Geologie en Mijnbouw 68: 221–235, 1989

Late Cenozoic geohistory of Seram, Indonesia

M.E.M. De Smet[1], P.A. Sumosusastro[2], I. Siregar[3], L.J. Van Marle[1], S.R. Troelstra[1] & A.R. Fortuin[1] [1] *Geomarine Centre Amsterdam, Institute of Earth Sciences, Free University – P.O. Box 7161, 1007 MC Amsterdam, The Netherlands;* [2] *Marine Geological Institute, Jalan Dr. Junjunan 236 – P.O. Box 215, Bandung, Indonesia;* [3] *Oil and Gas Technology Development Centre – Jalan Cipulir, P.O. Box 89, Jakarta Selatan, Indonesia*

Received 7 March 1988; accepted in revised form 6 October 1988

Key words: Seram, Banda Arc, Late Cenozoic, geohistory analysis, plate tectonics

Abstract

Two sections in Plio – Pleistocene basin deposits were measured and systematically sampled in the southwestern part of Seram, the largest island in the northern Banda Arc, Indonesia. A geohistory analysis of field and laboratory data allows for a reconstruction of the vertical movements of the area and demonstrates that the basement of southwestern Seram was subsiding at an average rate of 50 cm/ka during the Late Pliocene – Early Pleistocene, an episode of relative tectonic quiescence, following Late Miocene compressional deformation and uplift. In the Late Pleistocene, compressional deformation and uplift was renewed but differed in style and orientation from the Late Miocene events. This history of the basins in southwestern Seram is very similar to that of the oil producing basins in the northeastern part of the island.

The Late Cenozoic tectonic history of Seram can be interpreted in the frame of regional plate motions. Due to an anticlockwise rotation of the island and a relatively north to northwestward motion of Irian Jaya, the overall tectonic regime in the area gradually changed from compressional in the Late Miocene to strike-slip in the Late Pliocene. Finally, collision of Seram with the continental crust of Irian Jaya resulted in renewed compressional deformation from the Late Pleistocene onwards.

Introduction

One of the main objectives of the geological and geophysical research program of the Indonesian – Dutch Snellius-II Expedition (1984–1985) is to put constraints on the Late Cenozoic evolution of the Banda Sea region. The area is composed of several microplates, which are surrounded by the mutually converging Southeast Asian, Australian and Pacific plates. Subduction processes, accretion of sediments and collision of terranes lead to the present Banda Arc (Hamilton, 1979; Hartono & Tjokrosapoetro, 1984), but details of the evolution of this arc are still poorly known.

During Snellius Campaigns GF1A–2A Late Cenozoic sections on non volcanic islands in the arc were investigated to determine the history of vertical movements, providing information about timing and rates of ongoing accretion and collision processes. The method used to calculate these movements is the geohistory analysis technique (Van Hinte, 1978). Sections are measured and systematically sampled. Planktic and benthic foraminiferal assemblages are analyzed for chronostratigraphic correlation and interpretation of the paleobathymetry. Combined data allow for the re-

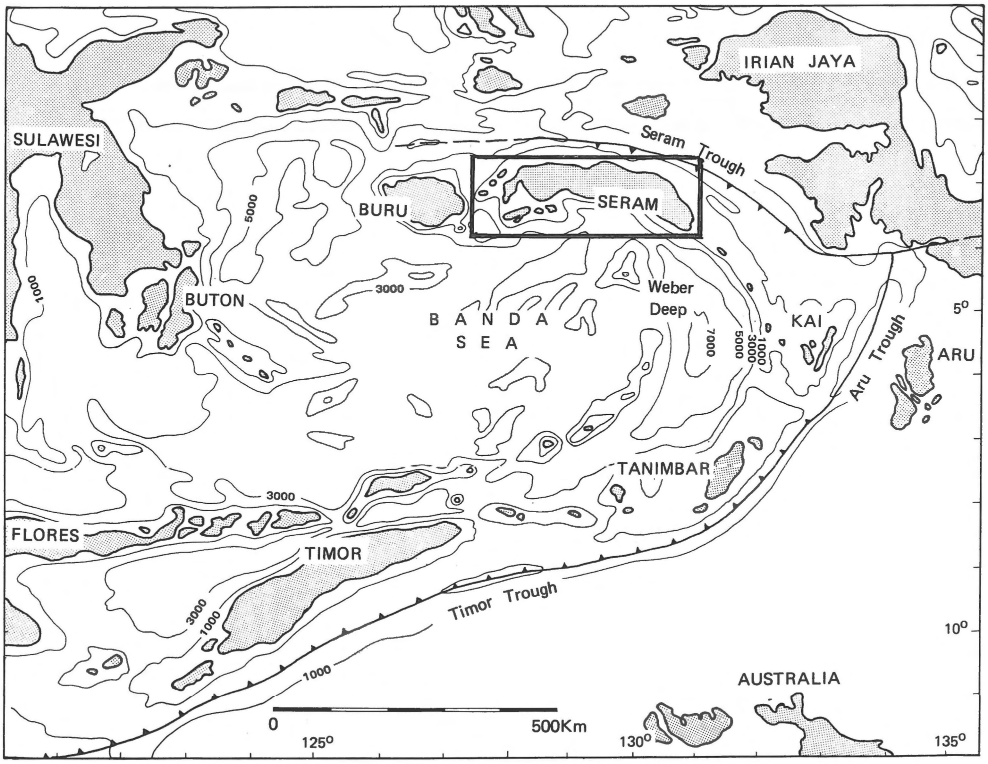

Fig. 1. Map of the eastern Indonesian Banda Arc, showing location of Seram. Toothed line is the deformation front of the Banda Arc subduction complex.

construction of the vertical movements of the basement. The expedition program comprised investigations on Timor in the South of the Banda area, on Buton in the West, On Buru and Seram in the North, and on Kai in the East (Fig. 1). This paper reports the results from Seram. The material discussed is filed in the Dutch National Museum in Leiden (Rijksmuseum van Natuurlijke Historie), under Van Marle's name. Results from the other islands are and will be published in separate papers (Fortuin et al., 1988; in press; De Smet et al., in prep.). Seram, one of the largest islands in the Banda Arc, is the emerged top of a thrustbelt, composed of Paleozoic, Mesozoic and Cenozoic rocks (Hamilton, 1979; Audley Charles et al., 1979). Its geological structure has much in common with that of Timor (Audley Charles et al., 1979). The Paleozoic rocks are low to high grade metamorphics, while the Mesozoic and Paleogene rocks in majority are of shallow to deep marine origin. During the Miocene, or perhaps already earlier, the rocks were deformed into a thrustbelt, on top of which Pliocene and Pleistocene sediments accumulated in local basins. The present study concerns the history of the southwestern part of island during and after the fill of such basins.

The geology of Seram is but partly known, due to the poor accessibility of this densely forested island. Valk (1945), Germeraad (1946) and Van der Sluis (1950), described the general geology of central, western and eastern Seram respectively, from notes and sample collections made by Rutten & Hotz during a Seram expedition in 1917–1919 (11 short reports, Rutten & Hotz, 1918–1920). Rutten (1927) summarized the geological characteristics of the island and this was repeated by Van Bemmelen (1949) and Hamilton (1979). Audley Charles et al. (1979) compared the geology of Seram and that of Timor and developed a tectono-stratigraphic scheme. Tjokrosapoetro et al. (1983a, 1983b) and

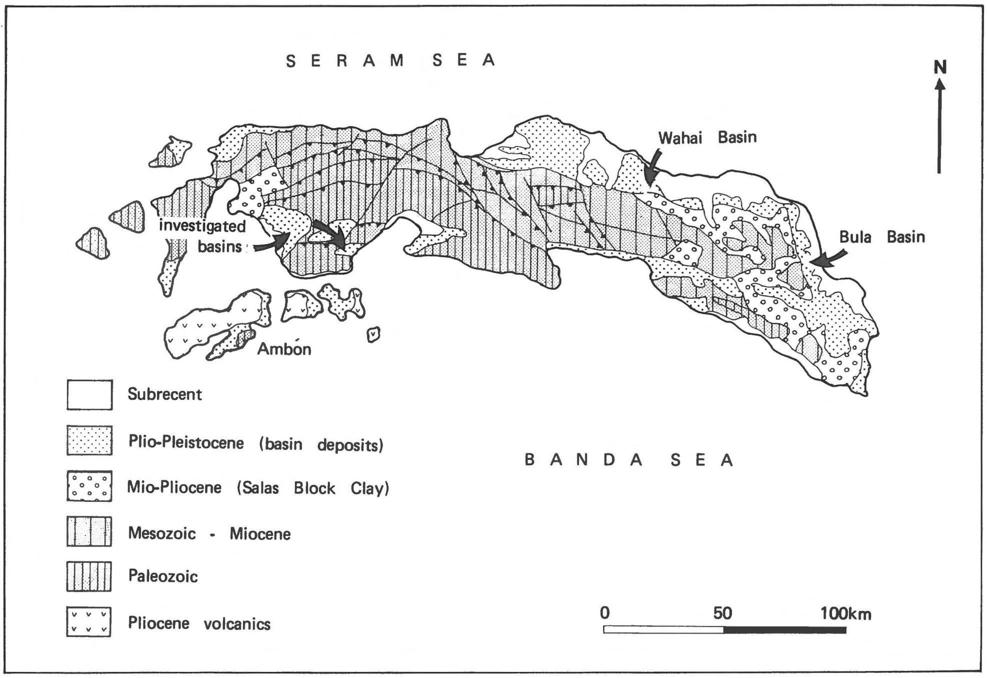

Fig. 2. Geological sketch map of Seram, simplified after Tjokrosapoetro et al. (1983a, 1983b) and Gafoer et al. (1984).

Gafoer et al. (1984) compiled preliminary geological maps at a scale of 1 : 250.000 (simplified version given in Fig. 2). Zillman & Paten (1975) and O'Sullivan et al. (1985) gave an overview of the Late Cenozoic deposits in the Bula and Wahai basins of northeastern Seram, which are relatively well investigated because of their oil potential.

Locality details

The Late Cenozoic deposits of SW Seram are the remnants of an E-W oriented basin fill on top of the metamorphic Taunusa and Tehuru Complexes of Paleozoic age (Tjokrosapoetro et al., 1983a). The metamorphic rocks are exposed as a belt of schists to the South of the basin and as a belt of phyllites to the North of it. Tjokrosapoetro et al. (1983a) interpreted the contact between the metamorphic complexes as a southward dipping thrust fault. We assume that the basin formed as a result of the thrusting and subsequent erosion in the Miocene, during a time of high tectonic activity in this segment of the Banda Arc (Hamilton, 1979). The deposits in the basin itself are hardly deformed. The oldest strata found in it are of Middle Pliocene age, but parts in the buried topography may contain older sediments.

Two Late Cenozoic sections have been investigated, the Masa Section at the eastern and the Rioeapa Section at the western side of the basin (Fig. 3). For details concerning the location and lithology of these sections the reader is referred to Fortuin (1986). The first section is along the river Masa, which discharges into Elpapoetih Bay near the village of Tala (Fig. 4). The section starts about 5 km upstream and ends near the river mouth. Rocks are intermittently exposed and slightly folded, with a dominant dip towards the East. The total stratigraphic interval that we covered is estimated at 600 m. The age of the sediments ranges from Chrono-Biozones N19/20 to N21 of Blow (1969) (Fig. 5).

The second section runs along the Rioeapa River and its tributary Kwa River. The Rioeapa discharges into Piroe Bay near the town of Kairatoe (Fig. 4). The section starts in the river Kwa, near the natural bridge described by Rutten & Hotz (in:

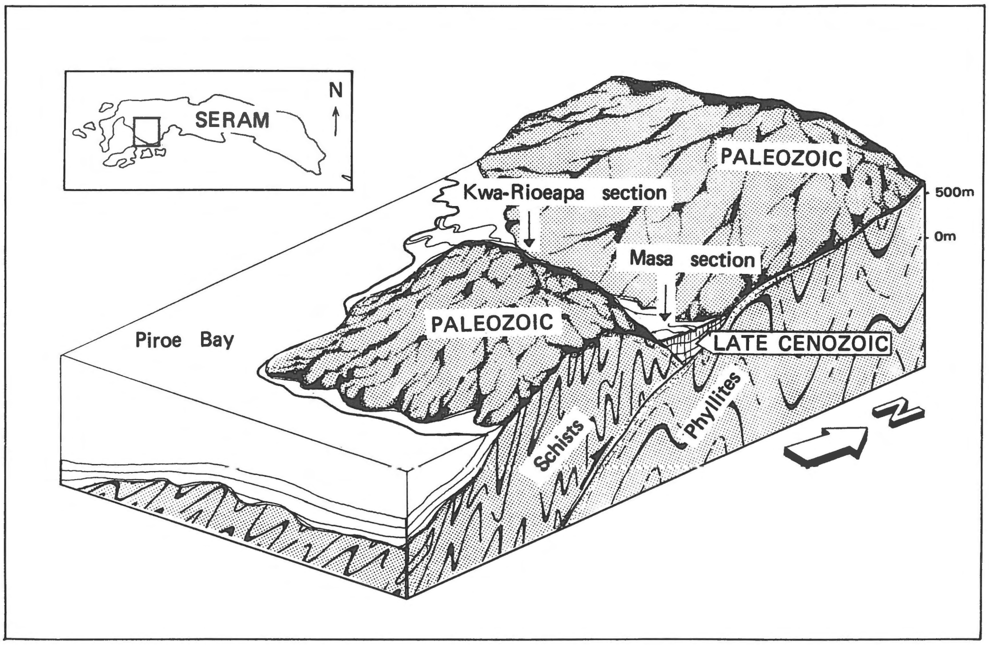

Fig. 3. Blockdiagram of SW Seram, indicating the unconformable position of the investigated Late Cenozoic basin deposits on top of Paleozoic metamorphic rocks. Length of block diagram is about 25 km.

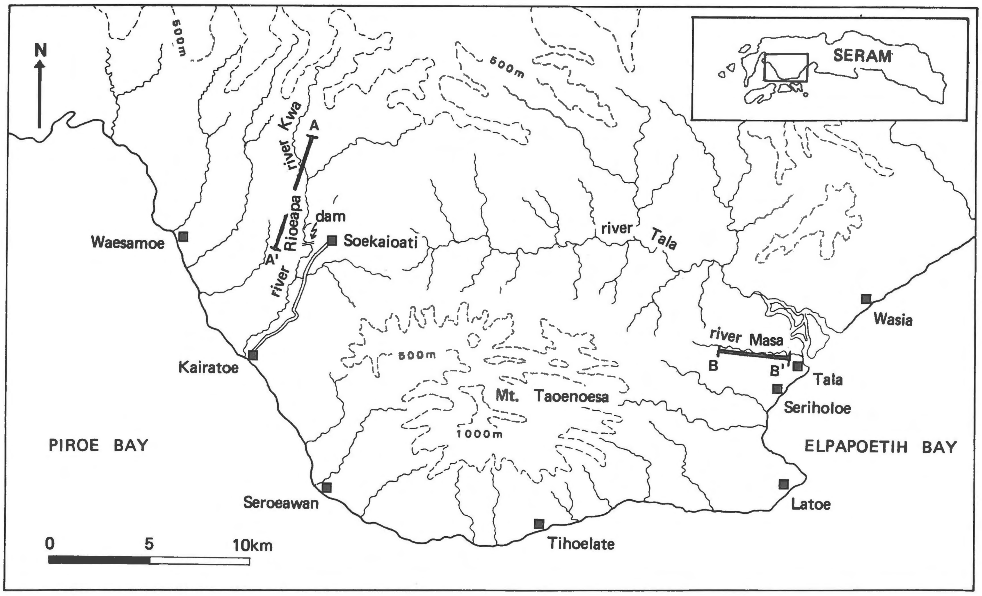

Fig. 4. Topographic map of SW Seram, showing location of Kwa-Rioeapa Section (AA′) and Masa Section (BB′).

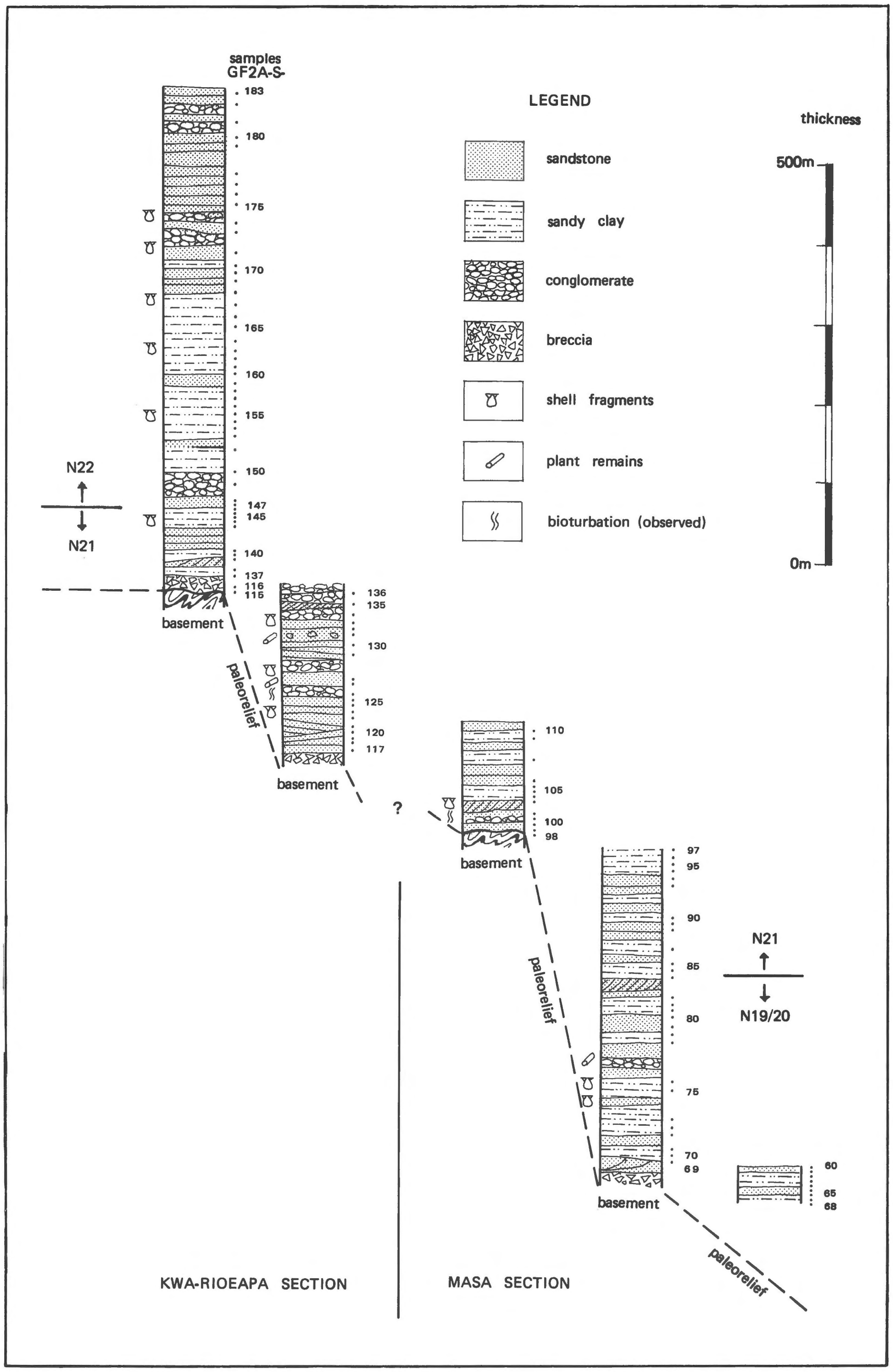

Fig. 5. Stratigraphy of the investigated sections. At the contact with the basement, the age of the overlying deposits varies as a consequence of the paleorelief. N19/20, N21 and N22 are Late Pliocene and Early Pleistocene biozones of Blow (1969).

Valk, 1945), which is about 3 km upstream from the confluence with the Rioeapa. It continues along the latter, down to an irrigation dam near the village of Soekaioati. Rocks are almost continuously exposed along the section and generally show a low angle tilt towards the West. The total thickness covered is about 800 m. The age of the sediments in this section ranges from Zone N21 to Zone N22 (Fig. 5).

Metamorphic rocks crop out at several places in the sections and show contacts with different stratigraphic levels in the Late Cenozoic deposits, which indicates that these deposits gradually covered an existing relief (Fig. 5).

Lithology

Exposures along both sections show a rather consistent picture of clastic fan deposits in a marine basin. Units of siltstones, sandstones and conglomerates, showing evidence of gravitational transport, irregularly alternate with units of silty clays (Fig. 5). Shell debris and plant remains are present throughout the sections, albeit in small amounts.

The finest grained units are up to tens of metres thick and consist of greenish to brownish, silty clays, in which horizontal lamination is a remarkably common feature. The preservation of such laminated sections indicates low biogenic activity, which may be due to low oxygen values in the bottom waters. Siltstones and sandstones in the sections also form units up to tens of metres thick, showing a well developed bedding at a cm-dm scale. They have greyish to brownish colours and are generally poorly sorted. Low angle truncation is sometimes visible between packages of strata, which suggests that deposition occurred in fan lobes. The limited size of the exposures, however, prohibited a study of the lateral development in the units.

Coarse-grained sandstones and conglomerates are intercalated in between the well bedded siltstones and sandstones. In general, the bedding type of these coarse-clastics is very irregular and lateral changes of the sedimentation pattern can be observed within single exposures. Channel structures were observed as well as lens shaped bodies. Mega cross-bedding, scouring marks, grading and lamination are common sedimentary structures. Part of the conglomerates is closely packed; another part consists of pebbles floating in a sandy matrix, indicating deposition from mass flows. The coarser fragments in all these sediments consist of metamorphics, while calcarenites are absent. The source area of the clastics therefore will have much resembled the present situation with almost exclusively metamorphic rocks surrounding the basin.

As the overall lithological composition and the sedimentary structures show the same characteristics throughout both sections, we assume that the depositional environment in the basin hardly changed during accumulation of the deposits. Because of truncation planes in the finer clastics and the rapid lateral changes in thickness and sedimentary structures in the coarser grained clastics, it is interpreted that the basin fill essentially consists of migrating submarine fan complexes with channels and overbank deposits. The clayey sediments partly belong to the latter, in part they have a pelagic origin.

Age

Qualitative analyses have been carried out on the planktic foraminiferal content of the samples. As can be expected in gravity flow dominated sediments, well preserved planktic foraminiferal assemblages are generally scarce. Nevertheless, sufficient reliable material was sampled to allow age assignments, following the zonal scheme of Blow (1969), as modified by Stainforth et al. (1975) and Bolli & Saunders (1985) and the geochronological correlation scheme of Berggren et al. (1985).

The lower part of the Masa section is correlated with plankton Zone N19/20, as indicated by the presence of well developed *Sphaeroidinella dehiscens* in sample 68 (Fig. 5). Other characteristic species in this sample are *Globorotalia tumida* (indicating base of Zone N18 or younger) and dextral *Pulleniatina obliquiloculata*. Thus an age within Zone N19/20 is indicated for the onset of sedimentation of the Masa section.

Upwards in the section, the transition to Zone N21 takes place at the level of sample 85, because of the first appearance of *Globorotalia tosaensis*. This species remains consistently present higher up in the Masa section. Because no *Globorotalia truncatulinoides* was found, whose presence would indicate the base of Zone N22 or a younger level, it is concluded that the top of the Masa section still belongs to Zone N21. The lower part (samples 117–136) of the Kwa-Rioeapa section can be correlated with plankton Zone N21. Here *Globorotalia tosaensis* is still consistently present, in the absence of *Globorotalia truncatulinoides*. *Globoquadrina altispira* and *Sphaeroidinellopsis* sp. occur in sample 136, thus showing an overlap in range with *Globorotalia tosaensis*. The first appearance of *Globorotalia truncatulinoides,* marking the base of Zone N22, is in sample 147. This species has an erratic range through the higher part of the section.

Globigerinoides sp. cf. *G. fistulosus* has its first occurrence in sample 75 in the Masa section and remains present till the top of the Kwa-Rioeapa section. The position of this first occurrence and the further range of the species through the sections is in general agreement with age assignments based upon the other marker species (Hays et al., 1969; Stainforth et al., 1975).

In conclusion, the combined Masa and Kwa-Rioeapa sections cover an age interval ranging from within Zone N19/20 to within Zone N22, which is from Late Pliocene to Early Pleistocene.

Paleobathymetry

The benthic foraminiferal content of all samples has been analyzed qualitatively and quantitatively. In general, the benthic foraminifera are well preserved and the species composition remains consistent throughout the Masa and Kwa-Rioeapa sections. As much of the sediment consists of gravitational influxes, much displaced faunal material can be expected. Therefore indicators of shallow marine environments, such as *Ammonia beccarii, Amphistegina lessonii* and *Elphidium* spp. are considered to have been displaced.

Based on published work on the upper depth limits and depth-ranges of benthic foraminiferal taxa (Frerichs, 1970; Pflum & Frerichs, 1976; Corliss, 1978; Moore et al., 1980; Van Morkhoven et al., 1986) and on comparison with modern benthic foraminiferal depth distribution in the Banda Arc (Van Marle et al., 1987; Van Marle, 1988), a paleodepth of 400–800 m is estimated for these Late Cenozoic deposits. Essential for this paleobathymetric interpretation is the presence of *Bolivina robusta, Heterolepa mediocris, Melonis affinis* and *Uvigerina proboscidea,* species indicative for upper to middle bathyal conditions. Absence of indicators of deeper waters, such as *Epistominella exigua, Laticarinina pauperata* and *Planulina wuellerstorfi* makes a lower bathyal environment unlikely.

Vertical movements

The assemblages of benthic foraminifera in the sections do not show any marked change in depth in the depositional environment during the Late Pliocene – Early Pleistocene. This apparent stability of the depth of deposition is in agreement with the sediment characteristics, which do not show any important change either. We therefore conclude that accumulation of preserved sediments in the basin was more or less in equilibrium with a general subsidence of the area, the rate of subsidence being of the same order as the sediment accumulation rate (Fig. 6).

The minimum rate of accumulation and subsidence can be estimated from the Early Pleistocene deposits in the Kwa-Rioeapa section. Since at least 550 m of sediments have been deposited during Zone N22, which has an estimated duration of 1.2 Ma (Berggren et al., 1985), the minimum rate is about 50 cm/ka. By extrapolation of this rate over the 250 m of sediments present below the N20/N21 Zone boundary in the Masa section, the minimum age of the base of this section can be estimated to be about 3.5 Ma. However, subsidence of the area must have started already before this date, because a certain time span is required to bring the paleorelief of the denudated Paleozoic rocks from above sea level to bathyal depth.

The rate calculated has to be considered as a

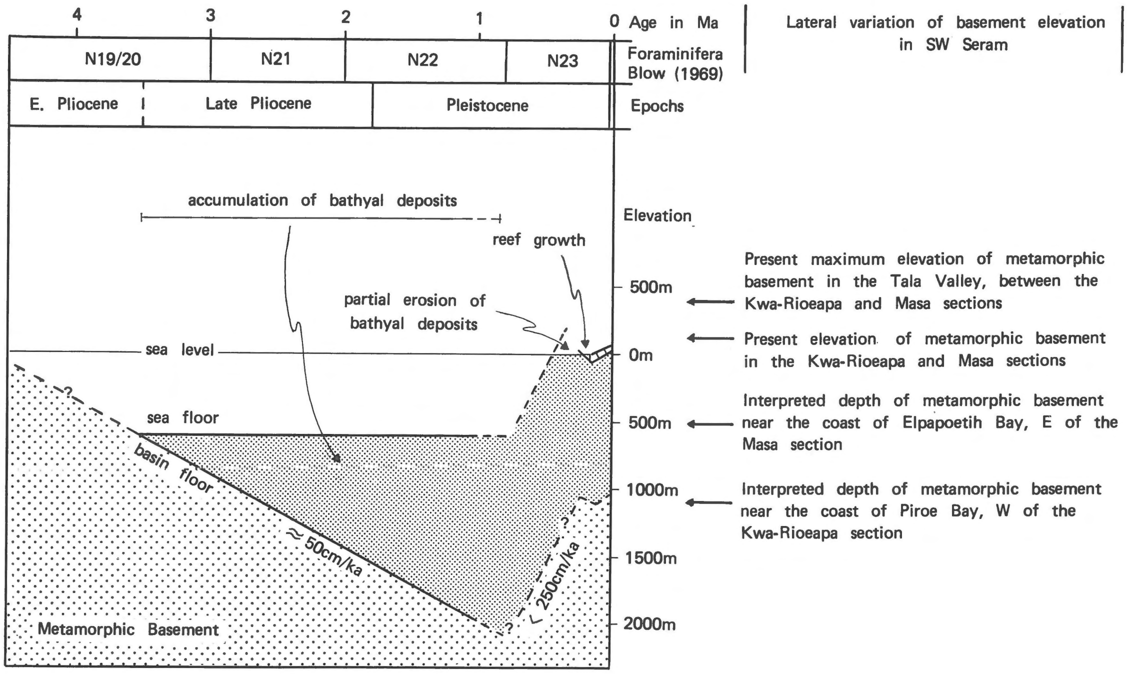

Fig. 6. Geohistory diagram for SW Seram. During the Pliocene and Early Pleistocene, the basement of the area was subsiding while sediments accumulated at upper bathyal depth. In the Late Pleistocene an uplift occurred, followed by erosion and reef growth. The present variation of basement elevation in the area is a consequence of the paleorelief below the basin deposits as well as of lateral differences in the history of vertical movements.

rough approximation because of the limited age constraints and because of a bias in the data set, used for its calculation. The bias is due to the wide lateral extent (up to 6 km) of the faintly dipping sections. Change of position in a horizontal direction may result in partially measuring lateral outgrowth of the sediment body, instead of its vertical accumulation. Furthermore, the vertical movements may have fluctuated from place to place within the basin, but trends thereof in our case remain undetected.

During the Pleistocene (N22–N23), subsidence of the area changed to uplift, which brought the basin deposits above sea level and finally into their present position. A model for this uplift in southwestern Seram (Fig. 7) illustrates that magnitudes of uplift differed laterally. In the central part of the area, the total amount has been in the order of 2100 m, while at the coastlines it has been less than 1000 m. Assuming that the uplift started late in Zone N22, the rates have been up to 250 cm/ka. We have no data regarding the offshore position of the basement, but, assuming that the trend of outward decrease of the rate of uplift continues beyond the coastline, then a line of zero uplift (hinge line) was located some 10–15 km offshore, while further away from the coast subsidence may have taken place. The only Late Pleistocene record consists of elevated reef limestones in the western part of the basin. They unconformably overly older basin deposits and show a slight tilt toward the coast. About 5 km inland they have a maximum elevation of about 50 m and near the coast the reefs occur just above sea level. Their inclination indicates that the youngest vertical movements in the area are still differential, but the small elevation of reefs near the coast as well as the partial erosion of the bathyal deposits before reef growth started, suggests that a change has occurred in the uplift pattern: The mean rate of uplift during the latest Pleistocene seems to have been less than during the initial uplift period (Fig. 6); some subsidence may even have

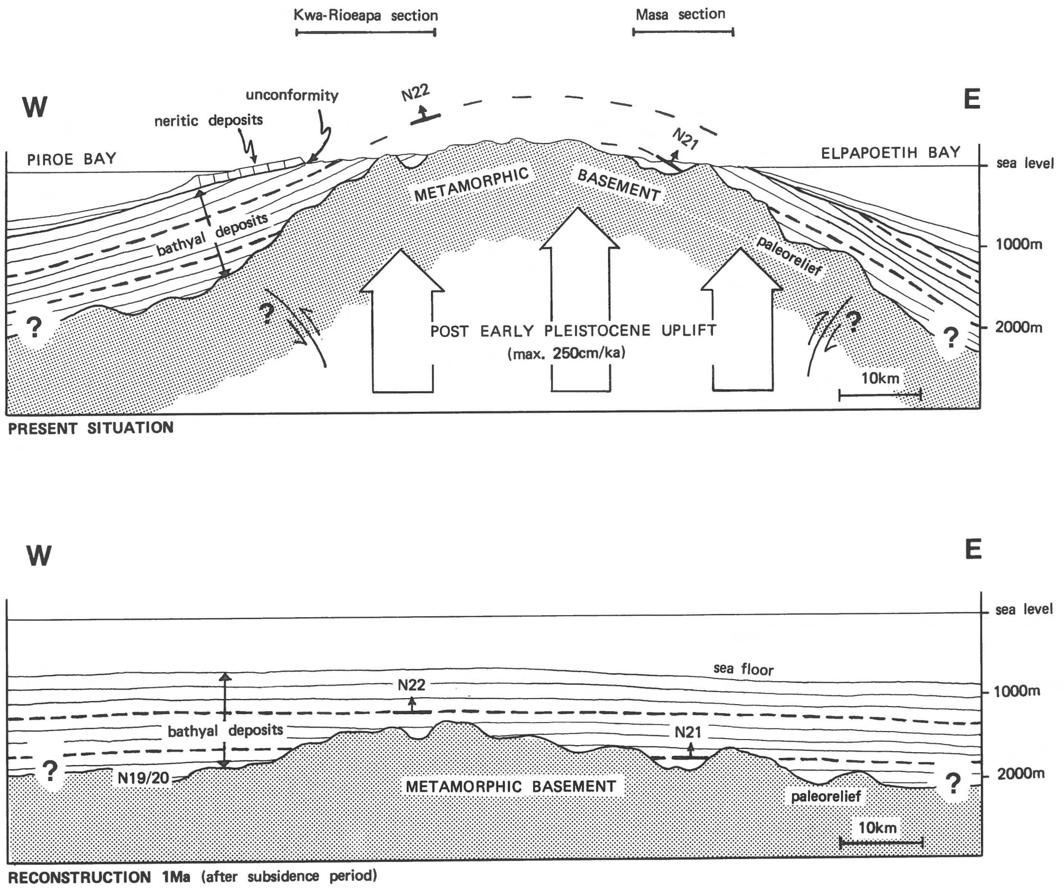

Fig. 7. Diagram, illustrating the lateral variation of Late Pleistocene uplift movements in SW Seram. Top: Present situation; the basin deposits are dipping away from the central area of Paleozoic rocks. Bottom: reconstruction at 1 Ma, when sediments accumulated at upper bathyal depth.

occurred before the start of reef growth. The fact that the reefs are hardly elevated near the shore, shows that the hinge line has shifted during the latest Pleistocene from some 10–15 km offshore towards the present coastline, and thus that the total area of uplift was reduced in size. Quaternary reef growth, however, is closely related with sea level fluctuations (Chappell, 1983; Shackleton, 1987), which have not been incorporated in the construction of the geohistory diagram (Fig. 6). The present stand of sea level is relatively high for the Late Pleistocene and this may result in an apparent decrease of uplift rate and an apparent reduction in size of the uplifted area.

Rutten (1927) already mentioned important young uplift movements in Seram. He suggested that the centre of the island, which now has an elevation of 3000 m, was just above sea level during deposition of the basin deposits. He furthermore noticed the presence of subhorizontal mountain crests at an elevation of 800–1200 m in the landscape of western Seram, which he interpreted as remnants of old peneplains. His data as well as ours indicate that Seram was a smaller island in the Early Pleistocene than it is now. However, during the Pliocene it had been subsiding and therefore it must have been larger before.

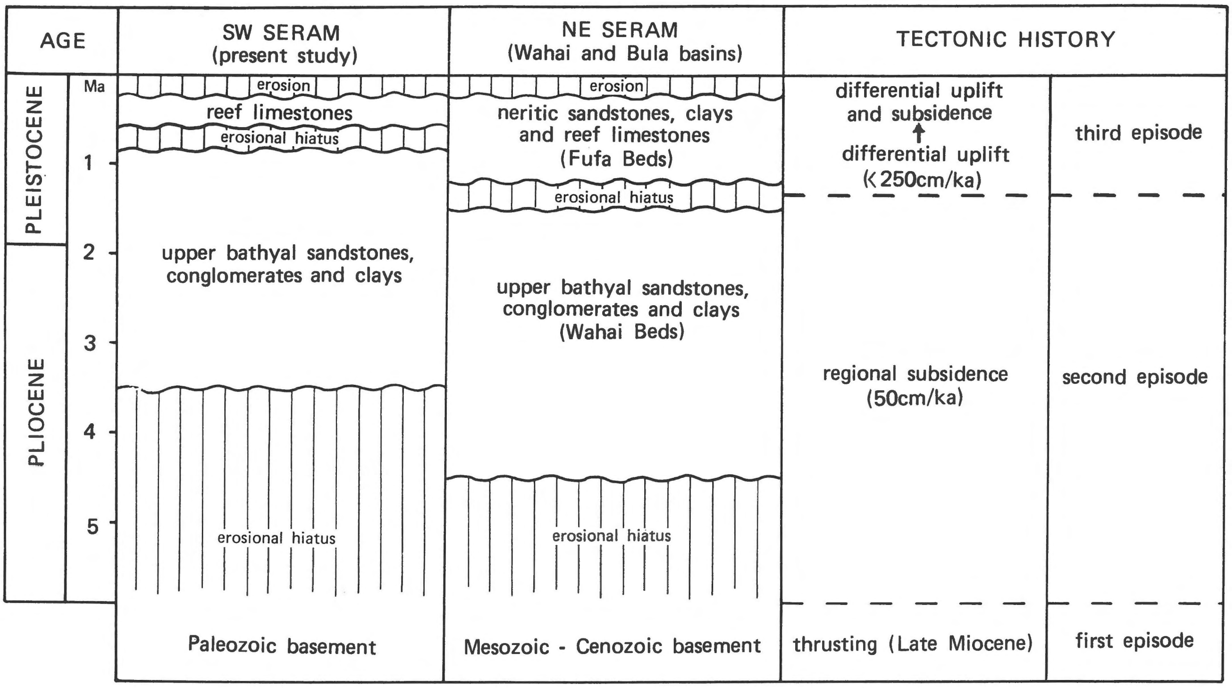

Fig. 8. Tectono – stratigraphic scheme for the Late Cenozoic of Seram, comparing the stratigraphic evolution of the basins in SW Seram (present study) with that of basins in NE Seram (Zillman & Paten, 1975). The tectonic history at both sides of the island is essentially the same.

Comparison to Northeastern Seram

Apart from the Late Cenozoic described in the previous sections, there are extensive Plio-Pleistocene deposits in northeastern Seram, in the oil producing Wahai and Bula basins (Fig. 2). Zillman & Paten (1975) and O'Sullivan et al. (1985) described these basins mainly from onshore and offshore drilling data and distinguished two stratigraphic units. A lower unit, the Wahai Beds, consists mainly of fine grained clastics of Early Pliocene to Early Pleistocene age, which were deposited at bathyal depths (Zillman & Paten, 1975). Age and environment of deposition of this unit are the same as the basin deposits in southwestern Seram (Fig. 8). The second unit, containing oil reservoir rocks, unconformably overlies the Wahai Beds and is named Fufa Beds. It consists of various shallow water deposits, including reef limestones, and is Late Pleistocene in age. In southwestern Seram a partial equivalent of this unit is found in the Late Pleistocene reef limestones.

Zillmann & Paten (1975) concluded that there has been an Early Pliocene to Early Pleistocene subsidence in NE Seram, which resulted in a marine transgression. Early Pleistocene uplift is recorded in neritic sediments. Finally, Late Pleistocene and recent uplift as well as subsidence, resulted in the present day basin configuration.

This history of vertical movements of northeastern Seram is essentially the same as the one we reconstructed for southwestern Seram. Both sides of the island show a marked change from subsidence to uplift during the Pleistocene, followed by unconformable deposition of neritic deposits and further tilting. The regional aspect of timing, magnitude and character of the movements indicates that the tectonic events on Seram in the Plio-Pleistocene were related to regional, large scale processes, such as the interaction of plates at the nearby plate boundary in the Seram Trough.

Tectonic episodes of Seram

Three tectonic episodes can be distinguished in the Late Cenozoic history of Seram (Fig. 8). The first one, before the origin of the basin deposits, concerns the thrusting of the basement. The youngest deposits involved in this thrusting are of Middle – Late Miocene age and belong to the Nief Beds, which are deep water limestones (Audley Charles et al., 1979). They are found in the central and eastern parts of the island. Deformation and uplift, however, may have started already before the Miocene, as is indicated by the presence of marine conglomerates in central Seram (Germeraad, 1946), which contain fragments of all rocks older than the Paleogene. Audley Charles et al. (1979) described similar clastic rocks under the name of Wai Tuh Beds, to occur steeply faulted in between high grade metamorphics. In both publications it is suggested that these clastics are of pre-Neogene age, which would mean that parts of Seram had already emerged and were eroded before the Miocene. Also on the island of Buru, just to the West of Seram, emergence and erosion occurred already in the late Paleogene and earliest Miocene (Fortuin et al., 1988), indicating that the area was tectonically active in that time.

The end of the episode of thrusting is marked by the deposition of a melange-like unit, known as the Salas Block Clay (Audley Charles et al., 1979). The unit consists of a variety of blocks, derived from the underlying thrustbelt, within a clayey matrix. O'Sullivan et al. (1985) assumed that the unit originated from a combination of tectonic mixing, olisthostromal processes and clay diapirism. According to Tjokrosapoetro et al. (1983b) and Gafoer et al. (1984), this melange is mainly restricted to eastern Seram. The Wahai and Bula basins lie partly on top of it.

The second episode contrasts with the first. It was a period of tectonic rest and subsidence, while erosional products accumulated at bathyal depths in the basins around the island. The episode started in the Early Pliocene and lasted till the middle of the Pleistocene. The Late Pleistocene constitutes the third and youngest episode in the tectonic evolution of Seram. It is characterized by renewed compressive deformation and uplift. The deformation is regional and has the character of tilting and folding with a wave length of several tens of kilometres and an amplitude of 1–2 km. The bathyal deposits were uplifted and unconformably covered by neritic sediments, which later became tilted in turn. This youngest episode not only differed in style of deformation from the first, Miocene episode, but also in the orientation of the stress field. The axis of folding in southwestern Seram is oblique to the thrustplane in the underlying metamorphic rocks (Fig. 7). Also in northeastern Seram the axis of uplift is somewhat oblique to older deformation directions, as indicated by the oblique relation of the basin deposits to the structural lines in the underlying basement (Fig. 2).

Plate tectonic aspects

Because the three tectonic episodes of Seram just discussed have regional significance, they must have been related to large scale tectonic processes. At the nearby plate boundary in the Seram Trough the continent of Irian Jaya presently is thrust against the island (Fig. 1). A reconstruction of plate motions at that boundary is necessary to interprete the history of tectonic events on the island itself (Fig. 9).

The present vector of motion of Irian Jaya with respect to the outer Banda Arc is not exactly known, but can be estimated: the motion of the Australian plate with respect to the Banda Arc area is about NNE, with an estimated rate of convergence of 70 km/Ma (Minster & Jordan, 1978). Irian Jaya is presently separated from the Australian plate by a left lateral strike-slip system, the so-called Tarera-Aiduna Fault (Fig. 9c; see discussions by Cardwell & Isacks, 1981; Nishimura & Suparka, 1986 and De Smet, in press). The resulting motion of Irian Jaya with respect to the Banda Arc area will be to the NW. This is about parallel to the axis of the Seram Trough and offers no simple explanation in terms of accretion for the origin of the island. In the Late Miocene, some 10 Ma ago, the Australian continent was situated further to the South, perhaps 700 km using the rate of conver-

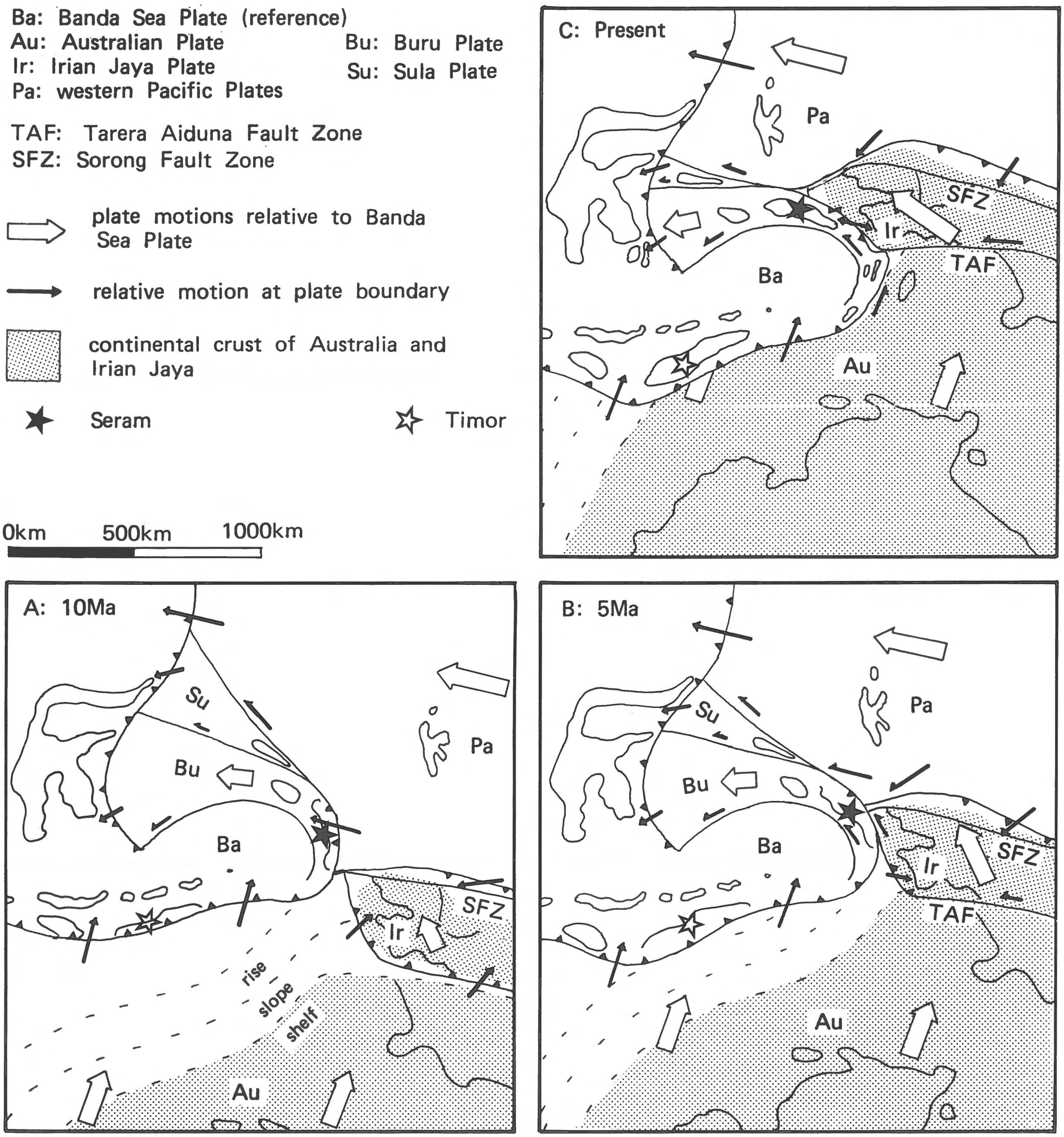

Fig. 9. Plate tectonic reconstruction of eastern Indonesia during the past 10 Ma, based on a geometrically consistent plate tectonic model by De Smet (in press). A: During the Late Miocene, Seram is in a compressive regime, caused by westward moving Pacific plates. B: During the Pliocene, the regime changes into a relatively quiet strike-slip system as a consequence of anticlockwise rotation of the island and northwestward motion of the Irian Jaya Plate. C: In the Late Pleistocene the continent of Irian Jaya collides with the Seram segment of the Banda Arc, leading to a new episode of compression.

gence given by Minster & Jordan (1978). At that time Irian Jaya was colliding already with the continent and therefore, very likely, was situated also further south (Fig. 9a). This means that the northern segment of the Banda Arc was not yet interacting with a NW moving Irian Jaya Plate, but with one of the western Pacific plates (or fragments thereof), such as the Caroline of Phillipine Plate. These plates have a motion vector similar to that of the large Pacific Plate, which, relative to Southeast Asia, is about WNW. The rate of movement can be estimated at about 100 km/Ma, on the basis of the global solutions for relative plate motions (Minster & Jordan, 1978).

Also the position of (proto-) Seram was different from the present one: Haile (1978, 1981) concluded an anticlockwise rotation of 74 degrees for Seram since the Late Miocene. We suggest that this rotation was the result of transport of Seram along the curved plate boundary of the Banda Arc (for details concerning the model behind this reconstruction, see De Smet, in press). If true, the Late Miocene position of the island was more to the southeast and its orientation about N-S (Fig. 9a), conform to Haile's findings (op. cit.).

This reconstruction of positions of Irian Jaya and Seram, indicates that in the late Miocene the tectonic regime at the plate boundary near Seram was compressional and that the materials, accumulating in the Seram belt, were derived from a plate at its eastern side.

In the Pliocene, the rotation of Seram resulted in a position more parallel to the plate motions in the western Pacific and also to the approaching Irian Jaya Plate (Fig. 9b). The tectonic regime at the Seram Trough changed from compressional to dominantly strike-slip and the compressional deformation of Seram decreased, as indicated by the second tectonic episode. We have no direct explanation for the subsidence during that episode, but assume that it was caused by regional crustal relaxation in response to the newly created situation.

Finally, the renewed compressional deformation of Seram in the Late Pleistocene is interpreted as caused by the push of the continent of Irian Jaya after it slowly moved alongside the island and started to collide (Fig. 9c). The new situation resulted in a reorientation of deformation on Seram, characterizing the third tectonic episode.

Volcanism near Seram is restricted to the islands of Ambon and Ambelau, just to the south of respectively Seram and Buru (Fig. 2). This volcanism is now extinct, but has been active in Middle – Late Pliocene times (K-Ar datings by Priem et al. 1978 and Abbott & Chamalaun, 1981 indicate magmatism between 4.5–3 Ma). This age is not in disagreement with subduction processes in the Late Miocene, as discussed above. For the Pliocene it shows that the subducted slab was still active, although deformation on the island of Seram had already ceased.

The plate tectonic reconstruction given above for the movements in the northern part of the outer Banda Arc differs fundamentally from that of the southern part of the arc, near Timor (Fig. 9a–c). Although the large scale structure as well as the stratigraphic compounds of Seram and Timor have much in common (Audley Charles et al., 1979), there are important differences in tectonic history. We speculate that these differences offer keys to the regional plate tectonic interpretation of the evolution of the Banda Arc and suggest that they should be given extra attention in future research.

Conclusions

Stratigraphic data from Late Cenozoic basin deposits on Seram show that tectonic conditions were relatively quiet during the Late Pliocene and Early Pleistocene and were associated with subsidence. In the basin studied, the rates of sedimentation and subsidence were in equilibrium, amounting to some 50 cm/ka. This situation reflects strongly decreased tectonic activity compared to the Late Miocene compressional deformation and is probably related to fundamental changes in plate tectonic conditions: rotation of Seram and northwestward movements of the continent of Irian Jaya changed a convergent regime in the Seram Trough into a dominantly strike-slip regime. The renewed compressional deformation of Seram in the Late Pleistocene has the character of open folding and tilting, as a result of which areas rise as well as subside.

Calculated rates of uplift for SW Seram are up to 250 cm/ka. Collision of the continent of Irian Jaya with the island arc is the probable reason for this youngest deformation.

Acknowledgements

Research has been carried out as a part of the Snellius-II Expedition, organized by the Netherlands Council of Oceanic Research (NRZ) and the Indonesian Institute of Science (LIPI). The research was financially supported by the Dutch Organisation for Pure Scientific Research (NWO), projects nrs. 76–83 and 76–85.

We are much obliged to Soebardjio Tjokrosapoetro who stimulated this research and to Jan E. van Hinte for his critical interest and his comments on the manuscript. This is a publication of the Geomarine Centre Amsterdam.

References

Abbott, M.J. & Chamalaun, F.H. 1981 Geochronology of some Banda Arc volcanics. In: Barber, A.J. & Wiryosujono, S. (eds): Geology and Tectonics of Eastern Indonesia – Geol. Res. and Dev. Centre, Bandung, Spec. Publ., 2: 253–271.

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 – Geol. Soc. London, 136: 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.

Blow, W.H. 1969 Late Middle Miocene to Recent planktonic foraminiferal biostratigraphy. In: P. Bronnimann & H.H. Renz (eds), Proc. First Int. Conf. Planktonic Microfossils – E.J. Brill, Leiden, 1: 199–422.

Bolli, H.M. & Saunders, J.B. 1985 Oligocene to Holocene low latitude planktic foraminifera. In: Bolli, H.M., Saunders, J.B. & Perch-Nielsen, K. (eds), Plankton Stratigraphy, Cambridge Earth Sci. Ser., Cambridge: 155: 262.

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

Cardwell, R.K. & Isacks, B.L. 1981 A review of the configuration of the lithosphere subducted beneath the eastern Indonesian and Philippine islands. In: Barber, A.J. & Wiryosujono, S. (eds), The geology and tectonics of eastern Indonesia. Geol. Res. Dev. Centre, Bandung Spec. Publ., 2: 29–47.

Chappell, J. 1983 A revised sea-level record for the last 300.000 years from Papua, New Guinea – Search, 14: 99–101.

Corliss, B.H. 1978 Recent deepsea benthonic foraminiferal distribution in the southeast Indian Ocean: inferred bottom water routes and ecological implications – Mar. Geol., 31: 115–138.

De Smet, M.E.M. 1989 A geometrically consistent plate tectonic model for eastern Indonesia. In: Van Hinte, J.E., Van Weering, Tj.C.E. & Fortuin, A.R. (eds): Proceedings Symposium Snellius-II Expedition (Jakarta, 1987), vol. I, Geology and geophysics of the Banda Arc and adjacent areas – Neth. J. Sea Res., in press.

Fortuin, A.R. 1986 Progress Report Snellius-II Expedition, Theme 1, Geology and Geophysics of the Banda Arc and adjacent areas, Campaign GF2, Buru, Seram and Kai islands; August – October 1985. Royal Neth. Acad. Arts and Sci. – Indon. Inst. of Sci.; 74pp.

Fortuin, A.R., De Smet, M.E.M., Sumosusastro, P.A., Van Marle, L.J. & Troelstra, S.R. 1988 Late Cenozoic geohistory of NW Buru, Indonesia. Geol. Mijnbouw, 67: 91–105.

Fortuin, A.R., De Smet, M.E.M., Hadiwasastra, S., Van Marle, L.J., Troelstra, S.R., & Tjokrosapoetro, S. 1989 Late Cenozoic sedimentary and tectonic history of South Buton, Indonesia – J. SE Asian Earth Sci., in press.

Frerichs, W.E. 1970 Distribution and ecology of benthonic foraminifera in the sediments of the Andaman Sea – Contr. Cushman Found. For. Res., 21(4): 123–147.

Gafoer, S., Suwitodirdjo, K. & Suharsono 1984 Geological map of Bula and Watubela islands quadrangle, Seram, scale 1:250.000, Geol. Res. Dev. Centre, Bandung (unpublished edition).

Germeraad, J.H. 1946 Geology of Central Seran. PhD thesis, Utrecht Univ. – De Bussy, Amsterdam, 136pp.

Haile, N.S. 1978 Paleomagnetic evidence for the rotation of Seram, Indonesia – J. Phys. Earth, 26, Suppl.: S191–S198.

Haile, N.S. 1981 Paleomagnetic evidence and geotectonic history and paleogeography of eastern Indonesia. In: Barber, A.J. & Wiryosujono (eds) The geology and tectonics of Eastern Indonesia – Geol. Res. Dev. Centre, Bandung Spec. Publ., 2: 81–87.

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

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

Minster, J.B. & Jordan, T.H. 1978 Present-day plate motions – Journ. Geophys. Res., Vol. 83, B. 11: 5331–5354.

Moore, G.F., Billman, H.G., Hehanussa, P.E. & Karig, D.E. 1980 Sedimentology and paleobathymetry of Neogene trench-slope deposits, Nias Island, Indonesia – J. Geology, 88: 161–180.

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

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

Pflum, F.B. & Frerichs, W.E. 1976 Gulf of Mexico deep-water foraminifera. Cushman Found. Foram. Res., Spec. Publ. 14: 7–125.

Priem, H.N.A., Andriessen, P.A.M., Boelrijk, N.A.I.M., Hebeda, E.H., Hutchison, C.S. Verdurmen, E.A.T., & Verschure, R.H. 1978 Isotopic evidence for a Middle to Late Pliocene age of the cordierite granite on Ambon, Indonesia – Geol. Mijnbouw, 57: 441–443.

Rutten, L.M.R. & Hotz, W., 1918–1920. Verslagen der Ceram Expeditie: I: Tijdschr. Kon. Ned. Aardrijksk. Gen, 2de serie, XXXV, 1, 1918. II: Tijdschr. Kon. Ned. Aardrijksk. Gen, 2de serie, XXXV, 2, 1918. III: Bull. Mij ter bev. v.h. Natuurk. Onderz. der Nederl. Kolon., 1918. IV: Tijdschr. Kon. Ned. Aardrijksk. Gen, 2de serie, XXXVI, 4, 1918. V: Tijdschr. Kon. Ned. Aardrijksk. Gen, 2de serie, XXXVI, 1, 1919. VI: Tijdschr. Kon. Ned. Aardrijksk. Gen, 2de serie, XXXVI, 1, 1919. VII: Tijdschr. Kon. Ned. Aardrijksk. Gen, 2de serie, XXXVI, 2, 1919. VIII: Tijdschr. Kon. Ned. Aardrijksk. Gen, 2de serie, XXXVI, 4, 1919. IX: Tijdschr. Kon. Ned. Aardrijksk. Gen, 2de serie, XXXVI, 5, 1919. X and XI: Tijdschr. Kon. Ned. Aardrijksk. Gen, 2de serie, XXXVII, 1, 1920.

Rutten, L.M.R., 1927 Voordrachten over de geologie van Nederlandsch Oost-Indie–Wolters Noordhof, Groningen: 732pp.

Shackleton, N.J. 1987 Oxygen isotopes, ice volume and sea level – Quat. Sci. Rev., 6: 183–190.

Stainforth, R.M., Lamb, J.L., Luterbacher, H., Beard, J.H. & Jeffords, R.M. 1975 Cenozoic planktonic foraminiferal zonation and characteristics of index forms – Univ. Kansas Pal. Contr., Art. 62: 425pp.

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

Tjokrosapoetro, S., Rusmana, E., Sukardi & Achdan, S., 1983a Preliminary geological map of the Ambon quadrangle, Maluku; scale 1 : 250.000 Geol. Res. Dev. Centre, Bandung (unpublished edition).

Tjokrosapoetro, S., Suwitodirdjo, K. & Achdan, A., 1983b Preliminary geological map of the Masohi quadrangle, Maluku, scale 1 : 250.000 – Geol. Res. Dev. Centre, Bandung (unpublished edition).

Valk, W. 1945 Contributions to the geology of West Seran. PhD thesis, Utrecht Univ., De Bussy, Amsterdam, 109pp.

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

Van der Sluis, J.P., 1950 Geology of East Seran. PhD thesis, Utrecht Univ. – De Bussy, Amsterdam, 71pp.

Van Hinte, J.E. 1978 Geohistory analysis – Application of micropaleontology in exploration geology – Am. Ass. Petr. Geol. Bull., 62: 201–222.

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

Van Marle, L.J. 1988 Bathymetric distribution of benthic foraminifera on the Australian-Irian Jaya continental margin, eastern Indonesia – Mar. Micropal. 13: 97–152.

Van Morkhoven, F.P.M.C., Berggren, W.A. & Edwards, A.S. 1986. Cenozoic cosmopolitan deep-water benthic foraminifera – Bull. Centr. Rech. Expl. Prod. Elf-Aquitaine (Pau, France), Mem. 11: 421pp.

Zillman, N.J. & Paten, R.J. 1975 Geology and petroleum prospects of Seram island, eastern Indonesia – J. Austr. Petr. Explor. Assoc., 15: 73–80.