated from the Hotong river outcrops by a fault, can be estimated as follows: Because the bathyal Hotong Formation is also found in SW Buru, it is very likely that these sediments were deposited in one and the same basin, which extended over the present Palatmada High. This implies that before sedimentation of the Leko clastics started, large quantities of Hotong sediments and of the younger Mesozoic series were eroded from the high. We estimate that the pre-Hotong basement in the Palatmada High was uplifted locally over 2000 m. Calculated for an interval of 10 Ma, the minimum rate of uplift varied from 10 cm/ka near Bara Bay, up to 20 cm/ka in the central parts of the Palatmada High.

4. Pliocene – Recent

Deposition of the Leko Formation and its subsequent tilting is the most obvious feature of this episode. Sedimentation in the Leko fan-delta system took place all around the Palatmada High and the trend of deformation, that originated during the Miocene must have continued, as the Leko deposits now all dip away from the Palatmada high at angles up to 10 degrees. Because of this continued uplift of the high, it is very well possible that the depocenter of the Leko Formation shifted away from the high during sedimentation. In that case, which is common for such types of sediment, the temporary subsidence during Leko deposition shown in the geohistory diagram (Fig. 9) is an artifact: The Leko depositional system would rather indicate outward displacement and growth of a sediment body (offlap), than vertical accumulation of sediments.

Another reference to rapid uplift of the Palatmada High may be an observation by Rutten (1927). This author was puzzled by the fact that the Leko conglomerates exposed on the island of Tengah, situated 25 km SW of Bara Bay (Fig. 2), include a considerable quantity of Paleozoic schists, which do not crop out in the neighbourhood. According to the geological map of Buru, the nearest outcrops of schists are at 50 km distance. Since nowadays the Mesozoic of the Palatmada High lies in between, this situation may well be due to the young uplift of the high.

The minimal uplift of the Palatmada High over the last 15 Ma has been in the order of 25 cm/yr, as can be concluded from the present maximum elevation of the pre-Hotong basement (+2500 m) and at the end of the Hotong depositional period (−1500 m), assuming that the Hotong deposits of Bara Bay area and in SW Buru were interconnected.

Plate tectonic aspects

Morphologically Buru clearly belongs to the outer Banda Arc, as it constitutes the northwestern end. Because a complicated thrus belt is lacking (Tjokrosapoetro & Budhitrisna, 1982) Buru may better disclose its whereabouts than the adjoining, more deformed parts.

The Banda outer arc is usually roughly interpreted as an accretionary wedge, resulting from the northward directed subduction of the Australian plate under the Southeast Asian Plate (Hamilton, 1979). Collision of the continent of Australia and the southern segment of the Banda Arc is still going on and has been well documented (e.g. Bowin et al., 1980; Nishimura & Suparka, 1986; Karig et al., 1987). The plate tectonic history of the northern segment is different, because of present interaction with several smaller plates, of which the Philippine Sea Plate and western Irian Jaya are the most important. Movements of these plates with respect to southeast Asia are not accurately known, but they are probably similar to the relative movement of the neighbouring large Pacific Plate, which is roughly to the WNW. The combination of these west-directed movements and the curvature of the northeastern Banda Arc (Fig. 1) leads to a laterally changing tectonic regime: Going from eastern Seram to Buru, left lateral transpression changes into the left lateral strike-slip. The plate tectonic model for Seram (De Smet et al., in preparation) shows Buru in a strike-slip regime for the last 10 Ma. This led to deformation of the island, but not to the development of a thrustbelt, as present in Seram

(Audley-Charles et al., 1979). The duration of such a strike-slip situation is still an open question. The first deformation episode discussed, of mid Tertiary age, may, however, have occurred in a completely different plate tectonic situation, because the development of the actual Banda Arc system is generally supposed to have started first in the Neogene (Hamilton, 1979).

Since its break-up from Gondwanaland in the Early Mesozoic, Buru formed part of a submerged microcontinent, till Eocene–Oligocene times. The stratigraphic succession accumulated during this time interval (Fig. 5) does not contain slope deposits. Therefore one may safely assume that the paleogeographic position of the block was beyond the direct influence of either the Austalian or the Southeast Asian landmasses. Furthermore, absence of deformation indicates that it was also outside the direct influence of an active plate boundary. Before deposition of the Waeken Formation started, in Oligocene times, the situation changed drastically. Buru was folded, uplifted and heavily eroded. If this was related to an approaching or newly formed plate boundary, to what plate system was it related? For the lowermost Oligocene, Le Pichon et al. (1986) reported a significant change in direction of motion of the Pacific Plate with respect to the Eurasian Plate, which would have resulted in changes of relative motion of the Philippine Sea Plate with respect to the Southeast Asia Plate. Related to these changes is the collision of India with Asia, which resulted in a series of extensive mid Tertiary tectonic events, such as the possible SE oriented extrusion of the Sunda shelf area (Tapponnier et al., 1986). If it was this geodymanic revolution around southeast Asia that brought Buru near an active plate boundary, than the island may have had a closer paleogeographic relation to southeast Asia or the Philippine area than to the Australian continent. The Australian continent, now in collision with the Banda Arc, was still far to the south in mid Tertiary times (Smith et al., 1981).

Besides Seram, the most closely related microcontinental pieces are Buton and Misool (Fig. 1; Pigram & Panggabean, 1984). If so, it is puzzling that these islands in mid Miocene times may have been widely apart: Buton became paleogeographically incorporated in SE Asia, because of its collision with Sulawesi (Katili, 1978). It should be noted that Buton shows a similar onset of clastic sedimentation on a deformed basement in the Early Miocene as in Buru (Smith, 1985; Fortuin et al., in preparation), though the timing is somewhat different. The island of Misool, in contrast, is situated on a salient part of the western Irian Jaya platform and clearly shows a Late Oligocene phase of deformation (Pigram et al., 1982). Paleomagnetic evidence (Wensink et al., 1987) shows that Early–Late Cretaceous rocks of Misool follow the Australian pole positions for that age, indicating that the Australian continent extended as far as Misool, at least until the beginning of the Tertiary. Since the Australian continent in the Oligocene was still widely separated from the subduction complexes around SE Asia, Australian-related tectonics are likely for Misool. We therefore doubt the suggested mid-Teriary proximity of Misool and Buru (Tjokrosapoetro & Budhitrisna, 1982).

The renewed deformation of Buru after the Early Miocene may be directly related to the evolution of the Banda Arc subduction system. The change in the direction of the compression of the island probably is related to overall changes in the plate tectonic configuration: In the middle Miocene another important change in plate motions in the western Pacific area took place. Collision was in progress in southern Irian Jaya and a new subduction zone formed north of Irian Jaya (Hamilton, 1979). In Seram important thrust tectonics took place (Audley Charles et al., 1979).

After the Miocene, the direction of deformation on Buru did not change anymore, although the uplift rates are discontinuous through time. We therefore assume that the plate movements around Buru remained more or less of the same type through the Pliocene and Pleistocene.

Conclusions

The Late Cenozoic sediments of NW Buru were deposited after a major mid Tertiary event of folding and uplift of the preexisting, fully pelagic Mesozoic–Eocene succession. Bathyal sedimentation was re-established during the Early Tertiary, at least for the 20–15 Ma interval, as shown by the Hotong Formation studied in its type section.

Prior to a poorly dated new depositional event in the Pliocene, strong uplift, tilting and erosion took place. This was associated with a changing deformation pattern, expressed by a shift of the centre of uplift from central East Buru to western Buru.

The studied Pliocene succession (Leko Formation) consists of coarse clastics that form part of the shallow marine prograding fan-delta system. To our knowledge these are the first reported fan-delta deposits with giant prograding sets for the Indonesian realm and as such merit careful sedimentological analysis (cf. Postma, 1984; Colella et al., 1987).

The mid Tertiary period of drastic changes in the Buru segment coincides with changes in direction of the Pacific plate with respect to the Eurasian plate. This reorganisation of movement may be related to interaction of Buru with SE Asia and not necessarily with the Australian continent, from which it initially derived.

In the course of the Middle–Late Miocene a change in the direction of deformation took place in Buru. This renewed deformation is probably directly related to the revolution of the Banda Arc system, which resulted in nearby Seram in important thrust tectonics.

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References

Audley-Charles, M.G., Carter, J.D., Barber, A.J., Norvick, M.S. & Tjokrosapoetro, S. 1979 Reinterpretation of the geology of Seram–Implications for the evolution of the Banda Arc and northern Australia – J. Geol. Soc. London: 547–568

Berggren, W.A., Kent, D.V. & Van Couvering, J.A. 1985 The Neogene: Part 2, Neogene geochronology and chronostratigraphy. In: N.J. Snelling (ed.): The chronology of the geological record – Geol. Soc. London. Mem. 10: 211–260

Bowin, C.O., Purdy, G.M., Lawver, L., Hartono, H.M.S. & Jezek, P. 1980 Arc-continent collision in the Banda Sea region – Am. Ass. Petr. Geol. Bull. 64: 868–915

Colella, A., De Boer, P.L. & Nio, S.D. 1987 Sedimentology of a marine intermonane Pleistocene Gilbert-type fan-delta complex in the Crati Basin, Calabria, southern Italy – Sedimentology 34: 721–736

De Smet, M.E.M., Fortuin, A.R., Tjokrosapoetro, S. & Van Hinte, J.E. 1987 Late Cenozoic vertical movements of non volcanic islands in the Banda Arc area – Abstract. Int. Symp. Results Snellius-II Expedition, Jakarta: 33

Fortuin, A.R. 1986 The Snellius-II Expedition Progress Report, Theme I Geology and geophysics of the Banda Arc and adjacent areas, Campaign GF2 Buru, Seram and Kai Islands, August–October 1985 – Roy. Netherl. Acad. of Arts and Sciences, Indonesian Institute of Sciences: 74 pp

Hamilton, W. 1979 Tectonics of the Indonesian region – U.S. Geol. Survey Prof. Pap. 1078: 345 pp

Hartono, H.M.S. & Tjokrosapoetro, S. 1984 Preliminary account and reconstruction of Indonesian terranes – Proc. Indon. Petr. Ass., 13th Ann. Conv.: 185–226

Karig, D.E., Barber, A.J., Charlton, T.R., Klemperer, S. & Hussong, D.M. 1987 Nature and distribution of deformation across the Banda Arc–Australian collision Zone at Timor – Geol. Soc. Am. Bull., 98: 18–32

Katili, J.A. 1978 Past and present geotectonic position of Sulawesi, Indonesia – Tectonophysics 26: 165–188

Le Pichon, X., Huchon, P. & Barrier, E. 1986 Pangea, geoid and the evolution of the western margin of the Pacific Ocean. In: Nasu, N., Kobayashi, K., Uyeda, S., Kushiro, L. & Kagami, H. (eds): Formation of active ocean margins – Terra Sci. Publ. Cy, (Tokyo): 3–42

Nishimura, S. & Suparka, S. 1986 Tectonic development of East Indonesia – J. SE Asian Sci. 1: 45–57

O'Sullivan, T., Pegum, D. & Tarigan, D. 1985 Seram oil search, past discoveries and future oil potential – Proc. Indon. Petr. Ass., 14th Ann. Conv. 3: 3–19

Pigram, C.J., Challinor, A.B., Hasibuan, F., Rusmana, E. & Hartono, U. 1982 Lithostratigraphy of the Misool Archipelago, Irian Jaya – Geol. Mijnbouw 61: 265–279

Pigram, C.J. & Panggabean, H. 1984 Rifting of the eastern margin of the Australian continent and the origin of some microcontinents in Indonesia – Tectonophysics 107: 331–353

Postma, G. 1984 Mass flow conglomerates in a submarine canyon: Abrioja Fan-delta, Pliocene, southeastern Spain. In: Koster, E.H. & Steel, R.J. (eds): Sedimentology of gravels and comglomerates – Can. Soc. Petr. Geol. Mem. 10: 237–258

Postma, G. & Roep, T.B. 1985 Resedimented conglomerates in the bottomsets of Gilbert-type gravel deltas – J. Sed. Petrol. 55: 874–885

Rutten, L.M.R. 1927 Voordrachten over de geologie van Nederlandsch Oost-Indie – Wolters, (Groningen): 839 pp

Smith, A.G., Hurley, A.M. & Briden, J.C. 1981 Phanerozoic paleocontinental world maps – Cambridge Earth Sci. Ser. (Cambridge): 102 pp

Smith, R.B. 1985 Sedimentology and tectonics of a Miocene collision complex and overlying late orogenic clastic strata, Buton island, eastern Indonesia – Unpubl. PhD thesis, Univ. California, Santa Cruz: 255 pp

Stow, D.A.V., Howell, D.G. & Nelson, C.H. 1985 Sedimentary, tectonic, and sea-level controls. In: Bouma, A.H., Normark, W.R. & Barnes, N.E. (eds): Submarine fans and related turbidite systems – Springer (New York): 15–22

Tapponnier, P., Peltzer, G. & Armijo, R. 1986 On the mechanics of the collision between India and Asia. In: Coward, M.P. & Ries, A.C. (eds) Collision tectonics – Geol. Soc. Spec. Pub. 19: 115–157

Tjokrosapoetro, S., Budhitrisna, T. & Rusmana, D. 1981 Report on the geology of Buru Quadrangle, Maluku. (with provisional geological map) – Unpubl. Rept. Geol. Mapping and Aerial Photo Interpret. Proj., Geol. Res. Devel. Centre, Bandung: 32 pp

Tjokrosapoetro, S. & Budhitrisna, T. 1982 Geology and tectonics of the northern Banda Arc – Geol. Res. Devel. Centre Bull. 6: 1–17

Van Bemmelen, R.W. 1949 The geology of Indonesia – Govt. Printing Office, The Hague: 732 pp

Van Marle, L.J., Van Hinte, J.E. & Nederbragt, A.J. 1987 Plankton percentages of the foraminiferal fauna in sea-floor samples from the Australian–Irian Jaya continental margin, eastern Indonesia – Marine Geol. 77: 151–156

Verbeek, R.D.M. 1908 Rapport sur les Moluques – Jaarb. Mijnw. in Nederl. Oost-Indie, XXXVII: 844 pp

Von John, C. 1906 Ueber die chemische Beschaffenheit der Asphaltschiefer der Bara Bai (Buru) – N. Jahrb. f. Min., Beilageband XXII: 686–691

Wensink, H., Hartosukohardjo, S. & Suryana, Y. 1987 Paleomagnetism of rocks – from the eastern Indonesian islands Timor and Misool – Abstract. Int. Symp. Results Snellius-II Expedition, Jakarta: 43