# The onshore West Baram Delta deposits: provenance and drainage in the Middle Miocene to Pliocene in NW Borneo and comparison to the Champion Delta

H. Tim Breitfeld[1,2], Juliane Hennig-Breitfeld[1,2], Gavin Anthony[2], Marcelle BouDagher-Fadel[3+], Pieter Vermeesch[3], Keno Lünsdorf[4], Delia Rösel[5], Matthias Konrad-Schmolke[5], Sabine Gilbricht[6]

*[1]Institute of Geology, TU Bergakademie Freiberg, Bernhard-von-Cotta-Straße 2, Freiberg D-09599, Germany*

*[2]SE Asia Research Group, Department of Earth Sciences, Royal Holloway University of London, Egham, Surrey, TW20 0EX, United Kingdom*

*[3]Department of Earth Sciences, University College London, London, United Kingdom*

*[4]Georg-August-University Göttingen, Geoscience Center, Department of Sedimentology and Environmental Geology, Goldschmidtstraße 3, D-37077 Göttingen, Germany*

*[5]Department of Earth Sciences, University of Gothenburg, Guldhedsgatan 5A, Göteborg SE-40530, Sweden*

*[6]Institute of Mineralogy, TU Bergakademie Freiberg, Brennhausgasse 14, Freiberg D-09599, Germany*

+ Marcelle BouDagher-Fadel passed away on June 30th, 2022 during preparation of this manuscript. She provided foraminifera identification and imagery of analysed specimens. Over many years her invaluable contributions to SE Asia Research Group projects redefined depositional ages across SE Asia, and we are very grateful to have had the opportunity to work with her.

ORCID

H. Tim Breitfeld: https://orcid.org/0000-0002-9563-1862
Juliane Hennig-Breitfeld: https://orcid.org/0000-0003-3866-4235
Marcelle BouDagher-Fadel: https://orcid.org/0000-0002-2339-2444
Pieter Vermeesch: https://orcid.org/0000-0003-3404-1209
Keno Lünsdorf: https://orcid.org/0000-0002-2738-8469
Delia Rösel: https://orcid.org/0000-0002-4140-0858

## Abstract

The Baram Delta province in NW Borneo forms a major hydrocarbon reservoir offshore northern Sarawak and Brunei. The delta sequence is thereby subdivided into the West Baram delta to the south and the Champion delta to the north. Onshore are the remains of the Neogene delta deposits exposed and provide the possibility to study the equivalent offshore successions in outcrop. This study focuses on the Neogene West Baram delta successions which were studied for sedimentological facies and provenance characteristics. The successions consist of the Lambir, Miri, Tukau, and the enigmatic southern Lambir/Belait-Sarawak formations. Deposition took place in various mixed-energy delta environments between the Langhian and early Pliocene. The sediments are all quartz-rich and heavy minerals are dominated by ultra-stable zircon, rutile and tourmaline. Dominant detrital zircon age clusters are in the Early Cretaceous and Permian-Triassic. Based on light mineral petrography, heavy mineral assemblages, and detrital zircon U-Pb geochronology, all formations are interpreted as derived from multi-recycled sources, likely the underlying Paleogene Rajang Group turbidites and the Oligocene to Lower Miocene Nyalau-Tatau delta deposits. Additionally, literature data of the Champion Delta and one sample from Labuan analysed for provenance in this study are used to demonstrate that the Champion Delta can be distinguished from the West Baram Delta by having higher chrome spinel and garnet contents and slightly different detrital zircon age populations. The Champion Delta deposits are interpreted as sourced by recycling of the Crocker Formation and older turbidites (e.g., Sapulut Formation) with potentially input from ultra-mafic basement rocks of Sabah.

Keywords: detrital zircon U-Pb geochronology, provenance, heavy minerals, Champion Delta, West Baram Delta, paleogeography, Borneo

## 1. Introduction

The continental margins of the South China Sea have a long history of hydrocarbon exploration and production. In the south, hydrocarbons have been discovered along the coasts of Vietnam, the Malay Peninsula, Natuna and Borneo in offshore sedimentary basins (Fig. 1a). Reservoirs include fractured granites (Cuu Long Basin), carbonates (Luconia Platform, Sarawak Basin) and Cenozoic (mostly Neogene) clastic sedimentary rocks (Fig. 1a). The northwest margin of the island of Borneo includes two major provinces, the Sarawak and Sabah basins, both of which are subdivided into various sub-basins.

On land in NW Borneo oil seeps had been known for many years by the local population (Sorkhabi, 2010) and the first shallow wells were drilled on the island of Labuan and in Brunei in the late 19th century (Redfield, 1922). The present-day Baram Delta region and Miri city in northern Sarawak close to the border of Brunei (Fig. 1b) were the sites of the first hydrocarbon discoveries and exploration in the area (Wannier et al., 2011) which led to commercial production in 1910. After successful exploration the Miri oil field was established and active until 1972 (Wannier et al., 2011) when exploration and production moved offshore. 624 wells were drilled in the field and 80 million barrels of oil were produced (Wannier et al., 2011). The adjacent Seria field in Brunei was discovered in 1929 and the only other successful onshore oil field (Sorkhabi, 2010) in NW Borneo.

Several large sedimentary basins formed in Borneo during the Cenozoic (e.g., Pieters et al., 1987; Doutch, 1992). Neogene exhumation removed at least 6 km of crust after the last uplift phase, and sediments were subsequently deposited in large basins around the island (Hall & Nichols, 2002). Large delta provinces developed in the Neogene with high sedimentation rates, which include e.g., the Mahakam Delta in east Borneo (Storms et al., 2005; Marshall et al., 2016; Morley et al., 2016) and the Baram Delta in NW Borneo (Tan et al., 1999; Lambiase et al., 2003), which is the focus of this study. Deltaic, fluvial, coastal and shelf successions were deposited from the Middle Miocene in tropical humid conditions (e.g., Sandal, 1996; Tan et al., 1999; Hall & Nichols, 2002; Lambiase et al., 2003; Morley & Back, 2008). Accommodation space was created by high subsidence rates in the Baram region (up to 3000 m/Myr; Sandal, 1996) related either to compression (e.g., Hazebroek & Tan, 1993; Morley et al., 2003; Morley & Back, 2008; Hesse et al., 2009; Cullen, 2010; Gartrell et al., 2011) or extension (Hall, 2013).

Most previous studies of the sediments in the Baram Delta region were concerned with detailed sedimentology, facies and environment interpretations (e.g., Lambiase & Cullen, 2013; Collins et al., 2017, 2018, 2020), but only very limited studies include provenance and sandstone petrography (e.g., Nagarajan et al., 2017). There are several contemporaneous lithostratigraphic units (e.g., Liechti et al., 1960) that have similar characteristics in terms of lithologies, depositional environments (e.g., Banda & Honza, 1997; Tan et al., 1999; Collins et. al., 2020), and geochemistry (Togunwa & Abdullah, 2017). This paper reports facies analyses, biostratigraphy, light mineral modes, heavy mineral assemblages and detrital zircon U-Pb ages from onshore sedimentary rocks of the Miocene West Baram delta to establish depositional environment and age, and to test whether these formations show differences in terms of provenance. The data is used to constrain uplift history and drainage evolution of the source areas, which provides further insights into the evolution of the Baram River system in NW Borneo.

## 2. Regional background

Western Borneo is subdivided into five tectono-stratigraphic zones that include SW Borneo, West Borneo, the Kuching Zone, the Sibu Zone and the Miri Zone (Fig. 2a; Haile, 1974; Breitfeld et al., 2017; Hennig et al., 2017). The Miri Zone is the northernmost zone and consists mostly of Oligocene to Neogene clastic sedimentary rocks (Haile, 1974). The Miri Zone extends offshore into various tectono-stratigraphic provinces (Fig. 2b) that include mostly shallow marine-deltaic clastic sediments of the Baram Delta province, Balingian province and Tatau province which form important offshore hydrocarbon reservoirs. The adjacent onshore Miri Zone successions, carbonates of the Luconia province, and deep-water siliciclastics of the Sabah Trough are contemporaneous stratigraphic equivalents and hydrocarbon reservoirs.

The study area is located in the northern part of the Miri Zone in North Sarawak (Fig. 1b), where Oligocene to Quaternary fluvial, tidal and shallow marine succession cycles (Liechti et al., 1960; Haile, 1974) that are unconformably above Paleocene to Eocene turbidites of the Rajang Group (Fig. 3) are well exposed and provide an opportunity to study onshore equivalents of the offshore reservoir sections in outcrop. The area onshore is also referred as Tinjar province (Fig. 1b and 2b).

### *2.1. Stratigraphy and tectonic evolution*

The oldest rocks exposed in the Miri Zone belong to the mainly Paleogene deep water Rajang Group, represented as small inliers of the Kelalan, Mulu and Belaga formations (Liechti et al., 1960; Wolfenden, 1960; Haile, 1962; Hennig-Breitfeld et al., 2019). The deltaic Oligocene to Lower Miocene Tatau-Nyalau formations are unconformably on top of the Rajang Group sediments (Fig. 3), and cover most of the southern part of the Miri Zone (Liechti et al., 1960; Hutchison, 2005; Hassan et al., 2013; Breitfeld et al., 2020a). The Setap Shale Formation and its equivalents (e.g., Sibuti Formation) in the northern part of the Miri Zone are interpreted to represent marine or prodelta mudstone facies of the Tatau-Nyalau system (Fig. 3; Breitfeld et al., 2020a), and for most parts directly underlie the deposits of the West Baram Delta (e.g., Liechti et al., 1960). In northern Borneo (Sabah) deep water sedimentation continued from the Late Cretaceous until the Early Miocene (Fig. 3; Hutchison, 1996; Burley et al., 2021), and is represented mainly by turbidites and debrites of the Crocker Formation (Hutchison et al., 2000; Jackson et al., 2009; van Hattum et al., 2013; Zakaria et al., 2013).

At c. 17 to 18 Ma, the Nyalau Unconformity (Fig. 3; Hennig-Breitfeld et al., 2019, 2020; Breitfeld et al., 2020a), EMU (Early Miocene Unconformity, Madon et al., 2013) and TCU (Top Crocker Unconformity, van Hattum et al., 2013; Burley et al., 2021) mark a major reorganisation of the drainage system. Sediment supply from the southwest was cut off and an emergent central Borneo supplied material (Hutchison, 2005; Hennig-Breitfeld et al., 2019). Open marine/prodelta deposition of northern Sarawak and deep marine deposition in Sabah terminated, and from the Middle Miocene onwards tide- or wave dominated deltaic successions were deposited near the present NW Borneo coastline. The Meligan, Miri, Lambir, Belait and Tukau formations were deposited in the northern part of the Miri Zone, and the Kakus, Balingian, Begrih and Liang formations in the southern part (Fig. 3; Liechti et al., 1960; Hutchison, 2005; Hennig-Breitfeld et al., 2019, 2020; Breitfeld et al., 2020a). North of the Miri Zone, the Belait, Seria and Liang formations form the dominant deltaic succession in Brunei, with the Belait Formation extending to western Sabah and Labuan (Fig. 3; Liechti et al., 1960; Wilson & Wong, 1964; Sandal, 1996; Hutchison, 2005). The Miri and Seria formations are associated with onshore oil and gas reservoirs; the Miri oil field and Seria oil field, respectively.

### *2.2. Onshore Neogene Baram Delta successions*

In the study area around Miri and near the present-day Baram Delta, Koopman (1996) subdivided the early delta development into three phases (Fig. 1b) related to uplift and erosion of the Sibu Zone. Phase 1 is represented by the poorly studied Lower Miocene Meligan Formation (Fig. 1b). Phases 2 and 3 are represented by the Middle Miocene to Pliocene East and West Baram Delta successions (Fig. 1b), which are hydrocarbon reservoirs in the offshore Baram Delta province (Fig. 2b). The East Baram Delta (also known as the Champion Delta) lies in Brunei and Sabah, and the West Baram Delta is situated in the present-day Miri River and Baram Delta area (Fig. 1b). The onshore Champion Delta deposits are formed mainly by the Belait Formation (Fig. 3), while the onshore West Baram Delta successions include the Miri, Lambir, Tukau and Belait (named here southern Lambir/Belait-Sarawak) formations (Fig. 3). Sources for the Champion/East Baram Delta sediments were assumed to be uplifted highlands in Sabah (Hutchison et al., 2000; Hutchison, 2005), while the West Baram Delta was interpreted to be sourced from the south by recycling of the Oligocene to Early Miocene Nyalau Formation (Hutchison, 2005). The East Baram Delta comprises Middle to Upper Miocene mostly shallow-marine sediments that are preserved in Brunei and on Labuan and extend offshore (Sandal, 1996; Hodgetts et al., 2001; Van Rensbergen and Morley, 2003). The West Baram Delta sediments are contemporaneous and include better preserved, thicker, and more widespread onshore deposits extending offshore into the present-day Baram Delta (Liechti et al., 1960; Van Rensbergen & Morley, 2003). During Phases 2 and 3 (East and West Baram Delta) approximately 9-12 km of coastal-deltaic to shelf sediments accumulated in Brunei over the past 15 Myr (Sandal, 1996; Collins et al., 2017).

Offshore the sediments are assigned to cycles (Fig. 3; Ho, 1978; Hageman, 1987; Hageman et al., 1987; Madon et al., 2013) or seismic sequences (Mat-Zin & Tucker, 1999) rather than lithostratigraphic units. Togunwa & Abdullah (2017) interpreted the West Baram Delta as prograding since the Middle Miocene. Morley et al. (2003) interpreted a Middle to Late Miocene fold and thrust belt offshore NW Borneo and Pliocene inversion of the basin.

## *2.3. West Baram Delta stratigraphy*

The stratigraphy of the Neogene to Quaternary West Baram Delta remains controversial as most of the successions were originally described from onshore wells and there is a significant lack of age-determining fossils. Middle to Late Miocene or Pleistocene ages were assigned to the formations (Liechti et al., 1960). Liechti et al. (1960) distinguished the Lambir, Miri, Tukau and Belait formations in the northern Miri Zone (Fig. 3 and 4) based on their different depositional environments, sand to mud ratios and calcareous content. The differentiation of the formations remains difficult due to interfingering relationships and inconsistent use of formation names. Detailed studies of the sediments have revealed that all formations were deposited in relatively similar environments, including wave-storm influenced, tidal and deltaic settings (e.g., Banda & Honza, 1997; Tan, 1999; Abieda et al., 2005; Jia & Rahman, 2009; Kessler & Jong, 2015; Cheng, 2019; Rahman & Tahir, 2019; Collins et al., 2020), emphasising the difficulties in formation assignment.

### *2.3.1. Basal contact of the delta system*

The Neogene deltaic sediments were deposited on top of the marine Setap Shale, Sibuti or Tangap formations that are distal parts of the older Tatau-Nyalau delta system (Fig. 3). An angular unconformity was not observed, but either a diachronous transition, disconformity or a sharp abrupt boundary was interpreted between the Neogene and Tatau-Nyalau systems (Liechti et al., 1960; Hutchison, 2005). Observed changes in sediment provenance indicate reorganisation of the drainage system, and Hennig-Breitfeld et al. (2019) and Breitfeld et al. (2020a) interpreted a major unconformity, the Nyalau Unconformity, between the systems. Madon et al. (2022) identified an angular unconformity on top of the Nyalau Formation, which might be the Nyalau Unconformity.

The dominant underlying shale unit is the Sibuti Formation. In contrast to the wider distributed Setap Shale Formation, the Sibuti Formation is interpreted to be more calcareous and cm-thick silt layers are abundant (Liechti et al., 1960; Banda & Honza, 1997; Peng et al., 2004; Hutchison, 2005; Breitfeld et al., 2020a). The Sibuti Formation is interpreted to be Late Oligocene to Early Miocene based on foraminifera (Haile, 1962; Banda & Honza, 1997; Simmons et al., 1999). A similar age range has been reported for the Setap Shale Formation (e.g., Kho, 1968; Breitfeld et al., 2020a).

### *2.3.2. Lambir and Miri formations*

The Lambir and Miri formations consist mainly of sandstones, shales and some limestones (Liechti et al., 1960; Tan et al., 1999) deposited in a mixed-energy delta (Collins et al., 2020). The Miri Formation is exposed in a small area around Miri city and Brunei, whereas the Lambir Formation is mapped at Lambir Hills (Fig. 4). In wells, the Miri Formation is divided into a lower shale-dominated and an upper sand-dominated part (Liechti et al., 1960). The Lambir and Miri formations grade laterally into the Belait Formation (Liechti et al., 1960; Haile, 1962). Liechti et al. (1960) interpreted the Miri Formation as conformably above the Lambir Formation, whereas Kessler & Jong (2015) assumed an interfingering contact. Collins et al. (2020) interpreted the Miri and Lambir formations as the genetically related first of several NW-prograding regressive deltaic wedges.

### *2.3.3. Belait Formation in northern Sarawak*

The formation consists of coarse, mostly cross-bedded, white sandstones, clay and sandy shales (Kirk, 1957; Haile, 1962). The basal part is interpreted to pass laterally into the Lambir Formation (Haile, 1962), suggesting a similar age range. The Lambir and Miri formations initially were distinguished from the Belait Formation by their shallow marine character, whereas the Belait Formation in its type section in Brunei was thought to be more littoral and deltaic-paralic (Liechti et al., 1960). In Sarawak however, the Belait Formation was found to be more paralic compared to the Brunei exposures and its difference from the Miri and Lambir formations was not clear (Liechti et al., 1960). Generally, the interior deposits in northern Sarawak have been mapped as Belait Formation (Fig. 4B) from 1960 onwards (e.g., Liechti et al., 1960; Wilford, 1961; Haile, 1962; Heng, 1992). Later, Banda & Honza (1997) suggested the abandonment of the term Belait Formation in northern Sarawak, as they assigned exposures in the interior to the Lambir Formation based on structural interpretation and detailed mapping and suggested that the deposits formed the southern limb of an anticline in which the Lambir Hills exposures (Fig. 4) were the northern limb. They used the informal term ‘southern Lambir Formation’. As the Belait Formation has its type locality in Brunei and is also exposed on Labuan, as part of the Champion Delta (Fig. 1B), whereas the southern Lambir section is part of the West Baram Delta, we therefore follow Banda & Honza (1997) in distinguishing between the Belait Formation in Brunei and Sarawak. However, it is uncertain if the ‘southern Lambir Formation’ really is part of the Lambir Formation as part of an anticline. Ramli & Padmanabhan (2011) reported various lithological differences between the Lambir and ‘southern Lambir’ formations, which questions the interpretation of Banda & Honza (1997) and we therefore use the term ‘Southern Lambir/Belait-Sarawak Formation’ for the interior deposits to differentiate them from the Lambir Formation in northern Sarawak and the Belait Formation in Brunei and on Labuan.

### *2.3.4. Tukau Formation*

The Tukau Formation is in parts the stratigraphic equivalent of the Lambir and Miri formations (Liechti et al., 1960) and was suggested to be part of the Lambir Formation (Banda & Honza, 1997) as there are no differences in lithology or facies. This conclusion was accepted by e.g., Rahman & Tahir (2019) and Collins et al. (2020), while others retained the Tukau Formation as separate unit (e.g., Hutchison, 2005; Kessler & Jong, 2015; Abdul Hadi et al., 2017; Nagarajan et al., 2017). We also retain the term for the deposits that form the top of the Lambir Hills above the Lambir Formation. Generally, the Tukau Formation consists predominantly of thick fine-grained sandstones interbedded with thin lignite layers and thick mudstones intervals deposited in a brackish-water coastal plain-shoreline environment (Wilford, 1961; Tan et al., 1999; Collins et al., 2020). It is conformable on top of the Lambir Formation (Haile & Ho, 1991) and in two exploration wells, the Tukau Formation supposedly conformably overlies the Miri Formation (Wilford, 1961). In contrast, e.g., Kessler & Jong (2015) interpreted an angular unconformity at the base of the Tukau Formation which separates the undeformed Tukau Formation from the slightly folded Lambir Formation, and correlates this with regional folding at c. 5.6 Ma (Morisson & Wong, 2003). However, Kessler & Jong (2015) also acknowledged that in some localities the Tukau Formation is slightly folded and apparently conformable on top of the Lambir Formation.

### *2.3.5. Ages of the successions*

The Lambir and the Miri formations possibly range from the Langhian to Tortonian based on sparse foraminifera assemblages (Liechti et al., 1960; Wilford, 1961; Banda & Honza, 1997; Tan et al., 1999; Hutchison, 2005; this study). Based on palynology Abdul Hadi et al. (2017) suggested a Middle to Late Miocene age for the Lambir Formation. The southern Lambir/Belait-Sarawak Formation was previously undated and a similar Langhian to Late Miocene or Pliocene age is indicated by foraminifera reported in this study. The Tukau Formation was assumed to be Late Miocene to Early Pliocene by Wilford (1961), and an Early Pliocene (Zanclean) foraminifera assemblage was identified in this study. In summary, there is a lower sequence with a Langhian base and a diachronous top between the Tortonian and Zanclean, and an upper sequence that is of Zanclean age.

#### *2.3.6. Thickness*

The thicknesses of the Lambir and Miri formations is estimated to be at least c. 1.5 km each, and the Tukau Formation is estimated to be about 2.5 to 3.0 km (Liechti et al., 1960; Hutchison, 2005). The southern Lambir/Belait-Sarawak Formation is thought to be approximately 1.1 to 2.6 km thick (Liechti et al., 1960). Since all the formations interfinger (Fig. 3), a precise total thickness for the onshore West Baram Delta sediments cannot be given with confidence. A present-day thickness of around 5.5 to 7 km is assumed here based on the stratigraphy and the published thickness estimates.

### *2.4. Champion Delta stratigraphy*

North of the West Baram Delta deposits lies the contemporaneous Champion Delta in Brunei, southwest Sabah and on Labuan (Fig. 1b). The Champion Delta deposits were interpreted to be related to a complex drainage system with multiple river mouths, therefore not representing a single delta succession (Lambiase & Cullen, 2013; Collins et al., 2017, 2018). At present this setting is still preserved with the Trusan, Limbang and Padas rivers draining into Brunei Bay.

The majority of the delta deposits are part of the Lower to Upper Miocene Belait Formation in Brunei and on Labuan (Fig. 3; Liechti et al., 1960; Wilson & Wong, 1964; Madon, 1994; Sandal, 1996; Abdullah et al., 2013; Hennig-Breitfeld et al., 2019). Kocsis et al. (2022) reported a Sr-isotope age of 12.1 + 1.4/-1.2 Ma (Serravallian) from calcareous fossils, and foraminifera data suggest a possible extension of the base into the late Early Miocene (Sandal, 1996). The Miri Formation of northern Sarawak also extends into Brunei (Liechti et al., 1960) and the differentiation between the West Baram and Champion deltas therefore becomes unclear. Some authors use the Baram Delta or West Baram Delta terms for all the successions in the area (Sandal, 1996: Collins et al., 2017, 2018). Calcareous fossils from the Miri Formation in Brunei yielded Sr-isotope ages of 8.9 to 10.5 Ma (Tortonian) (Kocsis et al., 2022). The tuffaceous Seria Formation overlies conformably the Miri and Belait formations in Brunei and is inferred to be at least partly Pliocene (Fig. 3; Liechti et al., 1960), while Kocsis et al. (2022) presented Sr-isotope ages of 7 to 7.9 Ma (early Messinian to late Tortonian) from calcareous fossils. The Liang Formation in Brunei is probably the youngest onshore unit within the Champion Delta. The formation is inferred to be Pliocene to Pleistocene, and based on subsurface data, unconformably on top of the Seria Formation (Liechti et al., 1960). In outcrop, no unconformity has so far been found. The white sand Jerudang Terrace forms the youngest deposit in Brunei (James, 1984), and similar sand terraces are found across western Borneo in Sarawak and Kalimantan related to sea level changes in the last 2 to 4 Myr (Liechti et al., 1960; Andriesse, 1970; Thorp et al., 1990; Thomas et al., 1999; Wannier et al., 2011; Breitfeld, 2021). The dominant depositional environments of the Champion Delta sediments (and the Belait Formation in Brunei and on Labuan) are shoreface, tidal and delta front settings with wave- and storm-influenced deposits and some shelfal mudstones (Lambiase & Cullen, 2013; Siddiqui et al., 2013; Fiah & Lambiase, 2014; Collins et al., 2017, 2018; Hennig-Breitfeld et al., 2019). Collins et al. (2017, 2018) identified a strong seasonality with distinct fair-weather and storm periods within the successions. Additionally, there are some fluvial conglomerates and sandstones deposited by braided river systems, which are poorly preserved (Drahaman, 1999; Tan, 2010; Lambiase & Cullen, 2013; Hennig-Breitfeld et al., 2019).

## 3. Methodology

### *3.1. Sampling*

A total of six sandstone samples were collected from the northern part of the Miri Zone SW of the West Baram River and around Miri city (Fig. 4), which include the Miri Formation (Mi-01, Mi-02), Lambir Formation (La-01, La-02), southern Lambir Formation/Belait-Sarawak (Be-01), and Tukau Formation (Tu-01). They were analysed for light mineral modes, heavy mineral assemblages, and detrital zircon ages. Heavy minerals from one sample from the Belait Formation from the island of Labuan (LL1) were also analysed in this study for comparison with the Champion Delta sequence (Fig. 1b). The sample is from the Layang-Layangan Beds, which are interpreted to belong to the Belait Formation unconformably above the Temburong Formation (Albaghdady et al., 2003; Gou & Abdullah, 2010; Abdullah et al., 2013; Hennig-Breitfeld et al., 2019). Additionally, two foraminifera-rich marls (Si-01, Si-02) from the underlying Sibuti Formation have been analysed for biostratigraphy. Samples are listed with coordinates in Supplementary Table 1.

### *3.1. Petrography*

Light mineral modal analysis was conducted on six stained thin sections, following the Gazzi-Dickinson method (Dickinson & Suczek, 1979; Dickinson et al., 1983). Sodium cobaltinitrite was used for staining alkali feldspar and barium chloride and amaranth solution were used for staining plagioclase. Porosity was not measured. The ribbon technique was employed over an evenly distributed grid. A total of 500 grains were counted for each sample. Grains smaller than 30 µm cannot be optically resolved and were assigned to matrix (Ingersoll et al., 1984; Pettijohn et al., 1987). Count numbers are listed in Supplementary Table 2.

Covered thin sections were analysed for biostratigraphy, following the approach described in BouDagher-Fadel (2015, 2018a). The approach primarily uses the Planktonic Zonation scheme (PZ) of BouDagher-Fadel (2018b), which is tied to the biostratigraphical and the radioisotope time scales (as defined by Gradstein et al., 2012 and revised by Cohen et al., 2013). The planktonic foraminiferal zonal scheme of BouDagher-Fadel (2015) is also correlated with the larger benthic foraminiferal 'letter stages' of the Far East, as defined by BouDagher-Fadel & Banner (1999) and later revised by BouDagher-Fadel (2018a).

### *3.2. Heavy mineral separation*

Sample preparation for heavy mineral analyses and zircon separates was carried out at Royal Holloway University of London. Heavy minerals were separated by using the funnel technique on a 63-250 µm fraction (Mange & Mauer, 1992) with the heavy liquid lithium heteropolytungstate at a density of 2.89 g/cm$^3$. The resulting heavy mineral fraction was poured and mounted into araldite epoxy resin. The resin mount surface was polished to ensure an even surface for Raman spectroscopy.

Part of the heavy mineral concentrates were further processed with a FRANTZ magnetic barrier separator and di-iodomethane heavy liquid with a density of 3.3 g/cm$^3$ was used to obtain zircon separates. Zircons were hand-picked and mounted into araldite epoxy resin. The resin mounts were polished to expose zircon mid-sections for analysis.

### *3.3. Heavy mineral analysis*

Raman spectroscopy was used for heavy mineral identification as it can achieve the most accurate heavy mineral assemblage identification (e.g., Ando & Garzanti, 2014; Dunkl et al., 2020). Polymorphs like rutile, anatase and brookite ($TiO_2$) can be differentiated, which is not possible with electron-based analytical methods.

Raman spectroscopy was conducted at the Department of Sedimentology and Environmental Geology, University of Göttingen, using a Horiba XploRa with a 532 nm laser coupled to an Olympus polarising microscope. The Raman spectroscope was calibrated with silicon prior to use, which is also measured every 200 grains. A detailed methodology description, as well as Raman setup parameters and sample preparation can be found in Lünsdorf et al. (2019). Acquired spectra were compared to the RRUFF database (Lafuente et al., 2016) to assign a 'best fit' coefficient. The coefficient describes how well a given spectra corresponds to its closest fitting spectrum in the RRUFF database with ‘0’ being a perfect fit and ‘1’ representing no fit to any spectrum. Results of 0-0.15 are classed as ‘good hits’ and were accepted. Spectra with correlation coefficients between 0.15 and 0.30 were classed as ‘medium hits’ and accepted after visual assessment. Spectra with coefficients over 0.30 were all visually checked, and those minerals were also optically assessed under the microscope. Furthermore, Mineral Liberation Analysis was conducted on a FEI Quanta 600FEG scanning electron microscope at the University of Freiberg to identify uncertain minerals.

Additional Raman spectroscopy was conducted for samples La-01 and La-02 at the Department of Earth Sciences of the University of Gothenburg using a Horiba LabRam HR Evolution Raman spectrometer. The analyses were performed with a 532 nm laser after calibration on silicon. Spectra were compared to the Horiba/Wiley internal database (KnowItAll software package) and to the RRUFF database (Lafuente et al., 2016). Sample LL1 was analysed using a Horiba XploRa Plus fitted with a 532 nm laser coupled to an Olympus BX43 polarising microscope at Chemostrat Ltd. Acquired spectra were compared to the RRUFF database (Lafuente et al., 2016) and an internal Chemostrat Ltd. database for identification. Supplementary Table 3 lists heavy mineral count numbers. Additionally to heavy mineral abundancies, commonly used heavy mineral ratios were used for differentiation. In particular the zircon-tourmaline-rutile (ZTR) value of Hubert (1962), the zircon-tourmaline (ZTi) ratio (Morton, 2007), and the rutile-zircon (RuZi), garnet-zircon (GZi) and chrome spinel-zircon (CZi) indices of Morton & Hallsworth (1994). Calculation of the ratios is explained in the Supplementary materials (document 1).

In addition to Raman spectroscopy, scanning electron microscopy based automated mineralogy (SEM-AM) with mineral liberation analysis (MLA) software was used at the Institute of Mineralogy, Economic Geology and Petrology, TU Bergakademie Freiberg after the methodology outlined in Schulz et al. (2020) to aid the identification.

### *3.4. Zircon geochronology*

Zircon geochronology was carried out at Portsmouth University using an ASI RESOlution 193 nm ArF excimer laser ablation system coupled to the ANALYTIK Jena Plasma Quant Elite quadrupole ICP-MS. Primary reference material was the Plešovice zircon (337.13 ± 0.37 Ma; Sláma et al., 2008). Secondary reference zircons included Temora 2 (416.8 ± 1.0 Ma; Black et al., 2004), 91500 (1065 Ma; Wiedenbeck et al., 1995) and BB9 (561 ± 2 Ma; Santos et al., 2017). A sample-reference material bracketing method was used to correct for instrumental drift and mass fractionation. Laser spot size was 20 µm, and measurements were taken with an energy density of 2.5 J/cm$^2$ at a repetition rate of 2 Hz. Data were processed using the software package IOLITE 3.31 (Paton et al., 2011). Sample Be-01 was also analysed from a second separate at the University of London with a New Wave NWR 193 nm laser ablation system coupled to an Agilent 7700x quadrupole-based plasma ICP–MS with a two-cell sample chamber. Plešovice zircon was used as a primary reference material (337.13 ± 0.37 Ma; Sláma et al., 2008) and Australian gem zircon GJ-1 (608.53 ± 0.59 Ma; Jackson et al., 2004) as secondary zircon reference material. Instrumental mass bias and depth-dependent inter-element fractionation of Pb, Th and U was corrected using the NIST 612 silicate glass bead (Pearce et al., 1997). Data reduction was achieved with the GLITTER software (Griffin et al., 2008).

The ages obtained from the $^{207}$Pb/$^{206}$Pb ratio is used for zircons older than 1000 Ma. For ages younger than 1000 Ma, the ages obtained from the $^{206}$Pb/$^{238}$U ratio are given. Concordance was tested by using a 10% threshold (90-110%) between the $^{207}$Pb/$^{206}$Pb and $^{206}$Pb/$^{238}$U ages for ages greater than 1000 Ma and between the $^{207}$Pb/$^{235}$U and $^{206}$Pb/$^{238}$U ages for ages below 1000 Ma. Laser ablation spots were selected after consideration of transmitted light and cathodoluminescence imagery to avoid cracks, mixed zonation or inclusions. Core or rim features were not targeted due to their low abundance. Uncertainties in age are reported as 2σ. Age histograms and kernel density (Vermeesch, 2012) plots were created using an internal R script and the IsoplotR package by Vermeesch (2018). Plots in the manuscript are split between 0-500 Ma and 500-4000 Ma for better visualisation of differences in the Phanerozoic. Analytical results are presented in Supplementary Table 4, and zircon reference data is listed in Supplementary Table 5 with illustration in Supplementary Fig. 1.

## 4. Sedimentology and facies of the West Baram Delta deposits

### *4.1. Sibuti Formation*

#### *4.1.1. Observations*

The Sibuti Formation is the dominant mudstone-siltstone sequence in the northern Miri Zone and is an equivalent of the upper part of the more widespread Setap Shale Formation (Liechti et al., 1960; Heng, 1992; Hutchison, 2005; Hennig-Breitfeld et al., 2019). The formation is well exposed along the road section from Bekenu to Beluru (Fig. 4) and commonly underlies the West Baram Delta deposits.

The dominant lithologies include fine-grained dark grey-coloured shales interbedded with thin siltstone to fine-grained sandstone layers and marls with subhorizontal bedding. The greenish-grey sandstone layers are partly calcareous and contain irregular carbonaceous mudstone bands. Shales are usually carbonaceous, and in contrast to the Setap Shale Formation, which is predominantly dark-coloured, the Sibuti Formation is commonly greyish in colour (Fig. 5a). Locally, thin stacked channel structures can be observed, outlined by cm-thick fine sandstone beds which form the base of these channels and appear to have eroded into the underlying mudstone to shale layers (see Breitfeld et al., 2020a). The thick shale layers are interbedded with siltstone to fine grained sandstones with coarsening up-section trends. South of Beluru (Si-02) the Sibuti Formation consists of shallow to moderately dipping thin rhythmically bedded siltstones and fine-grained sandstones which disconformably overlie a thick dark-coloured mudstone to shale unit interpreted as the Setap Shale Formation by Breitfeld et al. (2020a) (Fig. 5a, b).

#### *4.1.2. Interpretation*

Thick shale layers indicate an overall low energy environment. Carbonaceous mud indicates wash-in from coastal floodplains in a muddy shelf zone (Nichols, 2009), while limestone layers, marls and calcareous beds are related to input from nearby reef facies and wash-in from storm events. The coarsening-upwards patterns indicate episodic changes from a low to high energy domain. The formation is interpreted as an open marine carbonaceous shelf deposit (Reading, 2013; Hodgson et al., 2017) with limestone layers or calcareous beds representing inner shelf deposits, and channels observed within the formation are interpreted as distal tidal channels or as prodelta deposits. Based on the presence of marl beds and small oyster patch reefs, Nagarajan et al. (2015) suggested a deeper shelf to slope deposit. At Si-02 where the Sibuti Formation is folded, a sharp contact with the slightly older Setap Shale Formation is exposed which indicates rapid input of coarser material from the upper delta front or prodelta.

### *4.2. Southern Lambir/Belait-Sarawak Formation*

#### *4.2.1. Observations*

Approximately 2.5 km south of Beluru following the Tinjar and Bakung rivers, sand ridges are mapped as southern Lambir/Belait-Sarawak Formation (Liechti et al., 1960; Heng, 1992; Banda & Honza, 1997) (Fig 4). Towards the interior of the Miri Zone, the formation becomes the dominant stratigraphic unit and forms the large Dulit anticline farther to the southeast (Liechti et al., 1960). The highest peak which is in close proximity to the sample location is Bukit Balat. The exposures observed south of Beluru (where Be-01 was sampled) are composed of medium-grained to fine-grained sandstones forming amalgamated massive sandstone beds (c. 0.3 up to 2.5 m thick), which dip moderately to the southeast (dip direction/dip: 146/20) (Fig. 6a). The yellowish-brown sandstones show reddish and orange weathered surfaces which are likely related to limonitic and hematite alteration. The massive sandstone beds have erosive bases (Fig. 6b) and sedimentary structures include swaley (Fig. 6c) and trough cross-bedding, hummocky cross-stratification or parallel horizontal lamination (Fig. 6d), observable on fresh surfaces. The sandstone beds show lateral continuation, but pinch-out structures are also present. Heterolithic beds or mud-dominated intervals are very restricted and the dominant lithology in the analysed sections is sandstone. Bioturbation is also very restricted or absent and no plant material was observed. A few planktonic foraminifera were found within the sandstones and indicate a shallow marine environment.

#### *4.2.2. Interpretation*

The parallel laminations and erosional bases of the sandstone beds indicate a change from low energy to a moderate or high energy channel deposit environment and pinching-out structures are interpreted as large channel geometries. The high influx of sand and the moderate to good sorting indicates a high energy environment for most of the beds. The general absence of trace fossils may be a result of high input rates of clastic material accompanied by fast subsidence (Dashtgard, 2011). Foraminifera in the sandstones indicate a shallow marine environment for the channels, which suggests that these are tidal channels, which cut into and migrate over the delta plain, while isolated sand bodies represent tidal sand bars (e.g., Dalrymple & Choi, 2007). Ali et al. (2016) also interpreted a tidally-influenced delta succession with thick tidal channel and shoreface deposits. Hummocky cross-stratification indicates storm wave deposits in a shallow marine environment (shoreface, shelf) (Kumar & Sanders, 1976). Thick amalgamated sandstone beds may be attributed to increased rates and magnitude of storm-wave reworking and/or increased sand availability related to decreased water depth, increased storm-wave energy, and/or increased proximity to the sediment source (Swift & Thorne et al., 1991; Thorne et al., 1991; Storms & Hampson, 2005). Trough cross-bedded sandstone with unidirectional currents may suggest river-dominated distributary channels (Miall, 2013; Flood & Hampson, 2014; Ainsworth et al., 2015; Gugliotta et al., 2016). Three different facies that form several repeating cycles have been recognised (Fig. 6a). They are interpreted as a proximal storm-dominated delta front at the base to river-dominated distributive channels at the top of the outcrop (Tab. 1), which indicates a shallowing upward trend.

### *4.3. Lambir Formation*

#### *4.3.1. Observations*

The exposures of the Lambir Formation were analysed from a broadly E-W trending belt in the coastal area around Tusan Cliff to the area around Bukit Lambir at the Lambir Hills (Fig. 4). Sample La-01 was collected near the coastal road and La-02 southeast of Bukit Lambir. The coastal section from Tusan Cliff to the sample location La-01 is characterised by thick sandstone beds and sand-dominated heterolithic beds. The abundance of the latter increases towards the interior towards Bukit Lambir.

The Tusan Cliff section consists of thick amalgamated sandstone beds (up to c. 10 m) and minor interbedded heterolithic sandstone-mudstone alternations (c. 10 cm to 1.0 m) with cm-thin lignite/coal layers. Typical sedimentary structures are herringbone cross-stratification, planar cross-beds and trough cross-bedding. Conglomerates are also present in the succession and are best exposed south of Tusan Cliff in a beach section that is submerged during high water periods (Fig. 7a). Clasts are formed by intrabasinal well-rounded fine- to coarse-grained sandstone gravels. Beds at Tusan Cliff dip moderately to the NW.

Towards Bukit Lambir at location La-01, thicker heterolithic beds start to appear within the succession. Stratigraphically the section is higher up in the formation. A c. 10 m high outcrop along a smaller road was analysed (Fig. 7c). The base of the outcrop consists of c. 1.5 m thick wavy ripple-laminated heterolithic deposits interbedded with horizontally laminated fine-grained sandstone and siltstone layers (c. 0.5 m thick). Moderate to strong bioturbation was observed, dominated by vertical *Skolithos* burrows, up to 5 cm in length, and a few horizontal *Ophiomorpha* burrows (Fig. 7b). The middle section of the sequence consists of thick (up to 1 m) laminated, bioturbated sandstone interbedded with sandstone-dominated heterolithic beds and layers. The top of this alternation forms a c. 1.0 m thick carbonaceous mudstone bed (Fig. 7c). The upper part of the exposure is significant sandier, including several massive trough cross-stratified sandstone beds (c. 0.5 to 4 m thick), which alternate with c. 0.3-0.8 m thick laminated sandstone and a discontinuous mudstone-dominated heterolithic bed (Fig. 7c). The uppermost sandstone bed is channelised and shows pinching out structures. Crude swaley cross-stratification was observed in the higher section.

Southeast of Bukit Lambir (sample La-02) on the old road from Miri to Bintulu (Fig. 4) the stratigraphically highest section of the Lambir Formation in this study was observed. The section consists of multiple stacked channels with pinching out structures (Fig. 7d). The basal channel is formed by medium- to coarse-grained amalgamated sandstone with planar and trough cross-bedding. The bed is truncated by a succession of planar cross-stratified and horizontally laminated sandstones alternating with thin heterolithic siltstone-mudstone beds and lignite layers, which form undulating wavy ripple lamination (Fig. 7e). Mud drapes on foresets of planar cross-bedded sandstone are common (Fig. 7f). Locally, *Ophiomorpha* burrows were observed in the sandstones. Ripples are dominated by wave ripple laminations. The top of the section is formed by amalgamated trough-cross-stratified sandstone channels.

### 4.3.2. Interpretation

The presence of abundant trace fossils in planar and trough cross-bedded sandstones and sand-dominated heterolithic beds with subordinate beds of mudstone are interpreted as indicative of a shallow marine environment. *Ophiomorpha* indicates a high energy shoreface environment (Nagy et al., 2016) and *Skolithos* may indicate a sandy shore to shelf environment (Buatois & Mángano, 2011). Based on palynomorphs Abdul Hadi et al. (2017) concluded lower to middle shoreface, upper shoreface and offshore environments with pronounced storm, wave or tidal influence. The observed sandstone beds are interpreted as migrating tidal channels over mud-dominated tidal flats. Herringbone cross-stratification observed at Tusan Beach supports a periodic reversal in current direction in a tidal setting, often associated with a tidally-influenced sandy shoreface environment (e.g., Nichols, 2009; Ekwenye & Nichols, 2016). The undulating wavy laminations and mud drapes along with lignite on foresets may indicate a nearby coastal swamp environment or coastal floodplain, possibly delta plain. Heterolithic beds and wave ripple lamination indicate fluctuating water energy levels and weak currents with wave oscillations dominant, which are typical of tide-influenced environments (Vakarelov et al., 2012). The conglomerates composed of well-rounded sandstone gravel could represent an old beach section or a lag deposit. The dominant observed facies are trough cross-bedded sandstone and horizontally laminated sandstone (Fig. 7c and d) with bioturbation and amalgamated packages, interpreted as fluvial-tidal channels (Ali et al., 2016) interbedded with storm-dominated shallow marine deposits (Tab. 1). Intercalated carbonaceous muddy heterolithic beds are interpreted to indicate a low energy tidal environment or fluvio-estuary intervals (Tab. 1). The increase of heterolithic facies up-section, indicates a change from storm-dominated environment at the base to a tide-influenced shoreface at the top, suggesting shallowing water depths. The trough cross-bedded sandstones within the top section may represent fluvial channels. Collins et al. (2020) suggested a progradational to strongly aggradational deposition of the Lambir Formation in a large-scale, mixed-energy deltaic clastic wedge, where the lower delta plain was fluvial with superimposed tidal influence and the delta front was fluvial and wave dominated (storm-floods) with subordinate tidal influence.

## 4.4. Miri Formation

### 4.4.1. Observations

The Miri Formation outcrops along the coast section east of Lambir Hills (Mi-01) and at the eponymous city of Miri (Mi-02). Along the coastal road about 20 km south of Miri the formation outcrops in several smaller road cuts. At location Mi-01 a larger exposure was observed where c. 4.5

m of the succession is exposed in sub-horizontal beds (Fig. 8a). The outcrop is cut by a moderately SW-dipping thrust fault (Fig. 8a). Characteristic are amalgamated medium-grained sandstone packages (up to 1 m thickness), which show commonly planar or trough cross-bedding (Fig. 8b) interbedded with wavy ripple-laminated heterolithic mudstone-siltstone-sandstone alternations where crude flaser to lenticular bedding (Fig. 8c) is developed. Foresets of planar and trough cross-beds are formed by thin lignite laminae, and subhorizontally laminated lignite bands and undulated carbonaceous mudstone layers (up to c. 1 cm thick) are observed (Fig. 8b). Locally, abundant *Ophiomorpha* and *Skolithos* burrows are present in both sandstones and mudstone layers at a scale of several centimetres (Fig. 8b). The surface of the finer-grained sandstones is often reddish-brown due to iron oxide formed during weathering. The top of a hanging wall section consists of laminated mudstone, with crude lenticular bedding (Fig. 8d) that also form the base of the footwall of a thrust with a c. 2.5 m vertical offset. The footwall shows generally a coarser grain size with well-sorted cross-stratified pebbly sandstone layers interbedded with conglomerate beds (c. 1.2 m thick). Clasts consist of subrounded to well-rounded quartz, sandstone and shale fine granules and abundant angular to subrounded coal fragments (up to 3 cm in length) can be found in the lithofacies (Fig. 8e). Locally, the conglomerates are interbedded with thin irregular coal/carbonaceous mudstone bands (c. 3-10 cm). Parts of this conglomerate unit were dragged upwards into the fault zone (Fig. 8a). The top of the sequence (c. 0.4 m) is composed of a rippled sandstone interbedded with mudstone layers, which are both moderately bioturbated and show crudely-developed hummocky cross-stratification.

South of Miri city along the airport road is a c. 12 m high outcrop of the Miri Formation (sample Mi-02) in an abandoned quarry (Fig. 9a). The succession consists predominantly of massive cross-bedded sandstone and sandstone-dominated heterolithic deposits with bed thicknesses of c. 1.0-2.5 m, which show erosive bases into decimetre-scale mudstone-dominated heterolithic beds. The outcrop is cut by a series of normal faults (Fig. 9b) with abundant Fe-weathering and Fe-cementation on the sandstone and heterolithic bed surfaces. The fault geometry in the outcrop is discussed in detail in e.g., Sorkhabi & Hasegawa (2005) and Wannier et al. (2011). Typical lithology is a fine- to medium-grained massive, amalgamated sandstone that shows moderate to heavy bioturbation (*Ophiomorpha* and *Skolithos*) (Fig. 9c). Intercalated are parallel or wavy laminated mudstone-siltstone alternations (c. 10 cm thick) and cm-thin discontinuous lignite bands. Sandstone beds contain centimetre-scale elongated mud rip-up clasts (Fig. 9d), load casts and flame structures (Fig. 10a). The beds show cross-bedded foresets and hummocky and herringbone cross-stratification (Fig. 10b) in places. Sandstone-dominated heterolithic beds comprise irregular mudstone, lignite, and coal layers at millimetre- to centimetre-scale thickness. Ripple surfaces are asymmetrical, indicating transport direction towards the west (Fig. 10c). Mud-dominated heterolithic beds can develop lenticular bedding (Fig. 10d) similar to outcrop Mi-01. Gypsum coating on outcrop surfaces was observed (Fig. 10c). Fig. 9a illustrates a stratigraphic facies log of the upper part of the outcrop.

### *4.4.2. Interpretation*

The sedimentary structures within the Miri Formation include predominantly trough cross-beds and planar cross-beds with carbonaceous mud drapes on foresets as well as wavy to lenticular bedded heterolithic beds, which can be interpreted as tidal-influenced deposit (Reineck & Wunderlich, 1968). The presence of lignite and coal bands or clasts in the succession indicates a marshy environment nearby, especially the angular lignite clasts indicate a short transport distance without much reworking. The conglomerate layer with its sub-rounded to sub-angular clasts indicates periods of high energy, possible a shoreface storm deposit (Kumar & Sanders, 1976), and the conglomerates with rounded clasts could indicate a beach deposit. Offshore and lower shoreface to foreshore environments were also interpreted by Rahman & Tahir (2018). A tide-dominated environment is supported by the presence of *Ophiomorpha* and *Skolithos* that are common in tide-dominated estuaries with mixed tidal flat interaction (Buatois & Mángano, 2011; Ekwenye & Nichols, 2016; Nagy et al., 2016). This is also indicated with the occurrence of the trough cross-beds with carbonaceous mud drapes, which are interpreted as tidal channel deposits. The fine- to medium grained sandstone with wavy laminations might have developed in a low energy near tidal environment. Heterolithic beds with planar and ripple lamination and non-channelised layers are typically found in a tidal environment such as tidal flats (Feldman & Demko, 2015; Quijada et al., 2016). Flaser bedding is commonly observed in intertidal environments such as intertidal and subtidal flats, and tidal channels (Sellwood, 1972; Chakraborty et al., 2003; Dalrymple & Choi, 2007). Rip-up clasts are consistent with a storm endured environment and dewatering flame structures suggest high rates of sedimentation (Lowe, 1975) typical for a delta. Based on the hummocky and herringbone cross-stratification, abundant *Ophiomorpha* and *Skolithos*, and mud rip-up clasts a shallow marine deltaic to estuary environment is interpreted which was influenced by wave, tidal and sub-tidal mechanisms with sporadic storm events (Tab. 1) (Abieda et al., 2005; Ulfa et al., 2011; Siddiqui et al., 2017; Cheng, 2019; Collins et al., 2020). The high content of sand suggests a delta top environment. The facies log in Fig. 9a shows periodic changes between higher energy (trough cross-beds) and lower energy (rippled sandstone, heterolithic beds) typical for a tidally-influenced delta. Foraminifera reported by Tan et al. (1999) and Hutchison (2005) indicate a partially tide-dominated estuary environment. Syn-sedimentary extensional normal faults are related to a stress-releasing mechanism during folding of the Miri anticline possible associated with diapirism of underlying shale sequences (Wannier et al., 2011), as interpreted offshore from seismic (Clark, 2017; Chang et al., 2019; Morley et al., 2023).

## *4.5. Tukau Formation*

### *4.5.1. Observations*

The Tukau Formation is found only at an E-W trending ridge approximately 15 km southeast of Miri city with Lambir Hill (Bukit Lambir) as highest peak formed by the underlying Lambir Formation. The formation forms the uppermost succession at this ridge and was analysed in road cuts on the old road from Miri city to Bintulu. At Tu-01 the formation is dipping at a low to moderate angle towards the north (Fig. 11a). Dominant lithologies are thick mudstone beds (up to c. 1 m thickness) which are interbedded with wavy-laminated heterolithic mudstone-siltstone beds (c. 1-2 m thick) and fine- to medium-grained moderately sorted rippled sandstone layers (up to c. 30 cm thick) (Fig. 11b). The sandstone beds have erosional bases (Fig. 11c), show pinching out structures, and include cm-thin carbonaceous mudstone layers with wavy to ripple lamination in places (Fig. 11c and d), as well as coal fragments deposited on foresets of crudely developed planar cross-beds (Fig. 11d). Locally, there are sharp contacts between sandstone beds and heterolithic units. Ripple tops are rarely preserved and are dominated by asymmetric current ripples (Fig. 11b). Bioturbation is very sparse and restricted to a few *Skolithos* vertical tubes. The contact with the underlying Lambir Formation was not observed but appeared to be conformable with similar dip of beds.

### *4.5.2. Interpretation*

Thick mudstone beds suggest significant low energy periods like slack water conditions or flood plain environment (Ekwenye & Nichols, 2016; Gugliotta et al., 2016). Structureless mudstones likely record fluid mud deposition from high suspended sediment concentrations (Wright et al., 1988; Uncles et al., 2006), while laminated mudstones-siltstones record deposition by relatively low-energy suspension settling and minor traction currents (Collins et al., 2020). Planar cross-beds in the sandstones indicate migration of dune forms at low flow velocities, and in combination with erosional bases and channelized features indicate basal deposits of a delta channel (e.g., Miall, 2013; Johnson & Dashtgard 2014). The sandstones are interbedded with wavy laminated mudstone-siltstone heterolithic beds, including asymmetrical ripple marks, and indicate a fluctuating energy environment which is here interpreted as a tide-dominated delta plain dissected by tidal channels (Miall, 2013; Reading, 2013). Thicker sandstone beds were interpreted by Kessler & Jong (2017) as amalgamated tidal channel deposits interbedded with intertidal clastics. The unidirectional paleocurrents indicate a river-dominated environment up-section and are consistent with preservation of lateral or down-current migrating fluvial-tidal bars (Dalrymple & Choi, 2007; Legler et al., 2013; Gugliotta et al., 2015; Collins et al., 2020). The presence of coal flakes on foresets of planar cross-beds indicates a marshy environment which might have been periodically flooded. Thin lignite layers suggest coastal plain to shallow marine environments (Hutchison, 2005). Heterolithic facies may record high-frequency, low-magnitude river floods and interflood periods with a background tidal influence (Collins et al., 2020). A brackish water fauna was reported by sparse foraminifera (Wilford, 1961), and abundant carbonaceous material may indicate mangrove-rich floodplains and channel margins washed-in by fluvial–tidal currents. The *Skolithos* ichnofacies may indicate episodic sandy shore (littoral zone) to shelf (sublittoral zone) environment (Buatois & Mángano, 2011). The sparsity of bioturbation suggests a stressed environment with brackish-water conditions, probably related to mixed fluvial and tidal processes (Pemberton et al., 1992; MacEachern & Bann, 2008), and may support an overall fluvial-influenced character of the delta. Wilford (1961) reported foraminifera typically found in brackish water environment. The Tukau Formation outcrops are interpreted as delta plain deposition in muddy estuarines, interdistributary bays, or abandoned fluvial–tidal channels with an overall significantly reduced sand supply. A subtidal to intertidal environment of deposition was interpreted by Kessler et al. (2023), and Collins et al. (2020) interpreted the whole Lambir-Tukau sequence as fluvial-influenced and tide-influenced, coastal plain–shoreline succession.

### *4.6. Summary of depositional environments of the West Baram Delta deposits*

The base of the West Baram Delta deposits is formed by the Lower Miocene Sibuti Formation and consists of distal tidal channels, prodelta, inner shelf and slope deposits. Unconformably above are the deposits of the West Baram Delta, which have been subdivided by Liechti et al. (1960) into Belait, Lambir, Miri and Tukau formations based on minor variations in lithology, facies and age. This study identifies similar environments of deposition for all the formations which include storm, tidal, estuarine and river-dominated distributive channel deposits. This observation is consistent with detailed facies studies by e.g., Abieda et al. (2005), Ulfa et al. (2011), Ali et al. (2016), Siddiqui et al. (2017), Cheng (2019), and Collins et al. (2020). It is nearly impossible to differentiate the formations lithologically in the field; except for the lower sand content of the Tukau Formation. The Neogene successions in North Sarawak can therefore be subdivided into a lower part (consisting of southern Lambir/Belait-Sarawak, Lambir and Miri formations) dominated by storm- and tidal-influenced deposits with high input of sand-sized material, and an upper part (Tukau Formation) that shows a shallowing water depth dominated by estuarine and fluvial-tidal channels with high input of silt- and mud-sized material. The lithostratigraphic units of Liechti et al. (1960), already in question (Banda &

Honza, 1997; Collins et al., 2020), could therefore be further modified in future to simplify the stratigraphy.

# 5. Results

## 5.1. Biostratigraphy

Additionally, to the West Baram deposits, two samples from the underlying marine Sibuti Formation were analysed that yielded a foraminifera-rich assemblage. With the occurrence of *Catapsydrax dissimilis, Catapsydrax stainforthii, Globigerinoides trilobus, Globigeronides primordius, Paragloborotalia* sp. and *Globigerinoides subquadratus,* the Sibuti Formation samples can be assigned to Planktonic Foraminiferal zone N5b (20.4-18 Ma, lower Burdigalian, Early Miocene) and a shallow inner neritic environment. A similar assemblage was reported from the marine Setap Shale Formation in Sarawak by Breitfeld et al. (2020a) indicating the contemporaneous character of the marine mudstones. Samples from the Setap Shale Formation have been assigned an age range from N4 to N6 in Breitfeld et al. (2020a). As the Sibuti Formation is dated in Si-02 as lower Burdigalian (zone N5b), the underlying Setap Shale Formation is likely N4 to N5b (Aquitanian to lower Burdigalian).

Bioclasts in the West Baram Delta samples are poorly preserved in the analysed thin sections, and the samples yield only a few, mostly long-ranging specimens. Sample La-02 was barren of microfossils probably due to the fluvial-deltaic environment with high influx of clastic sediment. Be-01 from the southern Lambir/Belait-Sarawak Formation contained the most identified forms. With *Globoquadrina dehiscens*, *Planorbulinella larvata* (Fig. 12-1), *Hastigerinopsis* sp. (Fig. 12-2), *Orbulina universa, Orbulina suturalis* (Fig. 12-3), and *Calcarina* sp. (Fig. 12-6) an age from Planktonic Foraminiferal zones N9 (Langhian, Middle Miocene) to N20a (Early Pliocene) can be assigned for the succession. Additionally, reworked Upper Cretaceous foraminifera *Abathomphalus* sp. (Fig. 12-4) and *Globotruncana* sp. (Fig. 12-5) are present. These indicate recycling of the Rajang Group (Belaga Formation, Kelalan Formation) or even older sedimentary rocks (e.g., Pedawan Formation of the Kuching Zone). *Globotruncana* sp. has been recorded from the Kelalan Formation (Haile, 1962; Hutchison, 2005), which might be a lower Belaga Formation equivalent in the Miri Zone, making this a viable source of sediment. Hennig-Breitfeld et al. (2019, 2020) revised the stratigraphy of Belaga Formation turbidites in the Miri Zone, identifying metamorphosed sections previously mapped as the Eocene Bawang Member, which were suggested to be correlated with lower parts of the Belaga Formation, showing the possibility that there was Upper Cretaceous/Lower Paleocene Belaga Formation nearby at the time of deposition. La-01 from the Lambir Formation yielded *Paragloborotalia lenguaensis* (Fig. 12-7) and *Truncorotalia crassaformis* (Fig. 12-8), which can be placed in Planktonic Foraminiferal zone N17a (Late Tortonian, Late Miocene). Samples Mi-01 and Mi-02 from the Miri Formation yielded only long-ranging specimen *Calcarina* sp. (Fig. 12-9) and *Amphistegina* sp. (Fig. 12-10), along with rotaliid spp., which indicates a possible Middle Miocene to Holocene age. The Tukau Formation sample Tu-01 yielded *Quasirotalia guamensis* (Fig. 12-11), *Calcarina* sp. and *Elphidium* sp., along with small rotaliid. The assemblage indicates an Early Pliocene age for the succession. Age ranges of the West Baram Delta samples are illustrated in Fig. 13 alongside ranges presented in the literature. Sample LL1 from the Belait Formation on Labuan was barren.

### *5.2. Sandstone petrography of the West Baram Delta*

The analysed samples contain abundant quartz (up to c. 76% in sample La-02) with some feldspar (6-16%) and lithic fragments (9-13%). Matrix proportions are low with most samples being clearly below 10% and only the Miri Formation samples have around 10% matrix. The samples are sublitharenites (Lambir and Tukau formation samples) and those with more feldspar contents are subarkose (Miri and southern Lambir/Belait-Sarawak formation samples) (Fig. 14). Quartz grains are predominantly monocrystalline or monocrystalline undulose, with a few polycrystalline varieties and very rare volcanic quartz or chert grains. The feldspar is alkali feldspar, but a small number of plagioclase grains were also identified. Lithic fragments are dominated by sedimentary clasts, with some metamorphic and very rare volcanic clasts. Cementation is formed by limited thin quartz overgrowth, and feldspar shows low degree of dissolution into clay minerals. Based on the provenance diagrams (Dickinson & Susczek, 1979) the samples indicate a recycled orogenic and quartzose recycled to mixed source (Fig. 14). The Champion Delta samples presented by Hennig-Breitfeld et al. (2019) show significantly less feldspar content (Fig. 14) and can be differentiated from the West Baram Delta samples.

### *5.3. U-Pb zircon geochronology of the West Baram Delta*

No depositional age estimates can be given based on the zircon age assemblages as there are no Miocene zircons in the samples, indicating that no contemporaneous magmatism sourced the sandstones. CL imagery revealed that zircons are generally oscillatory or sector zoned, indicating a magmatic origin. A few core-rim structures were observed but not targeted with LA-ICP-MS. Convolute internal structure or homogeneous sites that indicate a metamorphic origin are also present, but are subordinate and are mostly found in Precambrian zircons. Individual sample plots can be found in Supplementary Fig. 2, and Supplementary Fig. 3 illustrates plots ranging from 0-4000 Ma.

### 5.3.1. *Southern Lambir/Belait-Sarawak Formation*

Zircon grains from the southern Lambir/Belait-Sarawak Formation sample Be-01 are subrounded with subangular and rounded varieties also common. 241 out of 317 zircon U-Pb analyses were classified as concordant. The most dominant zircon age cluster is in the Triassic with a tail that extends into the Permian (Fig. 15). The Triassic forms c. 23% (55 out of 241), and the Middle to Late Permian forms c. 5.4% (13 out of 241) of the whole age assemblage. There is an Early Permian age peak at c. 285 Ma. The second most prominent age cluster is Cretaceous with a very wide age range from c. 77 to 139 Ma (Fig. 15). The Cretaceous ages constitute c. 17% (40 out of 241) of the whole age assemblage. Other Phanerozoic ages are Middle Jurassic, Carboniferous, at the Silurian-Devonian boundary, at the Ordovician-Silurian boundary, and in the Cambrian. Around 37% of the ages are Precambrian with age peaks at c. 750 to 1000 Ma, c. 1.75 to 1.95 Ga, and c. 2.4 to 2.5 Ga. The oldest zircon is 2538 ± 18 Ma and the youngest is 37.6 ± 0.5 Ma.

### 5.3.2. *Lambir Formation*

The Lambir Formation was analysed in samples La-01 and La-02, and the combined zircon age histogram is displayed in Fig. 15. 180 concordant U-Pb zircon ages were acquired from 212 zircons. La-01 had 44 and La-02 had 136 concordant zircon ages. Grains are angular to subrounded, but a few rounded grains were also observed. Rounded grains usually show dark CL reflectance. The age distribution is bimodal in the Phanerozoic (Fig. 15). The most prominent fraction is Cretaceous, which forms 21.7% (38 out of 180) of the zircon assemblage and clusters around 110 to 130 Ma. Triassic ages form c. 16.7% (30 out of 180) and Permian 10% (18 out of 180), which makes this combined age cluster with 26.7% more abundant than the Cretaceous. There is a significant Early Jurassic cluster and a few Carboniferous, Ordovician to Silurian and Cambrian ages are present. There are 57 scattered Precambrian ages (c. 32%) that form smaller clusters including the most important one at c. 1.8 to 1.9 Ga. Other age cluster are at c. 800 Ma, 1.1 to 1.2 Ga, and at c. 2.5 Ga. The oldest grain is 3335 ± 24 Ma. The youngest grain is 40.8 ± 1.4 Ma which is significantly older than the depositional age.

### 5.3.3. *Miri Formation*

The Miri Formation was analysed in samples Mi-01 and Mi-02, and the combined zircon age plot is displayed in Fig. 15. 175 concordant ages were acquired from 207 zircons. Mi-01 had 130 concordant ages and Mi-02 had 45 concordant ages. Rounded to subangular grains are common. Rounded grains usually show dark CL reflectance. The Miri Formation samples show a bimodal age distribution in the Phanerozoic (Fig. 15). The most prominent age cluster is in the Cretaceous, which forms c. 26% (45 out of 175) of the age assemblages. The Cretaceous ages have a wide peak that cluster around 90 to 140 Ma. The second most prominent age cluster is Triassic, which extends into the Permian with an age range from c. 200 to 270 Ma (Fig. 15). The Triassic forms c. 14% (24 out of 175) and the adjacent mostly Middle to Late Permian 4% (7 out of 175). There is an Early Jurassic (around 195 Ma) and an Early Permian (around 282 Ma) age peak. Other Phanerozoic ages are scattered from the Cambrian to the Carboniferous. A few Middle to Late Jurassic zircons (around c. 170 Ma) form another small Jurassic age peak. Around 35% of the ages are Precambrian with major age peaks at c. 1.8 to 1.9 Ga, c. 2.5 Ga, and in the Neoproterozoic (c. 650 Ma, 800 Ma, 1.2 Ga). The oldest age is Paleoarchean at 3433 ± 14 Ma, and the youngest grain is 70.9 ± 1.5 Ma.

#### *5.3.4. Tukau Formation*

The Tukau Formation was analysed in sample Tu-01. 114 concordant ages were acquired from 150 grains. Most grains are angular to subangular. In contrast to the underlying formations that show a relatively heterogeneous zircon age distribution with bimodal Cretaceous-Triassic main age populations, the Tukau Formation sample is dominated by Cretaceous zircons with c. 30 % (34 out of 114) of the whole assemblage (Fig. 15). The majority of Cretaceous ages is between 110 to 120 Ma. Triassic ages represent c. 16% (18 out of 114) of the population, and Permian ages are very rare with c. 3.5% (4 out of 114). Other Phanerozoic ages are Jurassic and are scattered throughout the Palaeozoic. Devonian-Silurian and Cambrian zircons are the only other significant Phanerozoic grains. The Precambrian ages (c. 28%) are scattered with major clusters at c. 950 Ma, c. 1.2 Ga, between 1.8 to 1.9 Ga, and at c. 2.5 Ga. The oldest zircon is 2820 ± 25 Ma and the youngest zircon is 27.8 ± 0.8 Ma.

### *5.4. Heavy mineral analysis of the West Baram and Champion deltas*

Heavy minerals of the analysed samples are dominated by ultra-stable varieties with high numbers of zircon, tourmaline, and rutile that show variable abundancies (Fig. 16a). The mature assemblage results in very high zircon-tourmaline-rutile (ZTR) values (Hubert, 1962) from c. 85 (sample LL1) to 98 (sample La-01). Diagenetic $TiO_2$ phases anatase and brookite are also common, with brookite potentially also being an indicator of hydrothermal, pegmatitic or metamorphic primary sources (Mange & Maurer, 1992). There is also a number of anatase-rutile-brookite or anatase-quartz intergrowth composites. In total 2309 translucent heavy minerals were identified, with additionally

670 potentially diagenetic-related minerals, in particular anatase but also brookite and aluminium-phosphate-sulfate group minerals (APS).

The southern Lambir/Belait-Sarawak Formation sample (Be-01) is dominated by rutile (37.7%), zircon (33.6%) and tourmaline (21.9%) (Fig. 16a). Chrome spinel (5.4%) and REE-phosphates (1.5%) are the only other phases in the translucent assemblage. Authigenic $TiO_2$ phases anatase and brookite are both present. Additionally, some APS-group minerals (e.g., florencite, goyazite) were found related to breakdown of phosphate minerals like monazite and apatite.

The Lambir Formation samples (La-01 and La-02) are very different in rutile and zircon abundances with La-01 being dominated by rutile (79%) with low zircon counts (16%), while La-02 contains moderate zircon (38%) and rutile (47%) grains (Fig. 16a). Tourmaline is very low in both samples (c. 3-8%). Chrome spinel is present in both samples in low abundance (2.1-4.4%). La-02 also has a few REE-phosphates (monazite, xenotime) and one garnet grain was found. Anatase and brookite form the diagenetic heavy minerals in the Lambir Formation samples.

The Miri Formation samples (Mi-01 and Mi-02) show similar high variability in zircon, rutile and also tourmaline (Fig. 16a). Mi-01 is dominated by zircon (48%) with rutile (22%) and tourmaline (17%) being less abundant, while tourmaline and rutile are both the dominant translucent heavy mineral phase in Mi-02 (c. 35%) with zircon (22%) being less abundant. Chrome spinel values (5-10%) are relatively high compared with the other samples. Monazite and garnet are present in low numbers in both samples. Anatase as well as APS-group minerals form the majority of the diagenetic heavy minerals.

The Tukau Formation sample Tu-01 (Fig. 16a) contains predominantly zircon (66.9%) and significantly less abundant tourmaline (13.5%) and rutile (14.3%). Chrome spinel constitutes 2.3%, which is one of the lowest chrome spinel values in the analysed samples, and REE-phosphates (monazite) form 2.3% of the translucent assemblage. Authigenic $TiO_2$ phases are rare and dominated by anatase. However, some anatase intergrowth and $TiO_2$-Fe composite grains were identified.

Sample LL1 from the lower Belait Formation (Layang-Layangan Beds) from Labuan contains predominant rutile (45.3%) and zircon (30.1%). Tourmaline (9.8%), chrome spinel (9.1%), garnet (4.3%) and a few REE-phosphates (monazite, xenotime) form the rest of the translucent heavy mineral assemblage (Fig. 16a). The sample has a high number of authigenic $TiO_2$ phases, especially anatase.

The very mature assemblages prevent the use of the most commonly used heavy mineral indices (Morton & Hallsworth, 1994) as count numbers are too low. The ZTR (Hubert, 1962), RuZi (Morton & Hallsworth, 1994) and ZTi (Mange & Wright, 2007) indices are displayed in Tab. 3. ZTR is very high for the West Baram Delta samples, while the Champion Delta sample (LL1) has a slightly lower value. The RuZi and ZTi indices show mostly an inverse correlation, which might be a result of different energy environments of deposition (see Hennig-Breitfeld et al., 2019) as tourmaline is significantly less dense than rutile or zircon. The indices GZi (garnet-zircon) and CZi (chrome spinel-zircon) also show significant changes despite low numbers (Fig. 16b). GZi values show an increase from 0 at the base of the West Baram Delta deposits up to c. 6 in sample Mi-02. CZi shows a decrease up-section from 14 at the base to 11 in La-02, and a significant increase to 20 in the two Miri Formation samples (Tab. 3). The top deposit of the West Baram Delta represented by the Tukau Formation shows a sharp decrease in both GZi and CZi, reflecting the absence of garnet and low numbers of chrome spinel grains (Fig. 16b). The Champion Delta sample LL1 (Belait Formation) shows the highest values of both CZi with 23 and GZi with 13 (Fig. 16b), which may indicate a provenance different from the West Baram Delta.

## 6. Discussion

### 6.1. Differences in the onshore West Baram Delta deposits

Although the analysed samples all show slightly different abundances of quartz, feldspar, and lithic fragments, the number of samples analysed is not sufficient to characterise the formations with confidence. The heavy mineral assemblages and in particular the detrital zircon ages are better suited to differentiate between the successions. Translucent heavy minerals show some variability in zircon, rutile and tourmaline numbers (Fig. 16a), likely a result of hydraulic sorting as reported by Hennig-Breitfeld et al. (2019) for the Middle Miocene Balingian-Mukah Delta deposits in central Sarawak.

The uppermost sample Tu-01 from the Tukau Formation shows however a significant increase in zircon and decrease in rutile, garnet and chrome spinel (Fig. 16a), which is best reflected in the CZi and GZi indices (Fig. 16b) and could be related to a provenance change. Detrital zircons show an increase in Cretaceous ages and decrease in Triassic ages up-section, with the Tukau Formation sample having the highest abundance of Cretaceous zircons (Fig. 15). The only other published zircon data from the formations in northern Sarawak studied here is from the Tukau Formation (Nagarajan et al., 2017) and is similar to our sample (Fig. 15) with a dominant Cretaceous age cluster. In contrast to sample Tu-01, Nagarajan et al. (2017) reported very few Precambrian ages (12%), but those present resemble the age peaks of this study (Fig. 15). Although it is not certain, the high abundance of Cretaceous detrital zircons and differences in the heavy mineral assemblage may indicate a change in source and drainage, which may support an interpretation of an unconformity at the base of the Tukau Formation (e.g., Kessler & Jong, 2015). However, Togunwa & Abdullah (2017) concluded that there is no distinct difference in the geochemical characteristics of the Lambir, Miri and Tukau formations in terms of source input, depositional conditions, and thermal maturity, which shows the need for further detailed studies.

Deposition of the formations was in mixed-energy delta environments, with a wide range of facies ranging from shallow marine, storm, tidal to fluvial deposits. Abundance of tide-dominated deposits in the Tukau Formation and absence of shallow marine and delta front facies indicate a shallowing of the basin up-section.

### *6.2. Belait Formation(s)*

Figure 17 compares the detrital zircon U-Pb ages from the Belait Formation from Labuan presented by Hennig-Breitfeld et al. (2019) and Burley et al. (2021) with the southern Lambir/Belait-Sarawak Formation sample Be-01. The southern Lambir/Belait-Sarawak sample is dominated by Triassic zircons whereas the Belait Formation samples from Labuan are either dominated by Cretaceous zircons or a Cretaceous-Triassic assemblage (Fig. 17). Precambrian ages also vary between the samples. Be-01 has a heterogeneous age signature with various peaks at c. 550 Ma, 750 Ma, c. 850 Ma, c. 1.75-1.95 Ga and c. 2.4 Ga, while the Labuan samples show a more homogeneous pattern mostly dominated by a distinctive age peak at c. 1.8-1.9 Ga (Fig. 17). It can therefore be concluded that the southern Lambir/Belait-Sarawak Formation has a provenance different from the Belait Formation on Labuan, supporting the suggestion of Banda & Honza (1997) that they represent different formations. Figure 18 illustrates a multidimensional scaling (MDS) plot (Vermeesch, 2013) of the West Baram Delta and Champion Delta zircon age data of this study, which visualises the dissimilarity of Be-01 with the Belait Formation of Labuan and suggests an association with the Lambir and Miri formations. Currently only zircon U-Pb data from the Labuan succession of the Champion Delta are available, but since the Labuan deposits are the extension of the Brunei deposits (e.g., Wilson & Wong, 1962; Hutchison, 2005) it is expected that the Belait Formation in Brunei will be different from the southern Lambir/Belait-Sarawak Formation.

### *6.3. Provenance of the Neogene onshore West Baram Delta successions*

The analysed West Baram Delta sediments are all classed as subarkose or sublitharenite, with quartz-rich compositions, and a quartzose recycled orogenic character is indicated by the quartz-feldspar-lithic fragments (QFL) and monocrystalline quartz-feldspar-total lithic fragments (QmFLt) diagrams (Fig. 14), suggesting a multi-recycled provenance. In humid tropical conditions these plots should be considered with caution as feldspar dissolution and breakdown of lithic fragments is enhanced (Suttner et al., 1981; Sevastjanova et al., 2012). This is illustrated by the comparison with the two underlying potential source rocks, the Rajang Group turbidites and the Tatau-Nyalau Delta (Sunda River Delta) sediments, which are mostly slightly less quartz-enriched (Fig. 14). Only a few Rajang Group samples stretch into the quartz-rich field of the West Baram Delta sediments. This, however, shows that the Rajang Group sediments would potentially suit better as source candidate. The Belaga Formation of the Rajang Group was previously dismissed as potential source for the Miocene successions (Hutchison, 2005) due to a fine-grained often mud-dominated appearance (e.g., Baioumy et al., 2021). However, thick sandstone beds, debrites and high density turbidites with high contents of sand-sized material have been reported since, especially from the Kapit, Pelagus, Metah and Bawang members (Bakar et al., 2007; Galin et al., 2017; Kuswandaru et al., 2018; Hennig-Breitfeld et al., 2019; Ahmed et al., 2020, 2021), and Nagarajan et al. (2021) reported sandstone beds within the Kelalan Formation. However, the amount of polycrystalline quartz in the Rajang Group samples seems to be higher than in the West Baram Delta sediments, which would indicate some differences between eroded and preserved Rajang Group sediments. To the north in Sabah, the Oligocene to Early Miocene Crocker Formation and the possibly Paleocene to Eocene Trusmadi Formation also consist of very quartz-rich older turbiditic deposits (Fig. 14; van Hattum et al., 2013). In contrast to the West Baram Delta samples, they have higher contents of polycrystalline quartz and lower contents of feldspar (van Hattum et al., 2013). They are therefore better suited to be a source for the Champion Delta sediments (Fig. 14), although there is some overlap in the QFL diagram with the Lambir Formation samples.

The translucent heavy mineral assemblages are dominated by ultra-stable zircon, rutile and tourmaline. Rare chrome spinel, REE-phosphates and traces of garnet and APS-group minerals (e.g., florencite) are also present in the assemblages (Fig. 16a). This ultra-stable assemblage suggests multi-recycling in which unstable heavy minerals are not preserved. The near absence of apatite, which is commonly found in very low numbers in the underlying successions (Galin et al., 2017; Hennig-Breitfeld et al., 2019; Breitfeld et al., 2020a), suggests apatite dissolution as a result of acidic conditions (Morton, 1984) in the West Baram Delta. Garnet is also present in low abundances in the underlying Nyalau and Belaga formations (Galin et al., 2017; Breitfeld et al., 2020a), but almost absent in the West Baram deposits supporting an acidic environment interpretation (Morton, 1985). The underlying deposits of the Sunda River Delta (Tatau-Nyalau formations) and the Rajang Group turbidite fan are also composed of multi-recycled ultra-stable heavy mineral assemblages (Fig. 19) interpreted to be deposited under acidic conditions (Hennig-Breitfeld et al., 2019; Breitfeld et al., 2020a). Their assemblages were likely recycled into the West Baram successions with further removal of unstable varieties. Both, the Sunda River Delta deposits and the Rajang Group turbidites show comparable heavy mineral assemblages to the West Baram Delta, indicating recycling into the West Baram Delta (Fig. 19). Besides the slightly higher abundance of unstable varieties, it is notable that tourmaline is less abundant in the West Baram Delta (Fig. 19), which is probably a sorting effect due to its lower density. Cui et al. (2023) reported a similar ultra-stable heavy mineral assemblage from one sample of the Lambir Formation, dominated by zircon with small amounts of tourmaline, rutile and chrome spinel. The Champion Delta sample contains more chrome spinel and garnet (evident by CZi and GZi indices) as all other samples (Fig. 19), suggesting a different source. Although only a single sample was analysed the increased values appear to be significant.

There is no evidence for input of fresh material from e.g., the Plio-Pleistocene mafic volcanism in central Borneo exposed in Usun Apau, Linau Balui or Hose Mountains (Cullen et al., 2013) or from the Upper Miocene Kinabalu Granite in Sabah (Cottam et al., 2013) which would be expected to supply unstable heavy minerals, such as amphibole, epidote or pyroxene. It is therefore concluded that i) the magmatism in central Borneo post-dates the deposition of the West Baram sediments, ii) Sabah was not a source, and iii) uplift of the Kinabalu pluton post-dates the deposition of the West Baram sediments. Van Hattum et al. (2013) reported an ultra-stable heavy mineral assemblage from the Crocker and Trusmadi formations in Sabah with low numbers of apatite, chrome spinel, monazite, but also amphibole, pyroxene and significant numbers of garnet. Thus, the Paleogene Sabah turbidites were not a source for the West Baram Delta, but potentially a source for the Champion Delta.

Detrital zircon age signatures show a typical western Borneo pattern with variations in dominant Cretaceous and Permian-Triassic age peaks and a prominent Paleoproterozoic peak at c. 1.7-1.9 Ga (e.g., van Hattum et al., 2013; Galin et al., 2017; Breitfeld & Hall, 2018; Hennig-Breitfeld et al., 2019). While the Cretaceous age peak is related to the Schwaner Mountains granitoids, volcanics and metamorphic rocks (Williams et al., 1988; Hennig et al., 2017; Breitfeld et al., 2020b; Batara & Xu, 2022; Qian et al., 2022; Wang et al., 2022), the Permian-Triassic age peak is related to West Borneo (Williams et al., 1988; Setiawan et al., 2013; Breitfeld et al., 2017; Hennig et al., 2017; Wang et al., 2021a) and the Malay Peninsula (e.g., Liew & Page, 1985; Sevastjanova et al., 2011; Searle et al., 2012; Oliver et al., 2014; Ng et al., 2015; Basori et al., 2018; Cao et al., 2020; Quek et al., 2021). There are also Triassic basement rocks in eastern Sabah in the Segama Valley (Leong, 1974; Burton-Johnson et al., 2020; Wang et al., 2023), which could have contributed sediments, before the Crocker Range was uplifted to form a drainage divide in the Early Miocene (Hutchison, 1996). Rather than first-cycle sediments directly derived from the basement, the studied delta deposits reflect multi-recycling as evidenced by the light and heavy mineral compositions. The majority of material was likely recycled from the Belaga Formation (Rajang Group) as a Cretaceous-dominated age pattern with some Triassic zircons (Fig. 20) dominates the succession (Galin et al., 2017; Hennig-Breitfeld et al., 2019; Wang et al., 2021b; Zhao et al., 2021; Zhu et al., 2022). The southern Lambir/Belait-Sarawak Formation sample Be-01 with its high proportion of Triassic zircons also indicates recycling of the underlying Nyalau Formation, which is dominated by Triassic ages (Fig. 20) (Hennig-Breitfeld et al., 2019; Breitfeld et al., 2020a). As potentially the oldest of the West Baram Delta deposits, the southern Lambir/Belait-Sarawak Formation suggests a source of the Oligocene to Lower Miocene Nyalau Delta sedimentary rocks that were above the Rajang turbidites. Later unroofing of the deeper Rajang Group supplied sediment to higher parts of the West Baram Delta. An unmixing of the detrital zircon age cluster to assess contribution from potential source rocks (Sundell & Saylor, 2017) is illustrated in Supplementary Fig. 4 and listed in Supplementary Table 6. Contribution of upper Tatau-Nyalau deposits to sample Be-01 is c. 72% and decreases in the other West Baram samples to 25%, while the contribution of Rajang Group deposits increases from 8% to 42%.

### *6.4. Sedimentation rates*

The West Baram Delta formations in onshore northern Sarawak interfinger and a precise total thickness cannot be given with confidence. A minimum total thickness of 5.5 km for the Lambir, Miri, southern Lambir/Belait-Sarawak and Tukau formations is based on thickness estimates by Liechti et al. (1960). Based on available biostratigraphy data the sediments were deposited between the Langhian (Lambir, Miri and southern Lambir/Belait-Sarawak formations) and Early Pliocene (Tukau Formation), in a maximum period of 11 Myr. This corresponds to an average minimum sedimentation rate of 50 m/100 ka, by using the 7 km thickness maximum estimate the average sedimentation rate would reach 64 m/100 ka. Sedimentation rates calculated by Morley et al. (2016) for the Late Miocene to Pliocene in offshore NW Borneo based on well data are similarly high with values of 40 to 80 m/100 ka. The high values result from intense tropical weathering and high erosion rates of the uplifted central Borneo mountain range coupled with high subsidence rates that created the accommodation space. The area of Neogene sediments onshore covers at least 6000 $km^2$ in northern Sarawak (Fig. 4a). Assuming a minimum thickness of 5.5 km, a volume of 33,000 $km^3$ of sediment was removed from central Borneo and deposited in the onshore part of the West Baram Delta. Morley & Back (2008) determined more than 76,000 $km^3$ of Middle to Upper Miocene clastic sediments in the offshore region, suggesting that at least 100,000 $km^3$ of sediment was removed from uplifted central Borneo in the Miocene alone.

The onshore Champion Delta successions (Belait and Seria formations) are estimated to have a thickness of at least 8 km (Liechti et al., 1960) and up to 12 km (Sandal, 1996; Collins et al., 2017). Assuming deposition from the Middle Miocene to the Pliocene (c. 11 Myr), this gives an even higher sedimentation rate between 73 m/100 ka to 110 m/100 ka.

### *6.5. Drainage of West Baram and Champion deltas*

The West Baram and Champion delta systems initiated in the Langhian and after late Pliocene uplift both systems prograded westwards to their present-day location. Based on their different provenance characteristics, the Neogene drainage can be inferred. Morley & Back (2008) modelled the West Baram river to have drained highlands in central Borneo (present-day northern Sarawak), while the Champion 'river', consisting of Padas and Trusan paleo-rivers, drained highlands in southern Sabah and northernmost present-day Sarawak, similar to the present-day situation. Both systems have been interpreted to have formed a large delta province in the Miocene (Morley & Back, 2008).

The detrital zircon ages of the West Baram and Champion deltas are relatively similar, but differ slightly in Cretaceous age peaks. Besides the main age peak in both delta sequence at c. 110-120 Ma, there is a second prominent Cretaceous peak in the Champion Delta samples at c. 90-100 Ma (Fig. 20). The main age peak is well developed in most underlying sedimentary rocks (e.g., Rajang Group, Temburong Formation) and is originally related to the Sepauk Tonalite in the Schwaner Mountains (Breitfeld et al., 2020b). The Upper Cretaceous age peak is only found in samples from Sabah, in particular the in the Temburong Formation, but also subordinate in the Rajang Group equivalents (e.g., Sapulut Formation) and suggests a correlation (Fig. 20). Although there are granitoids of this age in the Schwaner Mountains (Hennig et al., 2017; Breitfeld et al., 2020b), the generally low abundance of zircons of this age range in Paleogene sedimentary rocks from Sarawak (Fig. 20; Galin et al., 2017; Breitfeld & Hall, 2018) indicates that those granitoids were not a main source for the underlying successions. Inherited Upper Cretaceous zircon grains in the Ranau ultramafic rocks of Sabah (Tsikouras et al., 2021) indicate a thermal event of Late Cretaceous age in Sabah, which could be the source of the upper Cretaceous zircons. The samples from the two deltas also differ in the Permo-Triassic zircon age population. The West Baram Delta samples have a Triassic peak at c. 240-250 Ma, whereas the Champion Delta samples have a peak at 230-240 Ma (Fig. 20). It is however not clear if this relates to the underlying successions. The Champion Delta samples also lack latest Archean to early Proterozoic (at c. 2.5 Ga) zircons, which are present in the West Baram delta samples (Fig. 20). An unmixing model of the detrital zircon data (Sundell & Saylor, 2017) illustrates the differences between the Champion and West Baram detrital zircon age record (Supplementary

Fig. 4, Supplementary Table 6). The Champion Delta samples show a contribution of c. 52% from Rajang Group equivalents in Sabah and a total contribution of c. 63% from Sabah. The West Baram Delta samples (Lambir, Miri, Tukau formations) in contrast, show a contribution of 42% from Rajang Group deposits in Sarawak (mainly Belaga Formation) and a total contribution of 68% from Sarawak source rocks with only about 32% potential contribution from Sabah.

There are slightly higher proportions of garnet and chrome spinel in the Champion Delta sample LL1, suggesting a different provenance compared to the West Baram Delta samples (Fig. 16). Both heavy minerals have been reported from the underlying Crocker and Trusmadi formations in Sabah as well as from the Sabah Setap Shale Formation below the Belait Formation on the Klias Peninsula (van Hattum et al., 2013; Cui et al., 2023). Although, those heavy minerals are also commonly found in samples from Sarawak (e.g., Belaga, Nyalau formations) (Fig. 19), their abundance in Sabah is higher on average (van Hattum, 2005). From the West Baram Delta, only the Miri Formation samples show CZi and GZi indices comparable to LL1 (Fig. 16), which could indicate a similar source of the Miri Formation and the Belait Formation.

The general similarities of the West Baram Delta and Champion Delta samples suggest similar sources for the majority of material. The West Baram Delta samples were all sourced mainly by uplifted Rajang Group turbidites (Belaga Formation) (Fig. 21) based on their similarities in detrital zircon ages, petrography, and heavy mineral assemblages. The Champion Delta samples were likely derived by recycling of Rajang Group sediments in northern Sarawak and the Rajang Group equivalents, Temburong and Crocker formations in Sabah with potential fresh input from the Sabah ophiolite and peridotites (Hutchison, 1975; Imai & Ozawa, 1991; Omang & Barber, 1996; Tsikouras et al., 2021) that would account for higher chrome spinel and garnet contents. Rather than a single large delta province as discussed by Morley & Back (2008), it is more likely that the West Baram and Champion deltas formed two separate provinces throughout the Miocene to Pliocene, with only the Miri Formation suggesting temporal overlap of the provinces.

## 7. Conclusions

The West Baram and Champion deltas were formed in the Middle Miocene after uplift of central Borneo, resulting in the Nyalau Unconformity/EMU or TCU. Deposition was in mixed-energy delta environments, ranging from shallow marine to fluvial. Tide-dominated deposits in the Tukau Formation may indicate a shallowing of the basin up-section. Provenance characteristics can be used to distinguish between the different delta systems, which is important as both systems extend offshore where they form major hydrocarbon reservoirs and the sequences potentially interfinger.

The West Baram Delta was sourced by recycling of the underlying Rajang Group sedimentary rocks with some input from recycled Nyalau Formation. The potentially oldest succession of the West Baram sequence, the southern Lambir/Belait-Sarawak Formation shows thereby the highest input from the Nyalau Formation, suggesting a gradually unroofing of the Rajang Group throughout the Miocene to Pliocene. The Kelalan Formation (Rajang Group) in particular could be a viable source based on the reworked foraminifera *Globotruncana* sp. in sample Be-01. Abundance of ultra-stable heavy minerals and quartz-rich character in the analysed deposits indicates multi-recycled sources. Lithologically, the southern Lambir/Belait-Sarawak, Lambir and Miri formations are very similar and further work could simplify the stratigraphy. Only the mud-richer Tukau Formation with potential slightly different provenance can be distinguished from the other formations.

The Champion Delta shows very similar characteristics in detrital zircon ages and heavy mineral assemblage. Its higher content of chrome spinel and garnet, and its additional Upper Cretaceous and Upper Triassic detrital zircon age peaks indicate a partly different provenance, which is interpreted to be turbidites and ultra-mafic rocks in Sabah along recycling of Rajang Group of Sarawak.

Sparse foraminifera identified in this study from the West Baram Delta deposits, in combination with literature data, indicates that the Lambir, Miri and southern Lambir/Belait-Sarawak formations all are relatively similar in age, possibly ranging from Langhian to Messinian. The Tukau Formation overlies the successions in the Early Pliocene. The adjacent Belait Formation of the Champion Delta was also contemporaneous possibly ranging in age from the Middle Miocene (Serravallian) to the Late Miocene (Tortonian).

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B34, B47 and B54 from the Tukau Formation of his study in 2017. Eldert Advokaat and an anonymous reviewer provided helpful comments for the manuscript.

## Figure Captions

Figure 1: a) Sedimentary basins of the southern South China Sea region in Southeast Asia (modified from Hennig-Breitfeld et al., 2021; Breitfeld et al., 2022). The red box indicates the area of b) along the NW Borneo coastline. b) NW Borneo delta systems and offshore tectono-stratigraphic provinces (modified from Sandal, 1996; Tingay et al., 2005). The Balingian-Mukah, Bintulu-Kemena (Kakus), West Baram, and Champion Delta systems of latest Early Miocene/early Middle Miocene to Pliocene age, were preceded by the Meligan Delta (grey) of potentially Early Miocene. The blue box represents the research area and is enlarged in Fig. 4.

Figure 2: a) Tectono-stratigraphic zones of western Borneo (modified from Haile, 1974; Hennig et al., 2017; Breitfeld et al., 2020b). The research area is located in the northern part of the Miri Zone. The red box indicates the location of the zoomed map. b) Offshore tectonic and hydrocarbon provinces of Sarawak, NW Borneo (modified from Hazebroek & Tan, 1993; Mazlan et al., 2013). The offshore Baram Delta province includes the West Baram and Champion Delta successions. The West Baram Delta deposits are the focus in this study.

Figure 3: Stratigraphic overview of the Miri Zone in northern Sarawak and units in Brunei, west Sabah and on Labuan (modified from Hennig-Breitfeld et al., 2019; Breitfeld et al., 2020a). The West Baram is separated by the Nyalau Unconformity from the underlying delta sequence consisting of the Tatau-Nyalau (delta-tidal deposits) and Setap Shale (marine deposits), while the Champion Delta is separated by the Top Crocker Unconformity from the Crocker and Temburong (turbidite slope deposits) formations. (* indicates the phases of Baram evolution from Koopmans, 1996).

Figure 4: Geological map (modified from Heng, 1992; Breitfeld et al., 2020a) with sample locations and stratigraphy (after Liechti et al., 1960; Banda & Honza, 1997; Hutchison, 2005; Hennig-Breitfeld et al., 2019; Breitfeld et al., 2020a).

Figure 5: Field photographs of the Sibuti Formation at location Si-02. a) and b) Rhythmically interbedded siltstones and mudstone from the Sibuti Formation (lower Burdigalian; sample Si-02) disconformably on top of dark shales of the Setap Shale Formation (Aquitanian; based on sample Set-01 in Breitfeld et al. (2020a) east of Bekenu).

Figure 6: Field photographs of the southern Lambir (Belait-Sarawak) Formation (Be-01). a) Facies log in outcrops of swaley-hummocky cross-stratified sandstone beds (F10) alternating with laminated

sandstone beds (F7). b) Massive sandstone bed with crude hummocky cross-stratification and erosional base (F10) on top of laminated sandstone (F7). c) Swaley cross-stratified sandstone (F10). d) Horizontal laminated sandstone (F7).

Figure 7: Field photographs of the Lambir Formation. a) Sandstone conglomerate south of Tusan cliff. Clasts dominated by sub-rounded to rounded sandstone and minor quartz. b) Bioturbation (mostly *Skolithos*) in wavy-laminated sandstone. c) Facies log of La-01 dominated by laminated sandstone interbedded with heterolithic beds and a carbonaceous mudstone bed. Top of the section is dominated by cross-bedded sandstones. Insert figure is a zoom of the carbonaceous mudstone bed, which is deeply weathered and colours the underlying laminated sandstones grey. d) Facies log of La-02 displaying a set of stacked tidal channels. e) Undulating wavy ripple lamination in laminated sandstone (La-02). f) Carbonaceous mud drapes on planar cross-stratified sandstone (La-02).

Figure 8: Field photographs of the Miri Formation (Mi-01). a) Facies log of a representative section along the road cuts, showing the thrusting of the left-hand side (hanging wall) over the right-hand side (footwall). Displacement may be approximately 2.5 m. The hanging wall is dominated by heterolithic beds dissected by a planar cross-stratified sandstone channel. The footwall shows several pebbly sandstone to conglomerate beds and layers. Parts of coarser units were dragged upwards into the fault zone. b) Planar cross-stratified bioturbated sandstone with abundant *Ophiomorpha* and *Skolithos* burrows. Three prominent undulated to rippled carbonaceous mudstone layers are intercalated. c) Flaser to lenticular bedding in bioturbated laminated sandstone. d) Crudely-developed lenticular bedding in mudstone-siltstone alternation the hanging wall to section. e) Pebbly conglomeratic sandstone composed of angular to subrounded coal clasts, and subangular to rounded clasts of quartz, sandstone and shale

Figure 9: Field photographs of the Miri Formation at the airport road old quarry (Mi-02). Exposed is the “456 Sands” of the Miri Formation that was a minor reservoir for the Miri field. a) Facies log of the upper section quarry, showing periodic changes between higher energy (trough cross-beds) and lower energy (rippled sandstone, heterolithic beds). b) Set of normal faults in a succession of tidally-dominated cross-bedded sandstone and intercalated heterolithic deposits. Faults Ft1 and Ft3-F5 show a displacement of c. 25-40 cm. Main displacement is along Ft2 with c. 5 metres (indicated by the yellow circle). Ft2 and Ft3 are antithetic faults forming a small graben structure. The whole outcrop extents farther with numerous other normal faults. c) *Ophiomorpha* and *Skolithos* bioturbation in amalgamated laminated sandstone. d) Sharp contact between bioturbated laminated mudstone and bioturbated amalgamated sandstone with abundant mud rip-up clasts.

Figure 10: Field photographs of the Miri Formation at the airport road old quarry (continued). a) Flame structures of upwelling mud into overlying sandstone. b) Crudely developed herringbone cross-stratification. c) Asymmetrical ripple lamination in sandstone-dominated heterolithic beds. Secondary gypsum weathering crust on the outcrop surface. d) Crudely developed lenticular bedding in bioturbated laminated mudstone-siltstone alternation.

Figure 11: Field photographs of the Tukau Formation (Tu-01) at Lambir Hills. a) Facies log displaying mud-dominated deposits overlain by rippled sandstone beds. The beds dip shallow towards the north. b) Well-preserved asymmetrical ripple tops. c) Erosional base on coarser sandstone overlying mudstone-siltstone alternations. Carbonaceous and coaly mud clasts and laminae in the upper part of the sandstone bed. d) Planar cross-stratification with coal fragments deposited on foresets, overlain by asymmetrical ripple laminae.

Figure 12: Plate of representative foraminifera. The arrows point to the foraminifera which are mainly poorly preserved, recrystallised and some are pyritised. 1) *Planorbulinella larvata* (Parker and Jones), Be-01. 2) *Hastigerinopsis* sp., Be-01. 3) O*rbulina suturalis* (Brönnimann), Be-01. 4) Reworked Upper Cretaceous, *Abathomphalus* sp., Be-01. 5) Reworked Upper Cretaceous, *Globotruncana* sp., Be-01. 6) *Calcarina* sp., Be-01. 7) *Paragloborotalia lenguaensis,* La-01. 8) *Truncorotalia crassaformis* (Galloway and Wissler), La-01. 9) *Calcarina* sp., Mi-02. 10) *Amphistegina* sp., Mi-01. 11) *Quasirotalia guamensis* Hanzawa, Tu-01. Scale bars on photomicrographs 0.3mm.

Figure 13: Stratigraphic age range for the onshore West Baram Delta deposits based on palaeontological and geochronological analyses. Sample numbers of this study are in bold. Literature age ranges are in italic. No age data from the southern Lambir/Belait-Sarawak Formation (abbreviated SL in the diagram) was previously available. The time scale is from Gradstein et al. (2012). (*Sr isotope age from Miri Formation by Kocsis et al. (2022) is from Brunei where differentiation between West Baram and Champion Delta deposits in the field becomes difficult).

Figure 14: Light mineral modal composition of analysed sandstone samples. Left panel QFL diagram display sandstone classification (after Pettijohn et al., 1987). Middle QFL and right QmFLt diagrams display light mineral provenance (after e.g. Dickinson & Suszek, 1979). Data of potential source rocks from the underlying Tatau-Nyalau delta system (Breitfeld et al., 2020a), Rajang Group turbidites in Sarawak (Galin et al., 2017, Hennig-Breitfeld et al., 2019), and from turbidite successions in Sabah (Crocker, Sapulut, Trusmadi formations) (van Hattum et al., 2013).

Figure 15: Detrital zircon age histograms with kernel density curves for the West Baram Delta samples in stratigraphic order. The southern Lambir (Belait-Sarawak) and the Miri Formation samples show a bimodal distribution in the Phanerozoic with main peaks in the Cretaceous and at

the Permian-Triassic boundary. The Lambir and Tukau formations only show a strong Cretaceous age peak. Precambrian ages vary throughout the samples with the Miri Formation having the highest abundance (c. 54%). The figure also includes a combined plot for Tukau Formation samples published by Nagarajan et al. (2017), which shows a similar distribution in the Phanerozoic but significantly less Precambrian ages. Bin size of 10 Ma for Phanerozoic ages and 50 Ma for ages > 500 Ma. Kernel density bandwidth 5 for Phanerozoic ages and 15 for ages > 500 Ma. X=number of samples.

Figure 16: Heavy mineral assemblages of the studied intervals, indicating zircon, rutile and tourmaline dominated assemblages. a) 100% stacked bar plot illustrating translucent heavy mineral species identified with Raman spectroscopy. Notable is the increase in chrome spinel and garnet in LL1, and the high zircon proportions in Tu-01. b) Critical mineral indices CZi (chrome spinel-zircon) and GZi (garnet-zircon) amplifying the settle differences. The Tukau Formation (Tu-01) shows the lowest values in both, suggesting a source change or a change in hydraulic conditions compared to the underlying deposits. The Champion Delta (LL1) has the highest values, indicating a different provenance where chrome spinel and garnet were widely available. The Miri Formation samples with the highest values for the West Baram Delta could indicate an episodic Champion Delta influence on the West Baram Delta.

Figure 17: Comparison of the southern Lambir (Belait-Sarawak) Formation with the Champion Delta samples from the Belait Formation on Labuan ($^1$LTB samples from Hennig-Breitfeld et al., 2019; $^2$LL1 from Burley et al., 2021), illustrating the differences between the West Baram Delta sample Be-01 (southern Lambir/Belait-Sarawak) and the Labuan samples. Only the upper Belait samples show a somewhat comparable age distribution, but differ in Precambrian ages.

Figure 18: Multidimensional scaling (MDS) plot of the West Baram and Champion Delta detrital zircon U-Pb age data of this study (created in IsoplotR, Vermeesch, 2018). Sample Be-01 (southern Lambir/Belait-Sarawak) can clearly be separated from the Belait Formation (Champion Delta) data, and suggests it is associated with the West Baram Delta and not part of the Belait Formation. The MDS plot can be used to distinguish between Champion and West Baram deposits. (Miri and Lambir samples are combined into their formations due to the low number of analysed zircons in La-01 and Mi-02).

Figure 19: Heavy mineral assemblages of the studied samples in comparison to the underlying potential source rocks (upper Tatau-Nyalau formations, Belaga Formation), illustrating similar assemblages that indicate recycling into the Neogene delta successions. Most available heavy mineral data from the Rajang Group is based on optical microscopy, only TB56 from Hennig-Breitfeld

et al. (2019) can be used for comparison, as it was analysed with SEM-EDS. Source rock data from Hennig-Breitfeld et al. (2019) and Breitfeld et al. (2020a). (Note: previous studies used SEM-EDS for mineral identification, which cannot distinguish the $TiO_2$ polymorphs. For comparison rutile, anatase, brookite, and $TiO_2$ intergrowth phases identified in this study for the West Baram and Champion delta samples have been summed to $TiO_2$).

Figure 20: Age distribution West Baram and Champion delta systems in NW Borneo, showing similar age cluster, consisting of Cretaceous (grey-yellow), Permian-Triassic (light blue), Paleoproterozoic (green), and Siderian-Neoarchean (light blue) populations. The Champion Delta samples differ slightly in having an additional Upper Cretaceous (red) and Middle-Upper Triassic peak (dark blue), while missing a prominent Siderian-Neoarchean age cluster. Potential source rocks of the Neogene NW Borneo delta systems in stratigraphic order, subdivided into Sabah and Sarawak. The characteristic Upper Cretaceous age peak of the Champion Delta is prominent in the Sabah samples. The Middle to Late Triassic ages that are prominent in the Champion Delta samples, show also higher proportions in Sabah, which indicates that the Champion Delta was sourced by rivers draining Sabah; while the West Baram Delta was sourced by uplifted sedimentary rocks (mainly the Rajang Group) in central and northern Sarawak. Pale shaded areas indicate typical NW Borneo zircon ages. Magenta coloured bar indicates the Upper Cretaceous age peak in Sabah, and blue coloured bar the Middle-Upper Triassic ages found in Sabah. Data from [1]Hennig-Breitfeld et al. (2019), [2]Burley et al. (2021), [3]van Hattum et al. (2013), [4]this study, [5]Nagarajan et al. (2017), [6]Breitfeld et al. (2020a), [7]Galin et al. (2017), [8]Wang et al. (2021b), Zhao et al. (2021) and Zhu et al. (2022), and [9]Zhang et al. (2023).

Figure 21: Paleogeography map at c. 12 Ma showing the Neogene delta provinces in NW Borneo after uplift of the Kuching-Rajang and Crocker ranges (modified from Hall, 2013; Morley & Morley, 2013; Hennig-Breitfeld et al., 2019; Breitfeld et al., 2020a). (S – Schwaner Mountains, BM – Balingian-Mukah Delta, BK – Bintulu-Kemena/Kakus Delta).

## Table Captions

Table 1: Facies table for studied outcrops of the Lambir, Miri, Tukau and southern Lambir (Belait-Sarawak) formations. The facies classes often include gradual variations between similar facies types.

Table 2: Foraminifera assemblage and biostratigraphy of the studied samples of the West Baram Delta deposits. Age based on first appearance Planktonic Foraminiferal zones, Shallow benthic zones and letter stages after BouDagher-Fadel (2018a) and BouDagher-Fadel (2015/2018b) relative to the

biostratigraphical time scale (as defined by Gradstein et al., 2012). Specimen listed in red are reworked Upper Cretaceous.

Table 3: Heavy mineral percentage of translucent species identified with Raman spectroscopy and heavy mineral indices (Hubert, 1962; Morton & Hallsworth, 1994; Morton, 2007).

## Supplementary Captions

Supplementary Table 1: Sample list with coordinates.

Supplementary Table 2: Light mineral modes as counts.

Supplementary Table 3: Heavy mineral count numbers.

Supplementary Table 4: Data table of LA-ICP-MS U-Pb zircon analyses.

Supplementary Table 5: Data table of LA-ICP-MS U-Pb zircon reference analyses.

Supplementary Table 6: Cross-correlation of detrital zircon unmix ages to analyse contributions of potential source areas, using DZmix (Sundell & Saylor, 2017). Literature source is listed and plotted in Fig. 20.

Supplementary Figure 1: Weighted mean age calculations for zircon reference analyses.

Supplementary Figure 2: Detrital zircon U-Pb geochronology individual sample plots.

Supplementary Figure 3: Detrital zircon U-Pb geochronology 0-4000 Ma plots per formation. Histogram bin width 50 Ma, kernel density bandwidth auto.

Supplementary Figure 4: Relative source contributions from Cross-correlation coefficient for Be-01, West Baram Delta and Champion Delta samples plotted with DZmix (Sundell & Saylor, 2017). Plot data can be found in Supplementary Table 6. Literature source data is listed in Fig. 20. Champion Delta data from Hennig-Breitfeld et al. (2019) and Burley et al. (2021).

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