Coupled ‘storm-flood’ depositional model: Application to the Miocene–Modern Baram Delta Province, north-west Borneo

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ABSTRACT

The Miocene to Modern Baram Delta Province is a highly efficient source to sink system that has accumulated 9 to 12 km of coastal–deltaic to shelf sediments over the past 15 Myr. Facies analysis based on ca 1 km of total vertical outcrop stratigraphy, combined with subsurface geology and sedimentary processes in the present-day Baram Delta Province, suggests a ‘storm-flood’ depositional model comprising two distinct periods: (i) fair-weather periods are dominated by alongshore sediment reworking and coastal sand accumulation; and (ii) monsoon-driven storm periods are characterized by increased wave-energy and offshore-directed downwelling storm flow that occur simultaneously with peak fluvial discharge caused by storm precipitation (‘storm-floods’). The modern equivalent environment has the following characteristics: (i) humid-tropical monsoonal climate; (ii) narrow (ca <100 km) and steep (ca 1°), densely vegetated, coastal plain; (iii) deep tropical weathering of a mudstone-dominated hinterland; (iv) multiple independent, small to moderate-sized (10² to 10⁵ km²) drainage basins; (v) predominance of river-mouth bypassing; and (vi) supply-dominated shelf. The ancient, proximal part of this system (the onshore Belait Formation) is dominated by strongly cyclical sandier-upward successions (metre to decametre-scale) comprising (from bottom to top): (i) finely laminated mudstone with millimetre-scale silty laminae; (ii) heterolithic sandstone–mudstone alternations (centimetre to metre-scale); and (iii) sharp-based, swaley cross-stratified sandstone beds and bedsets (metre to decimetre-scale). Gutter casts (decimetre to metre-scale) are widespread, they are filled with swaley cross-stratified sandstone and their long axes are oriented perpendicular to the palaeoshoreline. The gutter casts and other associated waning-flow event beds suggest that erosion and deposition was controlled by high-energy, offshore-directed, oscillatory-dominated, sediment-laden combined flows within a shoreface to delta front setting. The presence of multiple river mouths and exceptionally high rates of accommodation creation (characteristic of the Neogene to Recent Baram Delta Province; up to 3000 m Ma⁻¹), in a ‘storm-flood’-dominated environment, resulted in a highly efficient and effective offshore-directed sediment transport system.

Keywords Baram Delta Province, combined flow, delta front, gutter cast, humid-tropical, Miocene, shoreface, storm-flood.
INTRODUCTION

The transfer of sand offshore from land to shelf requires flows to overcome the ‘littoral energy fence’ (Allen, 1970), the mean landward-directed bottom stress induced by shoaling waves. Storm-driven mechanisms for sand transfer across this zone are intrinsic to facies models of ancient shallow-marine, wave-dominated and storm-dominated sedimentation (Walker, 1984; Walker & Plint, 1992; Johnson & Baldwin, 1996; Hampson & Storms, 2003; Clifton, 2006; Suter, 2006). In these models, sharp-based sandstone ‘event beds’ with hummocky cross-stratification (HCS) and/or swaley cross-stratification (SCS), sometimes associated with gutter casts (e.g. Myrow, 1992; Myrow & Southard, 1996), are synonymous with deposition by storm-generated combined flows. These result from superimposed oscillatory and unidirectional currents, with the latter commonly interpreted to reflect downwelling ‘relaxation currents’ (downwelling storm flows) related to storm coastal set-up (e.g. Duke, 1990). The HCS/SCS is associated with storm-generated oscillatory ripples in both lacustrine (Greenwood & Sherman, 1986) and shallow-marine settings (Yang et al., 2005, 2006). Synthetic HCS/SCS has also been generated from bed geometries formed under oscillatory-dominated combined flow (Dumas et al., 2005; Dumas & Arnott, 2006). However, other mechanisms have been proposed to form HCS/SCS in a spectrum of shallow to deep-marine environments, including: (i) wave-enhanced sediment gravity flows (Myrow et al., 2002; Lamb et al., 2008; Basilici et al., 2012), which are variably linked to hyperpycnal flows (Ogston et al., 2000; Puig et al., 2004; Traykovski et al., 2007); (ii) short-lived instability events in high-energy flows (Quin, 2011); and (iii) internal gravity waves propagating along the pycnocline (Morsilli & Pomar, 2012).

There is also uncertainty regarding the processes responsible for the shoreline to basinward transport of sand, particularly the interaction of climate, tectonics, fluvial hydrodynamics and shallow-marine processes (e.g. Levell, 1980; Walker & Bergman, 1993; Bhattacharya & MacEachern, 2009; Hampson, 2010), which are key components of source to sink systems (Allen, 2008). This partly reflects the difficulty of developing fully integrated studies of both modern systems and the geological record. However, it also reflects a limited number of suitable study areas where the present-day climatic, tectonic, fluvial and shoreline–shelf systems can be directly compared to equivalent ancient depositional systems.

In source to sink systems with relatively small drainage basins (10^2 to 10^3 km²), short river lengths (for example, ca 100 to 300 km) and subject to large (hundreds to thousands of kilometres diameter), high-intensity tropical storms (i.e. tropical to subtropical settings), ‘storms’ routinely affect both drainage basins and coastal regions simultaneously (Liu et al., 2008; Bhattacharya & MacEachern, 2009). Herein, the term ‘storm-flood’ (Bhattacharya, 2005; cf. ‘oceanic flood’ – Wheatcroft, 2000) is used to refer to the simultaneous occurrence of: (i) river floods caused by storm precipitation in the hinterland; and (ii) elevated wave height in the coastal–shelf region caused by strong winds and related meteorological conditions. This is discussed more fully in the interpretation sections of this paper.

The present-day Baram Delta Province (BDP) source to sink system (Fig. 1) has operated since the Middle Miocene (e.g. Hoggmascall et al., 2012). The BDP comprises a prolific petroleum province with ca 115 years of exploration and production history (e.g. Tan et al., 1999). Consequently, the Mi–Pliocene palaeogeography, tectono-stratigraphy and the gross biostratigraphic framework (Fig. 2) are extremely well-constrained, especially within the subsurface (e.g. Sandal, 1996; Saller & Blake, 2003; Torres et al., 2011; Hoggmascall et al., 2012).

This article presents a facies analysis of Middle Miocene (15.5 to 11.7 Ma, Serravallian–Tortonian) outcrops from the BDP and compares this to the subsurface and modern aspects of the BDP depositional system (Fig. 2). The objectives of this work are: (i) to document the facies character and stratigraphic architecture of ancient shallow-marine successions in the BDP, including the morphology, distribution and sedimentology of erosional gutter casts and event beds; (ii) to re-evaluate the mechanics of sediment supply and processes controlling ancient and modern, shallow-marine deposition; and (iii) to evaluate the factors conducive to wave and river coupling during storms and the subsequent impact on shoreline geometry and sediment distribution.

By providing a well-defined geological context to the formation and preservation of event beds and their stratigraphic successions, this study provides a comparative framework for the evaluation of other ancient shallow-marine and deltaic systems, which encompass a range of different...
geological settings. The geological setting and depositional model presented here may be close to a wave-dominated, fluvial-influenced end-member in the process-response spectrum of ancient shallow-marine and deltaic systems (e.g. Ainsworth et al., 2011). The rocks under evaluation also constitute the largest of all the hydrocarbon reservoirs within the subsurface of the BDP; this study provides new insight into their reservoir characteristics.

![Fig. 1](image1.png)

**Fig. 1.** (A) Location of the study area in north-west Borneo (box) on the southern margin of the South China Sea, Southeast Asia. (B) Simplified hydrological, topographic and bathymetric map of north-west Borneo and the Baram Delta Province (BDP). Coastal drainage systems have previously been grouped into two large catchment regions: the ‘Baram’ (western BDP) and Brunei Bay (‘Champion’, eastern BDP) drainage systems (e.g. Sandal, 1996). (C) Simplified north-west/south-east topographic cross-section across the Baram drainage basin (X–X' in B).

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![Fig. 2](image2.png)

**Fig. 2.** Geological setting for the Neogene Baram Delta Province (BDP). (A) Schematic map showing the tectonically active hinterland (unshaded), area of >5 km thick Neogene sedimentary rock (yellow shading) (Hamilton, 1974; Hall & Nichols, 2002), West Baram Line and approximate positions of seismically mapped shelf edges (coloured lines with ages in Ma) (Cullen, 2010). Outcrop data used in this study were collected primarily in eastern Brunei (pink box), supplemented by additional data from the Seria Field and outcrops in Labuan, Miri and Lambir. (B) Schematic north-west/south-east geo-seismic section, restored for Mi–Pliocene tectonic inversion, in the north-eastern BDP (Y–Y' in A), illustrating: (i) the dominant fault types; (ii) stratigraphic architecture styles; (iii) major structural folds; and (iv) approximate chronostratigraphic surfaces (coloured lines with numbers in Ma) (Gartrell et al., 2011; Torres et al., 2011). (C) The outcrop study area in eastern Brunei (after Morley et al., 1998) occurs in two main areas: (i) on the inverted limb of the syndepositional Jerudong counter a regional growth (antithetic) fault; and (ii) on the eastern limb of the associated Berakas Syncline (e.g. Morley et al., 2003). Outcrop successions correspond to strongly aggradational, late Middle Miocene stratigraphic successions (pink box in B).
GEOLOGICAL SETTING

Modern north-west Borneo coastal system

The present-day north-west Borneo coastline is generally linear (Fig. 1B), microtidal (average tidal range of 1-7 m) and has a low wave energy (average wave height of 0.7 m) (Sandal, 1996). The Baram delta has a cuspatate symmetrical geometry (Fig. 2A) and is interpreted to be wave-dominated (e.g. Sandal, 1996) with tidal-influence in the distributary mouth (Lambiase et al., 2002). The relative power of river and tidal processes is significantly increased within embayed and wave-protected coastal areas, such as Brunei Bay (Fig. 1B; e.g. Ainsworth et al., 2011).

A narrow (<80 km) and relatively steeply dipping (ca 1°) coastal plain is located between the shoreline and the north-east/south-west trending Rajang-Crocker Range hinterland (average 1800 m elevation; up to 4100 m). The coastal plain is partitioned into numerous small to moderate-sized (10^2 to 10^5 km^2) drainage basins (<300 km) north-west flowing rivers (Fig. 1B and C). Rivers invariably occur within incised valleys because of progressive and long-lived uplift (Sandal, 1996).

River runoff is dominated by suspended sediment (clay and silt; Milliman & Meade, 1983) which reflects deep tropical weathering of a mudstone-rich hinterland (Hutchison, 2005). Sand-grade bedload transport occurs during intense, high discharge events (>100 mm h\(^{-1}\) rainfall) linked to monsoonal storms (Dykes, 2000). The high suspended-sediment loads, presently enhanced by logging, form mud plumes at river mouths during heavy seasonal rainstorms.

The north-west Borneo shelf is generally 75 to 100 km wide, with a gradient of ca 0.1° (Fig. 1B and C; Hiscott, 2001). The Baram delta mouth bar slopes offshore at ca 0.8° (Hiscott, 2001) to ca 30 m water depth (Fig. 1C). Beyond the shelf edge, the continental slope steepens irregularly, due to toe-thrusts (King et al., 2010), into the ca 4 km deep North-West Borneo Trough.

Baram Delta Province

The Middle Miocene to Modern BDP in the south-west Baram-Balabac Basin (Cullen, 2010), is bounded by four tectonic features (Figs 1B and 2A): (i) the West Baram Line, a major, long-lived and basement-linked tectonic discontinuity (Cullen, 2014), to the south-west; (ii) the north-eastern margin comprises a series of north–south trending faulted anticlines (Sabah Ridges; Bol & Van Hoorn, 1980) (e.g. Fig. 2C); (iii) the North-West Borneo Trough (Fig. 1B), a possible deep foreland basin or remnant Miocene trench (e.g. Hutchison, 2010), to the north-west; and (iv) the south-eastern margin is defined by the Rajang-Crocker Range (Fig. 2A), which is a fold-thrust belt comprising deep-water Cretaceous–Palaeogene sediments. The latter was initiated by continental collision during the Eocene to Early Oligocene Sarawak Orogeny (Cullen, 2010), but has continued to be uplifted and eroded, providing sediment to the BDP throughout the Neogene to the present-day (Levell, 1987; Sandal, 1996; King et al., 2010; Gartrell et al., 2011; Hall, 2013). The erosional products of the actively uplifting hinterland compose the 9 to 12 km thick, progradational to strongly aggradational, Neogene basin-fill of the BDP (Fig. 2B; Hamilton, 1974; Bol & Van Hoorn, 1980; James, 1984; Sandal, 1996; Hall & Nichols, 2002). High subsidence rates in the BDP (up to 3000 m Ma\(^{-1}\); Sandal, 1996) reflect: (i) delta-related loading and deformation of the undercompacted, mud-dominated, prodelta substratum, including synthetic and antithetic growth faults (Fig. 2B) and mud diapirs; and (ii) continued movement along faults linked to the underlying fold-thrust belt, including the West Baram Line (Fig. 2A) and the mud-cored Jerudong Anticline (Fig. 2C; e.g. Morley et al., 2003; Gartrell et al., 2011).

Limited (ca 50 km) basinward shelf-edge progradation (Fig. 2A), coupled with inner shelf inversion and hinterland uplift (e.g. Torres et al., 2011; Hall, 2013), and an exceptionally high deltaic aggradation rate (Fig. 2B), indicate that coastal plain and continental shelf widths have remained narrow (<100 km) throughout the mid-Miocene to present-day (Fig. 2A). Reconstructions of Late Miocene palaeo-drainage systems in the BDP interpret multiple river systems with separate drainage basins, partly by analogy to present-day rivers (Lambiase & Cullen, 2013).

This study focuses on the early phase of basin-fill, with all of the outcrops being Middle Miocene age (15-5 to 11-7 Ma, Serravallian; Torres et al., 2011). This period is approximately coeval with the ca north-east/south-west oriented 12-2 Ma shelf-edge (Fig. 2A and B), which has been mapped from biostratigraphically
calibrated seismic data (Saller & Blake, 2003; Cullen, 2010; Gartrell et al., 2011; Torres et al., 2011).

DATA SET AND METHODOLOGY

Approximately 1 km of outcrop stratigraphy, comprising the Belait Formation and its lateral equivalents (Miri and Lambir formations), have been studied in eastern Brunei and Labuan (western Sabah) and Miri and Lambir (northern Sarawak) in north-west Borneo (Fig. 2A). Although outcrop exposures are relatively small (<100 m²), they are extremely high quality with moderate weathering and negligible vegetation cover (created during civil engineering projects). Outcrops were GPS-located and studied using integrated data sets comprising logged sections (1:50 scale), detailed photomontages, field sketches and mapping (1:50 scale). Detailed measurements of gutter casts include the long-axis orientation and dimensions (thickness and width).

FACIES ANALYSIS

Nine facies types (F1 to F9) have been identified (Table 1) and grouped into three facies associations (FA; see below). Three representative vertical stratigraphic successions from the Belait Formation, covering ca 200 m of stratigraphy (from ca 964 m of measured section), are illustrated in Figs 3 and 4.

Facies Association 1: Mudstone-dominated

Description

Facies Association 1 comprises mudstone-dominated facies (>80%) and minor (<20%), centimetre-scale, siltstone to very fine-grained sandstone layers (Fig. 5). Bioturbation in the mudstones varies from BI 0 to 2 (F3) to BI 5 to 6 (F2). Laminated mudstones (F3) contain centimetre-scale (typically <3 cm) sets of sub-millimetre to millimetre-scale, subtly lenticular, parallel laminae (or ‘streaks’) of quartz silt (cf. microfacies 2 of Plint, 2014) with a sporadic and impoverished ichnofauna dominated by millimetre-scale *Palaeophycus* and *Planolites* (Fig. 5A). F3 commonly sharply overlie layers (<3 cm thick) of muddy siltstone (F4) with symmetrical, lenticular, ripple laminae-sets of fine to coarse-grained quartz silt (cf. microfacies 1 of Plint, 2014). These layers are typically <5 cm thick, have sharp scoured bases, occasional basal lags of millimetre-scale mudstone clasts (Fig. 5B), and may show normal grading. Massive silty mudstones (F2) contain a relatively diverse ichnofauna, dominated by small-scale *Chondrites*, *Palaeophycus*, *Phycosiphon*, *Planolites* and *Thalassinoides*, which corresponds to a distal *Cruziana* ichnofacies (MacEachern & Bann, 2008). Plant debris is common throughout all of the mudstone facies.

Very fine-grained sandstone beds are typically 5 to 10 cm thick with sharp bases and tops, centimetre-scale gutter casts, wavy to low-angle planar lamination (F9) and SCS (F7), and, nearer bed tops, combined-flow and oscillation ripples (F6) (Fig. 5). Millimetre to centimetre-scale mud clasts and bioclastic material (disaggregated bivalves, bryozoans and shark teeth) and centimetre to metre-scale gutter casts may occur at bed bases. Beds may also show subtle normal grading, an upward increase in detrital clay matrix and lateral pinch out (on a metre-scale). Bioturbation is generally low (BI 0 to 1) but can be sporadically high (BI 2 to 3) and typically increases near bed tops (BI 1 to 2), and is dominated by small *Ophiomorpha* and *Palaeophycus* with minor *Asterosoma*. Facies Association 1 forms 8 to 20 m thick packages that sharply or gradationally overlie FA-2 and FA-3 and typically pass gradationally upward into FA-2 (Figs 3 and 4).

Interpretation

Facies Association 1 mudstone-dominated sediment is commonly interpreted to represent relatively quiet-water deposition by suspension settling, suggesting that deposition was dominated by fair-weather conditions (e.g. Howard & Reineck, 1981). However, mudstone deposition can also occur under moderate to high-energy conditions (e.g. Macquaker & Keller, 2005; Bhattacharya & MacEachern, 2009; Ghadeer & Macquaker, 2011; Plint et al., 2012; Plint, 2014). This is supported by studies of Recent muddy depositional environments that show mud deposition during periods of high wave and tidal energy (e.g. Nittrouer & Sternberg, 1981; Rine & Ginsburg, 1985). Laboratory experiments also demonstrate that mud, transported as bedload aggregates, can form migrating ripple bedforms at velocities capable of transporting sand (e.g. Schieber et al., 2007; Schieber & Southard, 2009). In Facies Association 1, the association of silty mudstone with siltstone and
Table 1. Summary of the main facies recognized in the Belait Formation in this study. Ichnogenera recognized include Asterosoma (As), Chondrites (Ch), Cylindrichnus (Cy), Ophiomorpha (Op), Palaeophycus tubularis (Pt), Phycosiphon (Ph), Planolites (P), Rhizocorallium (Rz), Rosselia (R), Skolithos (S), Teichichnus (T), ‘Terebellina’ (sensu lato) (Te) and Thalassinoides (Th) (for example, MacEachern & Bann, 2008). Grain-size abbreviations are: mst = mudstone; slst = siltstone; sst = sandstone; vf = very fine-grained; vfL = lower very fine-grained; vfU = upper very fine-grained; f = fine-grained.

| Facies | Grain size | Sand: Shale | Sandstone dimensions | Mudstone dimensions | Description | Biogenic structures | Process interpretation | Environment interpretation | Facies sketch |
|--------|------------|-------------|---------------------|---------------------|-------------|-------------------|----------------------|------------------------|----------------|
| F1     | mst matrix, sst boulders | 0.2–0.3 (unit) | Up to 1.3 m thick and 3.7 m wide boulders | 3–4 m thick, >300 m wide | Sub-rounded, internally deformed and fractured sst boulders (F7–F9), matrix (F2) supported | N/A | Rapid deposition, high energy lamellar flow | Delta front/shoreface failure, tectonic and/or excess sediment supply trigger |
| F2     | mst, minor slst and vfl. sst | <0.1 typically | N/A | 5–18 m thick, >300 m wide | Apparently massive and variably bioturbated, bioturbated F4 and F7–F9 interbeds, common carbonaceous debris | B10–5 | Suspension settling and/or high energy, sediment-rich transitional to lamellar flow deposition, variable physico-chemical stress | Storm-dominated prodelta to distal delta front (mud-rich)/ distal flanking shoreface to offshore–shelf |
| F3     | mst, minor slst and vfl. sst | <0.1 typically | N/A | 0.1–3 m thick, m-scale wide | Wavy parallel silt laminae/streaks and starved oscillation silt ripple laminae, carbonaceous material | B10–2 | Variable energy, distal storm-flow—low energy, suspension setting, high energy, wave-enhanced, sediment gravity flow and/or fluid mud | Storm-dominated prodelta to distal delta front (mud-rich)/ distal flanking shoreface to offshore–shelf |
| F4     | Silt, slst | 0.1–0.3 | mm to 1 cm thick layers, m-scale lateral discontinuity | 1–3 cm thick beds, m-scale lateral discontinuity | Wavy-parallel lamination, oscillation ripples, muddy laminae, normal grading into F3 or sharply overlain by F2 | B1–2, sporadic B14–5 | Oscillatory-dominated combined flow, low energy/mud rich (cf. F7–F9), waning flow, wave-enhanced sediment gravity flow | Storm-dominated distal delta front to proximal prodelta/distal flanking shoreface to proximal offshore–shelf |
| F5     | Silt, slst and vfl. sst | 0.3–0.5 | 0.1–10 cm thick, c. 10–100 m wide | 0.1–5 cm thick beds, 2–5 m thick & c. 1–10s m wide packages | Wavy-lenticular bedded, wavy to parallel laminated, oscillation to combined flow ripples, cm–dm-scale scours, minor mud drapes | B11–2, sporadic B14–5 | Fluctuating energy and/or sediment supply, erosional oscillatory-dominated combined flow, possible fluid mud drapes | Fair-weather-dominated, storm-influenced, distal delta front /distal flanking shoreface |

**Footnotes:**
- As = Asterosoma
- Ch = Chondrites
- Cy = Cylindrichnus
- Op = Ophiomorpha
- Pt = Palaeophycus tubularis
- Ph = Phycosiphon
- P = Planolites
- Rz = Rhizocorallium
- R = Rosselia
- S = Skolithos
- Te = Teichichnus
Table 1. Continued

| Facies                          | Grain size | Sand: Shale | Sandstone dimensions | Mudstone dimensions | Description                                                                 | Biogenic structures | Process interpretation | Environment interpretation | Facies sketch |
|--------------------------------|------------|-------------|----------------------|--------------------|------------------------------------------------------------------------------|---------------------|------------------------|---------------------------|-------------------|
| F6 Rippled sandstone           | vfu sst    | >0.8        | 0.1–0.3 m thick, 1 to >12 m wide | mm–cm thick, cm–dm wide | Combined flow ripples, minor wave and current ripples, common low angle (<10°) climb, carbonaceous and mud drapes | BI 0–3 | Wave to oscillatory-dominated combined flow, possible fluid/suspended mud deposition | Fainweather to storm-dominated, delta front/flanking shoreface | F2–F4 |
| F7 Swaley cross-stratified sandstone | vst        | >0.8 typically >0.9 | 0.1–1 m thick beds, 1.6–3.6 m thick and >300 m wide (ca km-scale) | F2–F5 interbeds <0.3–2.8 m thick and >300 m wide; mm thick, cm–dm wide | Low-angle (<20°), large-wavelength (>10 m) SCS, swaley discontinuities; common SCS downlap and asymmetry, sharp erosional bases, cm–m-scale gutter casts and sharp tops | BI 0–1, BI 1–2 near bed tops | Oscillatory-dominated combined flow, high sedimentation rates, frequent erosion | Storm-dominated (reworked) distal to proximal delta front/distal to proximal flanking shoreface | FA-1/FA-2 |
| F8 Hummocky cross-stratified sandstone | vst        | >0.8 typically >0.9 | 0.1–1 m thick beds, 1.3–6 m thick and >300 m wide (ca km-scale) | F2–F5 interbeds <0.3–2.8 m thick, >300 m wide; mm thick, cm–dm wide | As F7 but includes dm-scale convex-up laminae sets (hummocks) | BI 0–1, BI 1–2 near bed tops | Oscillatory-dominated combined flow, increased aggradation vs. erosion (cf. F7) | Storm-dominated (reworked) distal to proximal delta front/distal to proximal flanking shoreface | FA-1/FA-2 |
| F9 Laminated sandstone         | vst        | >0.9        | 0.1–0.75 m beds, 1–2.5 m thick and >300 m wide beds | mm thick, dm–m wide laminae | Low angle (<10°, typically <5°) lamination, subtle internal discontinuity surfaces, common downlap, minor mudstone and carbonaceous drapes and clasts | BI 0–1 | High energy, oscillatory-dominated combined flow | Storm-dominated, open-coast flanking shoreface or delta front | F7–F8 |

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sandstone layers with sharp scoured bases, normal grading, and wave and/or combined-flow ripples, is consistent with deposition by relatively energetic, oscillatory-dominated combined flows, probably related to storms. Plint (2014) interpreted a similar mudstone–siltstone association to reflect deposition by wave-enhanced sediment gravity flows in a wave-reworked and storm-reworked prodelta. The sporadic, impoverished bioturbation in laminated mudstones is also consistent with fluctuating sedimentation rates, typical of a deltaic setting, and may be related to dense, turbid, sediment gravity flows, and possibly fluid mud flows (e.g. Bhattacharya & MacEachern, 2009; Plint, 2014). Sediment gravity and fluid mud flows may also be linked to hyperpycnal river-flood discharge. Mudstone deposition may also have occurred by flocculation and fallout in flood-driven hyperpycnal flows. The latter is likely to have occurred both during the causal storm and during the immediately following fair-weather period. The abundance of plant debris also suggests proximity to river mouths. Facies Association 1 could therefore reflect deposition under a range of depositional energies, but is most likely to represent energetic conditions, perhaps in a storm-dominated prodelta or delta-influenced offshore/shelf (for example, along strike from an active river mouth).

Facies Association 2: Heterolthic

Description

Facies Association 2 sharply or gradationally overlies FA-1 and comprises 3 to 20 m thick
packages of interbedded mudstone-dominated facies (F2 to F5; 20% to 80%) and sandstone facies (Figs 3 and 4). Sandstones are typically very fine-grained and dominated by SCS (F7), and occur as non-amalgamated beds 0.25 to 1.15 m thick (average 0.37 m) and amalgamated bedsets up to 1.65 m thick. The bases of sandstone beds are sharp, undulatory and erosional, with abundant gutter casts and common mudstone clast and bioclastic lags (Fig. 6A and B); bed tops are also typically sharp (for example, Fig. 6A). Swaley cross-stratification typically consists of 0.2 to 0.6 m thick lamina-sets of exquisitely preserved, low-angle (<15°, typically <10°), millimetre-scale to sub-millimetre-scale, concave-upward laminae, which are separated by predominantly concave-upward, erosional bounding/discontinuity surfaces (Fig. 6A to D). Lamina-sets with convex-up laminae and bounding surfaces (HCS) are occasionally present (F8). Laminae are defined by subtle changes in very fine-grained sand size and, nearer the top of lamina-sets and beds, lamina-scale alternations of detrital clay matrix and carbonaceous detritus (Fig. 6E). The approximate half wave-lengths of SCS laminae sets can be up to 8 m. Lamina-sets are commonly asymmetrical and downlap the lower bounding surface (Fig. 6B). Near the top of sandstone beds, grain size decreases (to lower very fine-grained sand), the thickness of lamina-sets decreases (to <20 cm) and complex transitional relationships are observed between centimetre-scale wave ripples, combined-flow ripples and climbing current.

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ripples (F6) (Fig. 6C to F). Bioturbation in sandstone facies is generally low (BI 0 to 1), but both intensity and diversity increase near bed tops (BI 2 to 3). Bioturbation is dominated by *Ophiomorpha* and associated minor ichnotaxa include *Asterosoma*, *Lockeia*, *Palaeophycus*, *Planolites* and *Rosselia* (Fig. 6E); this constitutes an impoverished *Skolithos* ichnofacies (Buatois et al., 2008; MacEachern & Bann, 2008). Escape structures and collapsed burrows are also observed.

Muddy heterolithics (F5) form 5 to 20 cm thick packages between sandstone beds/bedsets and are similar to FA-1: they comprise a centimetre to decimetre-scale interbedding of predominantly mudstone facies (F2 to F3) and subordinate (20% to 40%), typically sharp-based, sandstone (F6 to F9) and siltstone (F4) layers. Bioturbation in F5 is sporadic and variable (BI 1 to 5) but relatively diverse. Intensely bioturbated areas resemble ‘mantle-and-swirl’ structures (Fig. 5; e.g. Lobza & Schieber, 1999; Bhattacharya & MacEachern, 2009). The main ichnifauna are *Chondrites*, *Palaeophycus*, *Planolites* and *Thalassinoides*. Less abundant ichnifauna include *Asterosoma*, *Cosmorhaphe*, *Cylindrichnus*, *Phycosiphon*, *Rosselia*, *Schaubcylindrichnus* and *Terebellina* (Fig. 5A). Together this constitutes a slightly impoverished *Cruziana* ichnofacies (MacEachern & Bann, 2008).

**Interpretation**

Facies Association 2 deposition was governed by storm waves and waning, storm-related flows, based on the following: abundance of SCS/HCS and oscillation/combined-flow ripples; erosional bases and gutter casts; mudstone clasts and bioclastic lags; upward decrease in lamina-set thickness and grain size; and upward increase in detrital clay content. Swaley cross-stratification and hummocky cross-stratification reflect aggradation and reworking of bedforms formed by large storm waves under oscillatory flow or oscillatory-dominated combined flow (e.g. Dott & Bourgeois, 1982; Harms et al., 1982; Yang et al., 2005, 2006; Dumas & Arnott, 2006). The common asymmetry and downlapping of the SCS, together with the occurrence of current ripples, supports deposition by combined oscillatory and unidirectional flows. The unidirectional component could reflect downwelling currents related to storm coastal set-up, geostrophic flow and/or downslope gravity flows, such as hyperpycnal flow (e.g. Lamb et al., 2008). The exquisite laminae preservation, low bioturbation, aggradational SCS/HCS, climbing to aggrading current and wave ripples, and biogenic escape structures, are indicative of high sedimentation rates. The abundance of internal erosional surfaces bounding lamina-sets and complex transitional relationships between SCS/HCS and ripple lamination, provide evidence for dynamic waxing and waning of flow. This could reflect either fluctuations in storm energy or variations in an additional sediment gravity flow component. The similarity between muddy heterolithic facies in FA-2 and FA-1 also suggests the influence of storm-wave generated and...
Fig. 6. (A) A characteristically sharp-based, erosional sandstone bed within Facies Association 2, displaying a gutter cast and swaley cross-stratification (SCS). (B) Interpreted and labelled line drawing of photograph shown in (A). Note the asymmetrical (anisotropic), low-angle SCS and localized mudstone intraclasts and bioturbation. See Fig. 4 for legend. (C) Photograph highlighting the exquisitely preserved fabric of SCS near the top of a sandstone bed in FA-2. (D) Line drawing of photograph in (C). The SCS fabric comprises multiple, erosionally amalgamated, lamina-sets; most show internal onlap onto the margin of their basal discontinuity, however, some show steeper downlap (purple arrow). (E) Sedimentary structures characteristic of the top of sandstone beds in FA-2. Ichnogenera (abbreviations listed in Fig. 5) also include Lockeia (Lo). Fugichinia (fu), indicating escape behaviour, also occur. (F) Line drawing of photograph in (E). Low-angle SCS containing carbonaceous drapes/concentrates changes gradually upward into an asymmetrical ripple set dipping northward. From left to right, ripple sets change from more current-like to wave-like, and start to aggrade and climb.
enhanced gravity flows; the sharp tops of sandstone beds may reflect collapse of a fluid mud layer (Plint, 2014).

Similar heterolithic facies associations, with alternating SCS/HCS sandstone and mudstone facies, are commonly attributed to deposition between fair-weather and ‘normal’ storm-wave base for sand (e.g. Plint, 2014). This environment is variously termed the ‘distal lower shoreface’ (Hampson & Storms, 2003), ‘lower shoreface’ (van Wagoner et al., 1990; Kamola & Van Wagoner, 1995), ‘middle-lower shoreface to offshore transition’ (Galloway & Hobday, 1996) and ‘offshore transition’ (e.g. Reading & Collinson, 1996). However, this interpretation masks significant variability resulting from different wave-energy and storm-energy regimes (Clifton, 2006), combined-flow processes (e.g. Lamb et al., 2008) and sediment supply (Plint et al., 2012). The increased bioturbation intensity and ichnofauna diversity near the top of sandstone beds reflects post-storm colonization (Pemberton et al., 2001; Buatois et al., 2008). However, the sporadic and slightly impoverished Cruzianna ichnofacies within F5 is atypical of fully marine shoreface to offshore successions (MacEachern & Bann, 2008). Instead, this reflects high but fluctuating sedimentation rates and/or physico-chemical stress, such as reduced and/or fluctuating salinity (Bann et al., 2008; MacEachern & Bann, 2008). Mantle-and-swirl structures suggest episodically dense, fluid mud, soupground conditions (Lobza & Schieber, 1999; Bhattacharya & MacEachern, 2009). Abundant carbonaceous detritus and coalified clasts also suggest proximity to river mouths. Overall, FA-2 was deposited in a storm-dominated, fluvial-influenced environment.

**Facies Association 3: Sandstone-dominated**

**Description**

Facies Association 3 typically sharply or gradationally overlies FA-2 (Figs 3 and 4), but occasionally sharply overlies FA-1. Facies Association 3 forms 2.5 to 18.0 m thick packages dominated by (>80%) amalgamated very fine-grained, swaley cross-stratified sandstone (F7; Figs 3 and 4). Many of the characteristics of F7 in FA-3 are shared with the sandstone beds in FA-2, particularly sharp scoured bases, gutter casts, normal grading, carbonaceous debris, bioclastic and mud clast lags, and the bioturbation. The main differences are the thicker beds (0.2 to 1.0 m) and bedsets (1.7 to 3.6 m), thicker lamina-sets and SCS with half wavelengths >10 m (Fig. 7A). The large-scale lamination may also display gentler concave-upward relief (F9), thus resembling ‘quasi-planar lamination’ (Arnott, 1993).

There is a hierarchy of discontinuity surfaces within the SCS (Fig. 7A). First-order discontinuities extend >10 m laterally and bound decimetre to metre-scale cross-sets and cosets (Fig. 7A). Second-order discontinuities are truncated by overlying first-order discontinuities, but may extend >10 m laterally and truncate underlying first-order, other second-order and all third-order discontinuities (Fig. 7A). Third-order discontinuities extend <10 m laterally and bound decimetre to metre-scale lamina-sets (Fig. 7A). Second-order and third-order discontinuities may become conformable with cross-laminae laterally (for example, blue arrows – Fig. 7A) and show subtle (centimetre-scale) shifts between sets (for example, green arrows – Fig. 7A). The transition from concave-up to convex-up lamination is invariably truncated by an overlying discontinuity surface, which enhances SCS preservation. Mud clasts and carbonaceous clasts (centimetre to millimetre-sized) frequently line basal bed boundaries and discontinuity surfaces (for example, yellow arrows – Fig. 7A).

Rippled sandstone (F6) comprises 0.1 to 0.2 m thick bedsets of 1 to 4 cm thick, very fine-grained sandstone beds, displaying oscillation and combined-flow ripple cross-lamination (Fig. 7B). Sandstone beds are commonly draped by 1 to 3 mm thick, mudstone layers that appear structureless or have subtle silty laminae (similar to F3) (Fig. 7C). Facies F6 is transitional with underlying F7: there is an upward decrease in the scale of SCS and increase in abundance of mud-draped ripples (Fig. 7B). Bioturbation in F6 is generally low (BI 0 to 1) but locally (centimetre-scale) intense (BI 4 to 5). Dominant ichnotaxa are Ophiomorpha, Planolites, Palaeophycus and Skolithos. Muddy heterolithic (F5) packages are typically <0.3 m thick but may reach ca 0.8 m in thicker FA-3 packages (Fig. 4); they show very similar characteristics to F5 in FA-2.

**Interpretation**

The increased thickness of F7 beds and bedsets in FA-3 probably reflects an increased magnitude and rate of storm reworking (Thorne et al., 1991). This is commonly interpreted to reflect a decrease in water depth. However, it may also represent an increase in mean storm strength or decrease in sediment supply, with increased...
reworking and progressive sorting (Swift & Thorne, 1991; Thorne et al., 1991; Hampson & Storms, 2003; Sømme et al., 2008; Charvin et al., 2011). The decametre-scale wavelength and low-angle SCS and quasi-planar lamination suggest higher energy combined flows that were capable of causing bedform degradation (Arnott & Southard, 1990; Arnott, 1993). The hierarchy of discontinuity surfaces and their complex lateral conformity and offset relationships, attests to highly variable spatial and temporal hydrodynamic conditions: there was repeated, laterally variable erosion and deposition both during and between successive waning-flow events.

Interbedded rippled sandstones (F6) and muddy heterolithics (F5) suggest periods of deposition under lower wave power. However, mud drapes/layer may represent fluid mud deposition under very high, near-bed suspended-sediment concentrations that developed as a result of storm-wave resuspension during coastal storms (e.g. Ichaso & Dalrymple, 2009; Baas et al., 2016). Furthermore, in tropical deltaic settings (such as the BDP), fluid muds may also develop near suspension-dominated river...
mouths, especially after hinterland storms. Fluid mud development would be optimized by simultaneous coastal and hinterland storms (Wheatcroft, 2000; Bhattacharya, 2005) and has the following consequences: (i) storm-associated precipitation increases contemporaneous river discharge, flushing large volumes of suspended sediment and phytodetritus to the delta front (Bann et al., 2008); and (ii) wave height and energy are dramatically elevated at the shoreline by strong monsoonal winds. Rapidly emplaced mud drapes may shield underlying sandstones from faunal colonization (Coates & MacEachern, 2007), which is consistent with the generally low but sporadically high bioturbation in F5 and F6.

GUTTER CAST ANALYSIS

Gutter casts are ‘downward-bulging erosional structures’ that may be isolated from, or connected (amalgamated) to, the base of sandstone beds (Whitaker, 1973; Myrow, 1992). The present study includes ‘pot casts’ (Myrow, 1992) within the range of gutter cast geometries. Analysis of gutter casts is presented separately because of: (i) their abundance throughout the three facies associations; and (ii) their significance for the overall depositional model.

Description

Distribution and architectural relationships

Gutter casts are an extremely common feature along the base of sandstone beds, especially in FA-2. In one case (outcrop JT-688; location 3 – Fig. 2C), at least 30% of sandstone beds (in FA-2) have basal gutter casts (Fig. 8).

Cross-sectional geometries and infill

Both amalgamated and isolated gutter casts exhibit a wide range of cross-sectional geometries (Figs 8 to 10): (i) symmetrical to asymmetrical; (ii) irregular (for example, Figs 9A, 10A and 10D); (iii) shallow to deeply rounded (Fig. 9B to F); (iv) rectangular (Figs 9F and 10E); (v) flat-based (Figs 9E, 9F and 10D); and (vi) complex bilobate (Figs 9A and 10D). Steep to overhanging walls are common (for example, Figs 9A, 9B, 10A and 10E). In all of the examples studied, the basal surface geometries are entirely erosional and are not associated with loading, dewatering or other soft-sediment deformation.

Gutter casts are mostly filled with very fine-grained sandstone with SCS (F7), commonly with centimetre-scale mudstone clasts (for example, Fig. 9B) and bioclasts. Swaley cross-stratification is typically asymmetrical with downlap onto one or both margins of the gutter (for example, Fig. 9A to C) and commonly apparently concordant with SCS in connected sandstone beds above (for example, Fig. 9B and D). Some examples have compound geometries and infill (Figs 9A and 10A).

Heterolithic gutter cast infills contain centimetre-scale mud-layers that sharply truncate underlying sandstone layers and may contain sandstone lenses (Fig. 9E). Millimetre-scale, muddy matrix laminae and mud drapes also occur within F7. More rarely, gutter casts possess a compound mud-dominated (F2 to F3) infill with internal erosion surfaces (Fig. 9F).

Fig. 8. Photomontage (A) and line drawing (B) of sandstone beds and extensive gutter casts in Facies Association 2 (Jalan Tutong Simpang 688; location 3 in Fig. 2C). The degree of vertical amalgamation of sandstone beds with basal gutter casts is ca 30% in two-dimensions but this is likely to be considerably higher in three-dimensions.
In the Belait Formation in Brunei, gutter casts are between 0.1 to 2.0 m thick (average 0.41 m) and 0.15 to 10.8 m wide (average 2 m) (pink points – Fig. 11A). There is a moderate correlation between thickness and width for all measured gutter casts ($r = 0.61$), which increases slightly ($r = 0.68$) for gutter casts <0.7 m thick and <4 m wide. Most of the measured gutter casts were amalgamated (71%) rather than isolated. Amalgamated gutter casts are on average wider (2.3 m) than isolated gutter casts (1.3 m). Although average thicknesses of amalgamated and isolated gutter casts are similar (ca 0.4 m), amalgamated gutters may be much thicker (up to 2 m) than isolated examples (maximum 0.7 m). Belait Formation gutter casts measured in Labuan (Fig. 2A) are 0.03 to 1.04 m thick (average 0.22 m) and 0.4 to 5.7 m wide (average 1.7 m) (purple points – Fig. 11A). Gutter casts measured in approximately time-equivalent formations in Miri and Lambir, northern Sarawak (Fig. 2A), are 0.13 to 1.7 m deep (average 0.47 m) and 0.28 to 5.9 m wide (average 0.28 m) (black points – Fig. 11A).

Based on over 200 measurements, gutter casts in the Belait Formation average 0.34 m thick and 2.0 m wide (Fig. 11A). This is significantly larger than previously published measurements ($n = 27$) of gutter casts, which average 0.17 m thick and 0.59 m wide (blue points – Fig. 11, Myrow, 1992): 24 gutters are also wider than the previously published maximum (3.8 m; Wincierz, 1973). Similarly large gutter casts in the Blackhawk Formation (Grassy Member) in Tusher Canyon, Book Cliffs, Utah (cf. Guilpain, 2006; Eide et al., 2015), were measured during the present study which are 0.2 to 2.7 m thick (average 0.6 m) and 0.2 to 5.5 m wide (average 2.0 m) (green points – Fig. 11A).

In Brunei, gutter casts with sufficiently exposed walls or basal axes are predominantly oriented north/north-west to south/south-east (Fig. 11B), approximately perpendicular to the general north-east/south-west palaeoshoreline in the Middle to Late Miocene (Fig. 2A; Sandal, 1996; Tan et al., 1999; Saller & Blake, 2003; Torres et al., 2011).

**Interpretation**

Gutter cast erosion is generally ascribed to powerful unidirectional currents generated during the rising and peak phase of storms (Myrow, 1992; Myrow & Southard, 1996). However, the abundance of SCS/HCS within gutter casts and connected sandstone beds above, is more consistent with deposition under oscillatory-dominated combined flow (e.g. Dumas & Arnott, 2006). The lack of soft-sediment deformation in gutter cast infill, even for steep-sided gutters, indicates a close linkage between initial erosion of a cohesive substrate and rapid deposition of the mainly sand-dominated infill (Goldring & Aigner, 1982; Myrow, 1992; Plint, 2014). This suggests a linkage between the processes controlling both the initial erosion and subsequent infill. In contrast to a combined-flow model,
gutter cast erosion by unidirectional currents and deposition by oscillatory-dominated combined flows (Myrow & Southard, 1996), requires an abrupt decrease and switch from a unidirectional flow to an oscillatory-dominated flow.

It is suggested here that unidirectional flow is unlikely to occur without oscillatory flow during the rising or peak storm phase. Firstly, coastal set-up requires elevated storm waves, and consequently oscillatory flows, to form. The resultant coastal set-up is a continuous process throughout the period of storm winds and elevated wave energy, which leads to downwelling offshore-directed flow (relaxation current or downwelling storm flow) that is gradually deflected shore-parallel by the Coriolis force (e.g. Plint, 2010). Secondly, wave-enhanced sediment gravity flows, formed purely by sediment resuspension, necessarily require storm waves (e.g. Myrow et al., 2002). Likewise, hyperpycnal flows or wave-enhanced sediment gravity flows caused by river floods would also be linked to storm-related precipitation and thus be coeval with, or slightly later than, elevated storm waves and oscillatory flow. Therefore, it seems likely that gutter erosion and infill probably occurred under oscillatory-dominant combined flow (Lamb et al., 2008) or purely oscillatory flow.

Fig. 10. Photographs (A) to (E) of gutter casts in the Belait Formation and time-equivalent formations. (A) A metre-scale gutter cast and attached sandstone bed displaying a compound geometry. (B) Bedding relationships between two gutter cast beds; the overlying larger gutter lies directly above the preceding gutter, suggesting the influence of erosional topography (cf. Fig. 9D). (C) One of the largest gutter casts observed in the study area (Lambir Formation). (D) A metre-scale, isolated gutter cast-sandstone bed complex displaying a compound geometry and infill: the gutter was initially underfilled and then subsequently reoccupied (Lambir Formation). (E) Steep-sided, rectangular-shaped, isolated gutter cast (Lambir Formation).
Within the boundary layer of a shallow-marine combined flow, the superposition of oscillatory flow and an offshore-directed current produces oscillatory-dominated, bottom shear stresses greater than the sum of each component and higher bed shear stress in the offshore direction (e.g. Grant & Madsen, 1986; Davies et al., 1988). Consequently, Duke (1990) argued that all shallow-marine storm-generated bottom flows and associated erosion and deposition (e.g. Duke et al., 1991) must be dominated by orbital waves. Combined-flow boundary layer dynamics also explain the general shoreline-normal orientation of most gutter casts (Myrow & Southard, 1996). This is consistent with the observations here in the Belait Formation.

The continuity between SCS within gutter casts and in contiguous overlying beds suggests that gutter erosion, infill and then deposition of the connected bed occurred during the same event. This suggests the mobilization and redistribution of a substantial volume of sediment (Robertson et al., 2007; Keen et al., 2012). Such an event would certainly be non-actualistic and record deposition by relatively infrequent, large-magnitude storms (e.g. Duke, 1985). For instance, despite its exceptionally large size (>500 km wide) and associated storm tide (Harris, 1963), Hurricane Carla resulted in the deposition of a sand bed only up to 9 cm thick in water depths <40 m, albeit across ca 800 km² of the Texas shelf (Hayes, 1967; Morton, 1981; Snedden & Nummedal, 1991).

Compound gutter geometries may reflect: (i) erosion by multiple flow events, with the succeeding events probably exploiting topography created by the preceding events; and/or (ii) instability within a single eroding flow. In support of the latter, seabed erosion and sediment entrainment may induce turbulence, perhaps generating vortices, which may cause more localized erosion, as in the case of pot casts (Myrow, 1992). The largest, compound gutter cast infills herein (for example, Fig. 10A, C and D) suggest reoccupation and multiple phases of deposition within the same gutter.

**VERTICAL AND LATERAL FACIES ASSOCIATIONS RELATIONSHIPS**

The architectural relationships of facies associations have been examined at numerous outcrops throughout the study area (Figs 3, 4 and 7), including the detailed analysis of representative facies successions in the eastern BDP (Tanjong Nangka, Fig. 12), and subsurface data in the western BDP (Seria Field, Fig. 13; from Atkinson et al., 1986).

**Description**

The facies associations form ca 5 to 50 m thick parasequences, comprising two end-member types (Fig. 14A): (i) Type A parasequences (5 to 35 m thick) are sandier-upward facies
successions (from FA-1 to FA-2 and/or FA-3) containing up to 4 m thick, SCS-dominated sandstone bedsets (in FA-3) with sharp tops and bases (for example, Unit 2 in Figs 3 and 12, and Units 1 and 3 in Fig. 4); and (ii) Type B parasequences are more heterogeneous and more gradual, sandier-upward facies successions (from FA-1 to FA-2 and possibly FA-3) (for example, Units 1 and 3 in Figs 3 and 12, and Unit 2 in Fig. 4).

Within both types of parasequence, and their subsurface equivalents (e.g. Atkinson et al., 1986; Johnson et al., 1987, 1989), trough crossbedding is absent and vertical transitions between facies associations may either be gradational or abrupt. Abrupt contacts are particularly common within Type A parasequences between FA-1 and FA-2 (for example, Unit 2, Figs 3 and 12), FA-1 and FA-3 (for example, Unit PS1, Fig. 13) and FA-2 and FA-3 (Unit 3, Figs 4 and 7A). These abrupt basal contacts often include gutter casts (for example, Fig. 7A) and can extend laterally across hundreds of metres (for example, Fig. 13B and C). Well log correlations show that sandstone units between 0-3 m and 1-5 m thick, which are interpreted in core to be individual beds (Atkinson et al., 1986), extend laterally for up to 500 m (for example, Sand H in Unit PS1 and Sand F in Unit PS2; Fig. 13B and C). In contrast, amalgamated sandstone bedsets, typically between 1-5 m and 3-0 m thick, extend for at least 1-5 km (for example, Sands I and C, Figs 13B and C) (Atkinson et al., 1986). Facies associations show gradational lateral relationships both parallel and perpendicular to the interpreted north-east/south-west palaeoshoreline (for example, Units PS1 and PS2, Fig. 13D). Parasequence type can also change laterally: Unit PS1 from type A to B northward and eastward, and Unit PS2 from type A to B eastward (Fig. 13D).

Boundaries between parasequences may be gradational or sharp. Gradational boundaries display a muddier-upward FA-2 package.
(typically <15 m thick) (for example, between Units 1 and 2 in Figs 3 and 12, Units 1 and 2 in Fig. 4, and Units PS1 and PS2 in Fig. 13B to D). These surfaces extend for hundreds of metres (Fig. 13B to D); however, their recognition is affected by the lateral discontinuity of sand beds (for example, Sand B in Unit PS2, Fig. 13B to D). Sharp boundaries are marked by an abrupt decrease in sand content and bed thickness, with occasional accumulations of bivalves, crustaceans and shark teeth (for example, Unit 2, Figs 3 and 12).

The spatial distribution of sandstone beds within and between parasequences is variable, but the following patterns were noted (Fig. 13E and F): (i) amalgamated sandstone bedsets

Fig. 13. Vertical and lateral stratigraphic architecture in a ca 2.5 km² area of the Seria Field, western Baram Delta Province (BDP) (location shown in Fig. 2A; after Atkinson et al., 1986). (A) Location and data coverage in studied area relevant for cross-sections (B) to (D). GR, gamma ray; SP, sonic potential. (B) NNW–SSE cross-section from X–X’ in (A) showing interpreted lithostratigraphic correlation of sandstone-dominated and mudstone-dominated units. Sand beds/bedsets are labelled from ‘B’ to ‘I’ (‘A’ and ‘J’ are missing from the studied sections). (C) Approximately east–west cross-section along Y–Y’ in (A). (D) Interpreted facies association architecture, sequence stratigraphy and parasequence (PS) types along sections X–X’ and Y–Y’ (legend in Fig. 12). Interpreted parasequences are labelled PS1 and PS2. (E) Isopach map across the full study area showing the maximum thickness distribution of Sand I and the immediately overlying Sand H. Note the thickness contours are in feet (based on original data in Atkinson et al., 1986). (F) Isopach map showing the maximum thickness distributions for Sands ‘H’, ‘C’ and ‘B’ within PS2.
Fig. 14. Synthesis of stratigraphic facies relationships. (A) A schematic proximal to distal, chronostratigraphic and gamma ray (GR) log correlation (Logs 1 to 3), illustrating the two end-member, sandier-upward parasequence types ('A' and 'B'): Type A is relatively sandy and displays relatively sharp facies association transitions, Type B is more heterogeneous and displays more gradual transitions between facies associations. (B) Simplified plan-view model for facies association and parasequence relationships in a wave-dominated, fluvial-influenced delta system. Type A are interpreted as relatively proximal and axial, Type B as relatively distal and lateral. (C) Schematic model for parasequence-scale facies association architecture and stacking pattern, illustrating complex spatial and temporal parasequence type relationships within a deltaic setting. During T1, a heterolithic to sandy delta lobe progrades in the western region (log 2), whereas the eastern area receives less sediment, especially sand; this creates a depositional topographic high in the west. Between T1 and T2, nodal avulsion occurred – or a relatively continuous shift in discharge partitioning occurs between west and east distributary channels – and a delta lobe progrades into the depositional topography in the eastern region (log 3). Note that the lithostratigraphic (Lithostrat.) and chronostratigraphic (Chrono.) correlations do not match.

(>1.5 m thick; for example, Sands I and C) and the thickest individual beds (for example, Sands H and B), form lobate bodies oriented west–WNW to east–ENE; and (ii) individual sand units tend to be thickest where underlying amalgamated bedsets are thinnest or absent (for example, Sand H above Sand I), which implies a non-random stacking pattern.
Interpretation

Type A and B parasequences are interpreted as progradational units deposited in a storm-dominated, fluvial-influenced setting (Fig. 14A and B). Abrupt parasequence boundaries are interpreted as marine flooding surfaces, associated with wave ravinement, condensation and winnowing. Gradational (muddier upward) parasequence boundaries reflect a more gradual increase in base level and/or reduction in sand supply, which may reflect a lateral shift in the sediment supply system. Both types of marine flooding surface are known to be laterally extensive from subsurface data, and are the main cause of the ubiquitous vertical stratigraphic compartmentalization within the BDP (e.g. Johnson et al., 1989; Hadley et al., 2006).

Abrupt facies association boundaries within parasequences are interpreted as higher frequency fluctuations in sediment supply and/or base level. Unit 2 at Tanjong Nangka (Figs 3 and 12) includes a possible sharp-based shoreface succession (i.e. the abrupt base of FA-2 with associated gutter casts), which resembles a ‘regressive surface of marine erosion’ (RSME) and could indicate forced regression (e.g. Plint, 1988). In general, however, gutter casts in the Belait Formation are not limited to ‘precursor facies’ related to an inferred RSME.

Overall, Type A parasequences are interpreted as aggradation-dominated successions that accumulated in proximal and/or axial, high sediment supply locations on subaqueous, storm-dominated, sand-rich delta lobes, also subject to frequent erosion (Fig. 14A and B). Sand supply was mainly through high-energy, relatively short-lived events under the combined influence of wave, storm and fluvial processes (storm-floods); they closely resemble storm-dominated parasequences that accumulated in the Columbus Basin as part of the palaeo-Orinoco Delta (Bowman & Johnson, 2014).

Type B parasequences are interpreted to have formed under lower energy conditions, in more distal and/or lateral delta lobe settings (Fig. 14A and B). Type B parasequences more closely resemble classic wave-influenced shoreface models (Hampton & Storms, 2003). In both parasequence types, the conspicuous absence of trough cross-bedding, which is widely considered as indicative of upper shoreface deposition (Clifton, 2006), probably reflects the dominant, very fine sand grain size in the Belait Formation, since angle of repose dunes only form in grain

Fig. 15. Depositional model for the Miocene–Modern Baram Delta Province. (A) During fair-weather periods, longshore drift predominates and sand is constrained to a relatively narrow, shallow-marine to coastal region by shoaling and breaking waves (the ‘littoral energy fence’; Allen, 1970). (B) Large-magnitude, high wave-energy storms are coincident with intense precipitation, which results in storm-floods (SF) and downwelling storm flows (DSF) that erode and transport sand in an offshore direction. Spatial and temporal variability in sand erosion and offshore-directed transport results in a laterally variable sand distribution above storm-wave base (SWB) (for sand). An example of the temporal change in geomorphology of a river-mouth (pink box) is shown in Fig. 16.
sizes greater than ca 0.15 mm (upper fine sand; Boguchwal & Southard, 1990).

Type A parasequences are interpreted to pass distally (Fig. 14A) and laterally (Fig. 14C) into Type B parasequences. Increased proximal/axial reworking will increase progressive sorting and mean event bed thickness (Thorne et al., 1991): sandstone and mudstone beds in Type B parasequences will correlate up-dip, proximally and laterally, with erosional discontinuities in Type A parasequences (Fig. 14A).

The sand thickness patterns (Fig. 13E and F) suggest compensational stacking and topographic avoidance during successive periods of sand deposition. This may reflect periodic reorganization of the sediment transport field (e.g. Straub et al., 2009), including river-mouth switching caused by upstream avulsion (Fig. 14C). Within an overall axial depositional setting, diminished sediment supply to abandoned lobes causes increased reworking, resulting in a sandier and thicker bedded FA-3 package (intermediate Type A to B parasequence; log 2, Fig. 14C). Sand and mud supply from tropical rivers coupled with high accommodation allowed accumulation of a relatively heterolithic succession.

Fig. 16. Google Earth satellite imagery (A), (C) and (E) and mapped sediment distribution (B), (D) and (F) for the Seria River mouth in western Brunei (location shown in Fig. 1B) from November 2011 (A) and (B), June (2012) (C) and (D) and October 2013 (E) and (F). For (B), (D) and (F), yellow shading indicates the approximate distribution of sand above low tide; green is vegetated strandplain, which is dominated by *Casuarina* trees. Note the different tide levels and wave-energy conditions.
in the succeeding lobe (Type B parasequence; log 3, Fig. 14C).

The highly repetitive vertical stratigraphic cyclicity in the BDP therefore reflects spatial and temporal fluctuations in the rates of sediment supply and accommodation space creation. It is not possible to deconvolve the interplay of the full range of autogenic and allogenic processes (e.g. Storms & Hampson, 2005; Gani & Bhattacharya, 2007) which, in the BDP, would have included: (i) active tectonics comprising basin margin uplift, basement-linked faulting and delta-related gravity tectonics, especially growth faulting (Hodgetts et al., 2001; Morley et al., 2003; Gartrell et al., 2011); (ii) eustatically and climatically-driven sea-level changes; and (iii) temporal and spatial variations in storm-related processes, which controlled both coastal–shelf hydrodynamics (for example, wave intensity and frequency) and fluvial hydrology (for example, river-flood discharge and frequency, avulsion, etc.).

DEPOSITIONAL MODEL

The facies analysis of the Belait Formation in outcrop has been integrated with the following aspects of the BDP: (i) modern coastal processes, fluvial hydrology and coastal plain geomorphology (Fig. 1); and (ii) regional geological studies (Fig. 2), including previous facies-related work (e.g. Atkinson et al., 1986; Johnson et al., 1989; Tan et al., 1999; Lambiase et al., 2002; Mulcahy et al., 2003; Hadley et al., 2006; Lambiase & Cullen, 2013; Lambiase & Suraya, 2013; Abd. Rahman et al., 2014). This has resulted in a model of storm-dominated/wave-dominated, fluvial-influenced deposition, which highlights strong seasonality with distinct fair-weather and storm periods (Fig. 15). It is inferred that this depositional system has operated in the BDP since the Middle Miocene (Fig. 15), with the modern providing a close analogue to the Mio–Pliocene period, and vice versa; the tectono-stratigraphic framework, climate, gross depositional environments and source to sink system have been in place for the past 15 Myr.

Fair-weather periods are defined by low to moderate wave energy and low river discharge (Fig. 15A). Fluvial discharge mainly comprises the transport and deposition of suspended sediment during infrequent and low magnitude flood events. Shoreline processes are dominated by relatively low rates of alongshore sediment transport, which is sufficient to nourish flanking shoreface and strandplain regions between river mouths and form accreting spits across river mouths and embayments (Fig. 15A). This is illustrated in many of the ca 20 river mouths that are located along the ca 150 km long, open coast that extends from Brunei Bay, across the modern Baram delta and towards the south-west (for example, the Seria River mouth, Fig. 16; location shown in Fig. 1B). Hence, a major aspect of fair-weather periods is the development and storage of significant volumes of well-sorted, fluvially supplied and wave-reworked sand within the coastal zone.

Storm periods coincide with the seasonal monsoons, particularly the dominant North-east Monsoon. The short distance (ca 50 to 100 km), and relatively steep gradient from the hinterland to the shoreline, results in storms that simultaneously affect both the coastal zone and the lower coastal plain (e.g. Bhattacharya & MacEachern, 2009). This is manifested in the following ways: (i) wave energy is significantly elevated, with wave height exceeding 3.5 m (Sandal, 1996; Ahmadi et al., 2011); and (ii) extreme rates of precipitation in the mountainous hinterland result in ‘storm-floods’, which cause peak sediment delivery (bedload and suspended) through river mouths to occur simultaneously with storm-wave reworking at the shoreline (Bhattacharya, 2005). At the shoreline, a combination of storm-floods, storm waves and downwelling flows related to coastal set-up may enhance both alongshore and offshore sand transport. Storm-generated offshore-directed downwelling flow may be further augmented by storm-flood linked hyperpycnal flows and purely wave generated and enhanced sediment gravity flows. Total river-mouth bypassing will be enhanced where there are multiple river mouths along the shoreline (Fig. 15B; cf. Fig. 1B). The intensity and effectiveness of this process controls the width of the storm-generated facies belt, notably the seaward extent of the lower shoreface and distal delta front (Walker, 1983; Emery & Myers, 1996; Clifton, 2006).

Storm-flood processes operate in the BDP on a range of timescales. The shortest time period (yearly to decadal) is shown by dramatic changes in river-mouth morphology along the north-west Borneo coast. The Seria River (western Brunei) serves as an example (Fig. 16). In 2011, fair-weather processes had built a river-mouth spit (Fig. 16A and B). However, by 2012,
an estimated 40,000 m$^3$ of sand had been removed from the river mouth and the river channel course had straightened (Fig. 16C and D); part of this sediment volume would have been transported offshore by combined storm waves and river floods. By 2013, fair-weather processes had rebuilt the spit complex (Fig. 16E and F). While the 2011 to 2012 erosion cannot be related directly to an individual storm, the period coincided with the passage of Tropical Storm Washi across the South China Sea during the North-east Monsoon in December 2011.

DISCUSSION

Single versus multi-sourced systems

Multiple river-fed coastal plains are most common along active margins with short source to sink distances (tens to hundreds of kilometres) and intense rainfall events (e.g. western Taiwan – Liu et al., 2008; Blum et al., 2013). In contrast, passive margin deltas with continental-scale source to sink distances (thousands of kilometres) most commonly have a single, river-fed coastal plain, such as the Mississippi (e.g. Frazier, 1967), Orinoco (e.g. Warne et al., 2002), Niger (e.g. Oomkens, 1974), Nile (e.g. Stanley & Warne, 1993) and Danube River deltas (Bhat-tacharya & Giosan, 2003).

For a single feeder channel (Fig. 17A), sediment will be deposited across a relatively large and lobate delta plain with a network of distributary channels and associated mouth bars. An increase in the effectiveness of alongshore sediment reworking, most notably when waves and storms increase in frequency and magnitude, will result in greater delta front elongation and longer (along strike) flanking shoreface systems (Weise, 1980).

As the number of feeder channels increases, shoreline geometry becomes increasingly arcuate to cuspate and eventually elongate (Fig. 17B to D). For a multiple river-fed coastal plain with one dominant drainage system, the plan-view shoreline geometry superficially resembles the morphology of a wave-dominated delta with a single, point-sourced distributary channel (for example, Fig. 17D). Early models of the BDP therefore postulated a single-sourced wave-dominated delta model, based on the analogous modern Baram delta (e.g. Ho, 1978; Scherer, 1980; Johnson et al., 1989). However, a wave-dominated delta front fed by multiple rivers is likely to show much greater lateral variability in sand distribution, as a result of spatial and temporal variations in sediment transfer across the multiple drainage basins; this is supported by numerous subsurface studies of reservoir-scale stratigraphic architecture (e.g. Atkinson et al., 1986; Johnson et al., 1989; Tan et al., 1999; Lambiase et al., 2002; Mulcahy et al., 2003; Hadley et al., 2006) and by more recent quantitative evaluation of the modern coastal system (Ahmadi et al., 2011).

The engineering-focused work of Ahmadi et al. (2011) used well-established numerical modelling software (Delft3D; Delft Hydraulics, 2009), calibrated to 17 years (1992 to 2008) of offshore hydrodynamic data (combining waves, storms and tides), to estimate sediment transport across ca 140 km of coastline, from the Baram delta (north-east) to Tanjong Payong (south-west) (Figs 1B and 18). At the regional scale (50 to 100 km), the results from Ahmadi

![Fig. 17. Schematic maps illustrating the differences in shoreline geometry and sediment distribution between a single-sourced system (A) and multi-sourced deltaic systems, with an increasing number individual drainage basins across the same area of coastal plain (B) to (D) (based on Weise, 1980). The present-day, multi-sourced, Baram Delta Province is used as a guide.](image-url)
et al. (2011) support overall coastline progradation, which is highest (23.6 m yr\(^{-1}\)) in the Baram delta region. However, at the sub-regional scale (ca 5 to 20 km; i.e. approximately the spacing of individual river mouths), and on short timescales (ca one year; i.e. annual monsoon cycle), individual coastal segments vary between net erosional retreat (0 to 6 m; average 0.7 m) and net accretion (0 to 24 m; average 3 m) (Fig. 18A). In addition to the Baram delta, the detailed nearshore bathymetry maps show several, shallow (<10 m water depth), lobate protrusions located downstream and seaward of the Miri, Niah and Sibuti River mouths (Fig. 18B). These lobes are all deflected south-westward, in the direction of the dominant longshore drift, which is driven by the stronger North-east Monsoon, as supported by wave height data (Fig. 18A). Based on seabed sampling, there is a nearshore belt, averaging 5 m thick and 2 km wide, of very fine to fine sand, landward of more cohesive sediment (Ahmadi et al., 2011); the lobate protrusions comprise very fine to fine-grained (locally medium-grained) sand (Sandal, 1996). The Baram River mouth dominates coastline sand supply (425 000 m\(^3\) yr\(^{-1}\)). However, although sand supply from other river mouths has not been quantified, the shallow lobate protrusions and overall coastline progradation during the Late Holocene (e.g. Caline & Huong, 1992), suggest active sand supply, albeit in lower volumes. Ahmadi et al. (2011) conclude that the present-day BDP coastal zone morphology reflects the interaction of seasonal north-east (dominant) and south-west (secondary) monsoon winds and waves, tidal currents and fluvial river discharge. Using the scheme of Ainsworth et al. (2011), the coast would be classified as wave-

**Fig. 18.** Coastal to shallow-marine, process and sediment transport modelling in the present-day, western BDP (location shown in Fig. 1B; after Ahmadi et al., 2011). (A) Synthetic wave roses derived from wave modelling (Delft3D: Delft Hydraulics, 2009) and 17 years (1992 to 2008) of offshore data. Also plotted are annual rates of net littoral sediment transport (black arrows), coastal accretion or erosion (red arrows) and coastal plain hydrology. (B) Bathymetry in a ca 15 km wide nearshore zone (between water depths of ca 20 to 60 m). Dashed black lines outline shallow (ca <10 m water depth) lobate protrusions located downstream and south-westward of contemporaneous river mouths.
dominated with fluvial-influence and tidal-influence. This compares favourably with the facies analysis presented earlier of the Miocene successions.

**Storm-flood processes in the Baram Delta Province**

Considering the geological timescale of the BDP, a number of interacting factors have been conducive to the development and preservation of storm-flood related facies during the past 15 Myr (Fig. 19): (i) a humid, tropical, monsoonal climate (ca 4° to 5°N latitude); (ii) moderately high (>2300 mm yr⁻¹) but intense and localized rainfall (Sandal, 1996); (iii) a tectonically active, rising, deeply exhumed hinterland (e.g. Morley & Back, 2008); (iv) deep tropical weathering and highly erodible hinterland lithologies (dominated by mudstone and sandstone); (v) high sediment supply; (vi) a narrow (<100 km) and steep (ca 1°) coastal plain between the shoreline and hinterland; (vii) small to moderate-sized (10² to 10⁶ km²) drainage basins; (viii) short (<300 km) river lengths; (ix) high rates of tectonic accommodation creation; and (x) long-lived, tectonically controlled, drainage basins (‘Baram’ and ‘Champion’; e.g. Sandal, 1996). Large and relatively long-lived intense storms typically affect several drainage basins and cause high discharge of >100 m³ sec⁻¹ in 10³ km² drainage basins and >1000 m³ sec⁻¹ in 10⁵ km² drainage basins (Fig. 1A; Sandal, 1996; cf. Blum et al., 2013). In contrast, smaller storms may only affect single drainage basins (Sandal, 1996). Consequently, there are large spatial differences in sediment transport, as suggested in the present-day (Ahmadi et al., 2011), which may contribute to the lateral discontinuity of storm sands seen in both the present-day (Fig. 18) and the stratigraphic record (for example, Figs 12 and 13B to D).

**Modern and ancient analogues**

Gutter casts are rarely documented from modern settings. One notable example is the modern shoreface of Sable Island, Canada (Amos et al., 2003), where shore-normal gutter-like features were observed forming within, and in close proximity to, storm-dominated shoreface–shelf deposits.
association to, large (50 m wide, 0.5 m deep) shore-normal channels in 10 to 40 m water depths. Gutter formation and infilling only occurred under strong coastal downwelling, which was related to onshore-directed storm winds and coastal set-up (Amos et al., 2003; Li & King, 2007), suggesting the importance of off-shore-directed downwelling storm flows (e.g. Myrow, 1992).

The importance of storm-related flooding and intervening fair-weather periods on delta evolution is illustrated by the episodic evolution of the Brazos delta, Texas (Rodriguez et al., 2000). During flood phases, sand is transported from the river and is initially deposited near the river mouth, forming a subaqueous channel mouth bar (Rodriguez et al., 2000), analogous to storm periods in the BDP (Fig. 15B). During the subsequent fair-weather period before the next major flood, alongshore wave reworking of near-shore sand results in emergence and alongshore accretion of the channel mouth bar to the shoreline (Rodriguez et al., 2000).

Gutter casts are widely documented in many ancient shallow-marine successions. The Upper Cretaceous Blackhawk Formation in the Book Cliffs of central Utah, USA, comprises sediments supplied by multiple river systems from an actively rising hinterland (Sevier Orogenic Belt), across a relatively narrow (ca 200 km) coastal plain, and deposited along the western margin of the Western Interior Seaway (e.g. Hampson, 2010; Eide et al., 2014). Within the Blackhawk Formation, the scale of storm-generated sandstone beds (5 to 50 cm thick) and gutter casts (average 0.6 m thick) are similar to those observed in FA-2 of the Belait Formation (Fig. 11) (Guilpain, 2006; Eide et al., 2015). A ca 10 m thick and 4 to 6 km wide zone of gutter casts is associated with interpreted subaqueous channels (e.g. Pattison et al., 2007), relatively steep bathymetric breaks (ca 0.5°) and near or seaward of distributary channels (Eide et al., 2015). Accordingly, gutter casts and related event beds are interpreted to have formed by river-fed hyperpycnal flows (Eide et al., 2015; cf. Mutti et al., 1996), which may have been initiated through storm-floods (e.g. Pattison et al., 2007; Pattison & Hoffman, 2008).

Gutter casts in the Upper Cretaceous Dunvegan (e.g. Plint, 1996) and Kaskapau formations (e.g. Varban & Plint, 2008) are generally up to 50 cm wide and deep; they formed on a typically very low gradient (<0.01°), storm-dominated shelf/prodelta within the western Canada Foreland Basin (e.g. Plint, 2000). The contemporaneous deltaic shorelines were fed by multiple short rivers that drained an active orogenic belt under humid-temperate conditions (e.g. Plint, 2000; Bhattacharya & MacEachern, 2009). Orientations of gutter casts range between shore-normal to shore-parallel through a spectrum of nearshore to offshore facies, attesting to a polygenetic origin involving storm-generated oscillatory waves, downwelling storm flows, geostrophic flow and storm wave-enhanced sediment gravity flows (e.g. Plint, 1996, 2014; Varban & Plint, 2008).

The different coastal–shelf geological settings of the Blackhawk (e.g. Hampson, 2010), Dunvegan (e.g. Plint, 1996; Bhattacharya & MacEachern, 2009) and Kaskapau formations (e.g. Varban & Plint, 2008) all share common factors conducive to storm-floods, and also hyperpycnal flows (Mulder & Svyitski, 1995), including: (i) steep, tectonically active hinterland; (ii) humid climate (intense rainfall events and weathering); (iii) relatively narrow coastal plain; and (iv) multiple, relatively small, fluvial drainage basins.

**Miocene palaeogeography and palaeoclimate**

Storms affecting north-west Borneo are likely to have been more frequent and of higher magnitude in the Miocene due to palaeogeographic changes in Southeast Asia (Hall, 2002; Doust & Sumner, 2007). At present, the Philippines attenuates the energy of fair-weather waves and tropical storms moving from the Pacific Ocean towards the South China Sea (e.g. Camargo et al., 2007). However, in the Miocene, the Philippines were further south relative to China and north-west Borneo (Hall, 2002); more tropical storms may have propagated into the South China Sea and closer to north-west Borneo. The warmer Miocene climate (e.g. Lunt et al., 2008) may also have increased the frequency of extreme El Niño events, which are associated with intense and long-lived typhoons (Camargo et al., 2007). Furthermore, the emergent Sunda Platform, from the Late Oligocene to Late Miocene (van Hattum et al., 2006), would have trapped wave and storm-wave energy, possibly leading to higher storm tides and stronger downwelling storm flows.

**CONCLUSIONS**

1 The Baram Delta Province (BDP), north-west Borneo, has been an active source to sink system
for the past ca 15 Myr (Middle Miocene to present-day). Preservation of up to 9 to 12 km of coastal–deltaic to shelf successions provides evidence of efficient sediment transfer from hinterland to shelf coupled with a high rate of tectonic subsidence.

2 Ubiquitous sandier-upward facies successions in the ancient (Middle–Late Miocene), proximal part of the BDP (the onshore Belait Formation) are interpreted to be the consequence of prograding, storm-dominated, storm-flood-influenced coastal–deltaic depositional systems; storm-floods are storm-enhanced river discharge events that occur during the causal storm. Decimetre to metre-scale, very fine-grained sandstone beds thicken and amalgamate-upward and are dominated by commonly asymmetrical, swaley cross-stratification. Deposition occurred under oscillatory-dominated combined flows during storms, with only minor preservation of fair-weather deposits. Offshore-directed sediment transport probably reflects downwelling flow related to coastal set-up and, possibly, augmented by wave-enhanced sediment gravity flows formed by storm-floods.

3 Exceptionally abundant gutter casts display a wide range of thicknesses and widths (centimetre to metre-scale) and consistent ca NNW–SSE orientation, approximately perpendicular to the palaeoshoreline. Gutter cast erosion and subsequent infill is attributed to offshore-directed, oscillatory-dominated combined flows during storms. Mudstone-dominated facies associations comprise structureless or silt-streaked mudstones and centimetre-scale siltstone and sandstone beds, which display scoured bases, normal grading and combined-flow ripples; deposition was from storm flows and wave-enhanced sediment gravity flows. Low intensity (BI 1 to 3) and impoverished ichnofauna in mudstones reflects a stressed environment with a high sedimentation rate, possibly combined with fluid mud, soupground and high turbidity conditions.

4 The BDP is a possible end-member model for storm-flood-influenced deposition, which is made more likely by the following aspects of the geological setting: (i) a humid, tropical, monsoonal climate; (ii) localized (10^2 to 10^3 km^2), high intensity rainfall events; (iii) high rate of sediment supply – deep tropical weathering, an actively rising hinterland and erodible (low hardness) hinterland lithologies; (iv) short (hundreds of kilometres) source to sink distances – narrow (hundreds of kilometres) and steep coastal plain (ca 1°) and shelf (>0-1°); and (v) multiple, small to moderate-sized (10^2 to 10^3 km^2), drainage basins. The combination of these factors makes this depositional model applicable to other comparable modern and ancient settings.

6 The north-west Borneo hinterland, coastal plain and the BDP coastal zone are all characterized by a suite of sedimentary processes similar to those that operated during the previous 15 Myr. Sedimentary process interpretations show consistency between the Miocene coastal–deltaic deposits and the present-day coastal zone. Hence, the present-day wave-dominated, fluvial-influenced and tide-influenced shoreline is a good, albeit partial, modern analogue. However, on a geological timescale, short-lived, high-energy, storm-flood events dominate, which link fluvial sediment supply and wave reworking more closely than has been appreciated in previous BDP models.

7 Source to sink systems with short (ca <1000 km) sediment transport distances and multiple river-fed deltaic shorelines (such as the BDP), contrast with single point-sourced deltas with continental-scale drainage (source to sink distances of thousands of kilometres), such as the Mississippi, Niger and Nile River deltas: (i) shoreline geometry is more elongate rather than lobate; and (ii) the coastal sand belt shows greater along-strike variability, reflecting spatial and temporal variations in storm-flood processes across multiple drainage basins.

ACKNOWLEDGEMENTS

The authors thank the Department of Earth Science and Engineering, Imperial College London for support of DSC via a NERC PhD Scholarship. Shell International Exploration & Production (Houston) is thanked for financial support for fieldwork and for technical guidance throughout the duration of the project, especially R. Smith and C. Hearn. We also appreciate the logistical and technical support provided by Brunei Shell Petroleum, particularly N. Hoggmascall, D. Gray, M. Mueller and S. Large. We thank C. D. Dean for fieldwork assistance and P. Crevello, C. T. Baldwin, M. van Cappelle, G. J. Hampson, C. A-L. Jackson, J. P. Tromp, P. Myrow and J. Lambiase for their insightful discussions. We are grateful to G. Plint and J. Bhattacharya for their critical reviews and editorial comments.
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\textit{Manuscript received 1 April 2016; revision accepted 1 August 2016}