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Cover Photograph:
Turbidites of Pitap Formation in Barito Basin, South Kalimantan.
Photo courtesy of J.T. van Gorsel from Houston, US.
Dear readers,

Our last issue of Berita Sedimentologi in 2013 will cover Borneo Island once again. We received nine high quality articles to be published in this Berita Sedimentologi No. 28, which include the most recent research findings on the Schwaner Mountains and its potential implication to sediment provenance on Sundaland (N. Setiawan et al.); and uplift history of the Meratus Complex as interpreted from palaeocurrents and sediment provenance (D. Witts). Other articles include geohistory analysis of South Makassar (P. Lunt & J.T. van Gorsel), Mass Transport Complex (MTC) control on deepwater sand deposition in Brunei offshore (H. Maulana & H.S. Hakimi) and Modern and Miocene ichnofacies in Mahakam Delta (E. Arifullah).

As usual we would like to appreciate all of our external reviewers for taking time to read through the articles prior to acceptance for publication in Berita Sedimentologi. In this issue, we specially thank Dr. Y.S. Yuwono for helping as invited external reviewer. We also have a new member in the editorial team. Farid Ferdian of ConocoPhillips Jakarta Indonesia joined us recently.

Coming up next is Berita Sedimentologi No. 29, which is planned for publication in April 2014. This is a special issue on SE Asia Biostratigraphy that will be championed by Han van Gorsel. We have received commitment to submit articles from various authors across the region and expect to give you a volume that will contain the most recent advances in understanding the biostratigraphy of this tectonically active region. We look forward to delivering the special issue to you all next year.

In the mean time, it is nearly time for year end holidays and we wish you Happy New Year 2014!

Minarwan
Deputy Chief Editor

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About FOSI

The forum was founded in 1995 as the Indonesian Sedimentologists Forum (FOSI). This organization is a communication and discussion forum for geologists, especially for those dealing with sedimentology and sedimentary geology in Indonesia.

The forum was accepted as the sedimentological commission of the Indonesian Association of Geologists (IAGI) in 1996. About 300 members were registered in 1999, including industrial and academic fellows, as well as students.

FOSI has close international relations with the Society of Sedimentary Geology (SEPM) and the International Association of Sedimentologists (IAS). Fellowship is open to those holding a recognized degree in geology or a cognate subject and non-graduates who have at least two years relevant experience.

FOSI has organized 2 international conferences in 1999 and 2001, attended by more than 150 inter-national participants.

Most of FOSI administrative work will be handled by the editorial team. IAGI office in Jakarta will help if necessary.

The official website of FOSI is:
http://www.iagi.or.id/fosi/

FOSI Membership

Any person who has a background in geoscience and/or is engaged in the practising or teaching of geoscience or its related business may apply for general membership. As the organization has just been restarted, we use LinkedIn ([www.linkedin.com](http://www.linkedin.com)) as the main data base platform. We realize that it is not the ideal solution, and we may look for other alternative in the near future. Having said that, for the current situation, LinkedIn is fit for purpose. International members and students are welcome to join the organization.
Late Triassic metatonalite from the Schwaner Mountains in West Kalimantan and its contribution to sedimentary provenance in the Sundaland

Nugroho Imam Setiawan¹,², Yasuhiro Osanai¹, Nobuhiko Nakano¹, Tatsuro Adachi¹, Lucas Donny Setiadji² and Joko Wahyudiono²

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ABSTRACT

This contribution presents petrography, geochemical characteristics and LA-ICP-MS U-Pb zircon dating from metatonalites in the Schwaner Mountains of West Kalimantan. The metatonalites mainly consist of plagioclase, biotite, quartz, apatite, muscovite, and titanite with relic clinopyroxene surrounded by hornblende. The geochemical characteristics show that the rocks have calc-alkaline affinities and were derived from subduction-related arc tectonic environment. Some of the metatonalites have adakite signature, which suggests the Schwaner Mountains were not formed by a consecutive subduction system. The result of LA-ICP-MS U-Pb zircon dating reveals that the metatonalite has magmatic age at 233 ± 3 Ma (Late Triassic), which is the oldest magmatic age in the Schwaner Mountains. Therefore, it strongly suggests that the Schwaner Mountains has significantly potential for important sedimentary sources in Sundaland not only from Cretaceous age but also from Triassic age as well as Tin Belt granites from Malay Peninsula.

Keywords: metatonalite, Schwaner Mountains, LA-ICP-MS U-Pb zircon dating, Late Triassic sedimentary provenance, Sundaland.

INTRODUCTION

Schwaner Mountains have been considered as Cretaceous granitoid pluton in the West Kalimantan (Haile et al., 1977; Williams et al., 1988; Amiruddin and Trail, 1993). It has important meaning to the tectonic setting of Sundaland and is considered as main supply of Cretaceous age detrital zircon to the surrounding areas through generated palaeogeography and palaeodrainage during the Late Cretaceous and Early Paleogene (e.g., van Hattum et al., 2006; Clements and Hall, 2008; Clements and Hall, 2011). The Permian to Triassic detrital zircon found in the sedimentary rocks around the Sundaland have always been considered as being derived from Tin Belt granite of Malay Peninsula (e.g., van Hattum et al., 2006; Clements and Hall, 2008; Clements and Hall, 2011).

This contribution reports the existence of metatonalite in the Schwaner Mountains, whose geochemical character can be differentiated from Cretaceous granitoids on the same location. Furthermore, we report LA-ICP-MS U-Pb zircon dating from metatonalite and discuss its correlation to the tectonic setting and contribution to sedimentary provenance in the Sundaland.

Mineral abbreviations in this paper follow Whitney and Evans (2010).

GEOLOGICAL BACKGROUND

Western to central parts of Indonesia region, particularly the Sunda Shelf, have low topography, free of seismicity and volcanism. The areas include Thai-Malay peninsula, NE Sumatra, Borneo, and NW Java. This tectonically quiet region forms the continental core of an area called Sundaland (Hamilton, 1979; Hall, 2002; 2009; Metcalfe, 1998).

Borneo Island has a Palaeozoic continental core in western part that is surrounded by ophiolite, island arc, and microcontinental crust accreted during Mesozoic (Hamilton, 1979; Wilson and Moss, 1999; Hall et al., 2008; Figure 1). The NW Kalimantan (Indonesian part of Borneo) domain includes fossiliferous Carboniferous limestones, Permo-triassic granites, Triassic marine shales, ammonite-bearing Jurassic sediments, and Cretaceous mélanges (Williams et al., 1988; Figure 1). NW Kalimantan domain was connected to the Thay-Malay Peninsula, which has a Proterozoic continental basement including tin-bearing Permian to Triassic granites and minor of Cretaceous granites (Hall et al., 2008). Williams and Harahap (1987) considered that NW
Kalimantan domain was allochtonously accreted continental terrane. In SW Kalimantan domain, Palaeozoic units are represented mainly by metamorphic rocks of Carboniferous to Permian age, although Devonian limestones have been found as river boulders in East Kalimantan (Hall et al., 2008; Figure 1). Cretaceous granitoid plutons, which associated with volcanic rocks, intrude the metamorphic rocks in the SW Kalimantan, which known as Schwaner Mountains (Williams et al., 1988; Figure 1). Apatite fission track ages indicate rapid exhumation of the granites in the Late Cretaceous (Sumartadipura, 1976).

Schwaner Mountains granitoids are composed of biotite-hornblende tonalite and granodiorite with minor mafic rocks and granite (Haile et al., 1977; Williams et al., 1988; Amiruddin and Trail, 1993). The granitoids formed a belt of 200 km wide and at least 500 km long, extending in an approximately E–W direction (Figure 1). Chemical analyses of typical rocks from the Schwaner Mountains indicate the I-type calc-alkaline nature of the suite (Williams et al., 1988; Amiruddin and Trail, 1993; Amiruddin, 2009). These rocks intruded into the low-grade metamorphic rocks during the Late Jurassic to Early Cretaceous and resulted in contact metamorphism (Williams et al., 1988). The progressive increase in metamorphic grade towards the intrusive contact with the tonalite bodies and the local occurrence of migmatite adjacent to them, indicate that the thermal metamorphism was due to the Early Cretaceous magmatism. The variations of contact metamorphic rocks are cordierite-andalusite-sillimanite hornfels, andalusite-biotite hornfels, and andalusite-sillimanite hornfels (Setiawan et al., 2013a). Foliations developed in the tonalite that was named as metatonalite (Amiruddin and Trail, 1993; Setiawan et al., 2013a). Strikes of schistosity of the metatonalites generally range from E–W to NE–SW (Amiruddin and Trail, 1993). It might be caused by deformation of the plutonic body after the magmatism (Setiawan et al., 2013a). Williams et al. (1988) particularly described as foliated tonalite.

Southward-directed subduction during Early Jurassic to Early Cretaceous is considered to be represented by granitoid plutons of the Schwaner Mountains (Haile et al., 1977; Williams et al., 1988; Amiruddin and Trail, 1993; Amiruddin, 2009). Furthermore, Zhou et al. (2008) proposed that the Mesozoic subduction belt in the North Kalimantan extends from Natuna Islands south-eastward along the Lupar River to the boundary between Sarawak and Kalimantan, turns eastward along the upper reach of the Kapuas River, and then turns northward in the Adio area of Sabah (Figure 1). The oceanic crust that subducted beneath West Kalimantan terranes, which has completely disappeared, is

Figure 1. (a) Situation map of Indonesia and the location of Schwaner Mountains in Borneo Island. (b) Simplified geological map of Borneo Island with the sampling localities of metatonalites. Modified from Wilson and Moss (1999); Hall et al. (2008).
considered to be beneath proto-South China Sea (Metcalfe, 1998; Zhou et al., 2008).

K-Ar ages of biotite and hornblende from the granitoids in the Schwaner Mountains range from 157 to 77 Ma (Jurassic to Cretaceous), while the northwest Kalimantan block ages are from 320 Ma to 204 Ma (Triassic to Carboniferous) (Haile et al., 1977; Williams et al., 1988; Amiruddin and Trail, 1993).

**SAMPLE DESCRIPTIONS**

The metatonalites are commonly found in the outer shell of the tonalite body (Figure 2a). Some of the metatonalite bodies are in layer with amphibolite. The rocks are predominantly composed of plagioclase, hornblende, biotite, quartz, apatite, muscovite, and titanite. Foliation texture is recognized from hornblende (0.5–2 mm), biotite (~0.2 mm), plagioclase (0.5–2 mm), and muscovite (~0.2 mm; Figure 2b). Some of the metatonalites have relict clinopyroxene that are surrounded by hornblende (Figure 2c). EPMA analysis confirms that the relict clinopyroxene has augite end-member in composition. Several samples contain numerous apatites.

Whole rock chemistries descriptions of metatonalites from the Schwaner Mountains have been discussed in detail by Setiawan et al. (2013b). Major, trace, and rare-earth elements composition of the metamorphic rock samples were analyzed by X-ray fluorescence spectrometry (XRF) using Rigaku ZSX Primus II and by laser ablation-inductively coupled plasma-mass spectrometry (LA-ICP-MS) with an Agilent 7500cx quadrupole ICP-MS with a New Wave Research UP-213 laser at Kyushu University, Japan on a fused glass disk (sample:flux ratio 1:2). The detail analytical conditions and procedure are given in Nakano et al. (2012).

Four samples of metatonalites were examined in comparison with Cretaceous granitoids data from Williams et al. (1988; Table 1). Based on the AFM diagram from Irvine and Baragar (1971), Setiawan et al. (2013b) suggested that the metatonalites are classified as calc-alkaline affinities rock similar with Cretaceous granitoids (Figure 3a). The rocks are plotted on the tonalite field in the CIPW normative An-Ab-Or ternary diagram from Barker and Arth (1979), whereas Cretaceous granitoids are scattered in the tonalite, granodiorite and granite fields (Figure 3b). Trace elements patterns (N-MORB norm. Sun and McDonough, 1989) obviously show negative anomalies of Nb and Ti as well as positive anomaly of Pb (Figure 3c), which are usually observed in subduction- and collision-related granitoids, and obviously have different pattern with Cretaceous granitoids. In the Rb versus Yb + Ta discrimination diagram from Pearce et al. (1984), the metatonalite are plotted on the volcanic-arc granite area similar with Cretaceous granitoids (Figure 3d). Moreover, Setiawan et al. (2013b) reveals that two samples of metatonalites were plotted on the adakite field using discrimination diagram of Sr/Y versus Y as proposed by Defant and Drummond (1990; Figure 3e). Whereas other metatonalites are plotted on island arc andesite–dacite-rhyolite (ADR) field together with Cretaceous granitoids from Schwaner Mountains (Figure 3f).

**Figure 2.** (a) Mode of occurrences of the metatonalite in the field. (b) Foliation of the metatonalite defined by biotite and hornblende. (c) Relict of clinopyroxene surrounded by hornblende. Scale-bars correspond to 1 mm.
LA-ICP-MS U-Pb ZIRCON DATING

Analytical method

One sample of metatonalite was examined by LA-ICP-MS U-Pb zircon dating. Zircon grains of analyzed rock were separated from #60 mesh whole-rock powder by panning, magnetic separations, and handpicking techniques. Internal structures of each analyzed zircon grains were observed using SEM (JSM-5310) equipped with cathodo-luminescence (CL) detector. U-Pb zircon dating was performed using Agilent 7500cx quadruple inductively coupled plasma mass spectrometer (ICP-MS), with a New Wave Research UP-213 Nd-YAG UV (213nm) laser ablation system (LA) installed at Kyushu University and with the analytical conditions following Adachi et al. (2012). Data acquisition and calibration of the isotopic ratios of zircon were processed using GLITTER software (Griffin et al., 2008). Data reduction and processing were conducted using the computer program of ISOPLOT version 3.75, provided by K.R. Ludwig at Berkeley Geochemical Center, University of California. Analytical errors shown in this study are 95% confidence levels (2σ error). Analyzed data for zircon grains are shown in Table 2.

Result of LA-ICP-MS U-Pb zircon dating

Zircon grains from metatonalite in Schwaner Mountains are 100–150 μm in size with elongated and subhedral shapes (Figure 4a). The grains show broad oscillatory and sector zoning under the CL images (Figure 4a). Most data are concordant and concentrated at ca. 230 Ma from 11 analyses of 11 grains and defined as concordant age of 233 ± 3 Ma (Figure 4b). The core domains (14 analyses from 10 grains) give older age of 246 ± 13 Ma and peak age distribution at ca. 235 Ma. The youngest age determined from the rim domain is 202 ± 13 Ma. It is true that oldest age is determined from core domain and youngest age is from rim domain of zircon grains. However, most of the data are overlapping each other regardless of the domain (Figure 4b). Therefore, the data are likely to give magmatic age of this rock at 233 ± 3 Ma (Late Triassic).

Table 1. Major (wt%), trace and rare-earth elements (ppm) value of metatonalite and granitoids from Schwaner Mountains of West Kalimantan

<table>
<thead>
<tr>
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<td>MgO</td>
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<td>99.56</td>
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</table>

FeO*, total Fe as Fe2O3. LOI and ash means loss on ignition and non-detection, respectively.
DISCUSSION

Implications for subduction-related felsic magmatism during Mesozoic

The youngest U-Pb age of magmatic zircon in the metatonalite from Schwaner Mountains is 202 ± 13 Ma. This age is older than K-Ar age of granitoids from previous researchers, which ranges from 157 to 77 Ma (Haile et al., 1977; Williams et al., 1988; Amiruddin and Trail, 1993). The magmatic zircon of metatonalite in this study has concordant age of 233 ± 3 Ma. The result gives older age than previous K-Ar granitoids in the Schwaner Mountains (Figure 4c) but in range with the K-Ar ages of granitoids from northwest Kalimantan domain, which ranges from 320 to 204 Ma (Williams et al., 1988; Figure 4c).

Figure 3. Geochemical analyses of the metatonalites and their comparison with Cretaceous granitoids (data from Williams et al., 1988) from the Schwaner Mountains (modified after Setiawan et al., 2013b). (a) The metatonalites are plotted on the AFM diagram from Irvine and Baragar (1971). (b) The metatonalites are plotted on the CIPW normative An-Ab-Or classification diagram (Barker and Arth, 1979). (c) N-MORB normalized (Sun and McDonough., 1989) trace element diagram of the metatonalites. (d) Discriminant diagram of Rb vs Yb + Ta (Pearce et al., 1984) for the metatonalites. (e) Two metatonalites fall on the adakite field in the discriminant diagram of Sr/Y vs Y from Defant and Drummond (1990).
Furthermore, the presence of metatonalite that have adakite signature in the Schwaner Mountains suggests that the Schwaner Mountains was not formed by a consecutive subduction system to build the granitoids because genesis of adakitic magma is different from that of non-adakitic rocks.

Table 2. LA-ICP-MS U-Pb isotope ratios and calculated ages of zircons

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<th>Samples/Analyzes No.</th>
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<th>Calculated ages and errors (Ma, 2σ)</th>
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<th>Dis (%)</th>
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<td>202 13 202 52 0.405 0.0621</td>
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Figure 4. (a) Cathodoluminescence representative images of magmatic zircon grains in the metatonalite. Circles indicate LA-ICP-MS U-Pb analysis spots. Scale bars indicate 30 µm. (b) U-Pb concordia plot for magmatic zircons of the metatonalites. (c) Probability distribution diagram of magmatic zircon in the metatonalite in comparison with Cretaceous granitoids from the Schwaner Mountains and NW Kalimantan domains.
Adakite was introduced by Defant and Drummond (1990) from the specific rock types from Adak Island in Aleutian Islands that are associated with subduction of hot, young (≤ 25 my) oceanic lithosphere. However, the magma genesis of adakite is still controversial and could be any of the following possibilities: slab melting due to the young (hot) oceanic plate; ridge subduction; and magma differentiation (Castillo, 2006).

Newly-found Late Triassic age metatonalite with adakitic signature strongly suggests that the subduction mechanism and felsic magma genesis changed between Early Triassic and Cretaceous. Some of the metatonalites, except adakite rock, show similar signature to Cretaceous granitoids, indicating that the subsequent metamorphism occurred during the Cretaceous subduction system. Metcalfe (1998) and Zhou et al. (2008) proposed that Mesozoic subduction of proto-South China Sea beneath West Kalimantan was responsible to build this formation. The proto-South China Sea has been regarded as a short-lived marginal sea, perhaps related to the backarc extension in Mesotethys domain (Zhou et al., 2008). Hence, the adakitic signature in the metatonalite might be derived from magma as a result of the melting of proto-South China Sea oceanic lithosphere, whose crust was still young (hot) at the time. As the southward-directed subduction of proto-South China Sea was taking place during Cretaceous to form granitoids in the Schwaner Mountains (Lupar-Adio belt; Williams et al., 1988; Amiruddin, 2009), the Indo-Australian plate came from southeast and subducted below southeast Kalimantan forming magmatic events in the Meratus Mountains (Meratus belt; Yuwono et al., 1988; Amiruddin, 2009). However, this double subduction event during Cretaceous in Borneo is still controversial and whether or not the Lupar-Adio belt is connected to the Meratus belt is unknown at present.

Other possibilities are the magmatism of granitoids in several parts of the Schwaner Mountains might be contemporaneous with the granitoids in the northwest Kalimantan domain (K-Ar: 320–204 Ma; Williams et al., 1988) or East Malaya province I-type granitoids (U-Pb: 267–80 Ma; compiled in Searle et al., 2012). This means that the subduction, which generated Triassic granitoids, was eastward below the Indochina or East-Malaya craton (Searle et al., 2012). The occurrence of adakitic magma might imply that the first generation of melting oceanic crust of Paleo-Tethys started from the Schwaner Mountains area and the next generations continued to the northward of Thai-Malay peninsula with or without adakitic signatures. However, detailed studies should be done to confirm the occurrence of adakitic rocks of Triassic age within Indochina or East-Malaya craton and NW Kalimantan domain.

Contributions for the sedimentary provenance in Sundaland

Many publications considered that the Permian–Triassic ages of detrital zircon from Sundaland particularly Borneo Island were supplied from sediment derived from Tin Belt granites in Malay Peninsula (e.g., van Hattum et al., 2006; Clements and Hall, 2008; Clements and Hall, 2011). Clement and Hall (2008; 2011) proposed early Cenozoic palaeogeography and palaeodrainage models in the southern Sundaland based on the study of detrital zircons from West Java. In their models, there is no contribution of sedimentary provenance from the Schwaner Mountains to West Java during Middle Eocene, however they found detrital zircons with Permian to Triassic signature in the Middle

![Figure 5. Suggested Middle Eocene palaeogeographical map of the Sunda Shelf region (modified after Clements and Hall, 2011). Schwaner Mountains might have contributed to sedimentary provenance together with Tin Belt granites from the Malay Peninsula.](image)
Eocene forearc sandstones. The contribution of Schwaner Mountains as sedimentary provenance in the West Java started in Late Eocene based on the appearance of Mid-Cretaceous detrital zircons in all Upper Eocene and Lower Oligocene formations (Clements and Hall, 2011). Result of LA-ICP-MS U-Pb zircon dating from metatonalite in the Schwaner Mountains in this study reveals the magmatic age of Late Triassic (233 ± 3 Ma). Therefore, it strongly suggests that the Schwaner Mountains has significant potential for important sedimentary sources not only from the Cretaceous age but also from the Triassic age as well as Tin Belt granites from Malay Peninsula (Figure 5).

CONCLUDING REMARKS

1. Schwaner Mountains were not formed by a consecutive Cretaceous subduction system. The subduction mechanism and felsic magma genesis changed between Early Triassic and Cretaceous and subsequent metamorphism occurred during the Cretaceous subduction system.

2. LA-ICP-MS U-Pb zircon dating from metatonalite in the Schwaner Mountains of West Kalimantan reveal magmatic age of Late Triassic (233 ± 3 Ma). It might contribute to the provenance of Triassic age detrital zircon in the Sundaland.

ACKNOWLEDGEMENTS

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Geohistory analysis of South Makassar

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ABSTRACT

A quantitative approach to stratigraphy shows how the south Makassar Straits Basin developed, and how the major sequences of sediment were deposited. After the initial rifting of the grabens there were four major, basin changing, unconformities that strongly affected sedimentation; at about 39 Ma (later Middle Eocene), 36 Ma (early part of Late Eocene), 34 Ma (almost on the Eocene to Oligocene boundary) and about 24 Ma (almost at the Oligocene-Miocene boundary).

The 39 Ma event saw accelerated rifting on the distal margins of Sundaland, in the south and southeast. The 36 Ma event is very strongly expressed in the South Makassar area and is shown by the geohistory analysis to have included basement subsiding by as much as 2 kilometres in 2 million years. The 34 Ma event led to the onset of the Berai Limestone to the west and very strongly reduced rates of deep-water sedimentation throughout the Makassar to Spermonde area. The c. 24 Ma event saw a substantial change in sedimentary conditions, with the end of condensed deep marine clastics and shallow marine Berai carbonates, and the start of a new clastic phase with high rates of sedimentation. These times of major change are all thought to be controlled by tectonism, because their magnitude exceeds rates of eustatic change, and also because their geographic expression is highly variable.

The first rift phase contains important lacustrine hydrocarbon source rocks in the inland areas of Sundaland, and large deltas with coaly sources on the fringes of the continent. The subsidence and associated sediment supply history has impact on the later deposition of coaly deltas, and the accumulation of overburden and thermal maturity of buried sediments.

Keywords: geohistory analysis, south Makassar, Makassar Straits.

INTRODUCTION

Many wells in the south Makassar Straits have their largest and most abrupt contrast in environmental indicators within the Eocene section, showing a sudden and strong transgression to fully bathyal conditions. This review examines four wells using geohistory techniques (van Hinte, 1978) to quantify this major change. It also considers the implications from this stratigraphic analysis on the large scale, basin architecture-related, sequences of sedimentation.

The Pangkat-1 [P-1] well drilled the complete Eocene to Recent sedimentary succession in the West Makassar Graben (Figures 1 & 3), reaching Cretaceous igneous basement at total depth [TD]. It is possible to carry out geohistory analysis from published well data (Djamil and Bani, 2005; Pireno et al., 2009; Kupecz et al., 2013). However this well was drilled on a palaeo-high, and therefore the lowest part of the graben-fill is condensed. To compensate for this, analysis can be extrapolated to the adjacent Makassar Straits-1 [MS-1] well, which drilled the graben centre but stopped within the Late Eocene sediments. Analysis at MS-1 can be extended to basement using published seismic data, tied to P-1, and a velocity - depth function to estimate the deepest section between drilled TD and basement.

The seismic line through P-1 and MS-1 crossing the West Makassar Graben, and almost reaching Takatalu-2 [TT-2] in the NE, was described in Djamil and Bani (2005), Figure 3 here. This graben is a branch of the south Makassar Straits rift basin (see Figure 10 in Kupecz et al., 2013). On the cross-section (cf. Figure 6 of Djamil and Bani, 2005, and Figure 3 here) some of the sedimentary facies, such as biothermal reefs and lacustrine muds, have been proven by wells, and these facies also have distinct seismic character. The inversion anticline drilled by the MS-1 has folded a deep marine carbonate debris-flow described in Pireno et al. (2009). This is the trap and reservoir for the Ruby Gas Field. However, in contrast to these seismically distinct facies, the majority of sediments in the wells, and in the plane of the cross-section, are deep marine claystones. It is the contrast between non-marine sediments, the first marine transgression, later deep marine deposition, and the shallow marine reefs outside the graben, that is the subject of this review.
This study also contrasts the geohistory of the West Makassar Graben with the Spermonde Graben to the south and, for comparison, the relatively stable horst of the Doang-Bone Platform. The Lombosang-1 [L-1] well drilled the centre of the Spermonde Graben, at present water depths of almost 1000m. Considerable and sudden subsidence is noted there, but at a different time to the main event in the West Makassar Graben. The Bone-1 well [B-1] on the stable, horst, platform naturally shows much less subsidence, and this well drilled shallow marine carbonate for almost the entire section. Even in this well there is an Oligo-Miocene stratigraphic anomaly that may be correlatable to the other wells.

GEOHISTORY ANALYSIS PANGKAT-1

The geohistory diagram (Figure 2) begins with lacustrine sediments deposited over Cretaceous andesitic basement. Oil, sourced from this lacustrine unit, was tested out of the weathered basement. At the well location these organic-rich beds were laid down over a palaeo-high within an ancient lake, which extended all around the P-1 location. The seismic response associated with this facies is a distinct package of strong, continuous parallel reflectors that is typical for organic lacustrine beds (cf. seismic in Sukanto et al., 1998; Asri Basin of SE Sumatra). This unit is difficult to date as the palynoflora is dominated by freshwater Botryococcus & Pediastrum algae, and has sparse age diagnostic flora or fauna. The presence of Meyeripollis naharkotensis down to the deepest sample (2414m) is indicative of an age no older than Late Eocene for a substantial part of this section (some caving into deeper samples is possible). The lowest marine beds immediately overlying the distinct lacustrine facies contain the dinoflagellate Duospheridium nudum (a Middle Eocene restricted taxon, Williams & Bujak, 1985), and Glaphyrocyusta ordinata (extinction near top Middle Eocene, Williams & Bujak, 1985), both present as single specimens. In the next 60m of mudstone above this initial marine transgression silty mudstones are found with Nummulites and Pellatispira that in this, and other wells nearby, are dated as Tb Letter Stage (see illustrations of advanced Pellatispira species in the TT-1 biostratigraphy report) The Tb Letter Stage is roughly equivalent to the Late Eocene (the Middle to Late Eocene boundary here being defined on the top Ta and P14 mass extinction of various larger and planktonic foraminifera at about 37 Ma; Berggren et al., 1995, Lunt & Allan, 2004, GTS04).

At about 2286m measured depth in this well there is a marked log-break reflecting an abrupt change in lithology. Overlying this event the mudstones are rich in planktonic foraminifera as well as fully

Figure 1. Location map.
bathyal benthic foraminifera such as Planulina wuellerstorfi, several Melonis species, Osangularia species, Gyroidina soldanii, Globobulimina pyrula, Uvigerina peregrina and Cyclammina species. Late Eocene planktonic foraminifera and nannofossils are found as high as 1963 m, which is the P17 / NP20 extinction event [c 33.8 to 34.0 Ma, Gradstein et al., 2004; Wade et al., 2011]. Deep marine (deep outer neritic to upper bathyal) conditions continued up hole and through time to at least the end of the Oligocene (about 1524 m MD), above this depth a new and accelerated influx of prograding clastic sediments (cf. Kupecz et al., 2013, Figure 18) led to gradually shallowing of the sea floor at this site, leading to the present mid-middle neritic setting (64 m water depth). In the younger section there is, at least, one uplift associated with inversion of the graben-defining fault, which is why the correlations deeper in the wells are referenced to the top Eocene event. At P-1 there could have been two slight uplift/relative sea-level fall events, at about 790 m and 427 m in the well. Later Pliocene to Recent sea-level fluctuations may have occasionally reduced this site to an inner neritic setting (less than 20 m).

Although it is sensitive to the sediment compaction calculation, the geohistory diagram shows the transition from bathyal to neritic conditions at P-1 (uphole; c. 200 m water depth; Hedgpeth, 1957 & Ingle, 1980) occurred during deposition of the Lower Warukin Formation. This matches the transition based on benthic fossil assemblages, which therefore gives some confidence in the proposed basement subsidence curve for the past 15 Ma (Figure 2).

Therefore the first problem for this geohistory review is to estimate how deep and how fast basement at P-1 subsided. This is a balance between an initial sudden subsidence and a later, longer, “sag” phase. As noted above, the contrast in environment-related benthic fauna is, by far, greatest at the 36 Ma event in the well samples. The noted changes at about 790 m and 427 m (mid Miocene) are slight regressive events, of much smaller magnitude than the 36 Ma contrast. A review of other wells reaching Eocene or basement in this area has found that many wells in the southern Makassar Straits have by far their largest palaeoenvironmental shift in their sedimentary history within the early part of the Late Eocene, soon after transgression over basement. This transgression led to bathyal deposition almost immediately overlying basement or very thin inner neritic facies.

**GEOHISTORY ANALYSIS OF MAKASSAR STRAITS-1**

The cross section (Figure 3) shows that during the first rift phase the organic rich lacustrine deposits, with their distinct acoustic/ seismc response, thickened into the graben, as did the subsequent transgressive, shallow marine clastics (it does not affect the calculation here if such thicker transgressive deposits are considered deeper marine in the graben than those sampled by P-1). The key feature at the MS-1 location is the correlation to the abrupt lithofacies change at 2286 m MD in P-1, the 36 Ma onset of bathyal deposition, which is between 2965 m and 3258 m ss in MS-1. The deeper number is the author’s preferred seismic correlation and will be used in the estimates below, but it should be borne in mind that there are alternative seismic correlations.

![Figure 2. Geohistory diagram for Pangkat-1 [P-1].](image-url)
Figure 3. A seismic line and cross-section between P-1 and MS-1 wells (seismic from Djamil and Bani; 2005).
that could be as much as 300m shallower. The next sequence boundary above this seismically correlated event is one that was reached by drilling in the MS-1 well. In most wells in the south Makassar area, the Eocene to Oligocene boundary (c. 34 Ma) is marked by a log break with lower gamma ray response above, coinciding with the first appearance, downhole, of Eocene foraminifera and nannofossils in cuttings. This is a good correlation horizon to hang older Eocene stratigraphy from, as it is necessary to remove the thickness-changing effects of the mid-Miocene inversion, which formed the anticline drilled by MS-1. In P-1 the thickness of sediment from this top Eocene event to the 36 Ma log break is 323m. In MS-1 it is 1508m, of which 1004m was proved by drilling. From the top of lacustrine seismic reflectors there are 1920m of sediment, entirely Late Eocene, Letter Stage Tb, in age.

This data means that at about 36 Ma [top of the shallow marine strata] subsidence occurred, and a depocentre made, so that by the end of the Eocene (c. 34 Ma) at least 1508m of sediment was deposited, but this sediment did not fill the depocentre to sea-level. In fact it did not even fill it to within 200m of sea-level, above which depth some outer neritic assemblages would be expected to have been seen in MS-1 palaeontological analysis. This simple arithmetic does not take into account the effect of sediment compaction, which is shown on the geohistory diagram (Figure 4). Removing the effects of post-Eocene overburden compaction, the actual thickness of Late Eocene sediments above the transgression, at the end of the Eocene, was about 1920m (calculation from Genesis modelling software). Therefore at the very least, basement subsidence must have been a minimum of about 2100m needed to keep the benthos out of neritic influences, within 2 Ma.

The cross-section from MS-1 to the carbonate platform in the NE (Figure 3) also gives geohistory data. While velocities in the limestone will differ from the claystone-rich data used to convert the P-1 and MS-1 area there is additional well control from the TT-2 well a short distance to the NE (Djamil and Bani, 2005; Kupecz et al., 2013). Both the seismic data and this well have the near end Oligocene termination of the Berai Limestone reef deposition (which is the same age as the terminal debris carbonate in the MS-1, Pireno et al., 2009)
at a depth of 990m. Also the first, intra Late Eocene marine transgression over basement is at 1585m in TT-2 (there are good photographs of Tb diagnostic fossils in the biostratigraphy report for this well). What this shows is that the correlative first marine transgression, occurring slightly before the 36 Ma subsidence, would have been offset in the graben by the amount of throw on the graben-defining faults. Being located on the horst, only much more gentle, long-term subsidence was seen at TT-2, like Bone-1 and Doang-1 wells to the south.

So the transgression horizon at 1585m in TT-2 is now displaced to 3258m at MS-1, a vertical offset of 1673m. The vertical displacement relative to Pangkat-1 is 700m. This simple subtraction is not affected by sediment compaction, and is just the fault displacement remaining now, without considering the fault inversion in the middle Miocene, which produced the fold at MS-1. Inversion uplift is thought to be negligible at Pangkat-1 as it is outside the fold drilled at MS-1, but sudden slight uplift in mid Miocene times is noted in the P-1 well section. It is hard to estimate, and thereby compensate for, the amount of inversion movement at MS-1 from the seismic and cross-section because the anticline at MS-1 is a “Sunda fold” (Eubank & Makki, 1981). In such structures shallower beds are folded more than deeper strata, and fault movement is not a simple vertical vector. The top Oligocene, top Berai equivalent, and overlying beds have been uplifted by about 640m to form the anticline at MS-1. Uplift of the unfolded Eocene horizons may have been only a fraction of this. The value of this cross-sectional estimate is that even though it cannot differentiate Eocene subsidence from any later sag, or the known inversion uplift, it supports the earlier estimation that the West Makassar Graben has seen over a kilometre and a half of subsidence, which the first method suggests was mostly within the short-lived Late Eocene.

GEOHISTORY ANALYSIS OF LOMBOSANG-1

The Lombosang-1 [L-1] was drilled in 2003 in 939m water depth (Figure 5). Samples at TD, not reached by logging tools, had high vitrinite reflectance indicative of metasedimentary basement. Seismic also indicates basement at TD. Like Pangkat-1 this well was drilled on a palaeo-high so the very oldest Cenozoic sediments were not sampled by the well. The deepest samples are non-marine and contain only general Cenozoic palynomorphs. About 200m above basement the well samples contain Middle Eocene and younger Florschuetzia pollen and there is a further 800m of lower coastal plain to paralic sedimentary section indicating both significant graben subsidence already within the Middle Eocene (see shallower age constraints below), as well as a high rate of sedimentation. At about 2550 to 2650m in the well there are well bedded coals, visible on wireline logs, and on the seismic data over a wide area in

Figure 5. Stratigraphic summary of Lombosang-1.

Above the coaly section there are less than 50m of marine beds with shallow marine limestone, containing larger foraminifera. At 2480m sample depth (1513m below the present sea floor) there is a log break at the top of shallow marine limestones. Above this there are deep marine mudstones (outer neritic to upper bathyal, passing
upwards into fully bathyal] with morphologically distinct P14 [and older] planktonic foraminifera. These are the keeled Morozovellid forams which became extinct at 38 Ma (Wade et al., 2011).

Bathyal sediments are present upheole for the rest of the well that was sampled, continuing to the present day sea-floor conditions at 939m depth. Put simply, paralic coals deposited slightly above sea-level were initially transgressed by shallow, photic limestones and then were suddenly subsided to very deep marine conditions, where they have stayed ever since.

The geohistory diagram for this well (Figure 6) is less constrained than MS-1 where a known thickness of later Eocene sediment requires a greater accommodation space to have been formed to contain it. At L-1 the ages of the non-marine sediments is poorly known, allowing the far-left data points on the geohistory diagram to have a range of possible positions along the X axis. The unknown thickness of the water column after subsidence (unlike MS-1) allows flexibility in how these points are plotted along the vertical, Y axis. It is possible to draw the diagram so the basement subsidence curve is a single trend, starting in early Middle Eocene times, and it was only at about 39 Ma that subsidence exceeded sediment supply resulting in rapid marine transgression. A comparative review of all geohistory plots for all the wells in the area is required to assess this, including such wells as ODB-1x which did not reach basement but shows high rates of later Eocene bathyal sedimentation after the 39 Ma transgression/subsidence.

Perhaps the most important indicator that an abrupt acceleration event in a longer term subsidence occurred at about 39 Ma is the fact that over more than 100kms across the Spermonde Graben, and also in a second graben some 200 kms to the southwest, where the ST Z-1, SG P-1, L 46-1 &-2 and Sawangan-1 wells are located (Figure 1), a marked environmental contrast from shallow marine to very deep marine occurs one or two hundred meters below the zone P14 extinction datum. There is no sign of diachronity, and some of the more distal wells such as ODB-1x and Kelara-1 have the thickest post 39 Ma Eocene marine sediment sections. It is also significant to note that Phillips et al. (1991) described, in this southwestern area, a marked angular unconformity between the older Middle Eocene “pre-Ngimbang” and the rapid marine transgression that occurred over it, post 39 Ma. The existence of such an angular unconformity at the same time as accelerated basement subsidence points to a regional tectonic event rather than a diachronous flood in steadily subsiding rifts.

**GEOHISTORY ANALYSIS OF BONE-1**

The Bone-1 [B-1] well drilled shallow marine carbonate, apart from a thin (<50m) clastic and coal section over weathered Cretaceous lavas at TD. As with other wells drilling the porous carbonate platforms (Doang-1, Manuk-1 and Paternoster-1) circulation was lost many times, and cuttings samples have numerous gaps. Unlike the other wells, the B-1 had a large number of sidewall cores taken and analysed for biostratigraphy, so a fairly reliable stratigraphic summary is possible.

The geohistory diagram (Figure 7) has been annotated with the main points of this analysis. In brief the later Miocene to Recent marker Alveolinella quoiyi is seen in cuttings and sidewall cores down to 809m, below which there is either a considerably condensed section or an unconformity. The distinct older Miocene marker Miogypsina is not found at all in this well and when an influx of older foraminifera are recorded at 887m and below, only older nephrolepidine Lepidocyclina are found, which in the absence of Miogypsina suggests the carbonates here are already Late Oligocene in age. The Oligocene carbonates continue for most of the rest of the well section although in the very lowest part of the limestone some Eocene Discocyclina are recorded, and in the very deepest carbonate sample the latest Eocene index Pellatispira. Sampling conditions are too poor in both Doang-1 and Manuk-1 to contribute any useful data.

The paucity of Te5 and lower Tf in Bone-1 appears, based on initial review, to be paralleled by a reduced number of limestone outcrops of this age, relative to thick Oligocene and Late Eocene sections, in the Tonasa Limestone to the east (Wilson, 1995; who also noted a peak in downslope reworking of carbonate at this time), and in the carbonates of Takatalu-1 and -2 wells to the north.

**SUBSIDENCE OVER A WIDER AREA**

As noted above, the transgression of bathyal deposition almost immediately overlying basement or very thin inner neritic facies is seen in many wells in the southern Makassar Straits. This includes the Siri-1, NSA-1C, NSA-1D, NSA-1F, JSS/Mojo-1, Crystal-1, JS 52-1, Igangan-1, Saibus-1, Pagerungan wells, Sakala-1, East Sakala-1, JS 25-1, Kiuau-1, Sepanjang Island-1, South Sepanjang-1, L-49-1, Sidulong-1, Sepapang-1, and West Kangean-1 and -2 (Figure 1). In all these wells this abrupt environmental change takes place within the Late Eocene Letter Stage Tb [c. 36 Ma], with thin larger foraminifera bearing limestones below and rich planktonic foraminifera in the mudstones above (see Phillips et al., 1991; Bransden & Matthews, 1992)

To the south, the Sawangan-1 well geohistory has a two-fold subsidence, from delta plain to middle or shallow outer neritic within later Middle Miocene times (with P14 or P13), followed by a subsidence to fully bathyal conditions within P15, early Late Eocene [the 36 Ma event].
The nearby wells L-46-1 & -2, ST Z-1 and SG P-1 also seem to have a later Middle Eocene subsidence and larger intra-Late Eocene subsidence to fully bathyal conditions. To the southeast the Lombosang-1, SSA-1, Soreang-1, Tanakeke-1, Kelara-1 and ODB-1 wells all show a marked and sudden subsidence to deep marine conditions within the Middle Eocene (intra P14 or P13) and this tectonic event is a different (Spermonde) rift and not discussed here. Soreang-1, ODB-1, SSA-1 and Tanakeke-1 also show additional subsidence at 36 Ma, but this is not clearly expressed in microfaunas that were already in a deep marine setting.

In the centre of the south Makassar rift the Kris-1 (K-1 in Bacheller et al., 2011) also contains this sudden intra Late Eocene transgression. However about 16 kms from Kris-1 the data on the Sultan-1 (S-1) well (Bacheller et al., 2011) suggests no
significant subsidence in Late Eocene times (although it might have been the event that first led to limestone being deposited over basement there). The S-1 site seems to have had continual growth of a reef from some time in the later Eocene until the latest Oligocene (after the evolution of Miogypsinoides at c. 25 Ma), as was also happening in TT to the west and Doang-1 and Bone-1 to the south. Unlike these other carbonate platforms the S-1 site then subsided, near the end of the Oligocene, to bathyal conditions, terminating reef growth. The current depth of the Eocene-Oligocene reef in S-1 is just a few hundred metres shallower than the deepwater deposits of the same age in K-1 (the good pick of the Eocene-Oligocene boundary is about 300m lower in K-1). On this limited data it is suggested that K-1 and S-1 were offset by 300m due to intra-Late Eocene fault movement, and K-1 became the site of condensed deeper water sedimentation through the Oligocene while S-1 hosted a reef, then both sites underwent additional subsidence at the end of the Oligocene. As a result of this K-1 was only onlapped by bathyal clastics during Middle Miocene times, and the slightly higher structure, and also reef, at S-1 onlapped in Late Miocene times. A reef similar to the one at S-1 is still not covered by sediment, and stands up to 300m above the surrounding seafloor, which is at about 2050m, some 35 kms north of S-1. This feature, called Snorkel in De Man et al. (2011) was sampled by dredge/piston core and a Late Oligocene Te4 biohermal carbonate fauna was found (van Gorsel and Helsing, in prep.).

THE TERMINAL EOCENE EVENT AND OLIGOCENE SEDIMENTATION

The geohistory method also reveals other important stratigraphic changes after the large 39 and 36 Ma subsidence events. As noted above and in Lunt (2013), the Eocene to Oligocene boundary across the southern Makassar area, from Spermonde in the SE to Kangean in the SW is characterised in deep marine sediments by a stepped reduction in gamma-log response (up-hole) at the same depth that the highest Eocene microfossils are first seen in cuttings (c. 34 Ma). The end of the Oligocene, close to the termination of the Berai Limestone and its deep marine equivalents, is dated as between 23 and 24 Ma, and is associated with several biostratigraphic datums in both carbonate and planktonic facies. This gives a reliable correlation for base and top Oligocene. The top Berai Limestone event (near top Oligocene) is a major boundary that, in seismic lines perpendicular to the one examined here, shows highly condensed sedimentation after the end Berai event and then large scale progradation of clastics down the West Makassar Graben, sourced from the west (Kupecz et al., 2013, Figure 18).

Using these two reliable correlation points (base and near-top Oligocene), of known ages, the geohistory analysis shows the 10-11 Ma of the Oligocene is represented by some 550m preserved in MS-1 or about 700m un-compacted at base Miocene times, which is equal to between 64 and 70m/Ma. This contrasts with about 2830m of compaction-corrected sediment between the base of the Oligocene and the onset of transgression some 4 Ma earlier, or about 710m/Ma: almost a ten-fold reduction in the rate of sedimentation. This ten-fold change is a maximum, as MS-1 was in a graben depocentre, while other locations such as P-1, over a palaeo-high with more condensed Eocene sedimentation, saw Oligocene rates of sedimentation reduce to about a quarter of what they were in the Eocene. This drop in sedimentation rate is clear on the geohistory diagrams.

The terminal Eocene event is extremely clear in the L-1 well in the Spermonde Graben where a sharp log break separates Late Eocene claystones from very highly condensed Early Oligocene hemipelagic chalks. The entire Oligocene is represented by less than 150m, but this is close to the top of the highest sample examined, and above this there is no detailed age breakdown. However the reduction in sedimentary rate is so severe that it stands out in spite of the lack of biostratigraphy (Figure 6). At about 1660m in the well a seismic event correlated to other wells, and the main inversion event of the Spermonde Graben, is known to be about mid Middle Miocene in age. It is only after this event that late Neogene uplift and erosion of Sulawesi Island introduces a new sediment source and the rate of sedimentation at the L-1 location becomes significant.

As noted above, it is hard to compare rates of sedimentation in the graben with those seen in Eocene and Oligocene carbonate platforms on the horst-ed highs as there was frequent loss of circulation while drilling these sections, and subsequently very poor samples. Only Bone-1 has sidewall core coverage enough to give a reliable stratigraphy. There is only minor correction for compaction in platform carbonates, compared to the mudstone sections in other wells, and basement subsidence until the unconformity/hiatus near top Oligocene is indicated to be just over 1100m in about 13 Ma, or about 85 m/Ma. The Oligocene section only is 950m in 11 Ma (also 85 m/Ma).

THE TERMINAL OLIGOCENE EVENT AND EVIDENCE FOR A MID-OLIGOCENE EVENT

Pireno et al. (2009) describe how the Ruby gas field (discovered by the MS-1 well) is trapped in a major carbonate debris flow facies that is associated with the termination of the long-lasting Berai Limestone. This debris flow travelled some 4 to 6 kms horizontally from the collapsed carbonate shelf edge. As noted above, the Berai Limestone is equivalent to about 11 Ma of condensed clastic sedimentation in the lows. Such large debris flows
are only seen at one other time in the underlying section, in about mid-Oligocene times (close to the extinction of Nummulites at about 28.2 Ma; Lunt & Allan, 2004). This is observed in MS-1, MS-3, and MS-4 wells (Pireno et al., 2009). The mid Oligocene debris flow does not appear to be as large (as thick or as widespread) as the terminal Berai slumps and flows. In addition the terminal Berai event is associated with a widespread change in lithofacies, while the mid-Oligocene event was succeeded by continuation of the same lithofacies. However they can both be interpreted as the products of large scale tectonic or eustatic changes.

Around the south Makassar region it has been noted that there are just two significant stratigraphic boundaries associated with Oligocene sedimentation, one in mid-Oligocene times and one virtually on the Oligo-Miocene boundary. Closest to the study area is the cluster of pinnacle reefs around S-1 (Bacheller et al., 2011) that were drowned out in very latest Oligocene times, and since then have subsided to a depth more than 3 kms below sea-level. While there was a temporary shift to clastic sedimentation above Te1-4 platform limestones at TT-2 and Berlian-1, those areas (c. 100 kms west of S-1) were again the site of biothermal carbonate growth soon afterwards, whereas the S-1 location was so deep it was not onlapped (by bathyal mudstones) until Late Miocene times. This suggests that at end of Oligocene times there were considerable differences in vertical movement between S-1 and TT-2 as well as Berlian-1 and also Doang-1 and Bone-1 (c. 190 kms & 230 kms to the south respectively). In these last two named sites carbonate platforms, growing near sea-level, continued from Oligocene and through the Miocene. It is this geographic variation in vertical displacement, relative to sea-level, that suggests a tectonic rather than a eustatic cause for the terminal Oligocene event in South Makassar. Similarly Lunt (2013) noted striking difference in vertical displacement in adjacent areas of the eastern Java Sea (around the NSA-1D well) in mid-Oligocene times, close to or just above the extinction of Nummulites at 28.2 Ma. For this reason a tectonic cause is favoured for both Oligocene times of stratigraphic change, but more data is required on the weaker mid-Oligocene event, mainly as intra Oligocene sections are hard to date accurately with biostratigraphy.

**SUMMARY OF THE MAIN SEDIMENTARY SEQUENCES, EOCENE TO BASAL MIOCENE TIMES**

Geohistory analysis indicates that there are 5 main sequences of sedimentation in the south Makassar Straits from Eocene to Early Miocene times, and that these sequences are separated by tectonic induced unconformities.

Within the Middle Eocene, or just a short while before it, rifting began on the south and southeast margin of Sundaland (the area now occupied by the Makassar Straits), with non-marine or paralic sedimentation slightly exceeding the rate of basement subsidence. At the same time lacustrine sediments were deposited in land-locked grabens in the more inland area around the West Makassar Graben.

At about 39 Ma (before the mass extinction of planktonic foraminifera at top P14) a tectonic event accelerated the rate of subsidence, particularly in grabens in the south and southeast. This led to the onset of very deep marine conditions in those areas. In the West Makassar Graben this event may have been the cause of the shallow marine transgression over the lacustrine deposits.

At 36 Ma (within the lower part of the late Eocene; about mid foraminiferal Zone P15), the main part of the southern Makassar Straits, including the West Makassar Graben, was subject to subsidence of an outstanding magnitude that can be measured using data from MS-1 (calibrated to P-1). This appears to be the event that created the south Makassar Straits as an open seaway. It completely changed basin architecture and sedimentation over a very wide area.

The fourth sedimentary phase was initiated by the severe drop in clastic sedimentary input on the Eocene to Oligocene boundary [34 Ma], which can be seen on geohistory plots from several wells. The mid-Oligocene event [roughly 28 Ma] produced a log-correlatable event near the extinction of Nummulites and Chiloguembeina cubensis (a small planktonic foraminifera not always recorded in older reports) in many wells, but it does not see a major shift in facies development. On the whole, the Late Oligocene in most areas did not receive significantly different sediments than the Early Oligocene.

The fifth major sedimentary sequence was the onset of renewed clastic sedimentation into the West Makassar Graben and other lows after the terminal Oligocene events. This early Miocene sediment was initially slow and distal, but rapidly prograded and filled, with sediment supply soon exceeding basement subsidence rate.

The importance of identifying these fundamental sequences is that Walther’s Law of facies relationships will apply within these major units, but across the bounding unconformities these relationships, and any palaeogeographic predictability, breaks down. The magnitude of the lithofacies changes across these sequence boundaries, as well as the geographic variation in amount of contrast, suggests these first-order sequence boundaries were tectonic events.
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INTRODUCTION

The Barito Basin in southeast Kalimantan contains a thick sedimentary succession of Middle Eocene to Pleistocene age. The basin is separated from the much smaller Asem-Asem Basin by the Meratus Mountains (Figure 1) – a complex of metamorphosed arc and ophiolitic rocks that record the accretion of East Java-West Sulawesi to Sundaland in the mid-Cretaceous. The complex was later uplifted during the Neogene. The uplift had a profound effect on the basin architecture, developing a foredeep along the emerging mountain front, and creating the present-day hydrocarbon plays of the basin. Thus, understanding the history of the Meratus uplift not only provides insight into the tectonic evolution of the basin, but also the development of the hydrocarbon system. Previous estimates of the age of uplift range from Middle Miocene (e.g. Letouzey et al., 1990; Satyana et al., 1999) to Late Miocene (e.g. Panggabean, 1991; Hutchison, 1996). Consequently, linking the uplift to regional tectonics has remained speculative. This article gives a very brief summary of palaeocurrent and provenance data, and makes reference to palynology results, that collectively aid our understanding of the Meratus uplift. The data were collected along the western flank of the Meratus Mountains during field and provenance studies between 2009 and 2011, as part of a PhD at Royal Holloway University, London. Additional data were collected during recent fieldwork with BP Indonesia, in association with Institut Teknologi Bandung (ITB), Indonesia.

A revised stratigraphy for the Barito Basin was published by Witts et al. (2012). Palynology showed the oldest part of the succession (Tanjung Formation), to be no older than Middle Eocene. The Berai and Montalat Formations overlie the Tanjung Formation conformably. They are laterally age equivalent (Upper Oligocene to Lower Miocene), but lithologically different. The Berai Formation comprises a near basin-wide shallow marine carbonate succession, whereas the Montalat Formation which is restricted to the northeast corner of the basin, comprises marginal to clastic shallow marine deposits (Bentot Member), followed by sandy and pebbly braid-delta deposits of the Kiwa Member. The Warukin Formation spans the Middle and Late Miocene, and probably extends into the Plio-Pleistocene locally. It overlies the Berai Formation conformably, and is thought to overly the Montalat Formation in a similar manner – although this latter relationship was not observed in the field. The formation comprises marginal marine (Barabai Member) to fluvo-deltaic (Tapin Member) deposits, recording a return to terrestrial deposition across the basin. The formation is overlain unconformably by, and locally interfingers, the latest Miocene to Pleistocene Dahor Formation, comprising alluvial material shed from basement rocks of the Meratus Complex. In my view, the Montalat and Warukin Formations contain evidence for a much earlier phase of Meratus uplift than previously speculated. They also suggest that the northern end of the Meratus was uplifting first. A brief summary of the results considered pertinent to this discussion is presented here.

MONTALAT FORMATION

Palaeocurrents recorded from cross-bed foresets through ~200 stratigraphic metres of the Lower Miocene Kiwa Member, indicate flow was directed towards the NW (Figure 2). We know that marine conditions were established to the east, north and southwest of where the measurements were recorded, which implies an uplifted source area to the SE. The only candidate for such an area is the Meratus. If this were the case, one would expect the material being supplied to the Kiwa Member to be reworked sedimentary rocks of the underlying Tanjung Formation.

This hypothesis was tested by comparing heavy mineral (HM) assemblages and zircon age populations of the Kiwa Member with those of the Tanjung Formation. The Kiwa and Tanjung datasets reveal a strong positive correlation in HM species and their relative proportions (Pearson Correlation -

Figure 1. Location of the Meratus Mountains and adjacent Basins. The Cenozoic sedimentary succession is also shown.
between 0.7 and 0.9). In addition, the oldest HM samples from the Tanjung Formation most strongly correlate with the youngest sample from the Kiwa Member, and vice versa. If the Tanjung Formation was reworked into the Kiwa Member, this inverse relationship would be expected. This interpretation is also supported by zircon geochronology (See Figure 3). When zircon age populations of the two units are compared statistically, using the degree of overlap and similarity (OS) test (Gehrels et al., 2006), the data correlate positively at 0.94.

**WARUKIN FORMATION**

Palaeocurrents from the Warukin Formation form two distinct trends. The first, directed towards the ESE (i.e. towards the present-day Meratus), indicates the Meratus were not yet elevated (rivers flow out of highland areas, not into them). The second trend records flow directed towards the WNW (i.e. away from the Meratus). This second trend was recorded from channel sandstones in the uppermost ~500 m of the formation. The 180° switch in palaeocurrents is clearly seen in Figure 4. I interpret this reversal to record the early stages of Meratus uplift – perhaps only to a few metres elevation – and the initiation of a new drainage system.

HM assemblages throughout the Warukin Formation showed very little variation (Witts, 2011). The same was true for zircon age populations. The sandstones were originally derived from the Schwaner Complex and the Central Kalimantan Ranges (Witts, 2011), located to the west and northwest respectively. Such a provenance is consistent with ESE-directed palaeocurrents, but inconsistent with those directed towards the WNW (away from the Meratus). Thus, it seems probable that the Warukin Formation essentially cannibalised itself when the Meratus was being uplifted.

**DISCUSSION**

Palaeocurrents and sandstone provenance of the Kiwa Member suggest the northern part of the Meratus was uplifting during the Lower Miocene. To the south, a reversal in palaeocurrent direction indicating the onset of Meratus elevation does not appear until the uppermost strata of the Middle – Late Miocene Tapin Member (Warukin Formation) – Figure 5. Due to limited age-diagnostic fossils in these coal-bearing Warukin strata, it was not possible to determine their exact age. However, some deductions can be made from palynology. We know from palynomorphs that the Tapin Member was deposited from between 16 Ma and 7.4 Ma (Witts, 2011).
The member is ~2 km thick, and the thickest coal-bearing horizons (which contain the palaeocurrent reversal) form the uppermost ~500 m. The up-section increase in coal bed thickness indicates conditions were becoming optimal for peat productivity. This most likely corresponds to the Miocene thermal maximum – between 15 Ma and 13.8 Ma (R. Morley, pers. comm., 2011) – when conditions were at their hottest and wettest for the Neogene. Therefore, it is unlikely that these horizons are older than 14 Ma. Secondly, the palynomorph assemblages from the youngest coal beds suggest the climate may have started to become more seasonal (R. Morley, pers. comm., 2011). Seasonal and drier climates began to develop across southern Sundaland from Late Miocene through Plio-Pleistocene (Morley, 2000). Therefore, it seems reasonable to suggest that the thick coal-bearing strata that contain the west-directed palaeocurrents were deposited during the Late Miocene.

These interpretations point to strongly diachronous uplift of the Meratus, initiating in the north during the Early Miocene, but not until the Late Miocene in the south. This corroborates many observations other than the data presented in this article. For example, thinning of the Lower Miocene marls of the Berai Formation over the ‘Meratus ridge’, reported from subsurface data by Bishop (1980); and the presence of reworked Eocene foraminifera (e.g. Pellatispira sp.) within those marls (Witts, 2011), that were probably reworked from marine-influenced horizons of the uplifting Tanjung Formation. In addition, the deepest/most developed part of the Meratus foredeep is at its northern end, suggesting longer lived uplift and loading in that area. Unfortunately, seismic coverage along the Meratus front is extremely limited, thus extrapolating subsurface observations is inherently problematic. Nevertheless, subsurface data do suggest a general thinning of the Warukin interval towards the Meratus flank in the north, with relatively constant thickness of the same interval to the south (BP Indonesia Exploration Team, pers. comm., 2013). Such a strongly diachronous uplift history suggests the Meratus Complex is, or has previously been, segmented; permitting piecemeal elevation of individual blocks at different times. NW-trending faults dissecting the Meratus, NW-trending stream alignments in the lowlands of the Barito Basin, and a similarly oriented basement fabric underlying the basin (BP Indonesia Exploration Team, pers. comm., 2013) all support this hypothesis. It could also explain the NW-trending ‘Tanjung Line’, which delineates an abrupt change in deformation style.

**Figure 3.** All concordant U_Pb ages of zircons analysed from the Kiwa Member (top) and the Tanjung Formation (bottom). Left histograms show complete range of ages. Right histogram is an expanded view of Palaeozoic and younger ages. T: Cenozoic, K: Cretaceous, J: Jurassic, T/P: Triassic and Permian, C/D: Carboniferous and Devonian, S: Silurian, O: Ordovician, C: Cambrian.
Figure 4. Palaeocurrent data from the Warukin Formation, recorded along the western flank of the Meratus. Warukin Formation outcrop is rendered orange, coastline shown as solid black line. Palaeocurrent data is divided into two sets. Those of the left are predominantly directed towards the WNW. Those on the right are predominantly directed towards the ESE.
Figure 5. Correlated sections of the uppermost horizons of the Tapin Member, recorded along the western flank of the Meratus. Palaeocurrents are also plotted. Solid red line indicates approximate stratigraphic position of last occurrence of ESE-directed palaeocurrents.
of the sedimentary succession in the subsurface. This interpretation warrants further study.

Collision between the Banggai-Sula microcontinent and eastern Sulawesi has been suggested as causing the Meratus uplift (e.g. Letouzey et al., 1990; Satyana & Silitonga, 1994). I feel it is highly unlikely that these collisions could have influenced tectonics in east Kalimantan. There is no evidence of collision in the Bone Basin, in West Sulawesi or in the Makassar Basins for this time interval. It seems more likely that uplift was a consequence of regional plate motions. The Miocene – in particular from 20 Ma – was a time of major reorganisation of plate boundaries throughout Southeast Asia, largely driven by the northwards push of the Australian plate against Sundaland (Hall, 1998). Therefore, we would expect most deformation to occur during this time. The Meratus Complex is a suture. Sutures are areas of weak crust that are commonly reactivated during episodes of tectonism (e.g. Busch et al., 1997; Bailey et al., 2000; Tikoff et al., 2001). The Meratus suture resembles a ‘flower’ structure (e.g. Sikumbang, 1986) positioned between two relatively rigid and undeformed continental fragments. It seems reasonable to argue that uplift of the Meratus merely reflects localised deformation along a pre-existing zone of weakness and controlled by a pre-existing basement fabric, in response to regional plate motions.

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The power of palaeocurrents: Reconstructing the palaeogeography and sediment flux patterns of the Miocene Sandakan Formation in eastern Sabah

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ABSTRACT

Up until now the use of palaeocurrents as a primary tool in establishing regional palaeogeography and sediment dispersal patterns in the ancient has been limited. An outcrop study on the Miocene Sandakan Formation in eastern Sabah identified five facies belts, subdivided on the basis of sedimentological and palaeontological data. More than 140 outcrop sections were measured over an area covering approximately 200 square kilometres. The facies were interpreted to have been deposited in settings ranging from ancient mangrove deposits through shoreface sediments and out into the open marine.

More than 200 palaeocurrent readings taken on these facies allowed a clear picture to be built up of the sediment dispersion patterns across the palaeo-Sandakan Basin, passing from northward directed flow through mangrove channels into a longshore drift dominated shallow marine, coastal shelf. This gave way northwards to a storm dominated belt of tempestite deposits cut by rip current channels. Flow through these channels, driven by frequent storms, was northward directed. Further to the north more quiescent, open marine, muddy conditions prevailed. A sediment budget can be estimated using the sand fraction of the various lithofacies identified during the study.

INTRODUCTION

Collision of a series of southward-moving microcontinents with the northern Borneo margin in the early Miocene led to Sabah undergoing compressive stress, uplifting the Eocene-Oligocene deposits of the Central Sabah Basin. Contemporaneous seafloor spreading in the Sulu Sea, to the east, rejuvenated the Basin, and formed a series of fault-controlled basins, trending NW-SE across eastern Sabah, in which relatively shallow marine to paralic sediments were deposited.

One of these basins is particularly well exposed on the Sandakan Peninsular in eastern Sabah (Figure 1), where the Sandakan Formation outcrops sporadically over an area of 200 km². More than 140 sedimentological sections (Figure 2) have been logged through sediments interpreted to have been deposited in a paralic to marine setting. Five lithofacies associations have been defined based on their sedimentological and palaeontological character: mangrove; tidally dominated flats and occasional tidal channels; stacked, upper shoreface sandstone beds; rip current dominated, heterolithic, middle shoreface deposits; and lower shoreface to open marine mudstone beds with thin, ironstone cemented siltstone beds.

A series of more than 200 palaeocurrent measurements were taken from outcrops across the Sandakan Peninsula. Combined with several occurrences of the basal contact of the Sandakan formation with underlying basement volcanics, it was possible to reconstruct a detailed picture of the palaeogeography and sediment dispersal patterns across the basin.

GEOLOGICAL EVOLUTION OF EASTERN SABAH

Sabah is thought to be floored by crystalline basement and associated ophiolitic rocks (Clennell 1992). These may represent uplifted slices of the continental crust underlying Borneo (Leong 1974, Holt 1998). Overlying the basement is the Rajang Group representing a North-facing accretionary prism (Hamilton 1979, Bol and Hoorn 1980, Bénard et al. 1990, Hutchison 1996) comprising mainly turbidites of the Trusmadi and Crocker Formations, extending up into the Miocene (Hutchison 1996).
Deposition in the Central Sabah Basin, the onland extension of the SE Sulu Sea Basin (Hutchison 1992), produced the Labang and Kulapis Formations, both dominated by shelfal turbidites (Noad 1999), though the Labang shallows upwards into shallow marine deposits (Clennell 1996, Noad 1999). These sediments are then unconformably overlain by extensive mélange deposits, related to (SE Sulu Sea) basin initiation in an intra-arc setting (Clennell 1991) during the late early Miocene.

Following deposition of the mélange, gentle regional uplift and quiescent conditions led to the deposition of shallow marine clastic deposits of the Tanjong and Sandakan Formations in the Neogene Sabah Basin (Rangin 1990, Noad 1999). General eastward progradation fed sediment from the Bukit Garam area towards the Sandakan Peninsula, and then out into deeper waters of the Sulu Sea (Clennell 1992, Noad 1999). Later regional structural realignment segmented these Miocene shallow marine deposits into a series of outliers (Noad 1999), separated from the Sulu Sea by a large transform fault (Hinz et al. 1989, Clennell 1996). Associated compression, possibly related to the rotation of Borneo (Hall 1996), and uplift raised these deposits above sea level.

SEDIMENTOLOGY OF THE SANDAKAN FORMATION

The Sandakan Formation was first described in detail by Lee (Lee 1970), but his interpretations have been superseded by the findings described in this paper. The total thickness of the Sandakan Formation is estimated at around 300 metres (Noad 1999, in strong contrast to earlier authors), with clear examples of lateral transition between facies passing from south to north across the Sandakan Peninsula. Frequent normal faults cut across the Peninsula, complicating the geological picture. Five lithofacies associations (Figures 3 and 4) were identified during the outcrop study, and these are described below:

Lithofacies association A: Mangrove facies

Lithofacies A1 is made up of thick, highly carbonaceous, dark grey mudstone, with an abundant fauna. The mudstone often contains logs, up to at least 2.5 m in length, as well as common rooted, buttressed trees in life position (Plate 1). The mudstone also contains whole leaves, from a variety of species. Scattered amber clasts are present at most localities, and can reach 15 cm in diameter. They are well rounded, and have yielded an insect fauna that includes millipedes, spiders, ants, and parasitic wasps. A dominantly molluscan fauna includes Corbicula and Hiatula, as well as Batillaria and cerithiid gastropods, and the fossil wood is bored extensively by teredinids. Several outcrops have yielded specimens of the mangrove lobster Thalassina sp., up to approximately 10 cm in length (Plate 2), as well as distinctive arthropod trackways.

Figure 2. Map of the Sandakan Peninsula, showing the locations of all sections logged in the area by the author.
Lithofacies A2 is represented by steep sided, fine grained, sandstone filled, channel deposits. Typically between 1 and 5 metres in width, they have a width:thickness ratio of 2 to 5. The sandstone shows few sedimentary structures beyond some trough cross-beds, partly due to a homogeneous grain size. However the top and basal channel surfaces may exhibit fossil bird, lobster and deer tracks. Lithofacies A3 comprises sheet-like sandstone beds up to a metre in thickness and at least 50 metres in width. They occasionally exhibit low angle planar cross-beds and wave rippled tops.

Interpreted depositional setting:
The juxtaposition of fully marine molluscs, a brackish water fauna, abundant mangrove lobsters (which are restricted to a mangal environment), rooted trees and amber yielding a rain forest insect fauna strongly suggests that this lithofacies was deposited in a coastal mangrove swamp environment. Many of the modern mangrove trees at Kuala Selangor, in the Klang Estuary, Peninsula Malaysia, are buttressed in a way similar to the Sandakan flora from Lithofacies A1.

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### Figure 3.

Summary section of the lithofacies identified in the Sandakan Formation, ranging from proximal mangrove deposits to distal mudstone beds.
Figure 4. Reconstructions of the major lithofacies associations identified in the Sandakan Formation.
The presence of gastropods such as cerithiids is also indicative of a mangrove setting (Plaziat 1995). The sandbodies of Lithofacies A2 are interpreted as mangrove channels, steep sided due to being confined by rooting, similar to channels observed in the mangroves of the Florida Keys. The abandonment surfaces have been extensively bioturbated by ambulatory organisms. Lithofacies A3 is interpreted to represent open areas ("broads"), also recorded as occurring within Floridian mangroves, and possibly swept clear of vegetation during storms.

**Lithofacies B: Thinly laminated heterolithic facies**

This lithofacies comprises thin, millimetre scale interbeds of silty sandstone and highly bioturbated mudstone, with abundant muddy drapes and occasional rootlets (Plate 3). Small fossil rooted plants with trunk diameters of less than 10 cm have been observed. This lithofacies directly overlies Lithofacies A1. Occasionally these laminae are capped by 20 to 40 cm thick, fine grained sandstone beds with planar cross-beds, and channelized margins, extending typically for 10 to 20 metres laterally in outcrop.

**Interpreted depositional setting:**

These sediments are thought to have been deposited on a mixed to muddy tidal flat, characterised by lenticular bedding, and by finely interbedded sands and thicker mudstones. The sandstone beds are interpreted to represent small, meandering tidal channels crossing the flats.

**Lithofacies association C: Thick stacked sequences of sandstone beds**

A series of stacked 2 m to 5 m thick sharp, flat based fine to medium grained sandbodies, with each sandbody featuring trough cross-beds, grading up from medium grained to fine grained sandstone (Plate 4). Each sandstone bed is capped with a very distinctive, ironstone-cemented, rippled horizon (Plate 5). Weathering exposes fields of, often strongly peaked, wave ripples that may extend for hundreds of metres, often showing interference patterns and ladderbacks. The rippled horizons are overlain by a thin grey mudstone bed, and then by the next sandstone bed. Rarely the sandbodies are made up of planar cross-beds, with sets on a metre scale, and cross-beds dipping at up to 30 degrees.

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**Plate 1.** Fossil mangrove tree with buttresses. Note that the core is petrified, the outer integuments vitrainised.

**Plate 2.** Fossil Thalassina (mangrove lobster, length 10 cm), collected from mangrove mudstone beds rich in plant debris.

**Plate 3.** Tidal flat deposits, with a well preserved rootlet.

**Plate 4.** Detail of upper shoreface sandstones with low angle bedset boundaries, rich in carbonaceous debris.
Few body fossils have been found in these beds, apart from moulds of bivalve and gastropod fragments, but *Ophiomorpha* burrows are very common, along with *Rhizocorallium*. Some outcrops have abundant *Skolithos* burrows.

**Interpreted depositional setting:**
The thick sandstone deposits are believed to have been deposited in the shallow marine, upper shoreface zone, supported by the textural maturity of the sandstone and abundant wave ripples. The trough and planar cross-beds are interpreted to indicate subaqueous dunes migrating across the seabed. The great lateral extent of many of the ripple fields shows that deposition occurred on a wide, flat, sand-dominated platform. The interference patterns suggest water depths of less than five metres, and the sharply peaked ripple crests may show semi-emergent conditions. *Ophiomorpha* may indicate a sandy shoreline sedimentary environment, and *Skolithos* generally indicates intertidal or foreshore deposits.

**Lithofacies association D:** *Heterolithic interbedded sand and mudstone beds on a sub metre scale*

One of the most prevalent lithofacies associations in the Sandakan Formation is a very heterolithic set of interbedded very fine grained, carbonaceous sandstone, siltstone and mudstone beds. The sandstone beds range from very laterally persistent, tabular sandstone beds with extremely flat bases (Plate 6), normally less than 50 cm in thickness, to channelised sandstone beds, commonly between 2 and 30 metres in width, and with an erosional relief of up to around 2 metres into the underlying beds (Plate 7). These are all interbedded with mudstone beds. The sandstone beds are generally very fine grained, and usually contain hummocky cross-laminae. The channelized sandbodies have very steep channel margins, and show reactivation surfaces. Each sandstone passes up into more silty beds, with climbing ripples often extending to the top contact. The overlying mudstone beds may feature starved sandy ripples (Plate 8). Dewatering and loading structures are particularly common in this lithofacies association.

A wide variety of ichnofauna are present through most of the beds that make up this lithofacies association. Apart from *Ophiomorpha* and occasional *Skolithos*, fifteen or more other ichnospecies have been delineated, some not previously recorded.
These include Phycodes, Scolicia, Scolichnia, Teichichnus, Olivellites, Rhizocorallium, Planolites and Thalassinoides.

**Interpreted depositional setting:**
The very heterolithic nature of these rocks suggests abrupt changes in energy levels. The thinner, tabular sandstone beds are hummocky cross-bedded, and are interpreted as storm deposits (Hobday and Morton 1984). The thicker, more channelized, beds have been cut by fairly strong currents, as evidenced by the steep channel margins, and the reactivation surfaces suggest several episodes of erosion. The presence of numerous dewatering structures, including the striking ball-and-pillow structures, supports relatively rapid deposition. This lithofacies is interpreted as representing tempestite deposits, with the tabular sandbodies indicating deposition on the storm to inner shelf transition zone. The channels are interpreted as having been cut by rip currents (Gruszczynski et al. 1993), with the steep walls suggesting rapid erosion and subsequent fill, supported by the presence of the dewatering structures. Overall these deposits are interpreted as storm dominated, middle shoreface deposits.

**Lithofacies E: Open marine mudstone facies**

This lithofacies is made up of thick grey mudstone, with occasional very thin but laterally persistent silty siltstone beds (Plate 9). These beds are generally less than 5 cm in thickness, but may extend for at least 300 m with no appreciable change in thickness. Internal structures of many beds are obscured by ironstone cementation during diagenesis but, where visible, hummocky cross-stratification can be discerned. A distinctive and varied fauna preserved as internal casts within the ironstone horizons includes 25 species of fossil crabs (Plate 10), as well as a very varied molluscan fauna and a previously unreported argonaut.

**Interpreted depositional setting:**
The thin ironstone cemented siltstone beds are interpreted as distal storm deposits, while the fine grained grey mudstone was deposited in open marine conditions in the lower shoreface to offshore transition zone. The very mixed fauna shows well oxygenated bottom conditions.

**Summary of facies belts**

Extensive fieldwork has enabled the delineation of the major facies belts (Figure 4) of the Sandakan Formation (Figure 6). To the south lie mangrove deposits, passing northward into shallow marine sandstones. The mangrove deposits are cut by fluvo-mangal channels. The interpreted position of the palaeo-coastline is believed to run along the boundary between these two facies belts, although this position is at best an approximation due to the limited outcrop and interpreted northward
progradation of the Formation. The humid conditions appear to have been conducive to rapid colonisation of any tidal flats, explaining their limited extent.

Northward the shallow marine sandstone beds pass into heterolithic tempestite deposits, indicating an unsettled climatic regime. Sheet-like storm deposits are cut by rip current channels, suggesting a significant fetch to the north, generating strong storm surges in a southward direction, which then reflect to the north. This facies belt passes gradually into more muddy, open marine sediments, with only thin storm siltstone interbeds and a varied fauna.

Plate 11 shows a classic coarsening upward sequence passing form open marine, mudstone beds with scattered distal rip current channels up into a heterolithic, lower shoreface succession with common hummocky cross-stratified sandstone beds and common rip current channels, overlain in turn by stacked upper shoreface sandstone beds. This succession is interpreted to indicate progradation of a sandy deltaic system, and demonstrates, using Walther’s Law, the lateral changes in lithofacies across the Miocene Sandakan Basin.

**SYNTHESIS OF PALAEOCURRENT DATA**

It has proved possible, through the measurement of more than 200 palaeocurrent orientations (Figure 5), to build up a detailed picture of sediment movement through the Sandakan sub-basin (Figure 6). The mangrove channels flow in a generally NNE-ward direction whereas the shoreface sandstones are characterised by a more complex palaeocurrent pattern. A coastline can be mapped, running WNW-ESE, where the mangrove deposits interfinger with, and pass laterally into, thick scarp-forming sandstone beds. Wave ripple crests covering the top surfaces of these beds run parallel to this coastal orientation, suggesting swash and backwash. However planar and trough cross-beds within the sandstone beds show a dominantly NW-ward vector, indicating along shore transport by strong tidal currents. The clastic material is sourced from an inferred major river system comparable in size to the Kinabatangan River, located to the east of the current Sandakan Peninsula.

Further offshore, the rip current channels are oriented NNE-SSW, parallel to the trend of the mangrove channels. This supports a model where storm currents are reflected back offshore towards the north, passing seaward into sheet-like tempestites, which pinch out northwards. The source of the large volumes of relatively fine grained, clastic sediment is thought to be from the river system described above, flowing NE-ward into the sea. Overall, the Sandakan Formation is interpreted to have been deposited in a supply-dominated regime, and provides an excellent example of the
utility of palaeocurrents in palaeogeographical reconstruction.

**DISCUSSION ON SEDIMENT FLUX**

The total sand volume stored in the Sandakan Formation on the Sandakan Peninsula is estimated at 29,040 million cubic metres (Table 1). The average total thickness of the Sandakan Formation at any one point is estimated at approximately 300 metres. Assuming that the Sandakan Formation was deposited over a period of around 4 million years (based on limited dating using microfossils), this gives a deposition rate of 90 metres/million years. This compares to a figure of 50 to 200 metres/million years of sediment accumulated on the Indochinese shelf (Kukal 1971). Using the figure of 4 million years this suggests that 7260 cubic metres of sand was trapped in the Sandakan depositional system every year, although almost certainly most of the sediment flux passed through the system, mainly through longshore drift to the west.

**CONCLUSIONS**

A total of five lithofacies associations have been delineated in the Sandakan Formation, namely: mangrove mudstones with sheet and channelized sandbodies; thinly laminated, heterolithic tidal deposits, rarely vegetated; upper shoreface sandstone beds; middle shoreface, heterolithic, storm dominated sheet-like sandstone beds cut by rip current channels; and open marine mudstone beds. These show a full transition from paralic deposits out into an open marine setting. The use of palaeocurrent data collected from all five lithofacies associations has allowed a detailed picture to be constructed of the sediment dispersal patterns in the Miocene Sandakan sub Basin. This methodology can be applied to many other basinal settings worldwide.

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*Table 1. Estimated sediment deposition parameters for the Sandakan Formation.*

*Figure 6. Palaeogeographic reconstruction of the Sandakan Formation, showing the depositional facies belts identified from field logging. An interpretation of the current pattern ('sediment flux') has been added, using palaeocurrent data collected from outcrop.*
REFERENCES


Mass Transport Complex (MTC) control on the basin floor stratigraphic succession and sand deposition: An observation from deepwater Brunei

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INTRODUCTION

Mass transport complexes (MTCs) are one of major geological features observed in many deepwater provinces, including in Deepwater Brunei, where MTCs are commonly initiated and deposited on the slope and basin floor settings. MTCs are broadly characterised on the basis of their internal characteristics and external morphological features. Posamentier (2004) offered a simple observational guideline for describing MTCs: 1) the underlying surfaces of MTCs usually are extensively scoured in a form of deep and linear grooves 2) the MTCs’ overlying surface is mostly irregular - hummocky relief bounded laterally by gentle to steep flanks 3) MTCs often have transparent to chaotic seismic reflections, amalgamation of MTCs stacks is not uncommon, and 4) MTCs could have a morphology of channel or lobes.

This guideline was utilised in describing the recent Brunei Mega MTC (i.e. McGilvery, et al., 2004) though in much larger scale. This paper uses the same set of guideline to describe older MTCs observed in study area. We will also attempt to investigate MTCs’ control on the overall basin floor stratigraphic succession and, in particular, sandstone deposition as it appears to be one of the key factors in delivering potential sands further into the basin floor.

METHOD AND WORKFLOW

Nearly 5,000 km² of 3D seismic data is available within the study area, all of which have been re-processed to enhance the seismic imaging. The reprocessed data reveals significant increase in imaging quality, which allows highly confident horizon interpretation and also enables reliable seismic attribute extraction work. Detailed interpretations were conducted on the zones of interest that is bound by the top horizon, referred here as the Blue Horizon, which lies approximately 300-600ms below the seafloor. One well was drilled within the study area. Despite it has only basic log data at the zones of interest, the well provides a good tie and a direct calibration for all geological features observed on the seismic.

Once the subsurface mapping was finished, proportional horizon slices were then applied and interpolated in between the horizons. We tried to extract several seismic attribute types from the proportional horizon slices, namely RMS amplitude, relative acoustic impedance, reflection strength and sweetness. Sweetness is a function of dividing reflection strength by the square root of instantaneous frequency. This mathematical definition captures qualitative attribute relationships, traditionally applied for isolated sand bodies in shale dominated successions; however, the density contrast of the lithologies within the MTCs appears to be shown distinctively by sweetness.

MASS TRANSPORT COMPLEXES

An analogue was taken from the recent Brunei Mega mass transport complex (MTC) described in details by McGilvery, et al. (2004) and Gee, et al. (2006). The MTC is a seismically chaotic and generally low amplitude body that extends for 120 km from the Baram Canyon head. The MTC base is irregular and truncates underlying reflections and the overlying surface is characterized by subdued topography.

Our observation on recent seabed also suggests multiple MTC systems within the deepwater Brunei, where it is uniquely bounded by two shelves and slopes, albeit the Luconia System is a less active system compared to the Baram System (Figure 1). These mega MTCs consist of slumps and debrites that indicate sediment gravity flows transported from the shelf/slope; the recent penecontemporaneous turbidite lobes which appear to have been restricted, suggesting the MTC control on the turbidite depositional lobes (Figure 1).

The Blue Horizon mass transport complexes consist of 3 distinguishable, stacked up MTCs of various thicknesses. The gross package appears to be thinning towards the inboard structural high. Each MTC shows distinctive similarities, they have chaotic seismic facies, generally low amplitude forming slumps and becomes significantly brighter toward the edge of the lobate body. The lobe extends ~ 80km from the same Baram Canyon head. Low angle thrust faults and pressure ridges are very apparent at the MTC termination in the basin floor environment (Figure 2a).
**Figure 1.** RMS amplitude extraction on recent mass transport complex (MTC) system in Deepwater Brunei shows the influence of two shelves/slopes systems with penecontemporaneous turbidite constrained by the MTCs.

**Figure 2a.** Sweetness extraction of the Blue Hz MTCs. The lobate form extends approximately 80km from the Baram Canyon to the South. Pressure ridges and low angle thrust faults are evident from the extraction which characterise this Blue MTCs. The red star indicates location of an exploratory well drilled in the study area.
The size of this Blue Horizon MTCs is smaller than the current Brunei Mega MTC, probably due to the more quiescence tectonic prior to Pleistocene thrusting and folding. Nevertheless, the size is significant enough to influence the basin floor geometry.

**INTERACTION OF MTCs – OVERLYING SEDIMENTS**

A well was drilled within the study area and it penetrated 8m sand sitting on top of the Blue Horizon MTCs. Sweetness extraction on the sand revealed a distribution that follows the underlying Blue Horizon MTC’s linear striation inboard and the sand becomes unconstrained toward the basin floor. The sand dispersal has NW-SE direction which is the direction of the underlying Blue Horizon MTCs (Figure 2b).

It is postulated that the sand’s preferential depositional route is controlled by the underlying MTC’s irregular surface. In deeper stratigraphic level, a similar sand deposition might have occurred as well, but it could have been eroded off by the subsequent MTC. Some sands might remain “sandwiched” in between these MTCs and it will be difficult to distinguish from the MTC rafted blocks which could be sandy (Figure 3).

**SUMMARY AND DISCUSSIONS**

The study area provides a calibrated MTC-controlled sand deposition in Deepwater Brunei and reveals the importance of the Baram Canyon.

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*Figure 2b. RMS extraction at the overlying sediment above Blue Horizon MTC calibrated by an exploratory well shows sand distribution toward the basin floor. The dispersal appears to be controlled by the underlying MTC.*
for the sediment dispersal. It appears that MTC provides preferential pathways for the overlying sand deposition through its irregular surface although not always distinct. Lithofacies A1 is made up of thick, highly distinguishable.

In one complete sediment cycle, there is a strong indication that MTC is overlain by turbidite which could be sandy and hemipelagites. This model is used to devise the succession of MTC stacks where the subsequent MTCs could erode off the whole turbidite and the hemipelagites, however, there will be some remnants of these turbidites though will not form a classical lobes or fan geometry. The key is to have detailed maps of various MTC facies that could have different influences on the sand terminal lobes.

Some questions remain unanswered such as major mechanisms for the MTCs emplacement (e.g. earthquake, methane degassing, pore water pressure), why some MTCs eroded off the tip of the anticline and some are gently draping over, and temporal and spatial relationship between MTC and sandstone turbidites. This study is expected to open up further detailed studies and to help understand and model sand deposition in this MTCs dominated environment.

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The ethological study of *Glossifungites* ichnofacies in the modern & Miocene Mahakam Delta, Indonesia

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**INTRODUCTION**

The *glossifungites* facies is an ichnofacies which represents an assemblage of burrows that occur in firm, not lithified ground. Some geologists believe that the presence of a *Glossifungites* ichnofacies surface is an evidence of a break between erosion and deposition (Pemberton & MacEachern, 1995), which is a boundary of a sequence stratigraphic unit.

Although firm ground assemblages are well understood, the *Glossifungites* ichnofacies concept generates some debates amongst geologists (Gingras et al., 2000), because firm grounds represent intermediate gradational state between soft ground and hard ground. The level of compaction and dewatering of substrate are varies and controlled by different factors.

Arifullah (2005) studied *Glossifungites* ichnofacies in East Kalimantan, which are common in both modern and Miocene Mahakam Delta system (Figure 1). The identification of *Glossifungites* ichnofacies in this study is based on ethology and morphological observation of the bioturbations.

![Observation Map of Modern Mahakam Delta, *Glossifungites* ichnofacies findings was located at Lantang-1 & 2 (After Arifullah, 2005).](image)
MODERN MAHAKAM DELTA

Lantang Island is located in the north of modern Mahakam Delta (Figure 1). In this island, Arifullah (2005) has identified a location with high index and diversity of bioturbations. On the ichnofacies distribution map (Figure 3), Lantang Island and surrounding area have the highest variability of ichnofacies. *Skolithos*, *Cruziana* and *Zoophycos* ichnofacies, which formed *Glossifungites* ichnofacies assemblages are common in here.

Correlation of Lantang-1 to Lantang-2 cores give a better picture of the lateral distribution of sedimentary facies. Figure 4 shows the mud flat had been exhumed after the sandflat eroded. This mud flat - “Hm” lithofacies unit- is dominated by lenticular, wavy or tidal rhytmites mud.

The burrows of *Glossifungites* ichnofacies respond to substrate stability of firm ground. Such condition has made erosion opened unlined burrows and filled it with sand and shell fragment, which characterized *Glossifungites* ichnofacies. Meanwhile the firm ground, and the cohesiveness of “Hm” lithofacies would not be eroded easily as burrows without lining (construction without strengthen material, i.e. coarser materials) despite the high energy condition on the water column (Figure 5).

MIOCENE MAHAKAM DELTA

*Glossifungites* ichnofacies assemblage studied at Kesejahteraan Quarry, is associated with erosional surface sedimentary structure (Figure 2). This ichnofacies cross cut the pre-existing Hm lithofacies unit. A hummocky cross stratification sandstone unit was deposited on top of this ichnofacies surface (Figure 6). The surface of the *Glossifungites* ichnofacies could be observed up to 300 meters laterally. It could be used as bounding surface or parasequence surface for stratigraphic correlation.

The “Hm” lithofacies unit looks like soft ground ichnofacies, but the biota built vertical or semi vertical burrows (“J”-like) without lining in the mud. Analogue studies on modern traces in Lantang Island with similar ethology indicated that the traces were developed in firm ground environment instead of soft ground.

CONCLUSION

The presence of *Glossifungites* ichnofacies indicate cohesive or more lithified substrate environment. Therefore, the presence of *Glossifungites* ichnofacies surface does not always related to sequence boundary, but more reasonable as event surface.
Figure 4. Lantang island lateral and vertical profiles. It was observed that ichnofacies index & diversity increase to the mudflat. It was strongly different between mudflat that colonized by Cruziana ichnofacies and Skolithos ichnofacies colonize the sandflat. Glossifungites ichnofacies localized at the mudflat only (After Arifullah, 2005).

Figure 5. Glossifungites ichnofacies: A) Skolithos, B) mud pellet excavation, C) Thallasinoides. Unlined construction used by organism to respond the firmground substrate (After Arifullah, 2005).

Figure 6. Prodeltaic ichnofossil. A) Thallasinoides (Th), Teichichnus (Te), and Planolites (Pl); B) Asterosoma (As) and Thallasinoides (Th); C) Erosional structure (blue line) between prodeltaic deposit (bottom) and shoreface deposit (top). Prodeltaic deposit totally bioturbated by Rhizocorallium (Rh), Planolites (Pl) and Thallasinoides (Th), they are part of Glossifungites ichnofacies. Shoreface deposit colonized by Ophiomorpha (Oph) specifically (look at figure 7C). (After Arifullah, 2005).
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REFERENCES

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