SEDIMENTARY FACIES AND STRATIGRAPHY OF THE PALAEOCENE TO MIDDLE EOCENE BELAGA FORMATION SIBU, SARAWAK

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ABSTRACT

A detailed facies analysis is presented for the Paleocene to middle Eocene Kapit and Pelagus members, Belaga Formation, exposed around Sibu, Sarawak, Malaysia. The Belaga Formation is compared to the West Crocker Formation of onshore Sabah and also the offshore, Neogene, hydrocarbon-bearing reservoirs of NW Sabah, in order to understand the variations in submarine channel, levee and lobe architecture in the Tertiary deepwater systems of Borneo. Facies in the Belaga Formation are characterised by thin to thick bedded turbidites, debrites and associated hemipelagic/pelagic mudstone. The facies are arranged into 6 types of facies associations. Channel-fill facies associations display channel geometries and are composed of interbedded debrites, turbidites and mudstone. The levee facies association comprises stacked, dm to m-thick, thinning upward cycles of interbedded turbidites, debrites and mudstone. The lobe (channel mouth) facies association forms thickening upward successions of interbedded high density turbidites, debrites, hybrid beds and mudstone. The lobe (fringe) facies association forms thinning upward successions comprising debrites with interbedded mudstone and thin turbidites. The basin plain facies association forms thick successions of interbedded mudstone and thin bedded turbidites. The slump facies association is composed of deformed levee deposits. The Belaga Formation is interpreted to represent deepwater deposits of a mud-rich, basin floor submarine fan depositional system, comprising elements of leveed channel systems and associated frontal splay-type lobes, which represent change in relative sea change. Tens of meters thick successions of lobe (channel mouth) deposits gradually overlain by channel fill deposits, or lobe (fringe) deposits overlain by thick levee deposits are interpreted as representing lobe progradation. Thick successions of interbedded levee and channel fill deposits are interpreted as representing leveed channel successions. The West Crocker Formation of NW Sabah also comprises leveed channel and lobe elements, but is a sand-rich system.

The Neogene deepwater deposits of offshore NW Sabah also share the same facies architecture with the Belaga Formation, but individual fans can be either sand- or mud-rich. The Belaga Formation and Neogene Sabah systems all share depositional trends which are perpendicular to the regional SW-NE structural trend of Borneo. However, the major difference is the common occurrence of laterally extensive mass transport deposits (MTDs) offshore NW Sabah, which are rare in the Belaga Formation, suggesting less frequent tectonic disturbance and/or lower sediment supply.

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ABSTRAK

Satu analisa facies yang terperinci untuk unit-unit Paleosen-Eosen Formasi Belaga (Ahli Kapit dan Pelagus) yang terdedah disekitar Bandar Sibu, Sarawak, Malaysia, dipersembahkan. Formasi Belaga dibandingkan dengan formasi West Crockeryang tersingkap di daratan Sabah dan juga takungan-takungan hidrokarbon berumur Neogen di lembangan sedimen perairan Barat Laut Sabah, untuk memahami kepelbagaian dari segi seni bina facies alur, tetambak lobus berumur Tertiar. Turutan Formasi Belaga mengandungi fasies turbidit nipis dan tebal, debrit dan batuan lumpur hemipelagik/pelagik. Fasies-fasies ini menjadi elemen kepada 6 jenis assosiasi fasies yang hadir dalam Formasi Belaga.Asosiasi fasiesalur menunjukan geometri cekung ke atas dan mengandungi fasies debrit, turbidit serta batu lumpur yang berselang-seli.Asosiasi fasies tetambak membentuk turutan menipis ke atas yang berselang-seli, dan mengandungi fasies turbidit, debrite dan batu lumpur yang berketebalan nipis (dm ke cm). Asosiasi fasies lobus (mulut alur) membentuk turutan menebal ke atas, yang mengandungi fasies turbidit berdensiti tinggi, debrit, perlapisan hybrid dan batu lumpur yang berselang-seli. Asosiasi fasies lobus (pinggir) membentuk turutan yang terdiri daripada fasies debrit, batu lumpur dan turbidit yang berselang-seli dan berlapisan nipis.Asosiasi fasies lembangan terdiri daripada turuan tebal batuan lumpur dan turbidit nipis yang berselang seli. Asosiasi nendatan terdiri dari mendapan tetambak yang telah tercangga. Formasi Belaga ditafsirkan sebagai sistem endapan pengenapan laut-dalam yang kaya-lumpur. Turutan di mana asosiasi fasies lobus (mulut sungai) berketebalan puluhan meter ditindih oleh asosiasi fasies alur, atau turutan di mana asosiasi lobus (pinggir) ditindih oleh asosiasi tetambak yang tebal ditafsirkan sebagai mewakili turutan progradasi lobus, yang mewakili perubahan pada paras laus. Asosiasi tetambak dan alur yang tebal ditafsirkan sebagai mewakili turutan pengenapan sekitaran alur-tetambak.Formasi West Crocker yang tersingkap di barat laut Sabah juga terdiri dari

sungai yang bertetambak dan lobus, tetapi formasi ini mewakili sistem yang lebih kaya-pasir. Turutan endapan di luar pesisir barat laut Sabah yang berumur Neogen, memiliki facies yang serupa dengan Formasi Belaga, tetapi menunjukkan seni bina fasies yang lebih pelbagai, dengan hadirnya sistem kaya-pasir dan juga kaya-lumpur. Formasi Belaga dan sistem Neogen luar pesisir Sabah juga memiliki kesamaan dimana alur mendapan menunjukkan orientasi yang serenjang dengan alur struktur barat daya-timur laut Borneo. Perbezaan yang paling jelas ialah kehadiran endapan angkutan massa, atau ―mass transport deposits" (MTDs) yang bertaburan luas di pesisir luar barat-laut perairan Sabah. MTD jarang berlaku di dalam Formasi Belaga. Ini menunjukan kurangnya gangguan tektonik dan/atau kurangnya mendapan yang tersedia.

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LIST OF SYMBOLS AND ABBREVIATION:

BI	:	Bioturbation Index
Ch	:	*Chondrites*
Cr	:	*Cosmorhaphe*
F	:	Facies
FA	:	Facies Association
FSST	:	Falling Stage Systems Tract
HST	:	Highstand Systems Tract
LAD	:	Lateral Accretion Deposit
LST	:	Lowstand Systems Tract
Pa	:	*Palaeophycus*
Pd	:	*Paleodictyon*
Th	:	*Thalassionides*
TST	:	Transgressive Systems Tract
WCF	:	West Crocker Formation

CHAPTER 1: INTRODUCTION

Sedimentary rocks in the Cretaceous-Eocene Belaga Formation, Sarawak, have been interpreted as deep water sediments, which show the characteristics of deep-water bathyal distal turbidites (Hutchison, 2005). The most recent facies analysis (Zainol et al. 2007) interpreted the Belaga Formation gravity flow deposits as elements of a submarine fan system. However, the facies analysis was of a general nature. New exposures of the Belaga Formation exposed around Sibu Jaya, Sarawak **(Fig. 1.1)**, preserve abundant sedimentary structures and provide an opportunity to conduct a detailed facies analysis of the unit, a comparison with previous work and constructing a depositional model.

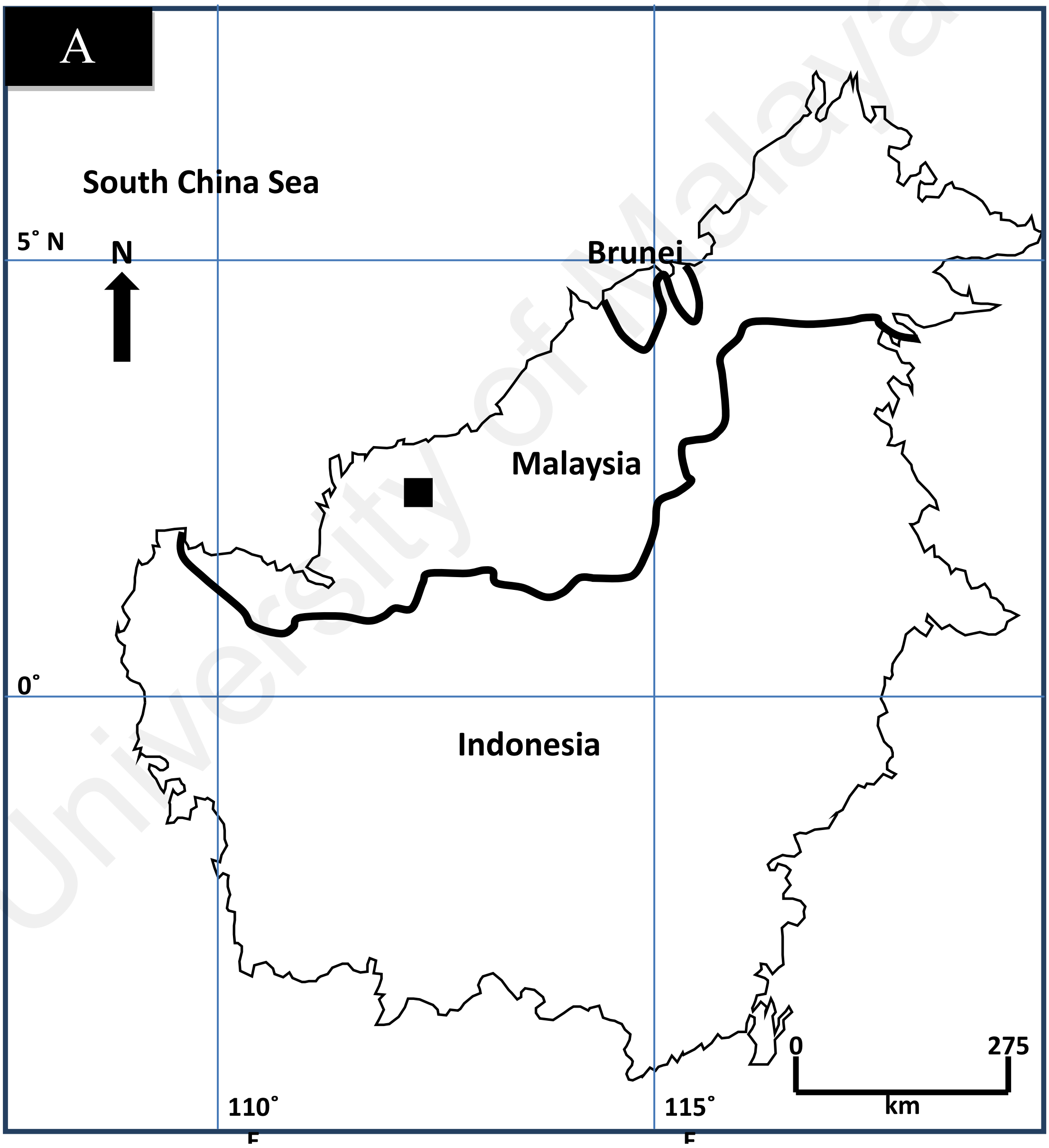
1.1. OBJECTIVES

This project will study the sedimentology and stratigraphy of the Belaga Formation exposed around Sibu Jaya, Sibu, Sarawak. The 3 objectives of this project are to;

1. Describe and interpret the sedimentary facies of the Belaga Formation exposed along Sibu Jaya road.
2. Describe the stratigraphic architecture of facies in the Belaga Formation exposed around Sibu, and
3. Reconstruct the geomorphology and depositional setting of the Belaga Formation around Sibu, Sarawak.
4. To discuss the evolution of the succession from a sequence stratigraphic perspective.

1.2. STUDY AREA

The study area is situated in the area of Sibu Jaya, Sibu, Sarawak (Latitude 2°14‘23.19 N and Longitude 111°58‘53.26). The outcrops are roadcuts exposed along a 12 km stretch of the AH150 road **(Fig. 1.1),** just south of Sibu Jaya and Sibu Airport. Based on the map of Wolfenden (1960), these exposed strata transect two different members of Belaga Formation **(Fig. 1.2)** (Kapit and Pelagus members).



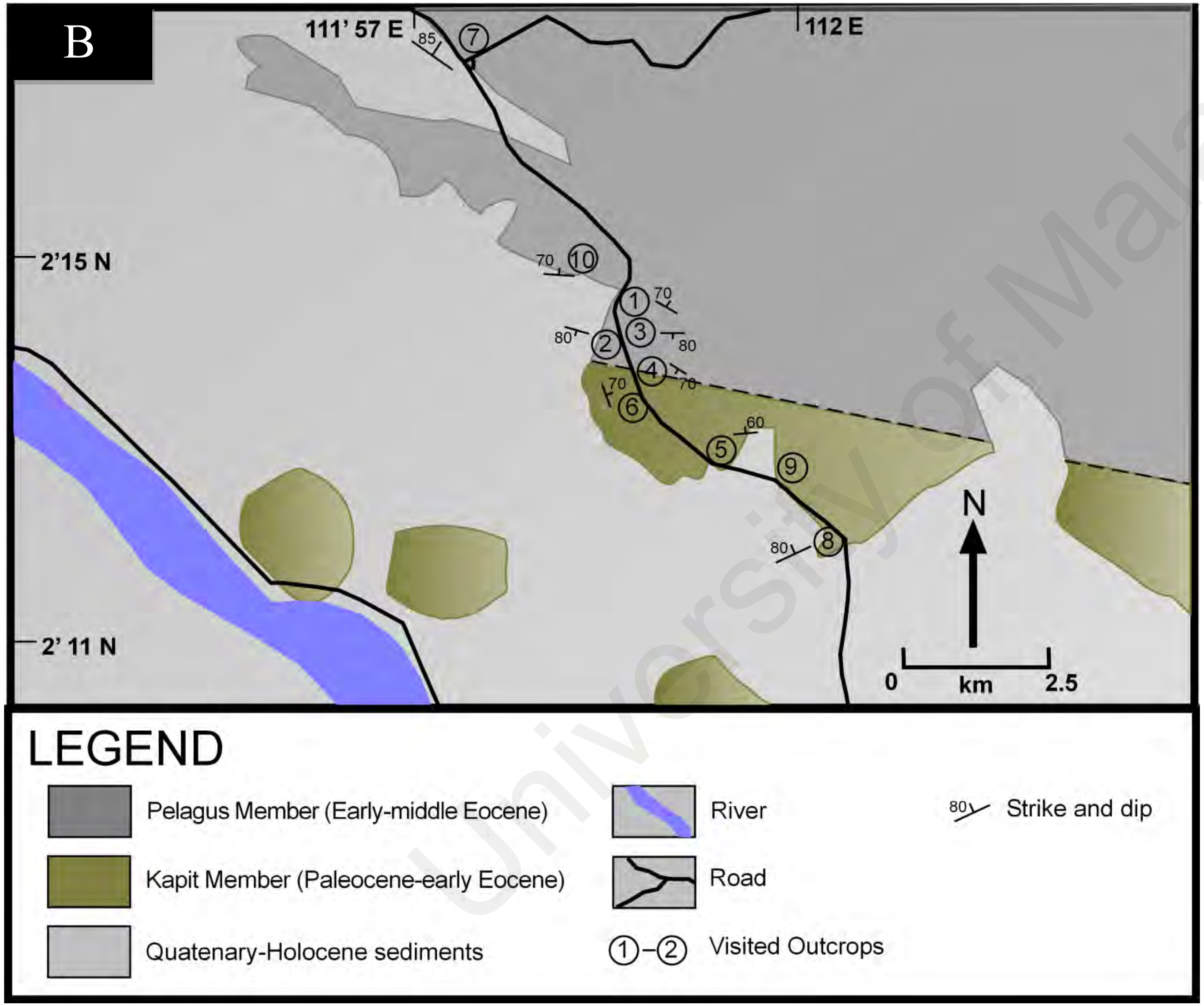


Figure 1.1. A) Location of study area is marked in black square. B). Enlarged map of study area. In the picture shows the location of studied outcrops (modified from Wolfenden, 1960)

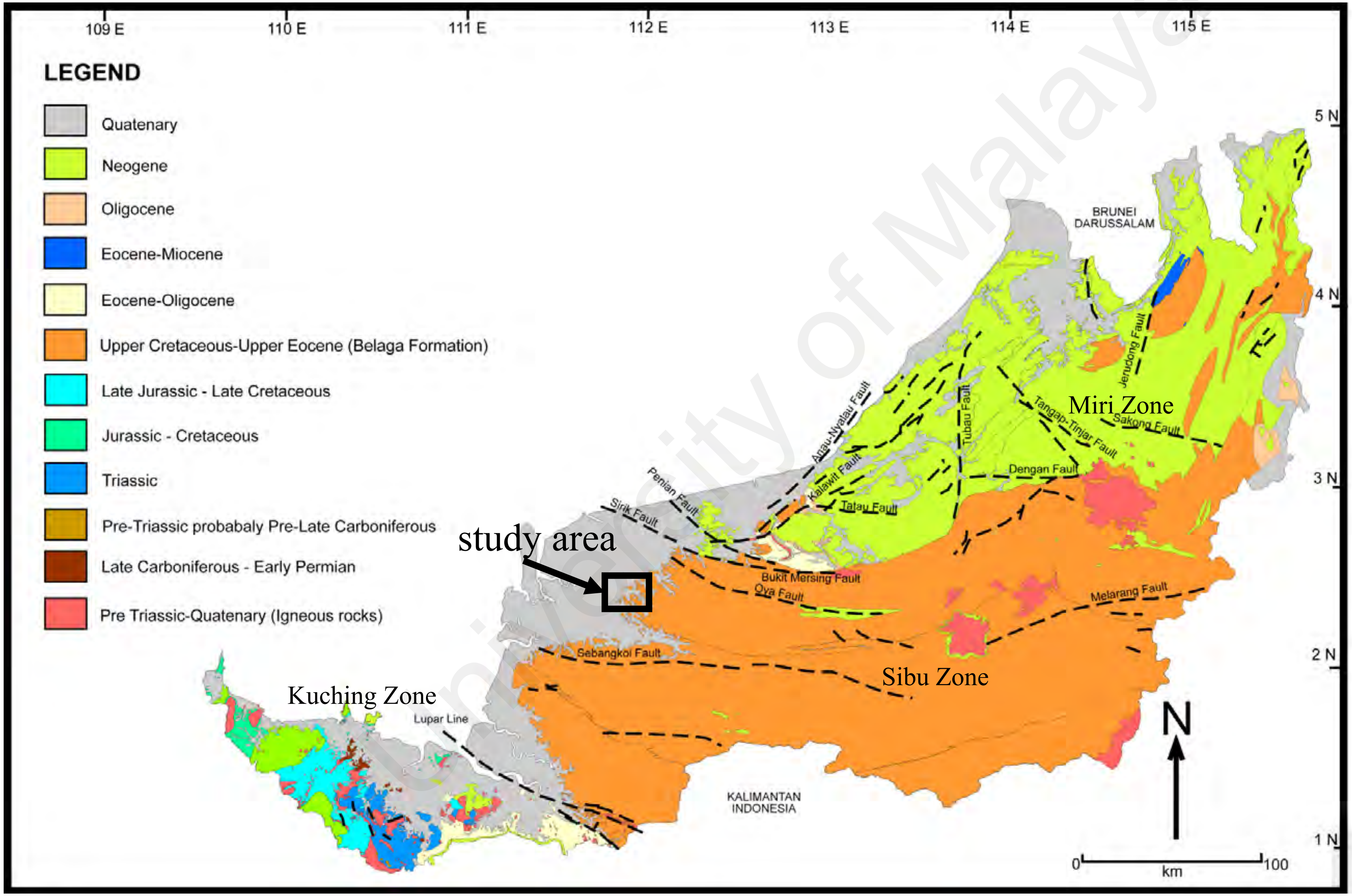


Figure1.2. Geology of Sarawak (modified from Geological Society of Malaysia, 2013)

1.3. METHODOLOGY

Ten outcrops along the AH150 road, just south of Sibu Jaya and Sibu Airport, were selected for detailed facies analysis. The strata generally dip steeply towards the NW and NE. No other outcrops were found within a 10 km radius from Sibu Jaya due to the limited road and land access. The field work was done within 30 days duration and completed at 3 different times.

The study presented in this project encompasses sedimentological, stratigraphic and ichnological techniques. Data for the study collected from the field outcrops include facies analysis-based sedimentary logging, photos of sedimentary structures, sketches, measurements of bed thickness, grain size, texture, bioturbation index level and type as well as paleo-currents readings. The outcrops were logged in meters using a Jacob‘s staff with attached-compass-clinometer for bed strike and dip measurement. The 10 logged sections range in thickness between 22 - 123 m. A total of 537 m of strata have been logged for this study. Paleo-current directions were measured using visible three dimensional sole marks (groove/flute casts) on the rotated bedding plane. The sandstone samples were collected for textural study, including sorting and grain size. The degree of bioturbation in the beds was measured using the Bioturbation Index (abbreviated BI) of Taylor and Goldring (1993) **(Table 1.1).**

Table 1.1. Bioturbation index (BI) where each grade is described in terms of the sharpness of the primary sedimentary fabric, burrow abundance and amount of burrow overlap (Taylor and Goldring, 1993)

Grade	Percent Bioturbated	Classification
0	0	No bioturbation
1	1 to 4	Sparse bioturbation, bedding distinct, few discrete traces and/or escape structures
2	5 to 30	Low bioturbation, bedding distinct, low trace density, escape structures often common
3	31 to 60	Moderate bioturbation, bedding boundaries sharp, traces discrete, overlap rare
4	61 to 90	High bioturbation, bedding boundaries indistinct high trace density with overlap common
5	91 to 99	Intense bioturbation, bedding completely disturbed (just visible, limited reworking, later burrows discrete
6	100	Complete bioturbation, sediment reworking due to repeated overprinting

1.4. GENERAL GEOLOGY

The Sibu Zone of central Sarawak forms an approximately 200 km wide and 150 km long, Northeast-Southwest trending belt composed mainly of sedimentary strata of the Rajang Group. The Rajang Group generally comprises highly deformed, slightly metamorphosed, deepwater clastic deposits of Late Cretaceous to Late Eocene age (Kirk, 1957).

The Rajang Group is stratigraphically divided into several units,based on foraminifera and nano-fossil biostratigraphy, from oldest to youngest: (1) Lubok Antu Melange; (2) Lupar Formation, and; (3) Belaga Formation (Liechti et al., 1960). The stratigraphy is younging northwards from the Lupar Fault Zone. The Lupar Fault Zone forms the southern boundary of the Sibu Zone with the Kuching Zone (Tan, 1979). It separates the Rajang Group from Eocene-Oligocene age, nearshore to continental deposits of the Silantek Formation further south. The Lubok Antu Melange is the oldest unit in the Rajang Group and is exposed north of the Lupar Fault Zone (Figure 1.2). It comprises a melange with a sheared, mudstone and shale matrix, with blocks of sedimentary rocks, serpentinite, gabbro, basalt, chert and metamorphic equivalents (Basir Jasin, 1996). Foraminifera and nanofossils from the matrix give a Late Cretaceous to Early Eocene age (Tan, 1979). The melange blocks are older, containing fossils of Jurassic (Tithonian) and Cretaceous (Albian-Cenomanian) age (Tan, 1979; Basir Jasin, 1996).

Lupar Formation is exposed north of and overlies the Lubok Antu Melange. It mainly comprises turbidites and associated igneous rocks of the Pakong Mafic Complex (Hutchison, 2005). The Pakong Mafic Complex can be further subdivided into a Gabbro

Zone composed of layered plutonic rocks, and a Basalt Zone composed of massive/pillowed volcanic rocks (Tan, 1979).

The Belaga Formation is the main focus of this project due to its wide distribution. The Belaga Formation is widely exposed in the area between Tatau and Sibu, in central and southwest Sarawak, Malaysia (Kirk, 1957; Wolfenden, 1960). The formation is estimated to be between 4.5-7 km thick (Hutchison, 2005) and comprises steeply dipping, strongly folded beds of sandstone/metasandstone interbedded with mudstone. The term flysch is commonly used to describe the strata of the Belaga Formation (e.g. Haile, 1969; Hutchison, 1996; Zainol et al., 2007), and the unit displays facies characteristics of deepwater turbidites (Hutchison, 1988, 1996, 2005). The Belaga Formation is Late Cretaceous – Late Eocene in age (Liechti et al., 1960). The formation is subdivided into five members, based on palaeontology, from oldest to youngest **(Table 1.2)**: (1) Layar Member (Late Cretaceous); (2) Kapit Member (Paleocene); (3) Pelagus Member (Early-Middle Eocene); (4) Metah Member (Middle-Late Eocene), and; (5) Bawang Member (Late Eocene) (Kirk, 1957; Wolfenden, 1960). Foraminifera indicate that the stratigraphic succession of the Belaga Formation youngs northwards away from the Lupar Line of southern Sarawak (Liechti et al., 1960).

The Layar Member is Late Cretaceous in age (Liechti et al., 1960) and is mainly composed of rhythmically interbedded/laminated slate/phyllite and metasandstone laminae, with minor thick beds (up to 3 m thick)(Tan, 1979). The sandstone beds display rhythmic bedding, graded bedding and ripple cross-lamination. Based on the characteristics the sandstones are interpreted as turbidites.

The overlying Kapit Member is Palaeocene to Early Eocene in age (Kirk, 1957; Wolfenden, 1960). The unit conformably overlies the Layar Member. Lithologically, the Kapit Member is very similar to the Layar Member and is mainly differentiated based on palaeontology. However, sandstone beds are more common in the Kapit Member, reaching up to 90 m in thickness. Conglomerates are also common. The Kapit Member succession appears to display a weak fining upward vertical trend, from a lower part with thick sandstone beds and conglomerate, into a middle part with thinner sandstone beds and an upper part comprising a monotonous succession of mudstone, slate, rare phyllite and occasional sandstone.

The Pelagus Member (Early to Middle Eocene, Kirk, 1957) forms an approximately 40 km wide belt extending East-Southeast from Sibu. The Pelagus Member is predominantly argillaceous, with common beds of sandstone and conglomerate.

The Metah Member conformably overlies the Pelagus Member and forms a belt parallel to, but northwards of the Pelagus Member. Foraminifera indicate Middle to Late Eocene age (Kirk, 1957). The Metah Member is very similar to the underlying Pelagus Member, but is less metamorphosed and the sandstone is more texturally mature (sub-greywacke). Sandstone beds are commonly 15-150 m thick, but thinly bedded turbidites are also present. The Bawang Member is Late Eocene in age and is characterised by a thick succession of amalgamated sandstone separated by thinner argillaceous units. Apparently the strata become more deformed southwards in older rocks (Zainol et al., 2007). The current study focuses on strata of the Kapit and Pelagus members exposed around Sibu, Sarawak.

Most of our understanding of the sedimentology and stratigraphy of the Belaga Formation is from older literature cited above, before the standardization of modern approaches to sedimentary rock analysis, such as the facies analysis approach and sequence stratigraphic concept (Vail et al 1977). The previous works were mostly interested in lihtostratigraphy, with little data regarding lateral facies variations and facies architecture. Thus, the model constructed by Wolfenden (1960), Liechti et al. (1960) and reviewed in Hutchison (2005) is very simplistic, with every Member of the Belaga Formation regarded as generally homogeneous with a uniform depositional setting. Current deepwater facies models tell us otherwise, and significant lateral changes in facies and facies architecture should be expected (e.g. Zakaria et al., 2013; Khan and Arnott, 2011).

Zainol et al. (2007) recently conducted a facies analysis of the Belaga Formation, based on 7 outcrops exposed around the Sibu and Tatau areas of Sarawak and encompassing 129 m of logged section. The studied strata included intervals of the Metah, Pelagus and Bawang members. They identified 3 facies: (1) thick-bedded sandstone interpreted as debrites; (2) heterolithic sandstone-mudstone interpreted as basin-floor or ―distal" turbidites, and; (3) mudstone interpreted as hemipelagic-pelagic deposits. They developed a generalised submarine fan depositional model for the Belaga Formation as a whole, based on Mutti et al. (2002), with the facies representing, middle fan, outer fan to basinal plain settings. This current project builds on this initial, and more regional work, on the Belaga Formation, and provides a more detailed facies and stratigraphic analysis of the Pelagus and Kapit members specifically. An example of the methodology planned for this study is Zakaria et al. (2013), where a detailed facies analysis was conducted for younger submarine fan deposits of the West Crocker Formation, further north in Sabah. The identification of facies, facies associations and facies succession will produce a higher resolution model of the submarine fan, with the potential identification of sub-environments and depositional elements such as channel levees, channelised/non-channellised lobes, mass transport complexes and slumps

(Mutti and Normark, 1991; Pirmez et al., 2000; Mutti et al., 2002). Also, despite the Belaga Formation being identified as a submarine fan, it is uncertain whether it was multiple sourced, point sourced or linear sourced, or whether the system was sand or mud-dominated (Reading and Richards, 1994; Stow and Mayall, 2000)

Another missing piece of the puzzle is a sequence stratigraphic interpretation of the Belaga Formation. The Belaga Formation is tectonically deformed, with tight folds and thrust faults with repeated strata (Hutchison, 2005). This makes it difficult to correlate between outcrops and to reconstruct a depositional history using sequence stratigraphy. However, integration of biostratigraphic dating from individual outcrops and facies analysis may provide a better understanding of the depositional history. Vertical trends in grain size (fining or coarsening upward) and the presence of certain facies or depositional elements may reflect fluctuations in relative sea level (Johnson et al., 2001; Haughton et al., 2009; Hodgson, 2009). Mass transport deposits, or MTDs are common during the falling stage systems tract, during falling relative sea level (Nelson et al., 2011). Thick, amalgamated sandstone intervals are common during the lowstand systems tract. Sand:mud ratio decreases during relative sea level rise (transgressive systems tract) (Posamentier and Walker, 2006). Fine grained sediments predominate during the highstand systems tract, when relative sea level rise slows down. However, it doesn't restrict the possibility of fine grained sediments to form during Lowstand system tract, which are common in levees (Bouma, 2000).

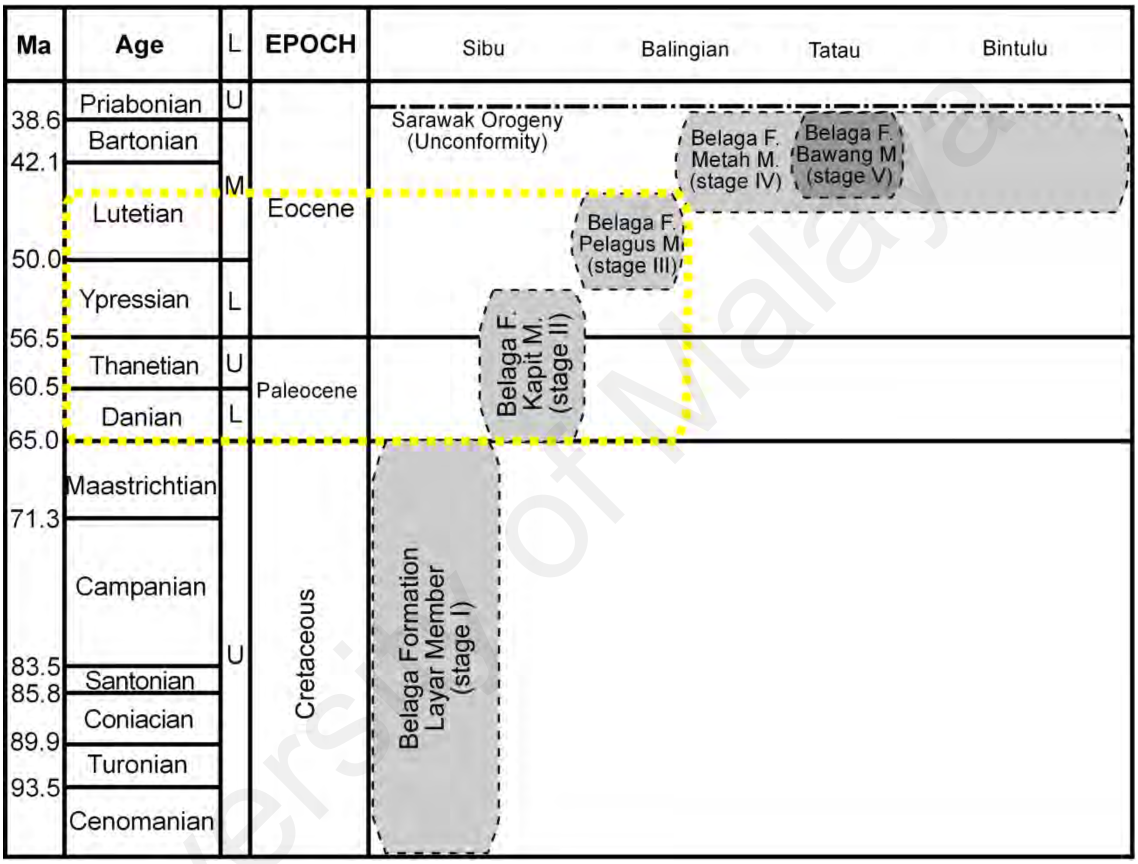
1.5. TECTONIC SETTING

The Rajang Group is segmented into imbricate thrust fault blocks up to 15 km wide (Tongkul, 1997, Honza et al., 2000).The Sibu Zone has also been referred to as the Rajang Fold-Thrust Belt (Madon, 1999) or Rajang Accretionary Complex (e.g. Honza et al., 2000). Despite a southward regional dip, the strata of the Rajang Group young towards the north. The fold and thrust belt has been interpreted as representing an accretionary prism of a South to Southeast dipping subduction zone beneath Northwest Borneo (Hamilton, 1979; Rangin et al., 1990; Tongkul, 1997). However, the younger units in the Rajang Group probably represent post-subduction collision and deformation of a marginal basin (Hutchison, 1996; Moss, 1998).

Accretionary prisms are associated with convergent margins from two different tectonic plates. During subduction, the sediments deposited on the down going oceanic crust are off-scraped and accreted onto the non-subducting continental crust, forming an accretionary prism (Huang et al., 1997; Karig and Sharman, 1975). Using paleomagnetic studies of Indochina and Borneo conducted by Fuller et al (1991), Honza et al. (2000) interpreted that Borneo has been bent into arc, convex towards the east. The compressional stress forming the arc is interpreted to have been provided by the northward movement of the Indian and Australian plate, with hard collision of India with the southern margin of Asia since 43 Ma (Lee and Lawver, 1995, Honza et al. 2000). This event might have accelerated the development of the Rajang accretionary prism of northern Borneo (Honza et al., 2000). Subduction ceased by the Late Eocene due to the collision of the Balingian-Luconia continent with the Rajang Accretionary Complex, initiating the "Sarawak Orogeny" (Hutchison, 1996), which consequently uplifted the Rajang Group and Silantek Formation. The presence of phylite and slate in the older units of the Rajang Group (Layar and Kapit members) (Liechti et al., 1960; Hutchison, 2005) provide supporting evidence for an arc system interpretation (e.g. Huang et al., 1997), where intensive deformation and metamorphism of earlier sediments are formed in the accretionary prism during the arc-continent collision.

Shallow marine deposits of the Late Eocene Tatau Formation unconformably overlie the Rajang Group (Madon, 1994; Almond et al., 1989).

Table 1.2. Stratigraphy of Belaga Formation (modified from Hutchison, 2005, age done by Kirk, 1957). Yellow square marks the age stratigraphy of study area



CHAPTER 2: SEDIMENTARY PROCESSES ASSOCIATED WITH DEEP MARINE DEPOSITIONAL SETTINGS

2.1. INTRODUCTION

Previous studies interpret the Belaga Formation as being composed of deep water facies deposited by debris flows and turbidity currents. Thus, it is practical to review our current knowledge of the sedimentological processes and facies present in such settings.

2.2. SEDIMENT TRANSPORT MECHANISM AND TYPE OF DEPOSITS

Deep marine, coarser grained deposits are affected by gravity and mass movement processes. Finer grained sediments such as silt and mud are deposited mainly from the fall out of suspended material (Johnson et al., 2001).

2.2.1. MASS MOVEMENT DEPOSITS

The term mass movement deposits is used to define deposits that are formed from the gravity-driven downslope movement of coherent to semi-coherent sediments particularly associated to plane failure (e.g. Arnott, 2010). They are also referred to as mass wasting deposits. These mass movements can travel distances of up to hundreds of kilometres, and are produced when the driving force such us gravity, exceeds the tensile strength of the former sediment pile. This tensile strength depends on the internal conditions of the sediment, which include sediment composition, consolidation, porefluid pressure, and mechanically weak layers (Locat and Lee, 2002). Mechanisms that initiate mass movement of the unstable layers include over-steepening, seismic loading, cyclic storm wave loading, rapid accumulation and under-consolidation, gas charging, gas hydrate dissociation and seepage (Locat and Lee, 2002).

Once mass movement is initiated, it will not stop until there is a resisting force. This usually happens when friction along the basal failure plane exceeds the gravitational driving force, leading to mass deposition. There are two-end member types of mass-movement: (1) slumps and (2) slides. The differences between these two are mainly in the intensity and nature of internal deformation. In slides, deformation is comparatively minor and brittle deformation is more dominant, in the form of detachment of bedding-parallel surfaces. Shear deformation may be intense in the thin zone near the unit base. In slumps, the deformation is more intense and ductile in character, and the deformation shows rotational movement. Folds in soft sediment slumps are abundant. Beds are normally tightly folded with the axial plane sub-parallel to the planes of the internal shear (Arnott, 2010).

2.2.2. SEDIMENT GRAVITY FLOWS

Gravity currents occur when more dense fluids move through and displace less dense fluid. Sediment gravity flow density contrast is produced by the presence of suspended sediment. Based on the flow properties and sediment-support mechanisms, the end member flows are classified into two types: (1) cohesive flows (debris flows) and (2) frictional flows (turbidity current) (Mulder and Alexander, 2001).

2.2.2.1 Debris flow deposits

A debris flow is a dense, viscous mixture of water and sediment where the sediment has higher mass as well as volume compared to the water (Major, 2003), and is flow having plastic rheology and laminar state from which the deposition occurs through feezing *en masse*(Shanmugam, 2006).A variety of grain sizes is present, from boulder to clay. The particles are principally suspended by cohesive forces provided by a matrix of fluid and fine grained sediment (silt clay mixture). Debris flows can occur on land (on alluvial fans), and also can occur in submarine environments, where sediments are transported down the continental slope and locally on coarse-grained delta slopes. Debris flow is triggered due to the slope failures, high sediment concentration from discharged rivers and oceanographic processes (Piper and Normark, 2009), causing the particles to be suspended by cohesive forces, generated by a matrix of fluid and fine grained sediments, which comprising of silt-clay mixture. The flow is stop and deposition is initiated when shear strength of the water mixture, grain-contact friction and the friction along the flow boundaries are exceeded, causing the flow to either freeze abruptly or gradually from low to high shear surface area (Kuenen & Sengupta, 1970; Kneller & Branney, 1995).

Debris flow deposits can be tens of kilometer in wide and over 100 meters thick, although thin beds are also common. Due to the rapid down flow reduction in the speed of the flow, toe-thrusts are common at their down flow terminus (Possamentier and Walker, 2006). The base of debris flow deposits is generally planar and non-erosional, due to the strength of mass moving and damping of large scale fluid turbulence. However, scoured bases are also observed, which may indicate that the erosive basal contact represents the pre-existing seafloor channel. It is also possible that the scoured base formed by ploughing of the rigid part of the flow through the underlying seafloor sediment (Posamentier and Walker, 2006). Dragged sediment on the surface can also form the linear grooves up to 40 m deep, hundreds of meters wide and longitudinally extending for over 20 km (e.g. Posamentier and Kolla, 2003). Gee et al. (1999) reported that modern debris flows could travel for a distance of up to about 700 km. Such long transport distances may be due to hydroplaning between the beds and overlying flow which reduced friction and enabled the flow to travel farther.

Debris flows can initiate the formation of turbidity currents (vice-versa) as the mixing of eroded sediment along its leading edge is lifted and turbulence suspension is developed above the flow (Hampton, 1972; Sohn, 2000). However, the amount of eroded and transferred sediment turned to turbidity current is very small due to the low permeability which causes minimum water infiltration (Mohrig, et al., 1998). Moreover, if the velocity of movement between the debris flow and the overlying turbulence flow is similar, the erosion along the interface between them will be negligible. But if the debris flow stops, the turbulence suspension can erode the top part of the debris flow deposit as it will detach from the debris flow deposit and continues to travel further downslope. The reverse can also happen, where turbidity current turns partially into a debris flow (Fisher, 1983). As the turbidity flow slows down rapidly and turbulence is reduced, the suspended sediment will fall off rapidly towards the bed of the cohesive mud particles, which eventually increases yield strength until the point where the particles are supported by matrix strength.

Mass transport deposits (MTD) are typically erosive based and occur in a variety of scales, from hundreds of meters square to kilometers square. The dimension (length and width) of mass transport deposits are controlled by geomorphology and position of the sediments source. The large scale deposit sare controlled by the instability initiated at the upper part of the basin margin or edge of the edge shelf, whereas the small scale is controlled by basin failure along local slopes (Moscardell and Wood, 2008). The internal structure of mass transfer deposits (MTD) consists of a complex, disorganised assemblage of slump, slide and debris flow deposits which are locally interstratified with channel and overbank strata. MTDs can be associated and interbedding with channelised turbidites sands (Algar et al., 2011), as MTDs can attain greater distance travel than turbidites (McGilvery et al., 2004). Large occurrences of mass transfer deposits in the stratigraphic column indicate a major change the basin sedimentation regime, which may indicate the initiation of Low system tract and Transgressive system tract (Possamentier and Kolla, 2003). These changes may relate to allocyclic processes which have sequence stratigraphic significance.

2.2.2.2 Turbidite Deposits

Turbidity currents are gravity-driven turbid mixtures of sediment temporarily suspended in water. It is a less dense sediment mixture than debris flow. The term turbidity current is also used in subaqueous sediment density flow having larger grains to settle, forming a layer by layer and displaying a normal grading fashion/trend (Talling et al., 2012). Turbidites were first generalised by Bouma (1962), which is now famous for the term ―Bouma Sequence" for the internal structure of turbidites. Turbidity currents consist of three distinct parts that are not sharply bounded: they are (1) head, (2) body and (3) tail (Kneller and Buckee, 2000). The head is the downcurrent part where rich sediment is mixed with ambient fluid. It is characterised by a sharp overhanging nose, above which the head slopes back in the upstream direction due to the resistance of the stationary overlying fluid. As a result, strong shear is generated that carries the sediment-rich fluid from the head towards the body of the current **(Fig. 2.1).** The body of the current moves faster compared to the head and continuously supplies sediments to the head in order to sustain the current. Due to differences in grain size within the flow, coarse grain sediments will decelerate first and will accumulate at the lower part of the head whereas the finer sediments will be brought backwards and upwards into the body of the current. With time, the flow becomes longitudinally differentiated in terms of grain size. The sediment concentration at the tail is low. As a result, the flow speed decreases and will eventually stop.

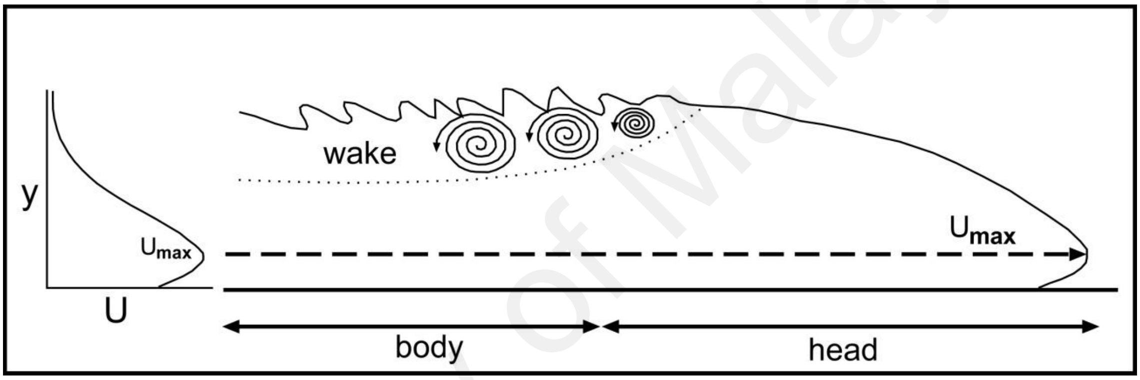


Figure 2.1. Line diagram illustrating the typical shape and velocity profile of a turbidity current (modified after Kneller and Buckee, 2000). The velocity maximum occurs in the lower part of the flow, unlike open-channel flow (river). The extensive mixing occurs along the upper part of the current.

Turbidity currents can be generated by slumping and normally occurs in deep lakes, continental shelves and deep marine environments. Turbidity current deposits (i.e. turbidites) are sharp-based and display a normal grading grain size trend. Turbidites also display a distinct and predictable vertical facies succession known as a Bouma Sequence (Bouma, 1962). The Bouma Sequence **(Fig. 2.2)** is vertically divided into 5 divisions (Bouma division A, B, C, D and E, from bottom to top respectively), each displaying different sedimentary structures. Bouma division A is the basal most division and consists of poorly sorted, structure-less sand. Bouma division B consists of parallel laminated sand. Bouma division C consists of cross-laminated sand, often displaying climbing ripples lamination as well as convolute lamination. Bouma division D is composed of fine sand and silt displaying poorly developed horizontal lamination. Bouma division E, which is the top most, consists of fine grained sediments from silt to clay sized. The fining upward grain size trend and vertical facies pattern is interpreted as representing waning flow deposition (Bouma, 1962, Blatt et al., 1984, Southard, 1991). However, the absence of cross-beds developing in between Bouma divisions C and D is problematic. Then, later years new models of turbidite division came up, characterized by different in grain size and internal structures; such as coarse grained turbidites or called high density turbidites (Lowe, 1982), which correspond to the distinct grains or clasts which are deposited differently from either individually or collectively. It has also other name like slurry flows (Lowe and Guy, 2000), concentrated density flows (Mulder and Alexander, 2001) and sandy debris flows (Shanmugam, 1996), and fine grained turbidites or called low density turbidites (Piper, 1978; Stow and Shanmugam, 1980) which correspond to the silty to muddy of Bouma division (Td-e), which are believed to have its own scale or order.

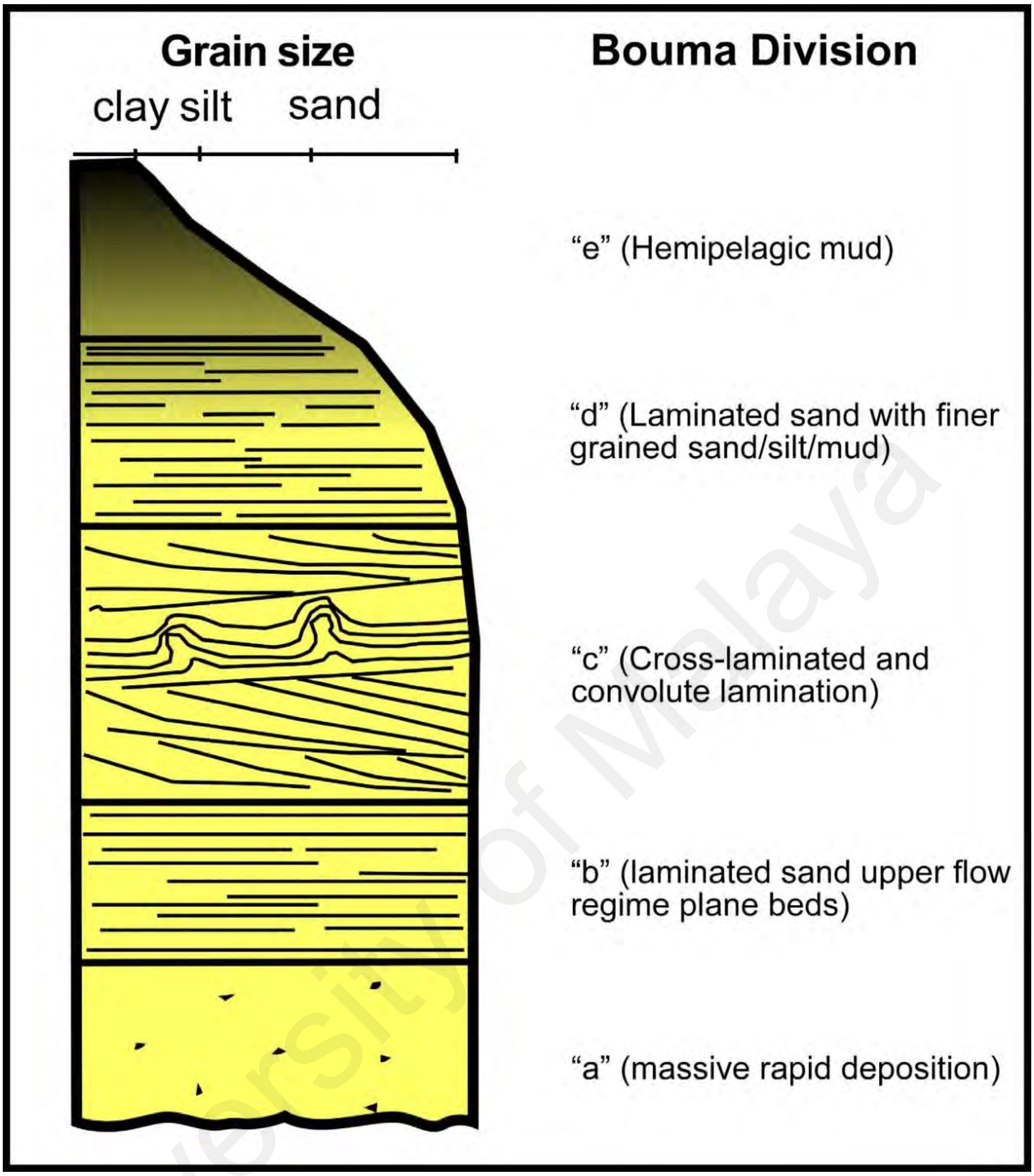


Figure 2.2. A completely ideal classic turbidite sequence (modified from Bouma, 1962)

Turbidite facies models can be divided into 2 types, based on the grain size and feeder-system type. They are: (1) sand-rich, point-sourced systems and; (2) mud rich, point-sourced systems (Reading and Richard, 1994). Similar models were also proposed by Bouma (1997) and Bouma et al. (1995), but given different names:(1) sand-rich trubidite system and (2) fine-grained, mud-rich, turbidite system. The grain size connotation refers to the general sand:mud ratio in the turbidite submarine fan. Sand- and mud-rich refer to the whole turbiditic succession which may be deposited on top of several fans separated by thick shales. The sand-rich system is associated with active margin settings whereas mud-rich system is associated with passive margin settings (Shanmugam and Moiola, 1988).

A submarine fan is a body of sediment, deposited at the sea floor that is deposited by mass flow, mainly from turbidity currents and have fan-shaped, channelized and sheet-sand complexes (Reading and Richards, 1994). Coarse grained/Sand-rich submarine fans have very high sand:mud ratio, have thinner and sand-rich shale between the individual fans than the regular component fan of the sedimentary complex. The source of sediment is relatively close to the shore, where coastal plain is not well developed, where deltas are rather small and often have a little or no space to switch back and forth. Long-shore drift transports most of the sediment that reaches the shoreline. Sand accumulating against a promontory or in an offshore-directed swale tends to move seawards due to the returning waves. The transported sediment at the bottom will gradually carve out submarine canyons. The sands in offshore basins are commonly medium to fine-grained. Coarse-grained fan systems may be semi-confined as they gradually build out, showing a down-dip decrease in grain size, thickness and sand/mud ratio.

Fine grained/mud-rich submarine fan systems are very sand rich, but are separated by thick accumulations of basinal shale, which are characterised by thin, very fine-grained sandy and silty turbidite with interbeded thin clayey/mud and some plant debris. The end member fan is most common in passive margin setting. The source of sediment is relatively far from the sea. This fine-grained submarine fans system is typically formed and influenced by a long and low-gradient fluvial system, abroad coastal plain with a major deltaic system, a wide shelf, and generally a large deep-water basin with comprises of other sub-basins (Bouma, 2001).

The deposition of sand in the basin is affected by sea level changes due to the climate change. River-transported sand has the tendency to settle down at the delta due to frictional deceleration and is difficult to transport beyond the continental shelf (Bouma, 2001) and onto the continental slope and basin, as deltas prograde directly on top of the shelf-edge (Porebski and Steel, 2003). This results in more rapid deposition of sand in deeper waters. Rapid deposition results in instability along the continental shelf edge and slope, and can result in sliding, slumping debris flow and turbidite deposition (Bouma,2001).

2.3. DEEP MARINE ENVIRONMENTAL ARCHITECTURE

2.3.1. Introduction

Deposition of deep marine sediments in turbidite system is mainly controlled by four factors interacting with each other, i.e. tectonics, climate, sedimentary characteristics and processes, and sea-level fluctuation (Bouma, 2004). A turbidite system is a body of genetically related turbidite facies and facies associations that were deposited in virtual stratigraphic continuity (Mutti and Normark, 1987, 1991). A turbidite system is either coarse grained or fine grained, represents an individual fan that consists of alternating sand and mud layers, and is characterised by a relatively high sand/mud ratio (Mutti and Normark, 1987, 1991).Reading and Richards (1994) classified turbidite systems based on morphology, recognizing a point source fan, multiple source ramps and line-source aprons **(Fig. 2.3).** They also classified turbidite systems based on grain size into; Gravel-rich, Sand-rich, sand and mud, and mud rich. Gravel-rich and sand-rich turbidite systems tend to be small in radius and laterally grade rapidly basin ward into fine grained deposits. Sand and mud rich turbidites are larger (tens to hundreds of km wide) compared to the gravel-rich and sand-rich deposits and show systematic changesintheir internal stratigraphy. The mud-rich deposits are the largest type of turbidite system and the most voluminous. Mud-rich turbidite systems range between tens to few thousands of kilometers in width. This grain size different also affect size, sediment supply mechanism as well as channel system between a point source fan, multiple source ramps and line-source aprons. The size of fan system within the single point source fan is larger, more wide-spread and radial, as well as more elongated, while the others are generally lobate and linear belt. Single point source fan also dominated by low-high density turbidites, with infrequent slumpings and are river generated as well as rework from shelfs. In term of the sediment supply mechanism multiple source ramps almost similar to the single point source, but only more frequent of slumpings. While the line-source aprons are majorly dominated by slumps, with minor turbidites. In term of channel system, single point source is the largest in channel system. Comprising with well developed levee and generally meander, except in the gravel rich. Multiple source ramps also develop channel and levee but not as large and well developed as single point source. The line-source apron is dominated by multiple and dominant chutes, with poorly developed channel.

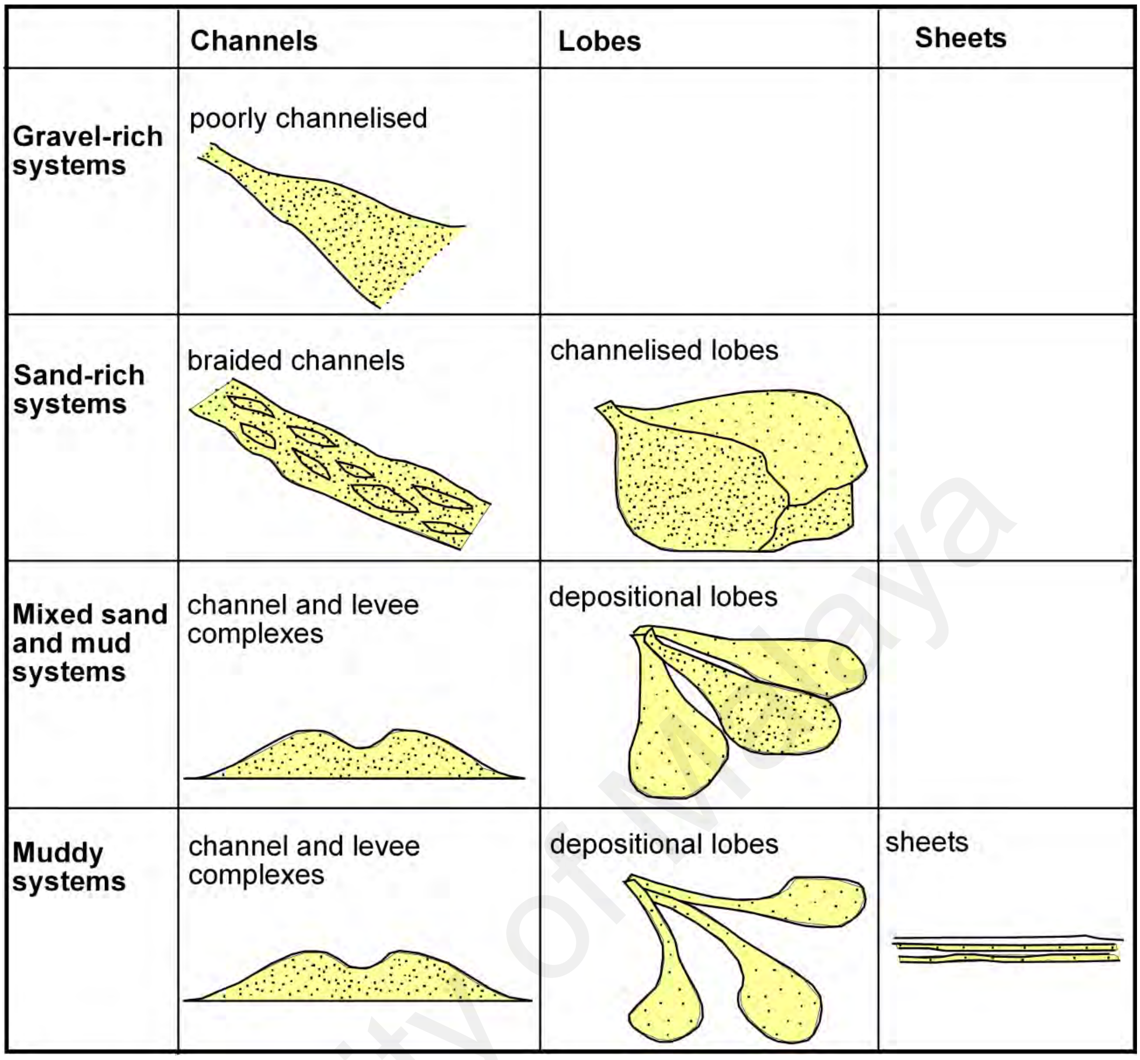


Figure 2.3. Submarine fan architectural different based on grain size dominant (modified from Reading and Richard, 1994)

Turbidite systems **(Fig. 2.4-2.7)** have also been generally described as deep sea fans, because of their common semi-conical shape. However, modern deep sea fans are observed to be elongated or irregular in shape, thus making the more generalized term "turbidite system" more suitable (Bouma et al., 1985). Pirmez (2000) divided turbidite systems into 4 zones, based on the spatial patterns of sediment erosion and deposition. They are submarine canyon, upper fan, middle fan and lower fan. The submarine canyon is dominated by erosion. The upper fan is a sedimentary zone of bypass. The middle fan is the zone of deposition from flow expansion from sediments that have been expelled from flow confinement. The lower fan is the zone of deposition, which is the continuation from middle fan, where the rate of deposition is gradually decreased.

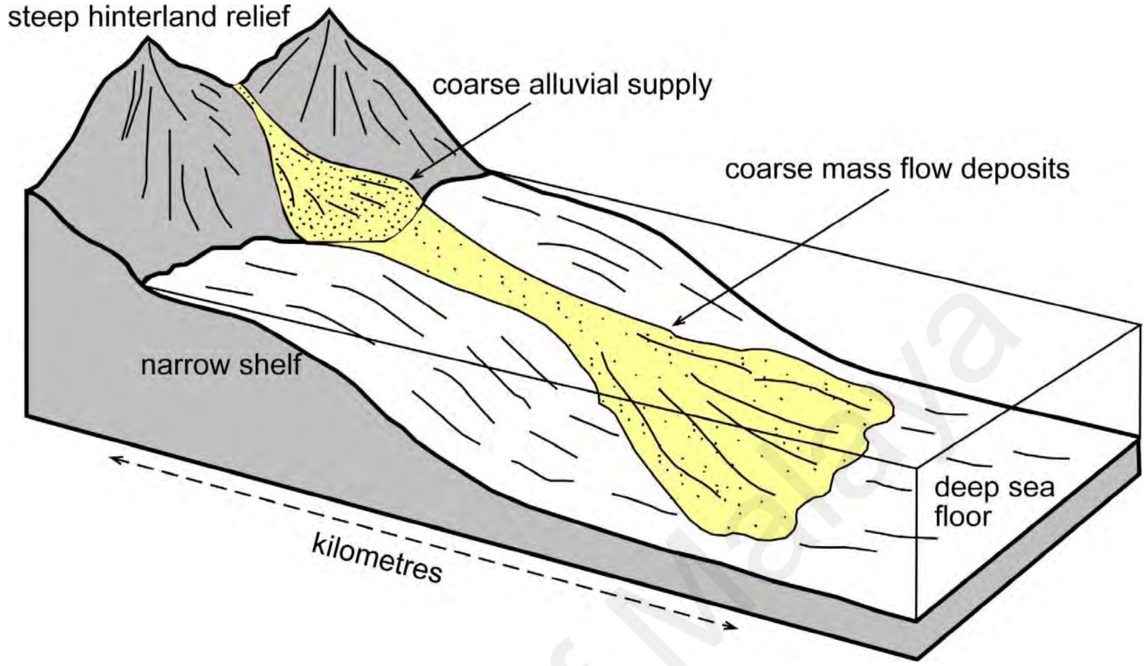


Figure 2.4. Gravel rich submarine fan model: typically found infront of coarse fan delta. It has small fan, associated with debris flow (modified from Reading and Richard, 1994)

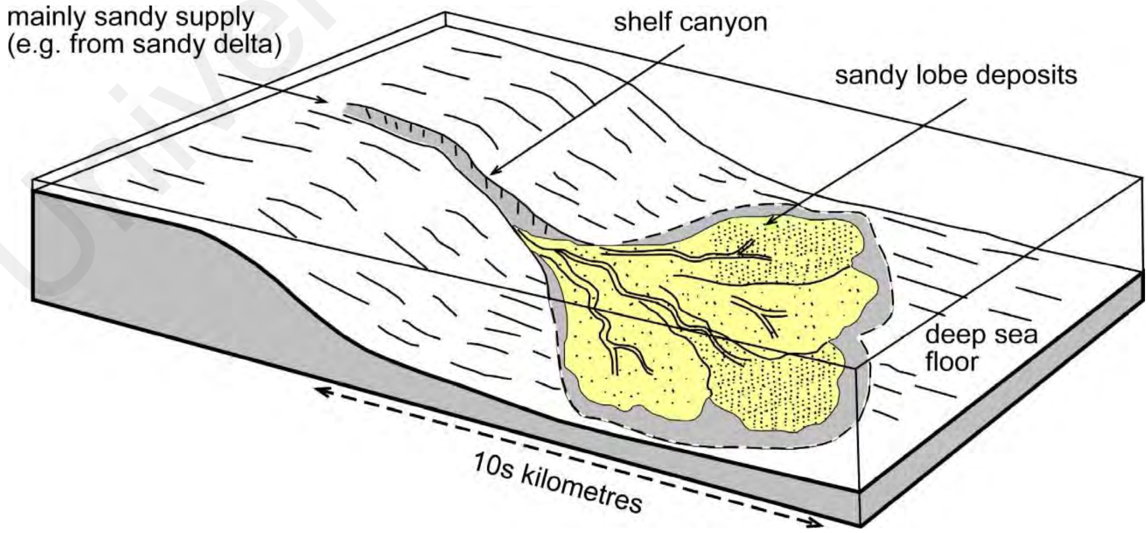


Figure 2.5. Sand rich submarine fan model: sand rich turbidites form lobes on the basin floor and switching of locus of deposition through time. (modified from Reading and Richard, 1994)

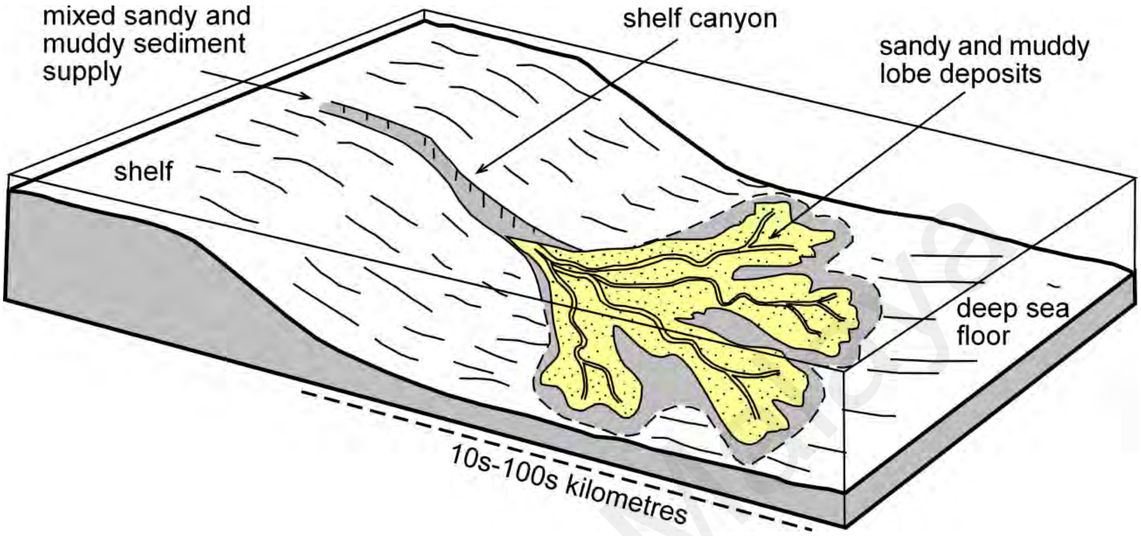


Figure 2.6. sand and mud submarine fan model: the lobes are mixture of sand and mud, and extent further out as turbidites travel longer distance (modified from Reading and Richard, 1994)

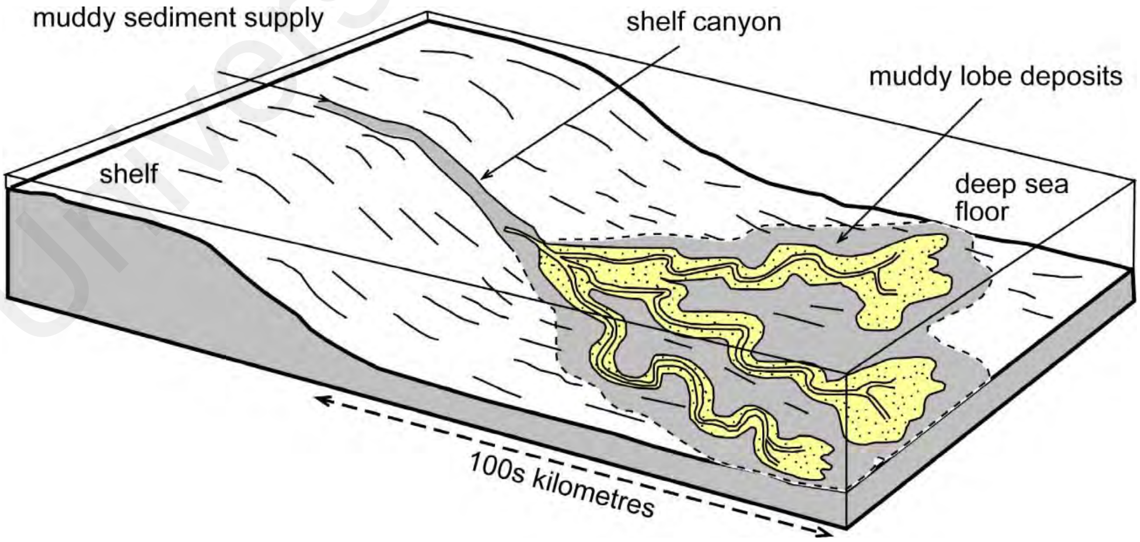


Figure 2.7. Mud rich submarine fan model: very extensive and sand is deposited near by the channel (modified from from Reading and Richard, 1994)

2.3.2. Channels

In deep marine turbidite systems, channels are formed from confined turbidity currents, which transport sediments along a major, long-lived pathway. These channels act as a place where erosion, bypass and deposition occur and are controlled by the characteristics and boundary conditions of the systems. Deep marine channels are similar to fluvial channels, which always search for a longitudinal profile graded to a base level, in this case gravity base (Posamentier and Walker, 2006), where the flow becomes unconfined at the up current end of the terminal lobe (Primez et al., 2000).

There are three kinds of deep marine channels **(Fig. 2.8)** based on studies of ancient and modern systems (Pickering et al., 1995).

A. Erosional channels: Channels bounded by a scoured surface and which truncates older strata.

B. **Depositional channels**: Channels bounded by well-developed channel margin levees and, as time passes, eventually become elevated above surrounding seafloor.

C. Mixed channels: Channels experiencing booth erosion and deposition. Mixed channels show a combination of levee deposition and channel axis erosion. The channel floor may lie above or below the adjacent seafloor.

Deep marine channels can also be classified based on the degree of confinement (Posamentier and Kolla, 2003). More confined channels are present up-flow of turbidite systems; this includes canyons and erosional channels. Erosional channels are replaced by leveed-channel systems further basinward. Channel levees form due to sediment expulsion from the top part of the flow that escapes the confined channel, creating channel-building-levees (Pickering et al., 1995). The levee system is succeeded down-flow by a complex of poorly confined channels associated with lobate sedimentary bodies, commonly known as lobes or fans, at the downstream end of the channels.

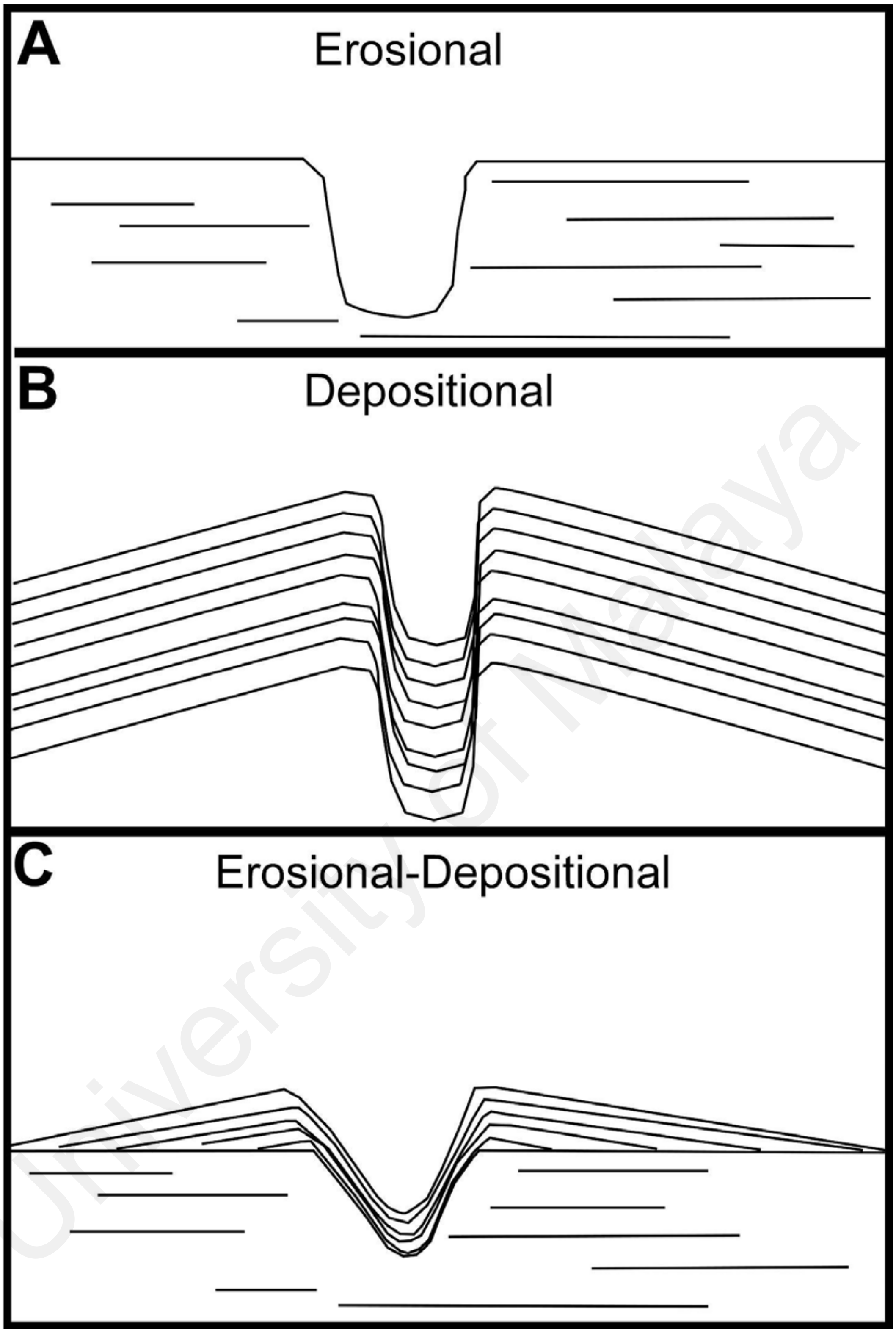


Figure 2.8. Nomenclature of deep marine channels (redrawn from Pickering et al., 1995)

2.3.3. Submarine Canyons

Submarine canyons can reach up to 2.5 km deep and 100 km wide (Normark and Carlson, 2003). The fill is stratigraphically complex and lithologically variable. In part, it relates to temporal and spatial differences in sediment source, which varies from canyon-wall-collapse, up dip feeder system sampling a shelf and/or continental sediment source (Galloway et al., 1991).

Submarine canyon fills dominated by local canyon-wall-collapse deposits are characterised by fine grained sediments, which is common in mud-rich mass movement and debris flow deposits (e.g. Galloway et al., 1991). Fine grained sediments are commonly deposited from suspension as drapes infilling part or the entire canyon. Coarse sandstone and conglomerate deposits occur as isolated elements in mud rich fills,but can be the dominant facies in canyons of tectonically active environments. Coarse sediments occur as thick bedded, structureless due to high concentration of turbidity current.

2.3.4. Erosional Channels

Erosional channel fills are characterisedby sand-rich and gravel-rich strata deposited by turbidity currents and frictional flows. There are four common stages to most models of erosional channel development (Samuel et al., 2003);

a. **Channel inception** : It is characterised by a period of successive flows with high transport efficiency that scours-out a through-going topographical feature that serves as the conduit for later flows. The transported sediment during this stage is brought farther basinward and deposited in more distal areas.

b. **Sediment bypass** : This stage succeeds the channel inception stage. Flow may display either complete bypass (no more erosion) or incomplete bypass. Incomplete bypass deposits are characterised by intercalation of laterally discontinuous coarse and fine grained sediments, the discontinuous nature of the deposits is due to the erosion of subsequent flows. Patchy occurrence of tractional sedimentary structures, dune cross stratification and coarse grained lags as well as thin drapes of fine grained sediments is also typical of this stage.

c. **Channel filling** : This stage is where transport efficiency is low. This initiates sediment deposition which can eventually fill a part of or the entire channel.

d. **Channel abandonment** : This stage is formed as a result of diversion of flow at the upstream point (avulsion) or decrease in flow that is probably due to sea level rise. The channel system is abandoned and fine grained deposits are laid down. These processes are repeated episodically, particularly the channel filling stage which temporarily rejuvenates the system, although the duration may vary.

2.3.5. Leveed Channels

A Leveed channel typically displays a sinuous planform with levee development along their margins. Channel fill strata terminate abruptly along a basal erosion surface defining the outer bed margin of the channel. The erosionally cut strata are either genetically related to the levee deposits or related to an older channel. Along the inner bend of the channel, the strata either grade continuously or stack on the older channel fills in highly confined channels. Levees and levee deposits are best developed along the outer bend of the channel due to the inertia of the fluid and result of continuous current along a straight line while the channel floor bends beneath it. Based on the channel confinement, leveed channel are divided into 2 types (Arnott, 2010).

2.3.6. Poorly confined leveed channels:

2.3.6.1. Channel deposits

The characteristics of a base poorly confined leveed channel are commonly asymmetric in cross-section, with steeper margin at one of the side. Often, the channel is characterised by a staircase-like geometry or called "step-flat" morphology, which indicate that the entire channel system has laterally undergone episodic, step-like lateral migration (Navarro et al., 2007). At the steep margin of the outer bend, the levee strata are either in erosional contact with adjacent channel strata or separated by a thin, fine grained unit of bypass deposits (Beaubouef et al., 1999). On the other hand, along the point bar of a sinuous fluvial channel, the channel-fill strata are either on-lap or grade laterally into levee deposits. The leveed channel fills consist of deposits of many smaller channels. Each individual channel can range from 1 to 10s of meters thick and 100s meter wide. The channels show a well-developed fining and thinning upwards vertical trend. The base of the leveed channel is characterised by the presence of coarse grained sediments and mudstone intraclasts. Occasionally, coarse amalgamated strata grade laterally into finer and more stratified deposits. Within the channel axis, strata are characterised by thick to very thick bedded, coarse-tail, normally graded or structureless, amalgamated conglomerate or sandstone beds, with lateral continuity between 10s to 100s meters.

The coarse grained, gravel and sand rich deposits represent Bouma (1962) division T-a. Channel margins with higher leveed channel fills are characterised by thinning of coarse grained deposits, grading from amalgamated to less amalgamated, which are more of thin to thick bedded sandstone beds characterised by T-a and T-b turbidites which are intercalated with thin bedded T-cd and T-cde of Bouma (1962) division, and intervening with mudstone (Arnott, 2010).

2.3.6.2. **Levee deposits**

The upward build-up of subaqueous levee deposits is due to the sediment flowing overtop the margin of the channel. The flow-overspill and levee aggradation can occur in three ways; Flow stripping, inertial overspill and continuous overspill (Arnott, 2010).

a. **Flow Stripping** : Flow stripping occurs at the channel bends, preferentially along the outer bank. It occurs when the fine grained sediment of the upper part of the flow becomes separated from the lower, coarse grained part that remains confined to the channel.

b. **Inertial overspill** : It occurs at the same place as the flow stripping but differs in that it occurs when an energetic flow is unable to follow the sinuous thalweg and runs up the channel margin, causing even the lower parts of the flow to escape the channel

c. **Continuous overspill** : Continuous overspill occurs where the thickness of the flow exceeds the depth of the channel, leading to loss of the flow above the height of the levee along both sides of the channel.

Levee growth along the inner bend and straight segments between channel bends is by continuous overspill (Posamentier and Walker, 2006). The flow escapes and expands rapidly resulting in elevated rates of sedimentation adjacent to the channel and rapid decrease in deposition away from the channel. A ―gull wing" pattern/profile is developed due to the lateral variation in average sedimentation (Hiscott et al., 1998). As the levee aggrades, only the fine and dilute portion of the flow will escape and overspill, thus creating thinning and fining upwards trend.

Levee deposits along the channel outer-bend consist of thick sand-rich sediment, while inner-bend deposits are finer grained and thinner. Palaeo-current measurements from Tc turbidites are generally obliquely-oriented to the main channel flow (Khan and Arnott, 2011; Hickson and Lowe, 2002).

Along the channel outer bend, the characteristics of the sediments are differentiated based on where they are deposited. The proximal part of the levee is characterised by thin to medium bedded, fine to medium grained sandstone. It is classified as Bouma (1962) Tb turbidites interstratified with thin bedded, fine grained sandstone or siltstone (Tbcd) and thick sandstone (Ta) turbidites. The distal part of the levee is characterised by thin bedded, very fine to fine grained sandstone or siltstone. These represent Tcde turbidites and the levee strata are intercalated with overbank or crevasse splay deposits (Arnott, 2007b).

Along the inner bend, the sediments are thinner and finer grained compared to the outer bend, and characterized by fine sandstone and siltstone classified as Bouma (1962) Tcde turbidites. Locally, the strata are interbedded with thicker, medium to thick bedded, fine to medium grained sandstone classified as Tc turbidites and have 3 or 4 sets of ripple cross stratification within a single bed. Inner bend strata are continuous with and grade laterally into channel fill strata, indicating a continuum between channel and levee deposits (Kane et al., 2007).

2.3.7. Highly confined leveed Channels

Highly confined leveed channels are much smaller in dimension compared to poorly confined channels, with widths of 10s to 100s meters and depths of 10s meters. The channels display higher sinuosity and develop disconnected channel fills (Arnott, 2007a). The channel fill deposits are locally clustered laterally or vertically. The base of the channel is generally horizontal and planar, although small localised scours are present. in places. However, step-like geometry, rising abruptly upward by several meters is observed. The top of the sandy channel fill interfingers with thinly bedded turbidites. A distinct and consistent inclination are shown within the channel fill strata forming 7 and 12 degrees and are interpreted as lateral-accretion deposits (LADs), which are similar to the features in meandering fluvial systems **(Fig. 2.9**). The channel fill is divided into a lower and upper part (Arnott, 2007a).

The lower part of the channel fill is characterised by thick to very thick, amalgamated beds. It comprises sharp-based, graded sandstone and less common granule or fine pebble conglomerate. The channel has a scoured based which erodes underlying beds. Mudstone appears as localized patches of intraclasts. The sand/gravel rich strata display an upward thickening pattern in bed thickness.

The upper part of the channel fill consists of mudstone and thinly bedded turbidites interbedded with the coarse grained sediments. The coarse grained sediments abruptly terminate at the slope of the channel bank, which is marked by gradual lateral trend in poorly confined-leveed channels. At the upper end of each bed of coarse grained strata, beds are normal graded, poorly sorted and capped by dune cross stratified or planar laminated sandstone (Arnott, 2007a). At the bottom end part of the coarse grained beds, fine grained beds are intercalated between the coarse grained beds. Nearby the end of the beds (right direction of Fig. 2.9), almost complete Bouma sequences are visible, up dip, thin, fine and eventually dominated by Tcde turbidites.

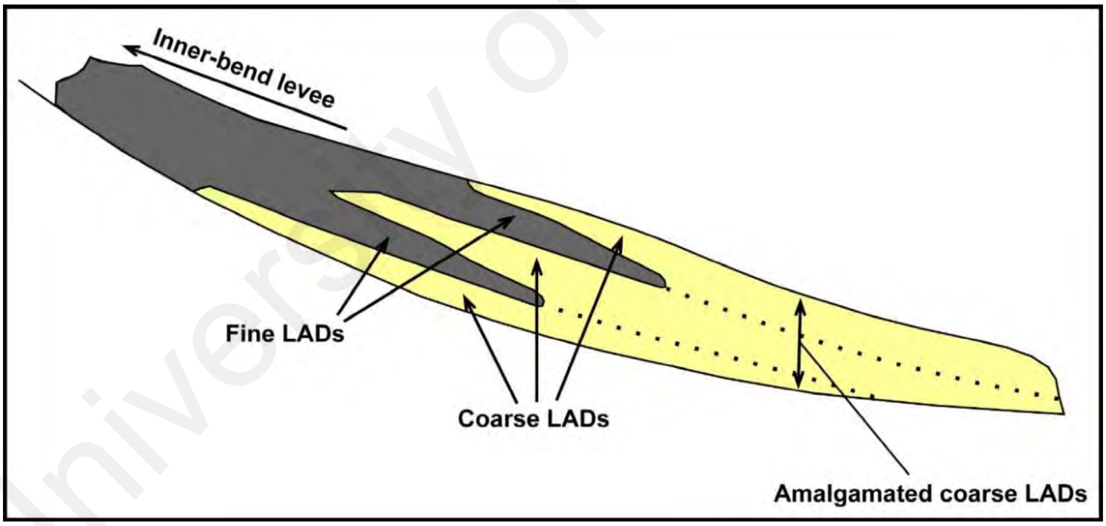


Figure 2.9. Idealized model of lateral-accretion deposits (LADs) formed by laterally migrating deep marine sinous channel (modified from Arnott, 2007a)

Along the inner bend levee part of the channel margin, the fine grained beds thin and fine abruptly beyond the terminus of coarse grained LADs. This represents the fine grained sediment at the inner bend levee part, where the coarse grained beds on-lap. Whereas the outer bend, coarse grained LADs terminate abruptly at the fine grained, thin bedded turbidites of a slightly older levee system. The interbedding of sand and mud at levee channels is caused by alternating periods of effective and ineffective turbidity flow, which bypasses the channel bends and transport sediments down the slope of levee channels.

2.3.8. Overbanks and Crevasse splays

Overbank splays are lobate and sheet like features formed by energetic flows that overtopped and escaped the channel in an unconfined manner, while crevasse splays are large scale, lobe shaped features formed after the downflow of a crevasse channel that is incised into the channel margin and proximal levee (Arnott, 2007b). Crevasse splay strata successions can be up to meters to tens of meters thick, consisting of decimeters to meters thick units of medium-thick bedded structureless sandstone with mud intraclasts (Arnott, 2007b) These strata are interbedded with Bouma Tcde turbidites. The structureless sandstone is characterised by poorly sorted and coarse-tail grading with a sandstone, siltstone and mudstone matrix, and are interpreted as having been deposited rapidly by capacity-driven deposition immediately downflow of an area of rapid flow expansion.

A crevasse splay is connected to the breach of the adjacent parent channel at its head-ward end by a crevasse channel. A crevasse channel fill is characterised by a distinct sharp based, coarse grained unit within a background of fine grained levee deposits. Crevasse channel deposits are thin up to a few meters thick and consist of amalgamated thick bedded, massive or normally graded sandstone or conglomerate and single-set-thick, discontinuous cross stratified dunes (Arnott, 2007b).

Overbank splays are formed from large magnitude, high concentration of coarse grained turbidity currents which overspill the unconfined channels. The beds are characterised by single to multiple beds with thicknesses of 2-5 meters, that possess internal units of amalgamated, thick bedded, medium grained sandstone displaying complete Bouma sequences (Arnott, 2007b).

2.3.9. Basin Floor deposits

Basin floor deposits are sediments that are deposited in water depths of more than 2000 m (Ailsa Allaby and Michael Allaby, 2015). Proximal parts of the basin floor are characterised by terminus of leveed channels which are characterised by a thick, laterally extensive sediment body (known as lobes), distributary channel complex, sand-sheet deposits, or frontal splay complex, where sediments escape due to the change of environment from confined to unconfined and depositional channels. The lobes can be tens of meters thick and 100 km wide (Posamentier and Walker, 2006). Further basinward, the beds become thinner and finer and the gravity flow sediments are eventually replaced by hemipelagic and pelagic sediments.

The composition of the transitional part from the channel to lobes depends on grain size. In coarse grained systems the channel connects directly with the lobes whereas in finer grained systems, channel and lobes are separated by a zone marked by large scours, sediment waves and sediment mounds (Posamentier and Kolla, 2003). But this transition zone could be also interpreted as sediment bypass and deposition.

2.3.10. Depositional lobes

Depositional lobe size and geometry varies with sand:mud ratio. Sand- and gravel-rich type depositional lobes tend to be area restricted due to rapid deposition. They form elements of a few kilometers wide and tens of meters thick which decrease in thickness as well as grain size rapidly down-flow. In mud-rich depositional lobes, the lobes are considerably wider than in sand rich types and form sheet-like features. These sheet-like features become thinner and finer down-flow but the change is more gradual. The stratigraphic architecture is more complex compared to the sand rich type. Lobe hierarchi consist of 3 architectural elements (Arnott, 2010);

a. **Deep channels** : Deep channels display incisions up to 10 m deep with locally steep margins. The channel fill is characterized by a heterogenous assemblage of coarse and fine grained strata related to incomplete bypass and thick-amalgamated sandstone beds (Johnson et al., 2001). It is a point of where sediments are being supplied to the depositional lobe complex.

b. **Shallow channels** : Shallow channels have minor relief with scours only a few meters deep, but can extend 100 m laterally. Shallow channels consist of an aggradational internal stratigraphy. Within the channel axis, there are thick to very thick amalgamated sandstone beds (5-10m) displaying Bouma Ta divisions, which gradually and show systematic change from thick to medium bedded of complete turbidites that increasingly thinner, finer and less complete turbidites, which eventually turned into single-set, thin bedded fine grained Tcde turbidites.

c. **Sheet-like splay deposits**: These deposits are composed of amalgamated sandstone that can reach several kilometers wide and display a random internal architecture.

2.3.11. Contourites

Contourites are sedimentary deposits which are products of thermohaline-induced deep water bottom currents and can be also be influenced by wind or tidal forces(Hollister, 1993). Contourites are normally formed in continental rise to lower slope settings, however they may occur in shallow waters, anywhere below storm wave base. These currents generally flow parallel to the slope. The flow can locally be diverted obliquely up or down the slope, due to topographical effects. These contourites are discovered to be the fundamental in large dimension within deepwater deposits, ranging from tens to hundreds km long, a few km wide and over 1 km thick, elongated, slope-parallel sediments bodies which are known as sediments drifts (Stow et al., 2002a)

Contourite characteristics are controlled by sediment supply (Stow et al., 2002b), which can comprise terrigeneous clastic, volcaniclastic or carbonate sediments. The grain size ranges from sand to mud and can be a mixture of both, but is most commonly mud and silt. Gravel sized particles are typically restricted to local areas that have high energy and attendant seafloor reworking and winnowing. The grain sorting is generally moderate to good, but otherwise in areas where the current speed is low and slow sedimentation. This enables the activities of organisms which mix the deposits and destroy the primary sediment layering. Bioturbation is dominated by *Nereites* ichnofacies and can be moderate to intensely bioturbated (Stow et al., 2002b). Traction structures are common and, depending on the flow speed and grain size represented by small or large scale cross stratification and also scour marks (Stow et al., 2002b) and paleo-currents are not strictly parallel to the slope. Contourites occur as composite units of 20-30 cm thick. Contourite successions are generally characterised by a basal coarsening upward interval which consists of muddy to silty to sandy contourites. This is overlain by fining upwards to muddy units of contourites (Stow et al, 2002b). This upward change indicates a systematic temporal change in flow speed or rate of sediment supply, which, based on Stow et al., (2002b), suggests that it occurs on time scales that closely parallel Milankovitch periodicities, suggesting a relationship between orbital forcing of climate and changes in bottom current velocity.

It is difficult to differentiate between turbidites and contourites **(Fig. 2.10).** However contourites can be identified based on the presence of reverse grading with cross-ripple stratification and intercalated with mud layers or drapes with ripple cross stratification of sandstone/siltstone units (Ito, 1996). Other than that, contourites show internal erosion surfaces. All these evidences indicate that there were fluctuations of bottom current speed **(Fig 2.11)**. Another interpretation is that contourites can also be formed from oscillation between traction and suspension sedimentation (Ito, 1996).

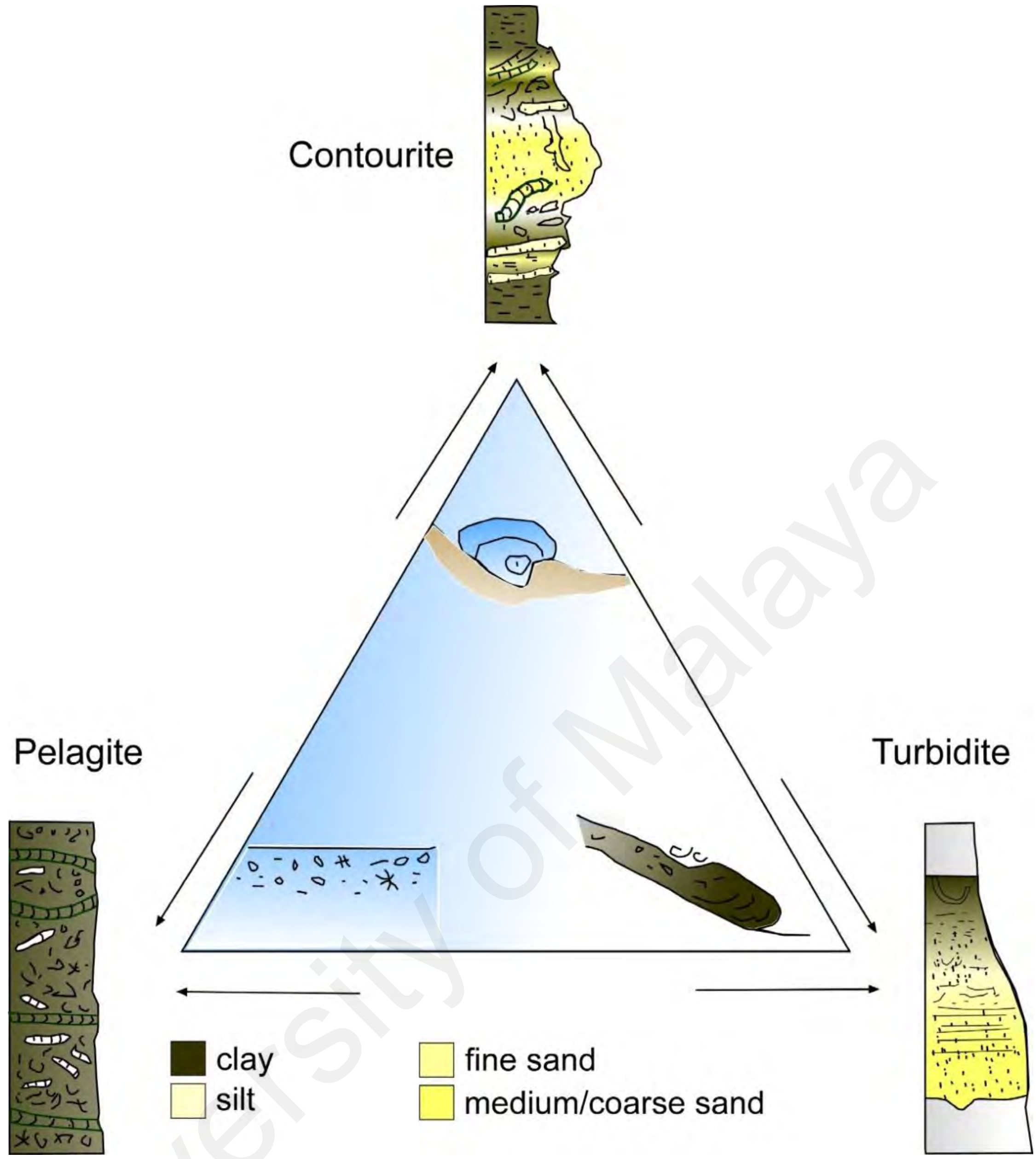


Figure 2.10. Three main types of conceptual diagrams of sedimentary processes in deep water and the facies models of the respective depositional (modified from Rebesco et al., 2014)

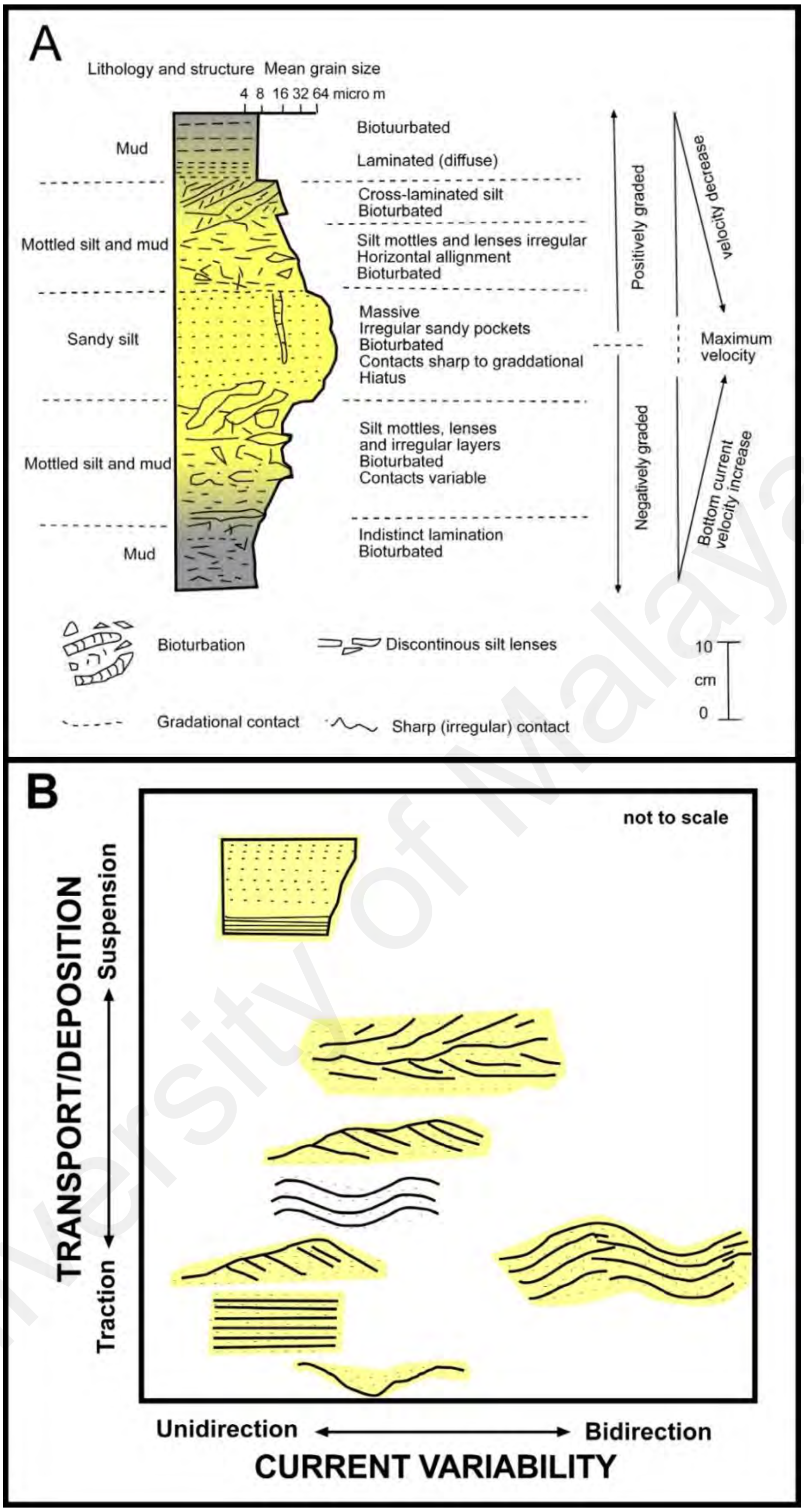


Figure 2.11. A) characterise of idealised contourite deposits (Gonthier et al., 1984; Rebescoe et al., 2014). B) Small scale characteristics of contourites deposits (Ito, 1996). Temporal variations in the speed and/or direction of the contour current cause many of the contourites structures to resemble features that occur in tidal deposits, although complete reversals are generally absent in contourite deposits

CHAPTER 3: FACIES CHARACTERISTICS OF THE PALEOCENE-MIDDLE EOCENE BELAGA FORMATION

Sixsedimentary facies are identified in the Belaga Formation logged sections near Sibu Jaya, Sibu, Sarawak. These lithofacies are defined based on variations in lithology, grain size, sedimentary structures and degree of bioturbation **(Table. 3.1):**

1) Mudstone (F1)
2) Thin graded sandstone (F2)
3) Medium graded sandstone (F3)
4) Thick graded sandstone (F4)
5) Structureless sandstone (F5)
6) Hybrid beds (F6)

3.1. Facies 1: Mudstone (F1)

Description

Facies F1 **(Fig. 3.1A-B)** is characterised by dark coloured mudstone up to 100 cm thick, displaying a planar bedform and sharp contacts with the overlying and underlying strata. The mudstone is commonly laminated. Also present are occasional intercalated layers of sandstone (<1 cm thick). F1 commonly grades upward into F2. The degree of bioturbation is generally low (Bioturbation Index, or BI, of 2) and mainly in the form of *Chondrites* and *Paleophycus* **(Fig 3.6B and D).**

Interpretation

Facies F1 is interpreted as pelagic or hemipelagic deposits. The sharp contact with overlying and underlying facies suggests post-flow background deposition. The mud was periodically deposited between turbulent flow events which provided sufficient time for bioturbation.

3.2. Facies 2: Thin graded sandstone (F2)

Description

Facies F2 **(Fig. 3.1C-E, 3.2A-D)** is characterised by thin (1 to 10 cm thick), tabular beds of very fine to fine grained sandstone commonly intercalated with 5-15 cm thick beds of F1 (sand/mud ratio between 0.7 to 0.5). The sandstone beds are sharp-based, tabular and commonly display normal grading from fine sand into silt with a sharp top. The graded beds display incomplete Bouma sequences, which include Bouma Tbcde and Tcde variants. Sandstone beds less than 3 cm thick tend to display ripple cross-lamination (Tc beds). The sandstone is sharply overlain by F1 mudstone, with some examples displaying intervals of starved ripples. The degree of bioturbation is sparse to moderate (BI 0-3) and is mainly in the form of *Paleodictyon, Cosmorhaphe, Thalassinoides and Chondrites* **(Fig. 3.6A, C, E and F).**

Amalgamation of 2-4 Tc beds forms 3 – 10 cm thick units. Td beds are rare, but when present, display sand/silt/mud lamination, with no gradation of grain size. Bouma Tb divisions are common, identified by interstratification of sand and silt within beds having thickness from 10-15 cm.

Interpretation

Facies F2 is interpreted as turbidites deposited from dilute, low energy, non-erosive and low density turbidity currents, based on the thin bedding, fine grained texture, and common absence of the lower Bouma divisions and the sharp but non-scouring base of the beds (Lowe, 1982; Talling et al., 2012). The presence of starved current ripples suggests reworking by later muddy flows that contain little sand (Talling et al., 2007, 2012). The origin of the Td division is still controversial (Talling et al., 2012). However, sand/silt/mud lamination in the Td division of F2 suggests that it was probably formed by rhythmic changes in flocculation/deflocculation and sediment settling within the viscous sublayer of a low energy turbidity current (Stow and Bowen, 1980).The rarity of the Td division in some beds was probably due to the presence of internal waves within the channelized turbidity current. This could have produced fluctuations in the hydraulic regime and grain size make-up of the over-spilling flow (Gervais et al., 2001; Hiscott et al., 1997). An alternative interpretation is that suspension occurred at the ending flow event, aggregating coarser grains with the suspended sediments, resulting in the deposition of a structureless layer by settling/fallout manner, rather than lamination, due the absence of flow.

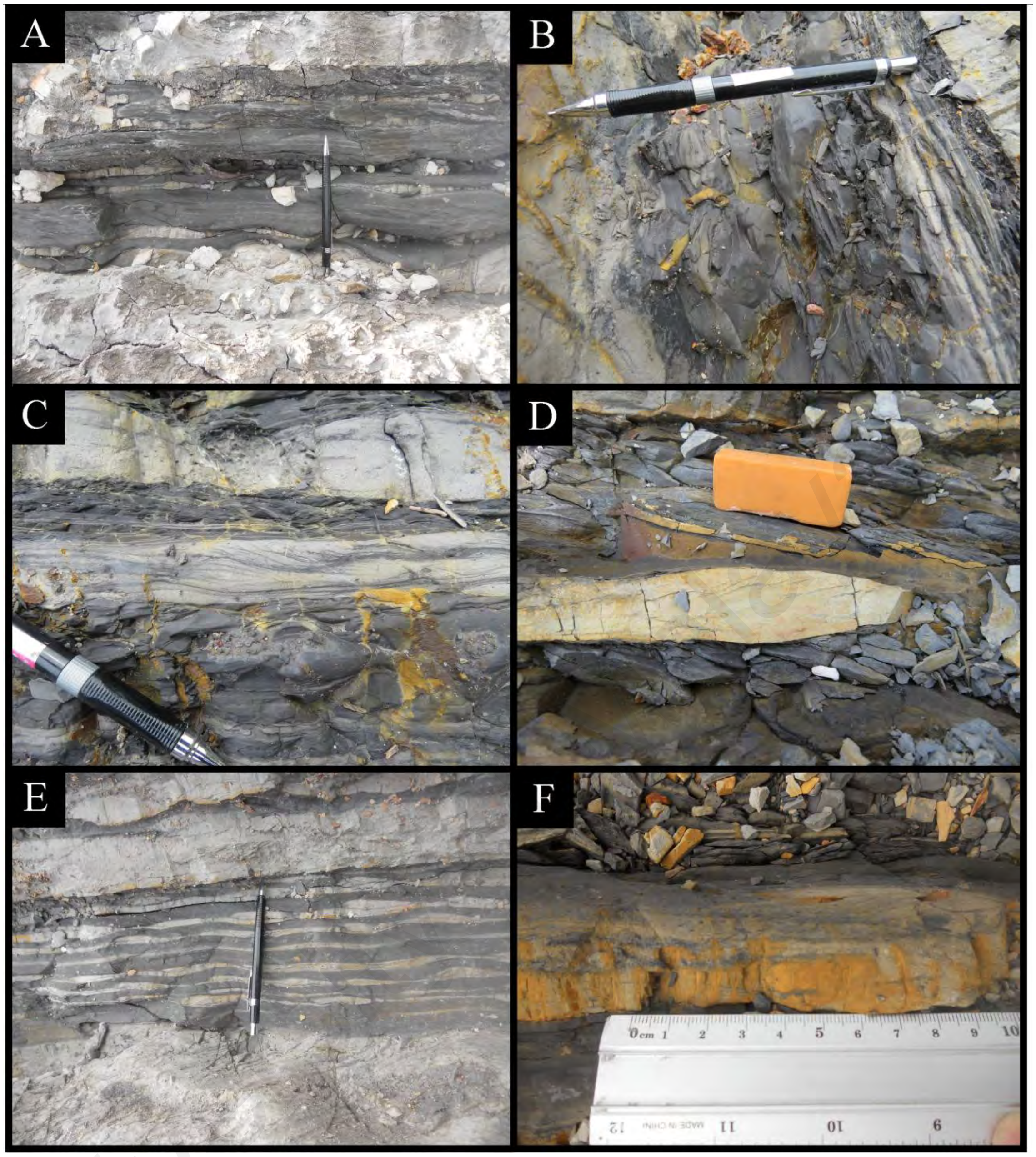


Figure 3.1. Photographs of mud (F1) and thin graded sandstone (F2): A) and B) display mudstone beds (F1) containing <10 mm thick sand lamination. C), D), E, F) typical of thin bedded sandstone (F2) occur as packaging, display cross ripple lamination. F) Thin bedded sandstone shows disturbed structures due the presence of trace fossils.

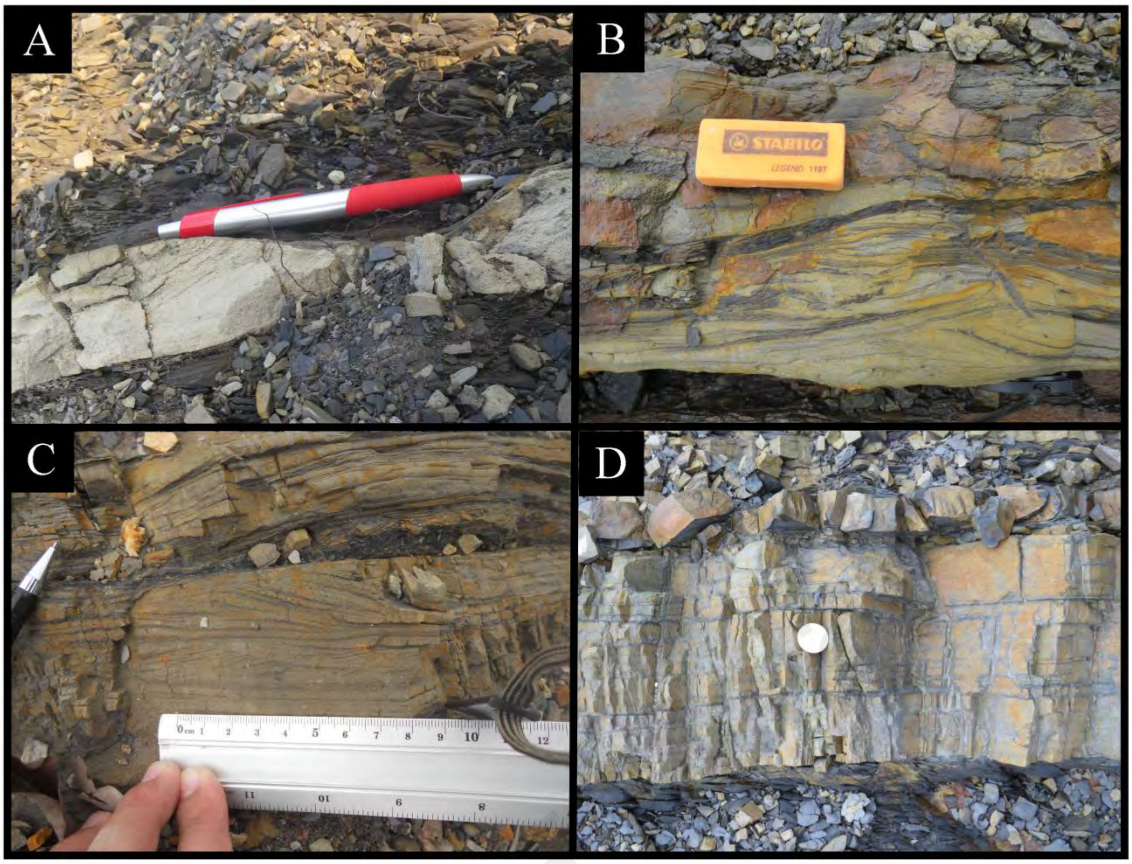


Figure 3.2. Photographs of thin graded sandstone (F2). A) Thin bed with sharp bases and undulating surface, containing ripple cross-lamination. B) Thin graded sandstone display Bouma (1962) Tc-e. C) Sandstone bed displays bidirectional cross-lamination. D) Stacked of sandstone beds, where each displays cross lamiantion.

3.3. Facies 3: Medium graded sandstone (F3)

Description

Facies F3 **(Fig. 3.3A-F)** is characterised by 15-30 cm thick, grey in colour, tabular beds of poorly sorted, medium to fine grained sandstone. Individual beds are sharp-based and display normal grading. The graded beds typically display Bouma sequences, mainly in the form of Tbcd sequences. However, some rare examples have a basal Ta division (Tabcd). Several beds of Facies F3 can form stacks up to 40 cm thick. Sedimentary structures resembling hummocky cross-stratification are observed in the Tb division of some F3 beds. Vertically adjacent ripple cross-lamina sets in the Tc division of some examples of F5 also display opposing dip directions (bidirectional ripples).

Interpretation

Facies F3 comprises Bouma (1962) Tb-d sequencess which are interpreted as the deposits of moderately high energy and moderately erosive, low density turbidity currents. The normal grading is interpreted as reflecting gradual flow reduction, which provides sufficient time for finer grains to be deposited from suspended load to bed load layers (Khan and Arnott 2011). The rare Ta division is interpreted as representing the suspension collapse portion of the flow (Rotzien, 2014; Bouma, 1962; Lowe, 1979, 1982). Planar laminated sandstone (Tb) is interpreted to be upper flow regime deposits, where the rate of suspension fallout was sufficiently low that there was significant bedform migration. The ripple cross- lamination (Tc) is interpreted as representing bedform migration at relatively lower sedimentation rates, as climbing ripple cross lamination is absent. The presence of bidirectional ripple cross-lamination and associated HCS-like sedimentary structures indicates the presence of reworking by bottom currents (e.g. Ito, 1996; Shanmugam, 2013; Rebesco et al. 2014).

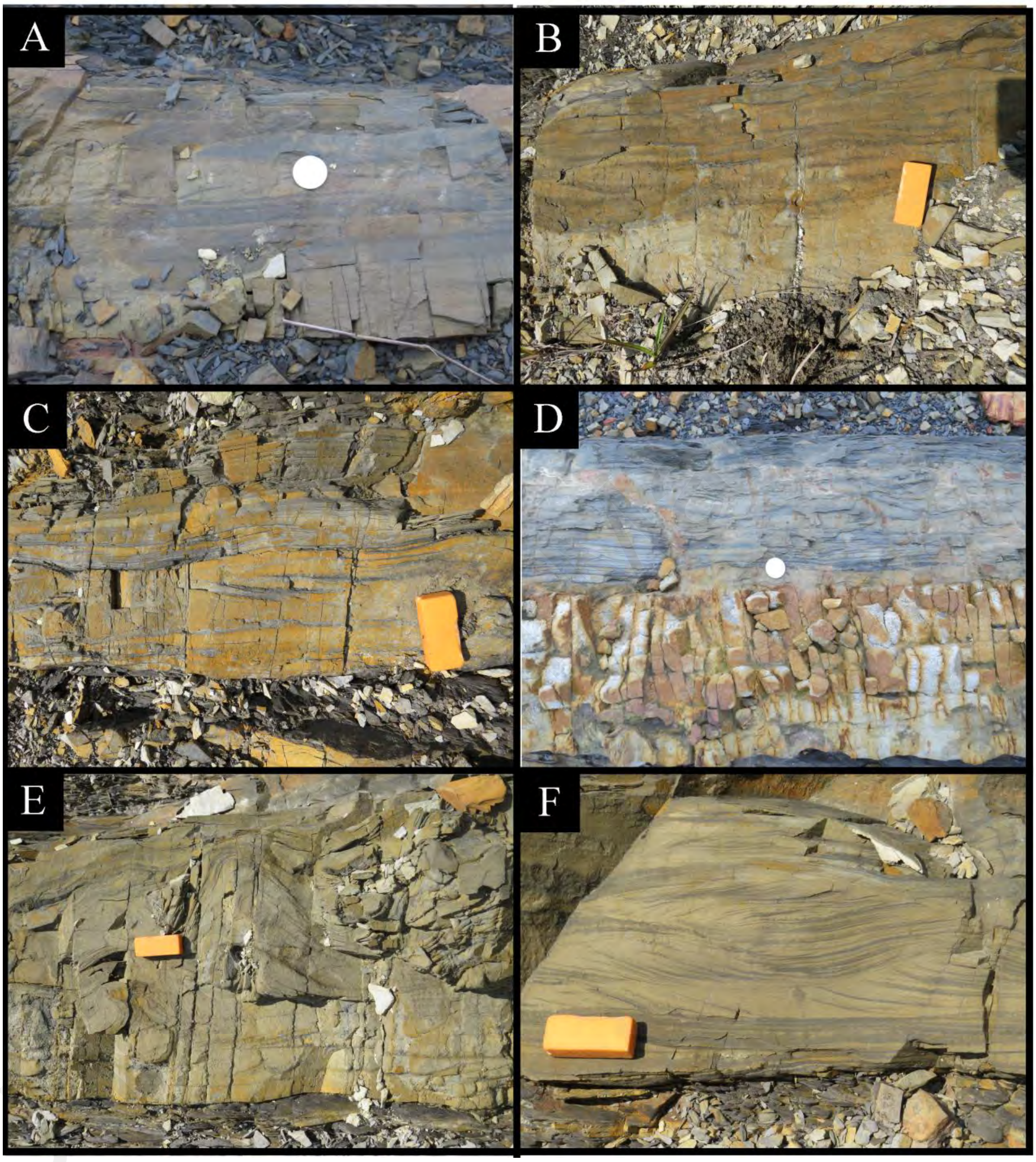


Figure 3.3. Photographs of medium graded sandstone (F3): A), B) and C) Normal graded sandstone beds with sharp bases, display Bouma Tb-d. D) Graded sandstone displays abrupt change in grain size. The medium grained size (rusty colour) appears structureless, interpreted as Bouma Ta, and displays erosive bases, overlain by this graded sandstone displaying Bouma Tb-e. This whole bed can be considered as perfect Boouma sequence (Ta-e). E) Graded sandstone with sharp bases and convoluted layer overlain by cross lamination. This bed comprises Bouma Tb-c. F) Graded sandstone displays bidirectional cross lamination.

3.4. Facies 4: Thick graded sandstone (F4)

Description

Facies F4 **(Fig. 3.4A-B)** is characterised by 50-150 cm thick, tabular beds of moderate to well sorted, medium to fine grained sandstone. The beds are typically sharp or scour-based and display normal grading. Individual beds of F4 commonly display complete Bouma sequences (Tabcde, with overlying F1 mudstone forming the Te division). However, many examples do not preserve the uppermost Te division, but are sharply overlain by younger Bouma sequences. Mud clasts can be present within some of the Ta divisions, ranging in size between 1- 10cm and are sub-rounded. Ripple beds in individual Tc divisions commonly display an upward fining in grain size, from fine to very fine grained sand. Some examples of F4 display an upward, incremental increase in mud content. These beds also display an upward increase in mud and sand rip-up clast concentration. Ball and pillow as well as fluid escape structures are commonly associated with facies F4. Scour marks and groove casts are also present at the base of the beds. The facies can form amalgamated stacks up to 7 m thick.

Interpretation

Facies F4 is interpreted to be the deposits of high energy, erosional turbidity currents based on the sand-dominated texture, Bouma sequences, sharp erosive base and the presence of sole marks. Mud clasts within the Ta division are interpreted as rip-up clasts eroded from underlying mudstone. The sharp base of the individual sandstone beds, with very little erosional relief, possibly indicate that the turbulence near the base of the flow was water saturated, and high sedimentation rates probably suppressed the erosional ability of the flow (Hickson and Lowe, 2002).

Planar lamination (Tb) is interpreted as representing upper flow regime deposition, where the rate of suspension fallout was sufficiently low that there was significant bedform migration (e.g. Paola et al., 1989, Khan and Arnott, 2011). Overlying ripple cross-lamination (Tc) indicates waning of the flow and deposition of fine grained sand from suspension. The fining upward grain size trend of ripple beds indicates slow suspension fallout. The parallel lamination (Td) of very-fine-sand/silt/mud is interpreted to have been deposited during the end flow of the waning stage. Te mudstone represents the finest grained sediments within the flow, deposited from suspension fall out.

The gradual, upward increase in mud content and mud clast abundance in some beds of Facies F5 is interpreted as reflecting progressive evolution of the flow behaviour with time, from non-cohesive (turbidity flow) to cohesive flow (debris flow) (Hodgson, 2009). The increase in mud clay content within beds was produced by destruction of mud clasts and incorporation of the clay content into the suspended load. This led to suppression of the flow turbulence ability and generation of a cohesive flow (Talling et al., 2007).

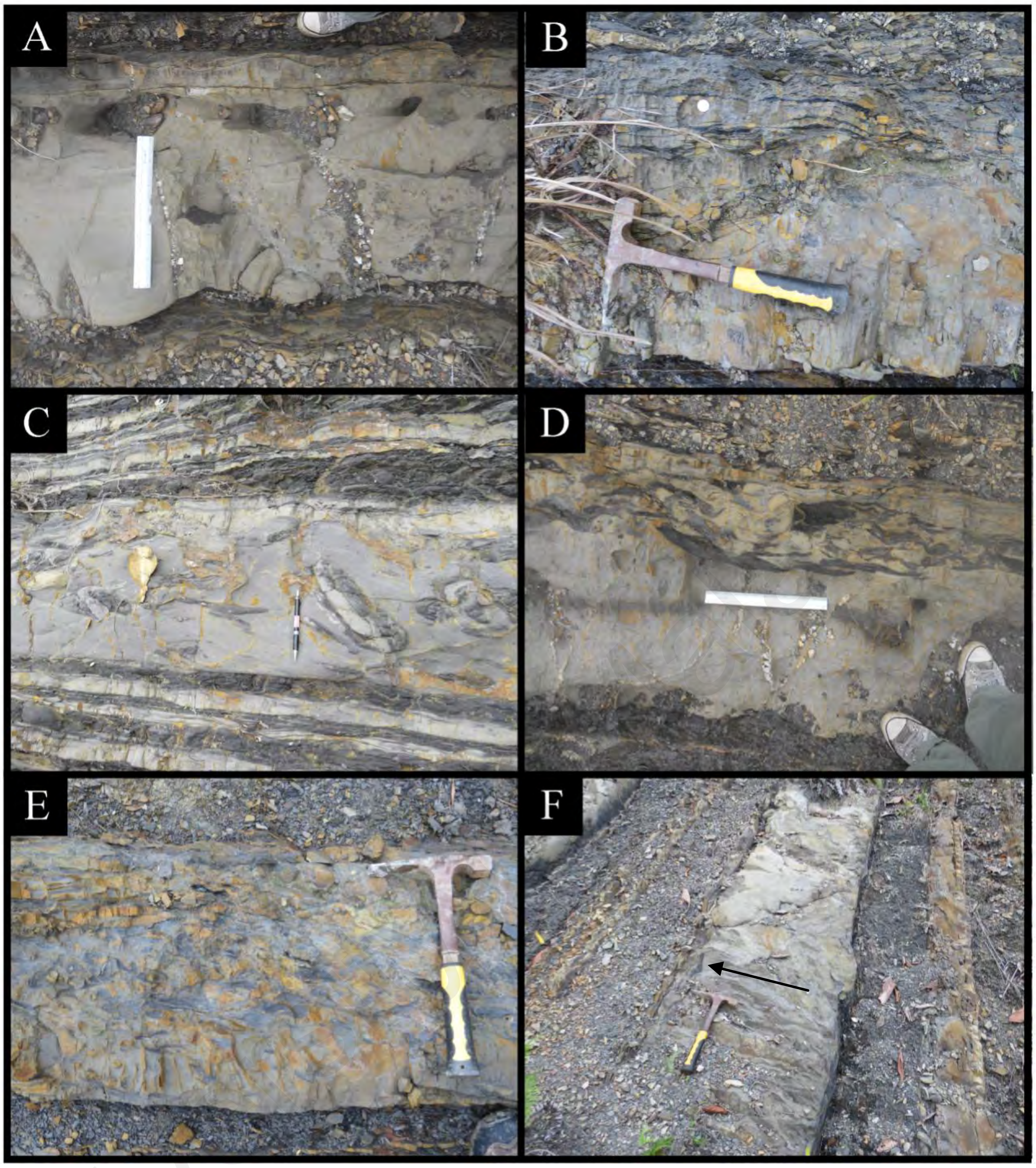


Figure 3.4. Photographs of thick graded sandstone beds. A) and B) Display thick graded sandstone, with scour and erosinal bases (F4).The beds display Ta-d/e. C-E are facies F5. C) Structureless sandstone displaying sharp and non-erosional bases with disoriented floating mud/sand clast. D) Structureless sandstone with erosional bases with elongated floating mud clasts at the upper part of the bed. E) Structureless sandstone bed displays disoriented mud clast. F) D3 "Hybrid bed" (Hodgson, 2009) structureless weakly-normal graded sandstone with floating mud clast at the upper part of the bed (F6).

3.5. Facies 5: Structureless sandstone (F5)

Description

Facies F5 **(Fig. 3.4C-E, 3.5A-F)** is characterised by 40 to 300 cm thick, tabular beds of grey to dark grey coloured, poor to well sorted, medium to fine grained sandstone. Some beds are more argillaceous and contain mudclasts. The base of the sandstone is scoured. Facies F5 commonly contains abundant floating sandstone and mudstone clasts. The sandstone clasts display parallel and cross-ripple lamination. The size of the clasts varies, with sub-rounded clasts being comparably smaller in size than angular clasts. The angular clasts size ranges between 1-30 cm in size, while the sub-rounded clasts range from 1-10cm. The percentage of clasts within the beds varies between 10% - 70%. Facies F5 typically does not display any grading. The boundary with overlying mudstone facies is also sharp.

Some of the structureless sandstone in beds form stacked successions up to 4.5 m thick, with each bed separated by cm-thick mudstone.Other examples form stacks of 2 beds, with the bottom bed being clean sandstone, while the overlying bed is more argillaceous. The bottom clean sandstone is 30-60 cm thick, with a sharp, erosional base overlying mudstone. The overlying, more argillaceous sandstone bed is thicker (50-150cm), sharp-based and is normal graded, with upward increase in mud content. The argillaceous sandstone beds contain abundant, 1-50 cm long, elongated to subrounded, floating mud and sand clasts. Some of the sand clasts display lamination. The clasts become more abundant upwards.

Interpretation

F5 is interpreted as mainly representing debris flow deposits (debrites). This is based on the presence of argillaceous matrix supporting chaotically larger components with no preferred orientation, indicating inefficient sorting and *en masse* deposition from a cohesive flow (Hodgson 2009). Downslope flow of sediment-water mixture can scour and rework underlying beds, resulting in broken up clasts being transported together further downslope. Flow reduction and eventual depletion downslope stopped the movement of the fluid when the frictional force was greater than the flow of the fluids (Kneller & Branney, 1995).

The presence of angular clasts in the structureless sandstone bed indicates short transport distance and deposition from weak turbulent flows which limited clast collision (Hodgson, 2009). The presence of elongate and rounded mud clasts in some beds indicates deposition from high density turbidity currents (Saito and Ito, 2002) and/or debris flows (Shanmugam, 2000).

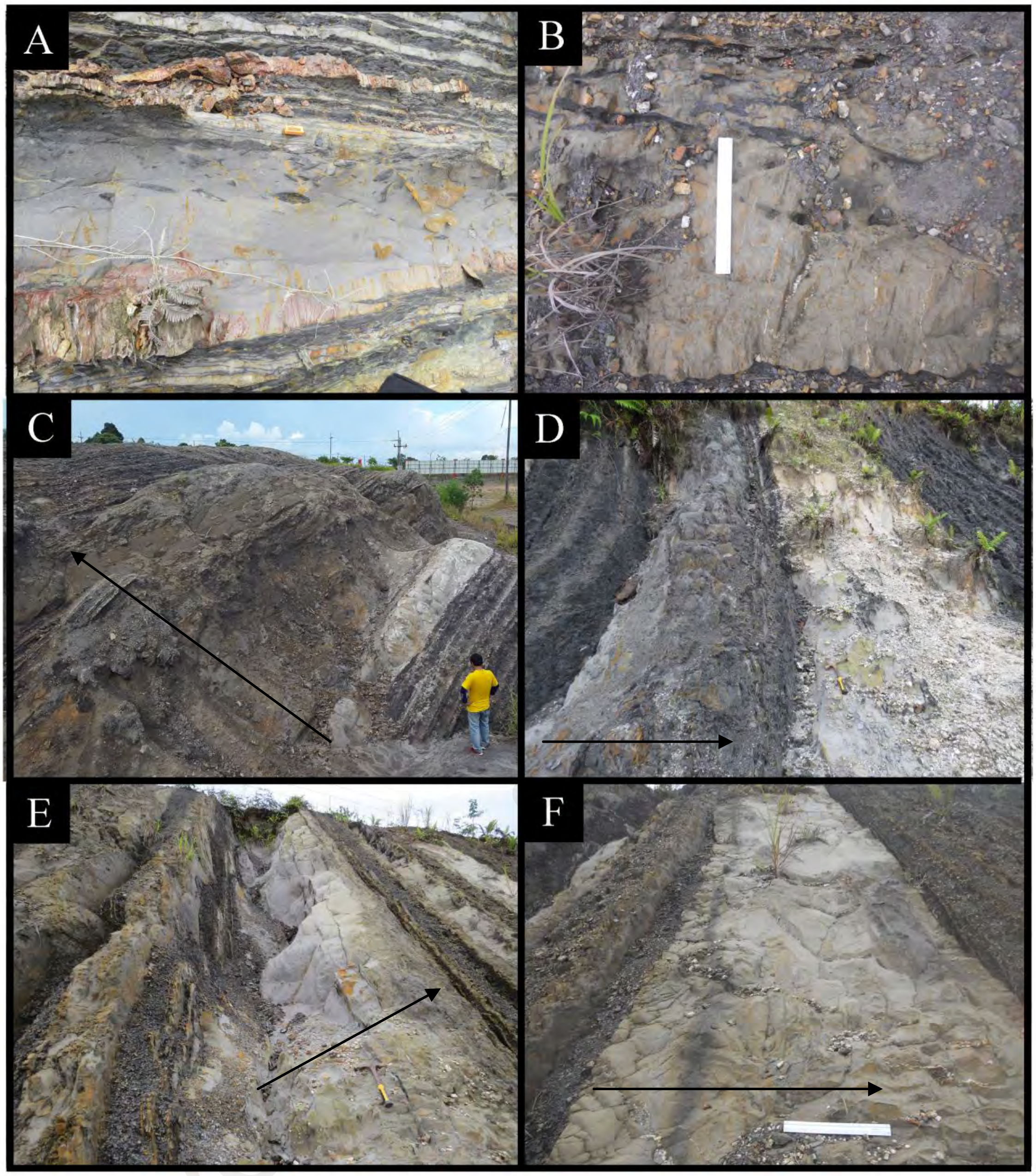


Figure 3.5. Photographs of thick bedded structureless sandstone (F5). A) Structureless sandstone display sharp bases, normal grading by the presence of increase in mud content and clast upward. B) Erosinal structureless sandstone displaying elongated mud clast at the middle to upper part of the bed. C) Massive structureless sandstone displays normal grading identified by the increase in mud content upward. This bed display massive mud and sand clast. D) Stack of amalgamated structureless sandstone, displaying the one is more argillaceous and displayed weakly normal grading with floating mudclast, while the clean sandstone displays elongated mud clast. E) Stack of amalgamated structureless sandstone, where the top most bed display normal grading Bouma Tb-d. F) Thick clean structureless sandstone.

3.6. Facies 6: Hybrid beds (F6)

Description

Facies F6 **(Fig. 3.4F)**is characterised by 30-60 cm thick, medium to fine grained, poor- moderately sorted sandstone beds. The beds display sharp upper and lower contacts with overlying and underlying beds. The base can either be straight or erosional, with less than 1 cm relief. A basal mud clast lag is commonly associated with the erosional base (clast diameter less than 3 cm). Weak normal grading is faintly visible in some examples of Facies F6. The upper part of the sandstone bed contains 1-10 cm long, floating, disk-shaped mud clasts. The floating mud clasts have smooth edges and are aligned with their long axes parallel to bedding. Beds of Facies F6 are commonly amalgamated to form stacked successions 10-15 m thick.

Interpretation

Facies F6 resembles the D3-type hybrid beds of Hodgson (2009). Hybrid beds are deposits with a bipartite structure, which display evidence of deposition from more than one type of flow (e.g. cohesive and non-cohesive flows) (Hodgson, 2009). In D3-type hybrid beds, both the lower sandstone bed and overlying mud clasts are suspension deposits of turbidity current. The mud clasts were deposited at the rear of a non-cohesive flow, resulting in the concentration of mud clasts at the top of the bed.

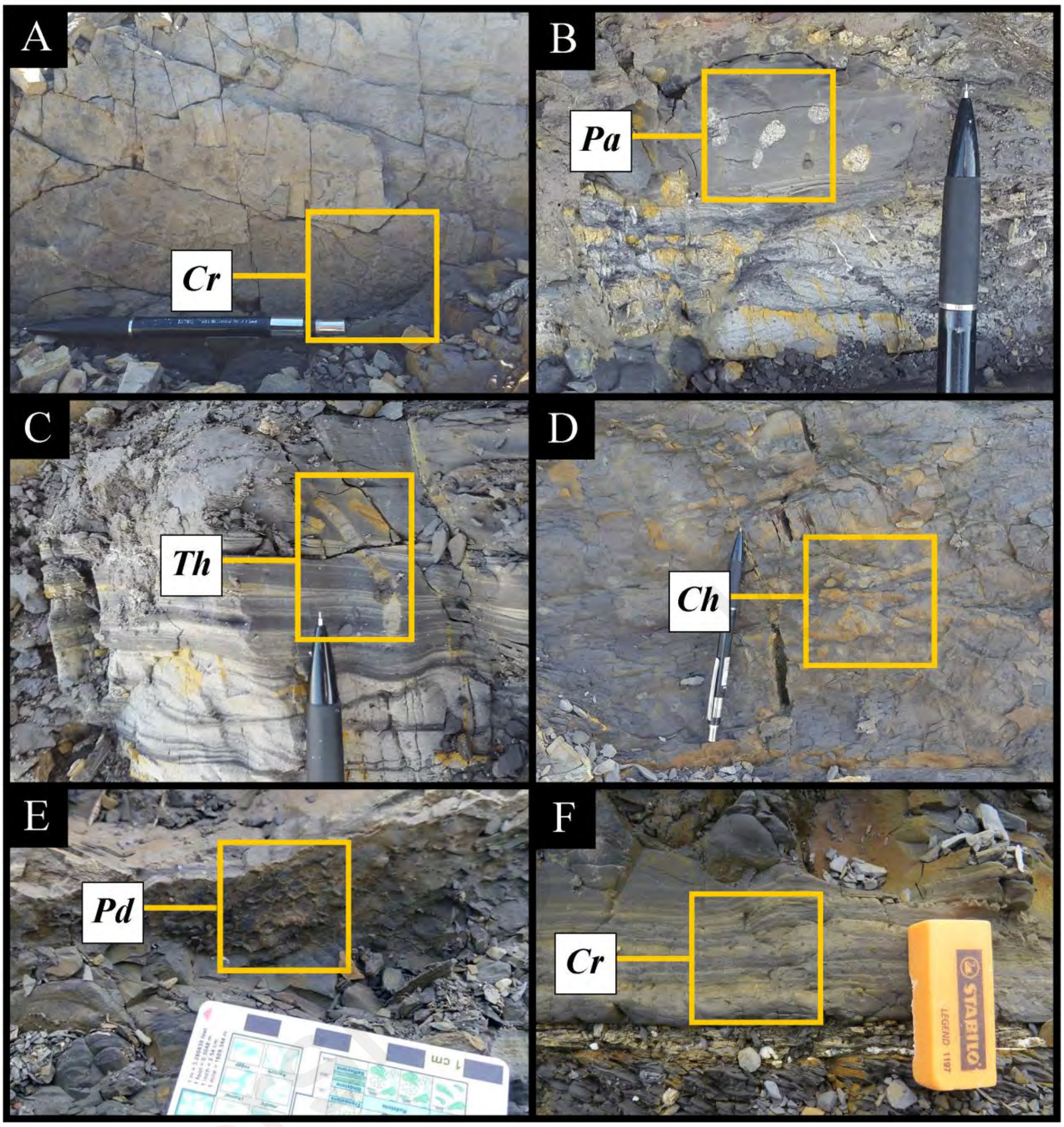
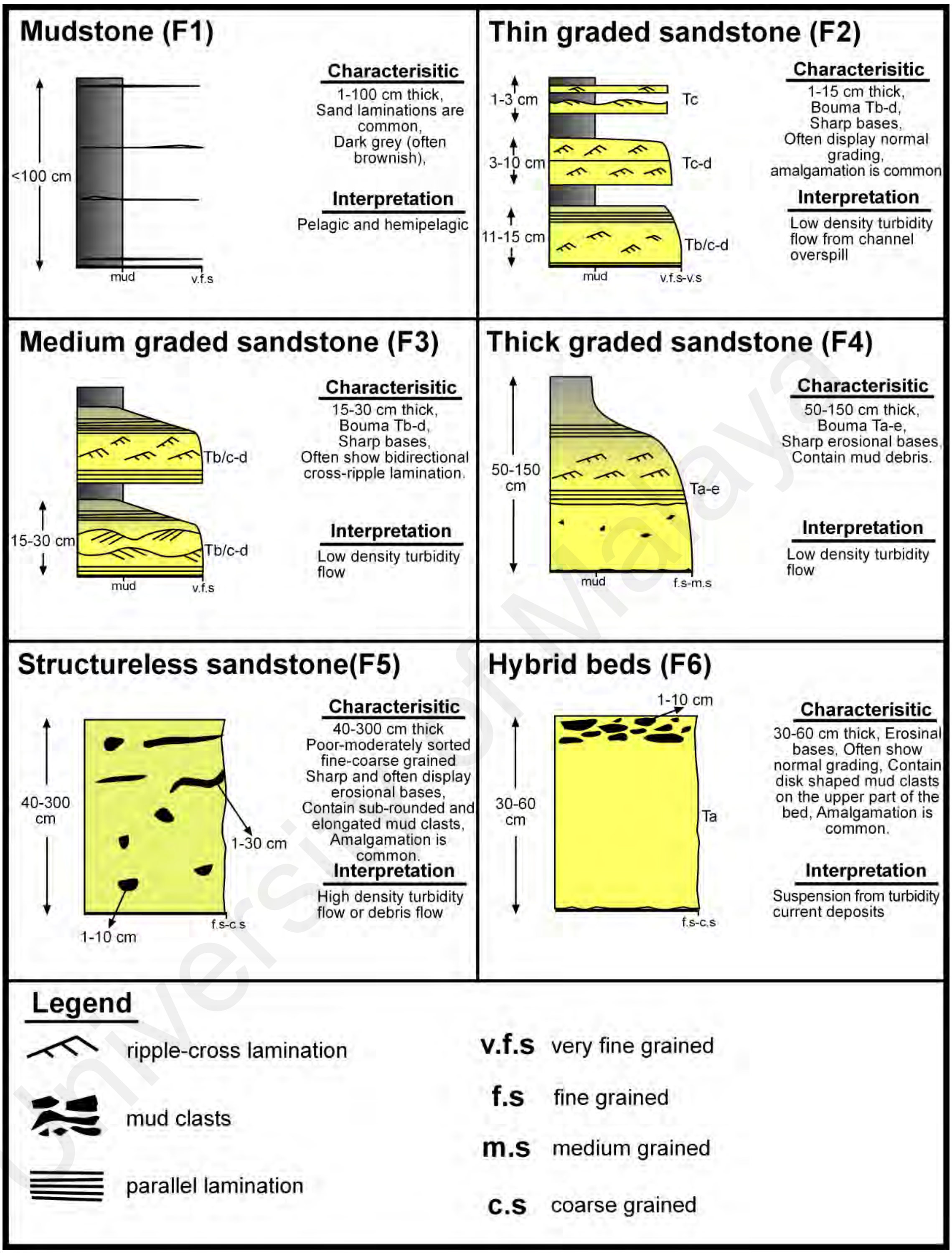


Figure 3.6. Photographs oftrace fossils characteristic of the *Nereites* ichnofacies.*(Cr) Cosmorhaphe, (Th) Thalassionides, (Ch) Chondrites, (Pd) Paleodictyon, (Pa) Palaeophycus.*

Table 3.1. Summary of the facies characteristics of the Belaga Formation, Sibu.



CHAPTER 4: FACIES ASSOCIATIONS (FA) OF THE PALEOCENE-MIDDLE EOCENE BELAGA FORMATION

Six facies associations (Table 1B) are recognised in the study area. They are; Channel fill (FA1); Levee (FA2); Lobe (channel mouth, FA3); Lobe (fringe, FA4); Basin Plain (FA5) and; Slump(FA6). Of the total thickness of 536 m of the logged sections, the percentage thickness of the individual facies associations are as follows: FA1= 14.7%, FA2= 50.6%, 330 FA3= 6%, FA4= 13.8%, FA5= 5.9%, and FA6= 9%. Therefore, the studied sections of the Belaga Formation around Sibu are dominated by strata interpreted as levee deposits (50.6%). It should be noted that the study transects two members of the Belaga Formation, i.e. the Kapit and Pelagus members. The differences and similarities in facies association composition between the 2 members are summarised in Table 4.2.

4.1. Channel fills Facies Association (FA1)

Description

The channel-fill facies association (FA 1)(Fig. 4.1) forms up to 11 m thick successions, which generally consist of debrite beds (F5) with minor interbedded mudstone and thin to medium bedded turbidites (F1, F2. Individual facies F5 beds display concave-upward erosional bases with 10-30 cm relief. Facies F5 also commonly displays a wedge-shaped geometry, with abrupt thinning and on-lap/off-lap relationship with adjacent F5 facies beds. The erosional base commonly truncates underlying facies and result in lateral thinning of beds. Groove casts and flute casts are common at the base of the sandstones. Facies F5 can form amalgamated stacks up to 7 m thick.

Intervals of Facies F1 and F2 between 1 to 4.5 m thick are typically intercalated between amalgamated beds of F5. The grain size tends to be medium to fine at the bottom of the succession and becomes finer vertically upward. Palaeocurrent readings indicate a general northward flow direction (Fig. 4.2).

Interpretation

FA1 is interpreted as gravity flow channel-fill deposits based on facies composition dominated by debrites and turbidites, the sharp, erosional base and fining upward succession. The erosional bases and deep multiple scours indicate deposition of high energy flows, with the concave upward profiles of the basal erosion surface representing channel scour and incision (Rotzien et al., 2014). The non-tabular amalgamated beds indicate the high energy erosive power of gravity-driven sediment flows in channels due to a greater flow thickness, higher flow velocity and turbulence between turbidites (Mattern, 2002). These combined characteristics are commonly associated with large channels (Rotzien et al., 2014). The wedge-shaped geometry with on-lap and off-lap feature characteristics may represent complex, within-channel erosion, which implies that the channel was frequently shifting its position.

The intercalation of facies F1, F2 and thinner F5 between the thick amalgamated sandstone beds is interpreted as representing channel abandonment deposits represented by thin to very thin bedded turbidites (Rotzien et al., 2014). However, it is also possible that facies F1, F2 and F5are channel-levee deposits (e.g. Khan and Arnott, 2011; Rotzien et al, 2014). This is supported by the fact that at some outcrops, these facies have oblique palaeocurrent directions **(Table. 4.1)** from the main channel, which suggest that they were deposited from channel overspill flows at channel levees (Khan and Arnott, 2011; Hickson and Lowe, 2002; Hubbard et al., 2008).

This facies association occurs in both the Kapit and Pelagus members, and they share many similar characteristics. However, FA1 in the Pelagus Member is relatively thinner (<10m) compared to Kapit Member (9-20m).

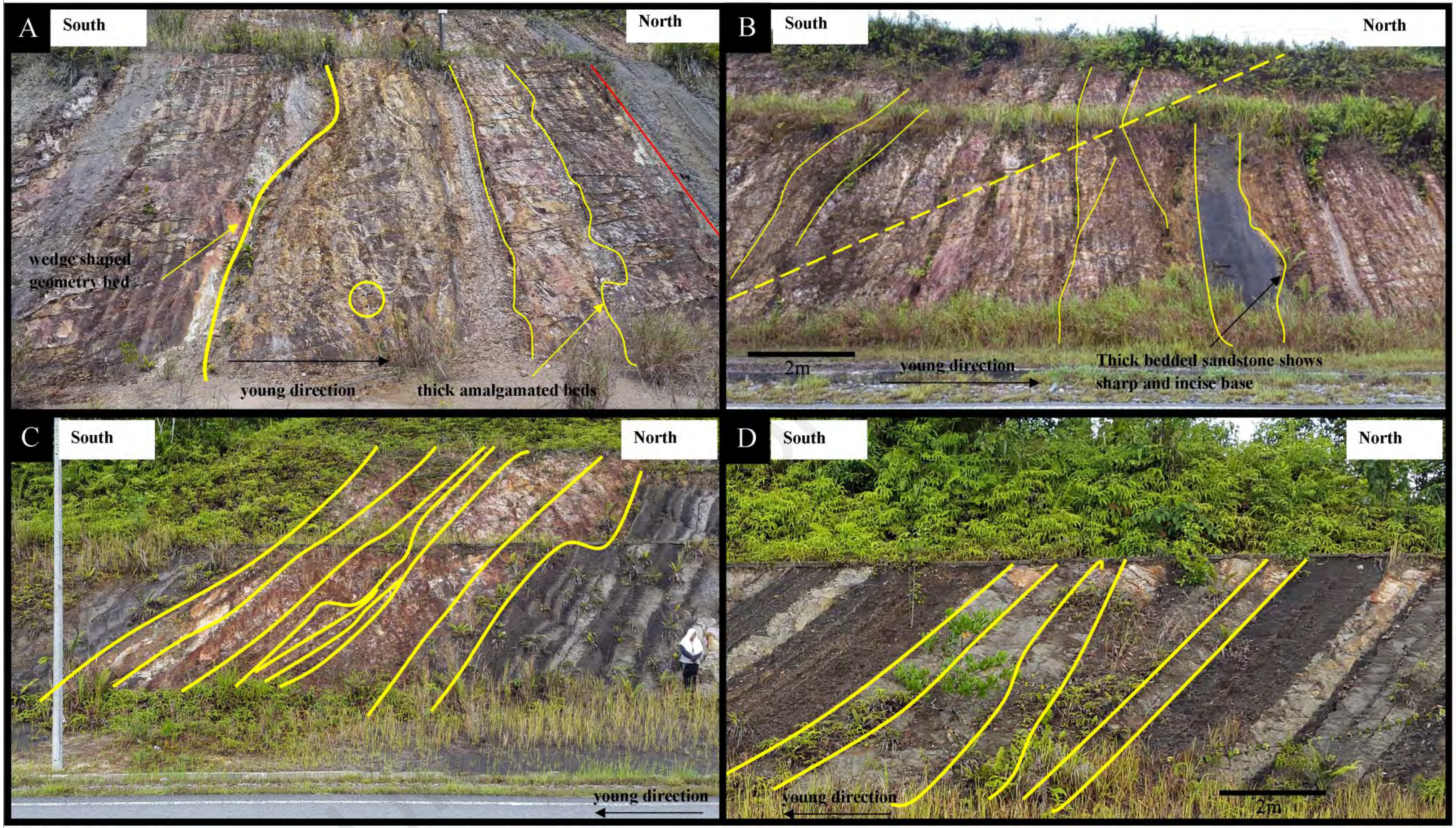

A
South
North
wedge shaped geometry bed
young direction
thick amalgamated beds
B
South
North
2m
young direction
Thick bedded sandstone shows sharp and incise base
C
South
North
young direction
D
South
North
young direction
2m

Figure 4.1. Photographs of channel fill facies association (FA1). A and B are from the Kapit Member (Durin road, outcrop G8), while C and D are from the Pelagus Member (Durin road, outcrop G2) (Wolfenden, 1960). A) Displays thick bedded (3m) sandstone with wedge shaped geometry, intercalating with thin graded sandstone, overlain by thick succession amalgamated sandstone. B) Thick succession of amalgamated sandstone displays on- and off- lap beds. Thick succession of amalgamated sandstone incises the underlying mudstone. C) Thick succession of amalgamated sandstone displays multiple incisions. D) Thick succession amalgamated sandstone displays incision, where some beds display lateral pinching out.

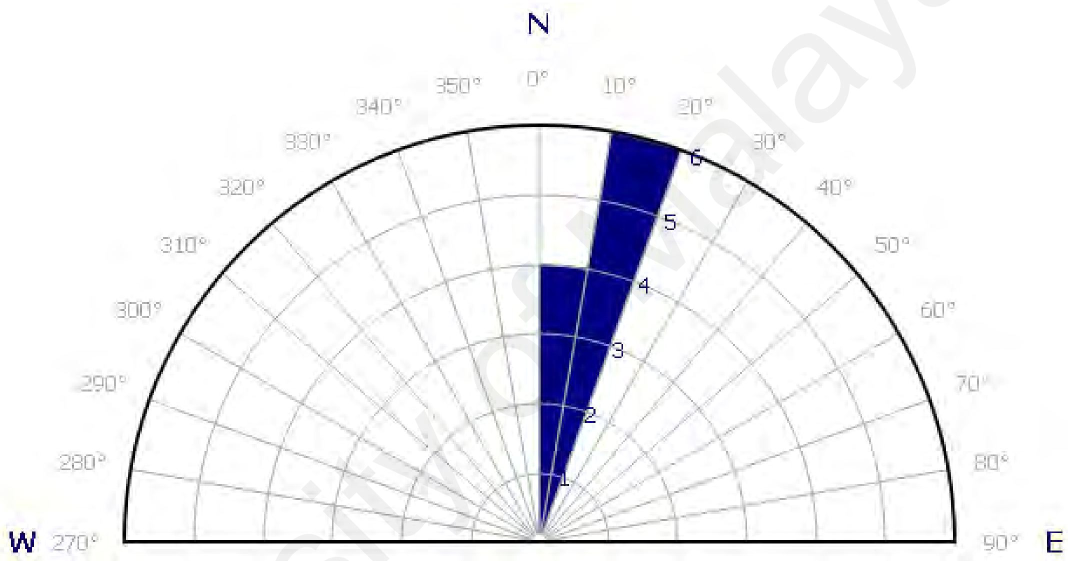


Figure 4.2. Rose-diagram shows the paleocurrent direction of channels (n=10) obtained from flute casts(outcrop G2 and G8).

4.2. Levee Facies Association (FA 2)

Description

The levee facies association (FA 2) is represented by successions up to 15 m thick, composed of multiple (40-250 cm thick) fining and bed thinning upward cycles of tabular bedded sandstone and mudstone **(Fig. 4.3).** The basal part of these fining upward cycles typically comprise 20-40 cm thick beds of medium bedded turbidites

(F3) or 20-30 cm thick, debrites (facies F5). Facies F5 commonly fines upward into thin bedded turbidites (facies F2). Rarer examples form complete cycles from facies F3/F5 into F2 and F3, with intercalated mudstone (F1). Ripple cross-lamination in F3 associated with structureless sandstone of F5 and heterolithic beds of F2 and F1 facies occasionally display opposing foreset dip orientations (bidirectional ripple orientations)**.**

One meter thick debrites (F5) with wedge-shaped geometry occasionally cut through thin-very thin bedded sandstone/mudstone**(Fig. 4.3C)**

Palaeocurrent readings obtained from sole marks indicate a general northeast palaeocurrent direction, compared to the northward palaeocurrent direction of the channel fill deposits (FA 1). Slump intervals (FA6) 2-30 m thick are commonly interbedded between FA2 units at some localities. The levee facies association is present in both the Kapit and 384 Pelagus members. The Kapit Member examples display thinner beds and successions, with consistent repetitive thinning upward from F3,F2, and F1, while the Pelagus Member examples display thicker beds and more facies variation, with repetitive successions of F5, 387 F4, F3, F2 and F1, and the occurrence of interpreted crevasse splays.

Interpretation

Facies association FA2 is interpreted as levee deposits originating from channel overspill, based on the dm-thick, repeating thinning upwards successions of F3 or F5 overlain by 1-2 m thick succession of interbedded F2 with F1. Alternating medium bedded turbidites (F3) and debrites (F5) indicates that the energy within the flow was inconsistent. The absence or minimal degree of erosion along the base of sandstone beds and the tabular geometry of the beds suggests a relatively low energy, unconfined environment (Mattern, 2002).The abundance of medium bedded turbidites (F3), the absence of climbing ripples in the Tc divisions of the turbidites, and absence of Bouma Td in some F2 beds are typical characteristics of channel levee deposits (Khan and Arnott, 2011). The thin to very thin bedded turbidites of Facies F2 are interpreted as being deposited by levee overspill (Hickson and Lowe, 2002; Jackson et al., 2009). The abundance of ripple cross-lamination within F3 and F2 suggests overbank levee deposition (Lyons, 1994; Coleman, 2000; Beaubouef, 2004; 401 Hickson and Lowe, 2002; Cronin et al., 2000).The reduced thickness of the beds relative to FA1 is also consistent with an unconfined setting. Despite the erosive conditions, confined channels contain more suspended material above the sea floor, resulting in relatively greater sediment fall out rate per unit area, whereas in unconfined environments, the thickness of the flow is reduced, and with it the amount of suspended material per unit area leading to a relatively smaller sediment fall-out rate (Mattern, 2002). The different variations of FA2 successions may represent different parts of the levee. The succession characterised by sandier elements comprising of facies F5 and F3, possibly represent outer bend levee deposits. These sandier elements can be interpreted to have been deposited from higher energy flows from overspill of the outer bend levee margin due to inertial effects (Imran et al 1999; Straub et al., 2008). Sand beds within the inner bend levee have relatively less sand (F2 turbidites may represent inner bend levee deposition).The oblique orientation of ripple palaeocurrent relative to main channel axis is also consistent with an outer bend levee setting **(Table. 1B)**(Hickson and Lowe, 2002; Khan and Arnott, 2011). The thickness of beds may have been affected by the area of deposition, where thicker sediments deposited at the more proximal levee slope, while the thinner beds are more distal.

The thin, fining upward cycles possibly indicate cycles of gradually smaller depositional flows (e.g. Rotzien, 2014) or produced by levee aggradations, with overspill deposition becoming more dilute with time due to increase in relief between channel floor and levee crest (Arnott, 2007b). This thinning upward succession is also common in levee deposits (e.g. Braga et al., 2001).

The intercalation of slump intervals (FA6) also supports a levee interpretation for FA2, since levees are inherently unstable and slumping is common (Kassi et al, 2011; Audet, 1998). The presence of thick bedded sandstone cutting through the interbedded thin-very thin bedded sandstone/mudstone may be interpreted as crevasse splay deposits, which are commonly associated with the outer bend of leveed channels. The presence of bidirectional ripple cross-lamination indicates significant reworking of the levee deposits by bottom currents (e.g. Ito, 1996; Shanmugam, 2013).

The levee facies association is present in both the Kapit and Pelagus members. The Kapit Member displays thinner beds and consistent, repetitive thinning upward F3,F2, and F1 cycles, while the Pelagus Member displays thicker beds and more facies variation of repetitive successions of F5, F4, F3, F2 and F1, with the occurrence of interpreted crevasse splays.

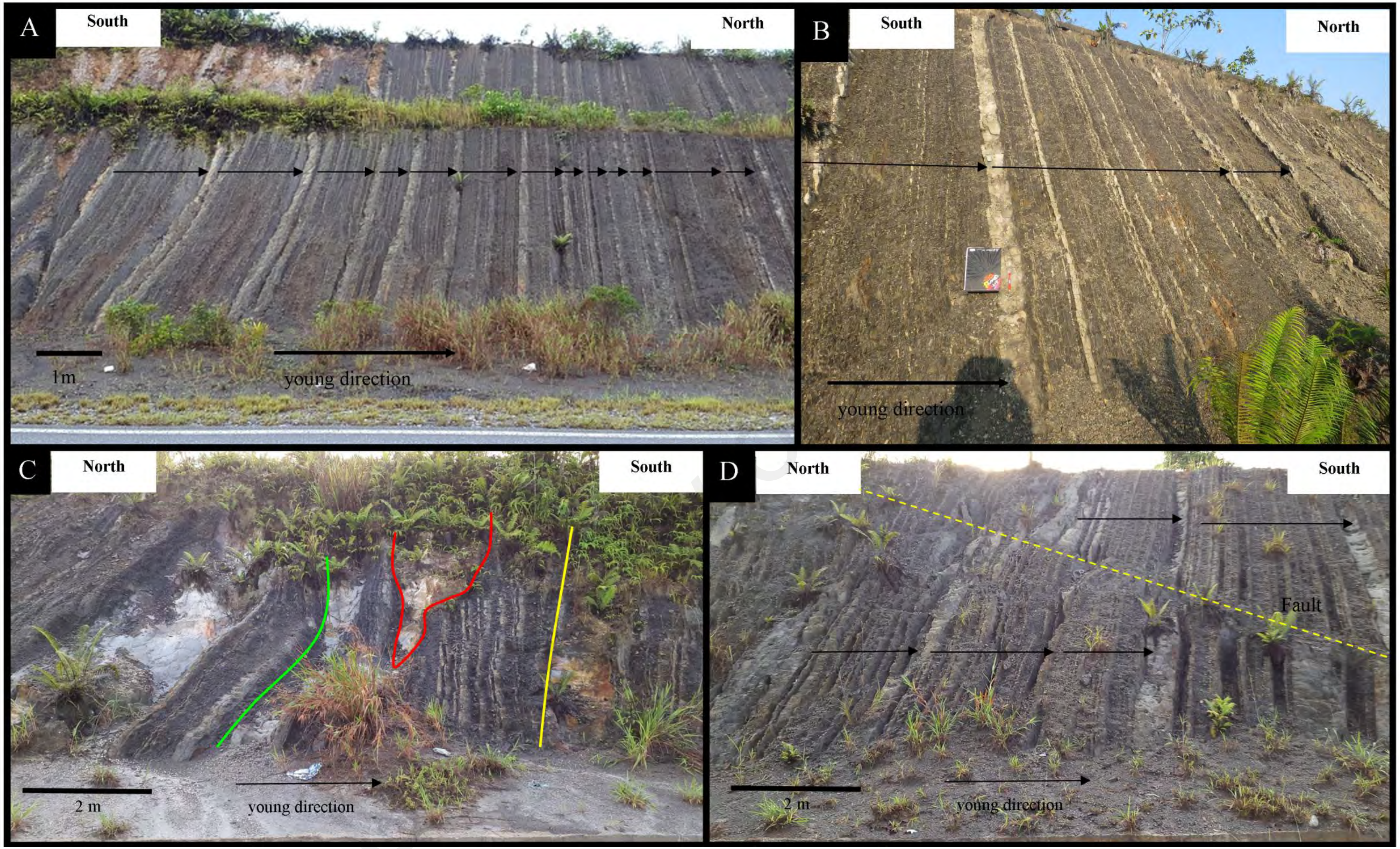

A
South
North
1m
young direction
B
South
North
young direction
C
North
South
2 m
young direction
D
North
South
Fault
2 m
young direction

Figure 4.3. Photographs of the Levee facies association (FA2) from Durin Road. A) Displays packages of repetitive thinning upward cycles (outcrop G8). B) Repetitive thinning upward packages (outcrop G6), having thinner bed thickness than photographs A. C) Thinning upwards cycle (green-yellow), displaying sandstone interpreted as crevasse splay, incising the underlying interbedded thin graded sandstone/mudstone (outcrop G3). D) Repetitive thinning upward packages (outcrop G3), displaying thicker bed thickness than photographs A and B.

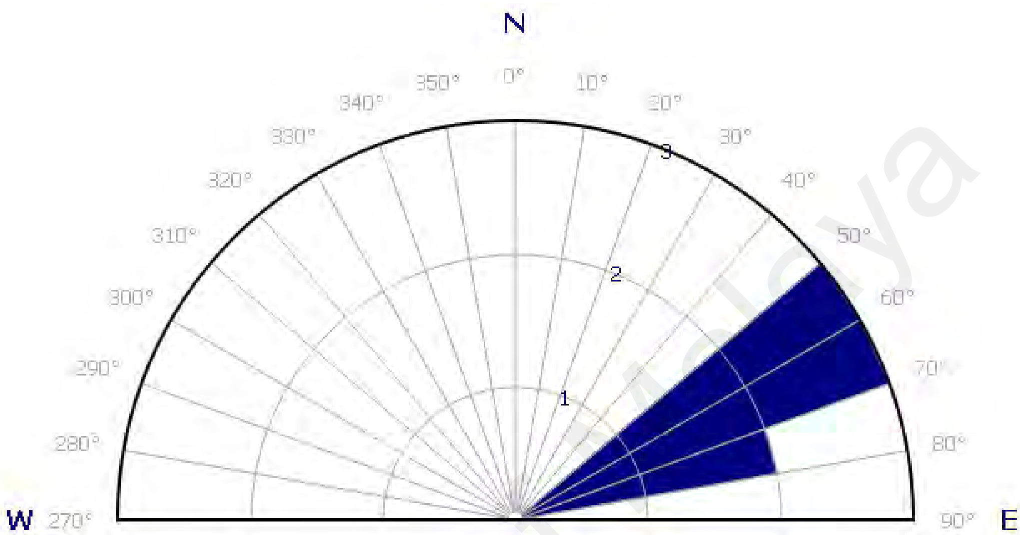


Figure 4.4. Rose-diagram shows the paleocurrent direction obtained from flute castof levee deposits from outcrop G8 (n=8)

4.3. Lobe (Channel mouth) (FA 3)

Description

The Lobe (Channel mouth) facies association forms 8-10 m thick, bed thickening upward successions comprising interbedded debrites/high density turbidites (F5), hybrid beds (F6) and mudstone (F1) facies(**Fig. 4.5**).

The F5 sandstones are tabular and erosionally based. They form amalgamated stacks up to 10 m thick. Weak normal grading is observed in some beds, with floating subangular- subrounded mud clasts (diameter less than 5 cm) on the upper part of the bed. A gradual decrease in mud clast abundance in individual sandstones is also observed to be associated with the bed thickening upward trend of FA 3. Hybrid beds also decrease in abundance upwards, while mudstone beds become gradually thinner, as the succession thickens upward. The thickening upward succession is typically overlain by FA1. Flute and groove casts are common and display a northward paleocurrent direction **(Fig. 4.6).**

Interpretation

FA3 is interpreted as lobe channel mouth deposits based on the presence of structureless sandstone (F5) and hybrid beds (F6) with tabular bedform, displaying a thickening upward succession, with upward thickening of F5, diminishing number of F6 beds and thinning of F1. The tabular beds indicate that facies F5 was deposited in an unconfined environment. The channel mouth interpretation is indicated by bedform geometry and amalgamated beds with the stacking up pattern of thickening sandstone (F5 and F6) beds (Mutti, 1977; Mutti and Ricci Lucchi, 1972), where F6 is decreasing upwards, combined with the occurrence of FA1 overlying the FA3 strata. The erosional and floating mud clasts are typical of D3 "hybrid beds", which indicate deposition from a flow that has erosive ability, in this case a turbidity flow. D3 hybrid beds are common at the fringe of lowstand submarine fan systems (Hodgson, 2009). This facies association is observed only in the Pelagus Member.

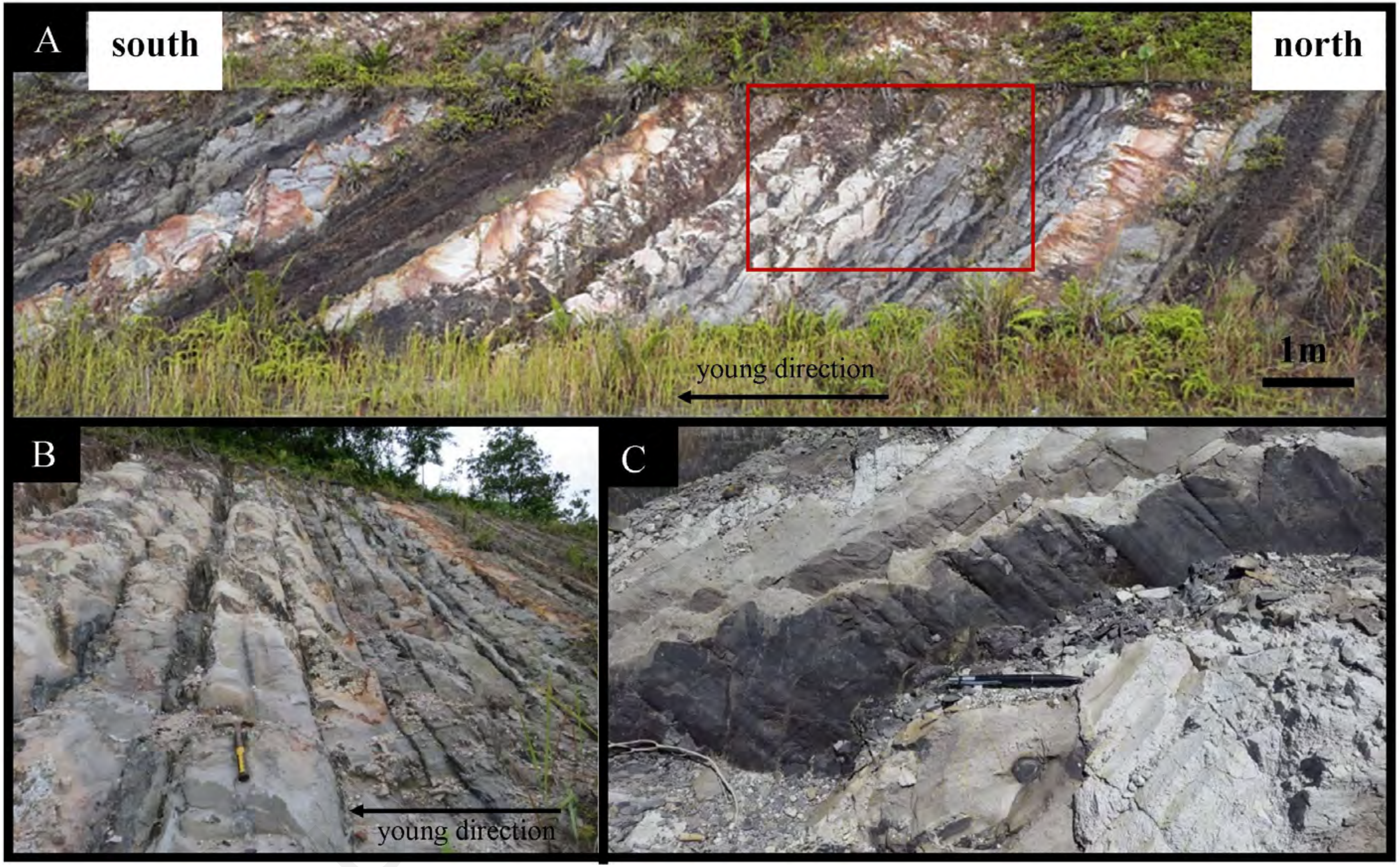

A
south
north
1m
young direction
B
young direction
C

Figure 4.5. Photographs of Lobe (channel mouth) facies association (FA3) from Durin road (outcrop G3). A) Thickening upward succession where thickness of sandstone bed thickens upwards. B) Close up view of the red-square area in Fig. A, displaying a thickening upward succession. C) Groove casts and flute casts along the base of a bed.

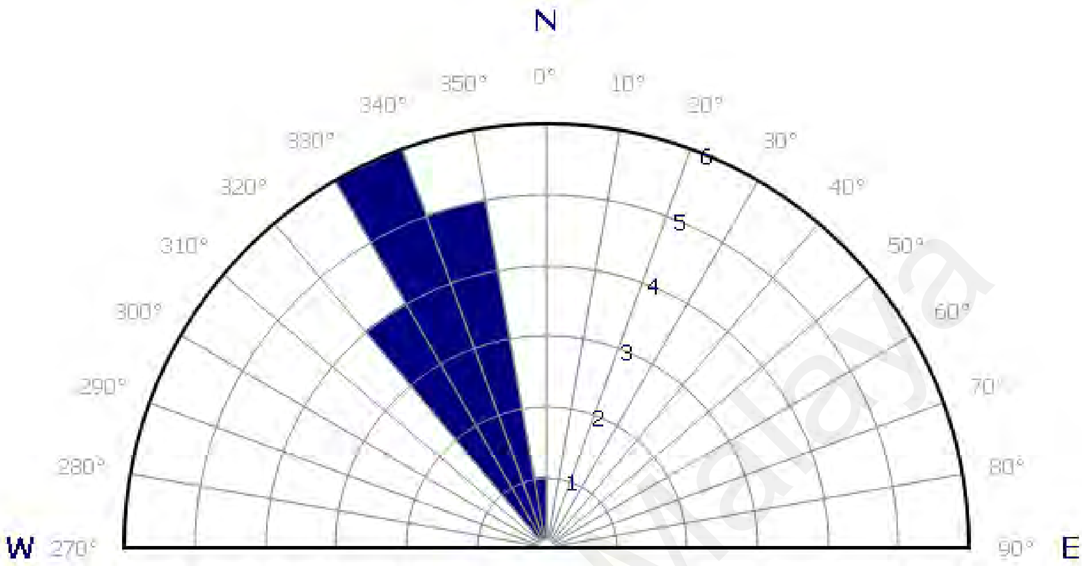


Figure 4.6. Rose-diagram shows the paleocurrent direction of lobes (n=16), taken from groove casts and flute casts at outcrop G2.

4.4. Lobe (fringe) (FA 4)

Description

The Lobe (fringe) facies association forms 15-65 m thick, upward bed thinning successions. FA4 comprises debrites (F5) with interbedded mudstone (F1) and thin bedded turbidites (F2) (Fig. 4.7). Hybrid beds (F6) in this facies association are generally more than 30 cm thick. The beds are commonly tabular with no significant basal erosional scours. However, scours do occur locally within the lobe deposits and are mainly tabular scours (i.e. flat basal scour with less than 2 cm down-cutting). F5 commonly displays floating mud clasts on the top part of the beds. Chaotically arranged, angular mud and sand clasts with diameters of up to 30 cm, are concentrated near the top of beds. The upward increase in clasts in individual F5 beds is also closely associated with an upward increase in mud content. F5 is commonly amalgamated with clean, less thick-bedded structureless sandstone, which are either underlying or overlying the muddier beds (F6). These resemble the D2-type "hybrid beds" of Hodgson (2009) but are thicker. Interbedded, thin-bedded turbidites (F2) and mudstone (F1) become more common upsection and also thin upwards in FA 4.

Interpretation

FA 4 is interpreted as representing lobe fringe deposits, based on the presence of a thinning upwards succession of amalgamated "hybrid beds" (F6) with thick structureless sandstone (F5) beds, which were deposited in a broad and unconfined setting, such as a lobe environment, since lobe deposits do not display evidence of significant scouring (Mutti, 1979). The local occurrence of tabular scours is consistent with a lobe interpretation, as they are typically present in the upper intervals of lobe successions (Mutti et al., 2002). "Hybrid beds" are also a common element of lobe deposits (Hodgson, 2009). The presence of "hybrid beds" indicate that FA4 probably represents deposition during the initial growth of a submarine fan, as ―hybrid beds” are common during periods of relative sea-level fall (Goldhammer et al., 2000; Johnson et al., 2001; Hodgson et al., 2006; Hodgson, 2009). Sediment from the slope channel was eroded and entrained within the turbulent flow as angular sand/mud clasts. As the flow travelled basinward, the entrained clasts collided/corroded and became more rounded. Broken up clasts became mixed with the flow, enriching it with fine grained sediments which eventually changed the flow behaviour from non-cohesive to cohesive. As the flow reached an unconfined setting (basin floor), a hybrid bed is deposited. This facies association occurs only in the Pelagus Member and is absent in the Kapit Member.

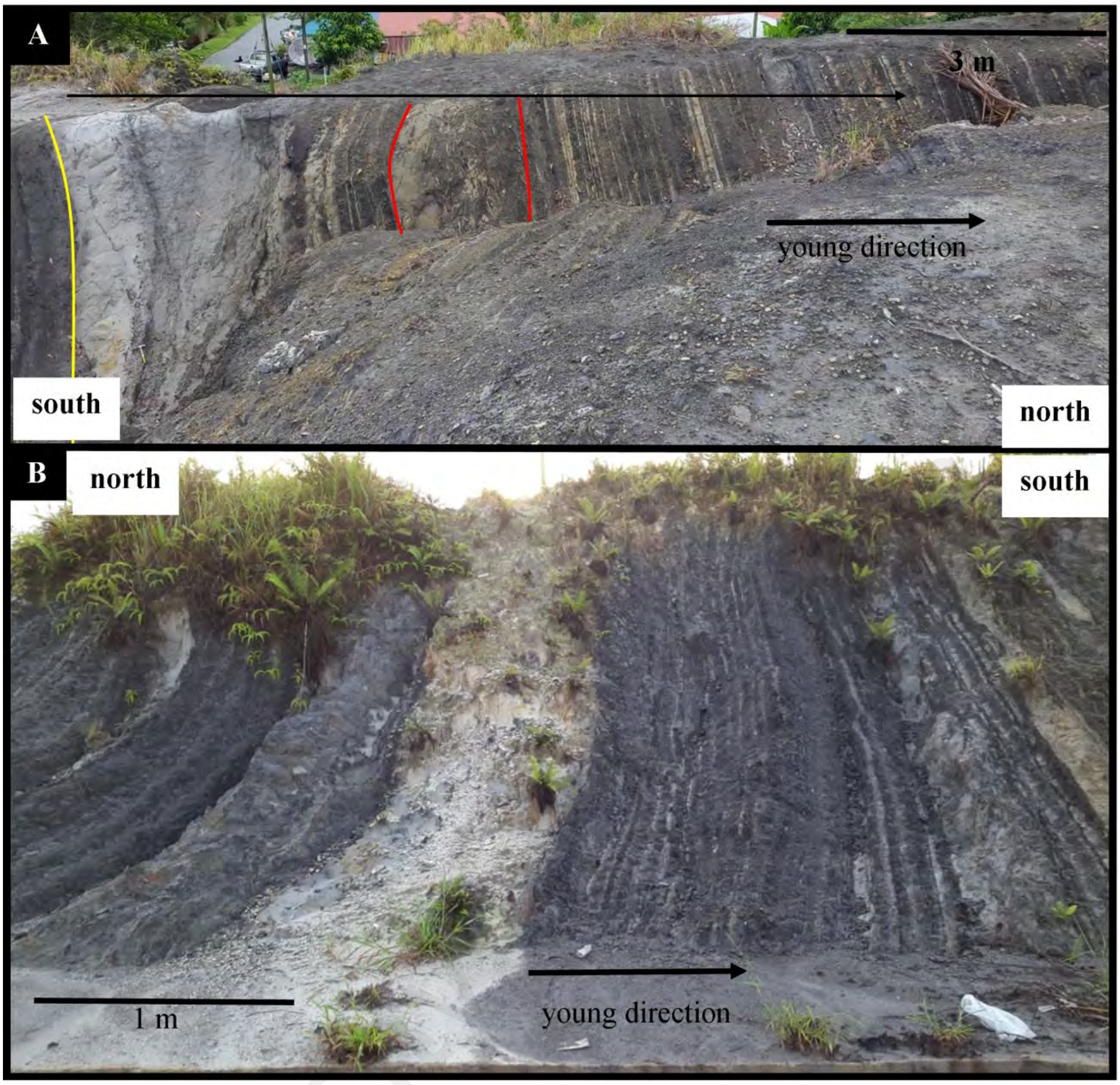


Figure 4.7. Photographs of the Lobe (fringe) facies association (FA4) from Durin road. A) Displays thinning upward succession (outcrop G10). Yellow line marks the base of the succession of thick amalgamated sandstone overlain by thin succession of thinning upwards, interbedded thin graded sandstone and mudstone. B) Thick amalgamated sandstones displaying dirty (argillaceous) sandstone, overlain by clean sandstone, overlain by thinning upward interbedded thin graded sandstone and mudstone (outcrop G3).

4.5. Basin Plain Facies Association (FA 5)

Description

The basin plain facies association is characterised by 5-20 m thick successions of interbedded mudstone (F1) and thin bedded turbidite (F2) facies **(Fig. 4.8).** FA 5 displays some variability in turbidite bed thickness, with numerous bed thinning/thickening trends throughout the succession. *Trace fossils are common, in the form of* (Cr) *Cosmorhaphe, (Th) Thalassionides, (Ch) Chondrites, (Pd) Paleodictyon, (Pa) Palaeophycus***(Fig. 3.6).**This facies association occurs only in the Kapit Member.

Interpretation

The fine grained texture, tabular and laterally extensive, sheet-like turbidites and the absence of significant cyclicity in FA5 suggest a basin plain depositional environment (Kassi et al., 2011). The thin bedded turbidites of F2 were probably deposited from the end-tail of long-lived, high energy and high density turbidity currents, which extended from a submarine fan system and thinned out in the distal, basin plain setting.

Figure 4.8. Photograph of the basinal plain facies association (FA5) from Durin road (outcrop G5). Interbedded sandstone thin graded sandstone and mudstone, display no pattern, such as thinning or thickening upward (scale human=1.6m)

4.6. Slump Facies Association (FA 6)

Description

FA6 is characterised by 1-15 m thick units of interbedded sandstone and mudstone of faces F1, F2 and F3, displaying intraformational, recumbent and asymmetric folding **(Fig. 4.9).** Large rotated sandstone blocks (F4 and/orF5, 30-120 cm long) are commonly associated with the facies association, as well as intraformational faults. The base of FA6 lays in oblique orientation with the underlying strata.

Interpretation

The intraformationally deformed strata, with recumbent and asymmetric folds in between undisturbed strata are interpreted as slump deposits formed by mass-movement (Stow, 2005) Slumping can be triggered by over-steepening, seismic loading, cyclic storm-wave loading, rapid accumulation and under-consolidation (Locat and Lee, 2002). The facies composition indicates that FA 6 most likely represents slumped levee deposits, being a result of levee failure and collapse (e.g. Ballance, 1964; Gregory, 1969).

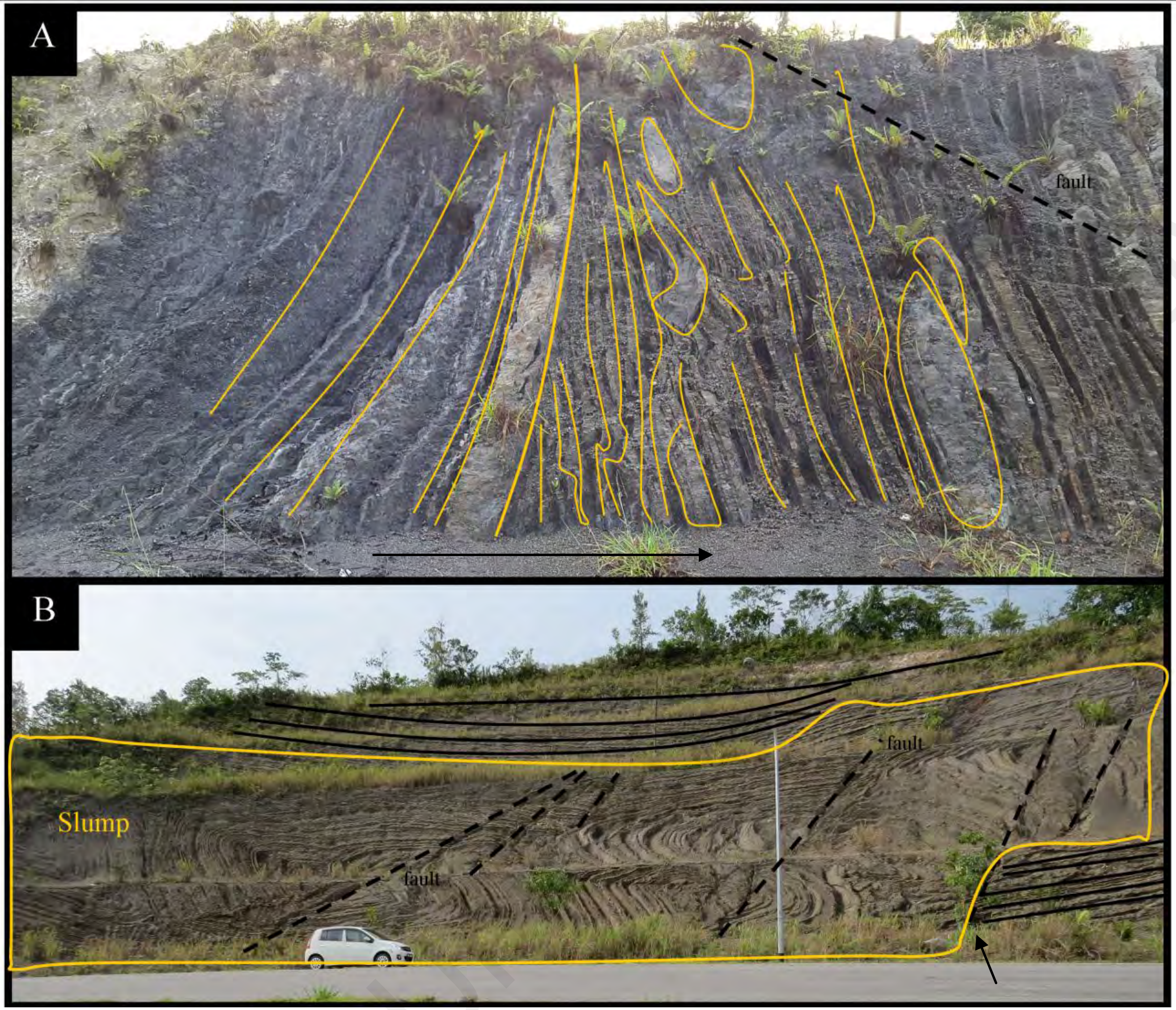


Figure 4.9. Photographs of slumps Durin Road. A) Displays disoriented strata with the occurrence of thick sandstone chunks (outcrop G3). B) Massive slump approximately 15 meters thick, comprises of recumbent and isoclinal foldings, with multiple occurrence of fault (outcrop G9)

Table 4.1. Summarised facies association for both Kapit and Pelagus member.

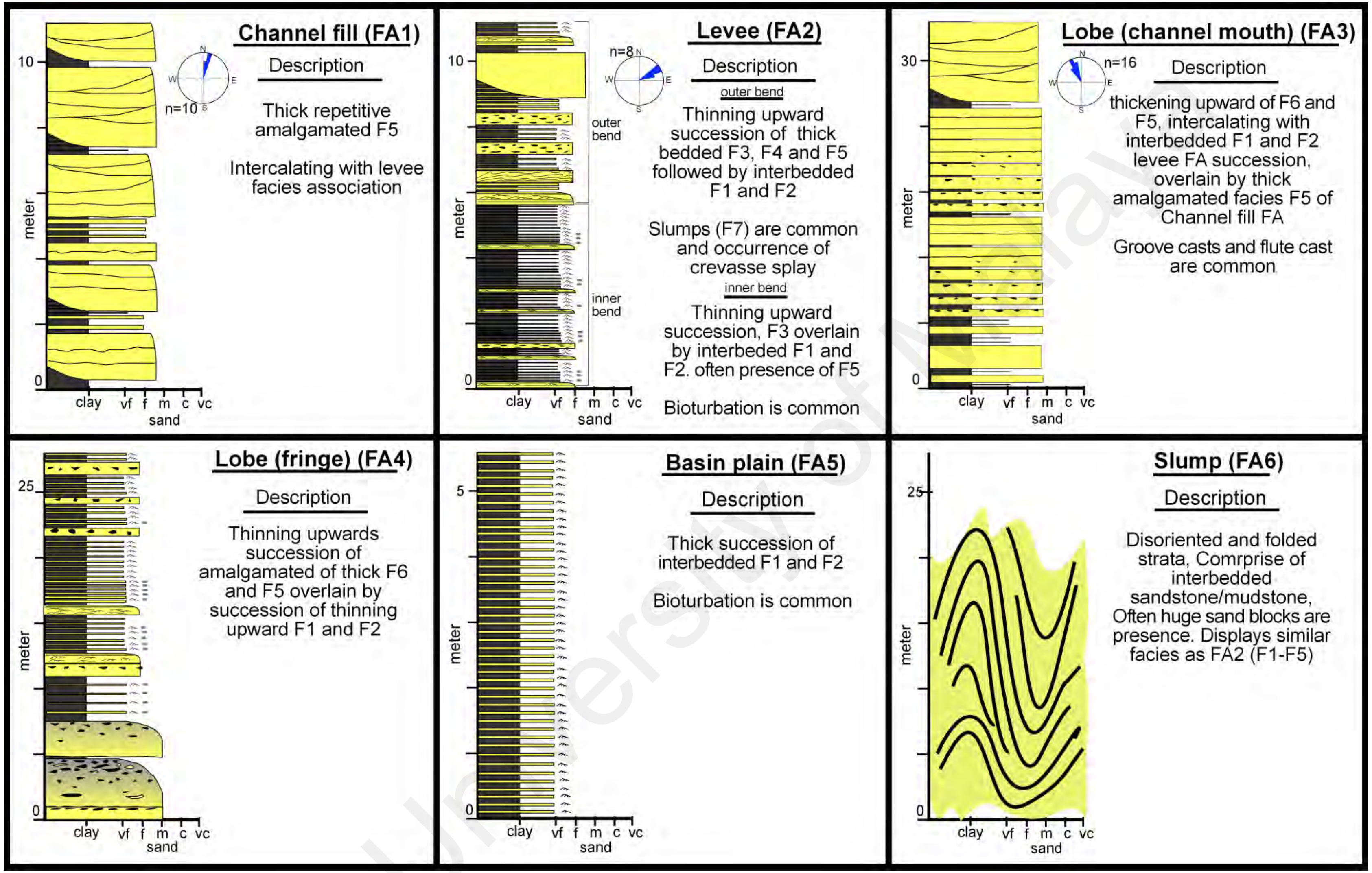


Table 4.2. Summarised Facies Association difference of Kapit and Pelagus Member.

Facies Association	Kapit Member	Pelagus Member
Channel fill (FA1)	Comprises F5, intercalated with interbedded F1/F2. Wedge shaped geometry, on-/off- lap bed relationship. Amalgamation is common with deeply scoured. Forming 50 m succession	Comprises F5, intercalated with F1. Amalgamation is common. Wedge shaped geometry and on-/off-lap bed, deeply scoured bases. Form 5-10 thick succession.
Levee (FA2)	Mainly comprising Facies F1, F2 and F3. Thinning upward packages of F3 overlain by interbedded F2/F1. Rarely F5 displaces F3.	Mainly comprising facies F1, F2, F4, F5. Thinning upward packages of F5/F4 overlain by interbedded F1/F2. Thicker F2 be compared to Kapit member
Lobe (Channel mouth) (FA3)	Absent	Mainly comprising F5,F6 with minor occurrence of F2 and F1. F5 and F6 form thickening upward successiosn, where F6 diminish upsection.
Lobe (Fringe) (FA4	Absent	Mainly comprising F6,F5,F3,F2 and F1. Thinning upward succession of F6 and F5, often appears amalgamated, overlain by thinning upwards interbedded F3,F2 and F1
Basin plain (FA5)	Interbedded F2 and F1 and common in *Nereites* trace fossils	Absent
Slump (FA6)	Rare. One outcrop (G6) display 1-2 m and one massive slump up to 15 m thick.	Common. Thickness ranges from 2-30 m.

CHAPTER 5: DEPOSITIONAL ARCHITECTURE OF PALEOCENE-MIDDLE EOCENE BELAGA FORMATION

The steep dipping strata, tectonic deformation and limited lateral extent of exposures prevent the precise reconstruction of the Kapit and Pelagus member depositional systems. The same restrictions prevent correlation between outcrops around Sibu. However, the distinctive vertical facies patterns of the identified facies associations provide a general view of the depositional system. Zainol et al. (2007) described only 3 facies from the Belaga Formation, all of which are observed in our study (debrites, thin bedded turbidites and hemipelagic/pelagic mudstone). This project reports 6 identified facies, which includes previously undescribed hybrid beds. The identification of a higher number of facies is mainly due to the more detailed facies analysis conducted on thicker successions and a higher number of outcrop exposures.

This project also builds further on the work of Zainol et al. (2007) by identifying several facies associations which have not been previously reported, including channelized sand bodies (FA1), associated levee and slumped levee deposits (FA2; FA6), prograding channel mouth lobes (FA3), lobe fringe (FA4) and basin plain deposits. All 6 facies associations are here interpreted as elements of leveed channel and frontal splay depositional environments of a submarine fan.

5.1. OUTCROP DESCRIPTIONS

5.1.1. Outcrop G1

G1 is 23 m thick and is interpreted as a thick succession of levee facies association **(Fig. 5.1).** The bottom part of the log is characterised by medium bedded sandstone overlain by a thick of succession of interbedded thin bedded sandstone and hemipelagic mudstone, showing a fining upward trend. Overlying the levee facies association is another levee deposit with different characteristics. Thick bedded high density turbidites are overlain by thin-very thin interbedded sandstone/mudstone showing a fining upward trend. The levee succession is overlain by slump deposits, and then overlain by the more levee deposits, identified by the oblique relation with the underlying strata rather than parallel. The interbedded succession in this outcrop displays relatively thicker beds compared to the other outcrops.

5.1.2. Outcrop G2

G2 comprises a 58 m thick succession of lobe, channel fill and levee facies associations**(Fig. 5.2).** The lowermost 9 m of the section comprises a thick succession of lobe facies association, characterised by repetitive interbedding of thick bedded, high density turbidites and thick hemipelagic mudstone. This is then overlain by a thick succession of intercalated channel fill and levee facies associations. The overlying strata comprises repetitive thickening upwards successions of amalgamated sandstone deposited from high density turbidites. In the outcrop, two thickening upward lobe (channel mouth) facies associations (7 and 10 m thick respectively) are recognised and are separated by 5 m of intercalated channel and levee facies associations. Another 5 m of intercalated channel fill and levee facies associations overlies the lobes facies associations, which is then successively overlain by a thick succession (10 m) of a channel fill facies association, followed by another 5 m lobe facies association.

5.1.3. Outcrop G3

The G3 section is 63 m thick and characterised by a thick succession of lobe fringe, levee and channel facies associations**(Fig. 5.3).** The basal part of the section comprises a fringe facies association, characterised by thick bedded, high density turbidites overlain by interbedded thin bedded turbidites/mudstone. Bed thickness thins upward. Slump deposits overlie the lobe deposits. This is then successively overlain by a thick levee facies association, characterised by thick repetitive succession of thick bedded high density turbidites overlain by interbedded thin-very thin bedded turbidites with hemipelagic mudstone. The levee deposits are divided into smaller-scale, thinning upward cycles. Meter-thick channel-fills are sporadically interbedded between levee deposits.

5.1.4. Outcrop G4

G4 is comprises a 55 m thick stack of levee facies associations**(Fig. 5.4).** The succession is characterized by thinning upward cycles of medium bedded turbidites and interbedded very thin bedded turbidites with mudstone. Rare thickening upward cycles are also observed.

5.1.5. Outcrop G5

The G5 section is a 31 m thick succession of basin plain deposits, comprising interbededd, thin-very thin bedded sandstone and hemi pelagic mudstone **(Fig. 5.5).** The lowermost 23 meters comprises mainly mudstone with thin bedded turbidites. The thin bedded turbidites sandstones display pervasive bioturbation, which increases upwards. Overlying this succession is a 1 meter thick unit of hemipelagic mudstone with sand lenses. Above this mudstone is an interval of interbedded thin bedded sandstone with and mudstone, displaying an upward increase in sand layer frequency. The beds also display pervasive bioturbation.

5.1.6. Outcrop G6

The G6 section is 45 m thick and comprises interbedded channel fill and levee facies associations**(Fig. 5.6).**The section is divided into 3 bed thinning upward successions, which can reach up to 15 m thick. The succession grades upward from amalgamated channel fill deposits with thin intercalated levee deposits, into thicker levee deposits.

The first lower two successions display similarity in thickness of beds and succession. However, the third successions displays a single channel fill overlain by thick succession of levee facies association. The channel beds in this section are less amalgamated, with a lower sand proportion, compared to the channels from other outcrops (e.g. G2 and G8). The beds in the leveed deposits are also comparatively thinner compared to those of other outcrops (G1, G3, G4 and G8).

5.1.7. Outcrop G7

The G7 section is 46 m thick and is characterized by a levee facies association overlain by thick slump deposits **(Fig. 5.7).** The levee deposits are divided into smaller scale, thinning and fining upward cycles. The slump deposits are also characterised by thin-thick bedded low-high density turbidites or debrites.

5.1.8. Outcrop G8

The G8 section is 123 m thick and is divided into a lower part (60 m thick) characterized by levee deposits, and a 51 m thick upper part comprising channel fill deposits with thin intercalated levee deposits **(Fig. 5.8)**. The lower levee succession comprises smaller scale, thinning upward cycles of F3/F5, F2 and F1. The channel-fill deposits are commonly composed of amalgamated sandstone beds, with abundant erosiional surfaces. Successive channel-fill deposits become muddier upsection and are overlain at the top of the section by thick levee deposits.

5.1.9. Outcrop G9

The G9 section is composed of levee deposits with a thick interbedded slump interval **(Fig. 5.9).** The leveed deposits are characterized by intercalated thin bedded mudstone and sandstone displaying thinning and thickening upward trends. The slump interval is 15 m thick, is displaying disoriented folded strata and associated with faults.

5.1.10. Outcrop G10

The G10 section is 67 m thick and is characterised by a thick succession of lobes (fringe) facies associations**(Fig. 5.10).** The base of the facies association is characterised by thick bedded sandstone, composed of floating mud clast deposited form high density turbidity flow or debris flow. This is then overlain by a thinning upwards succession of interbedded thin-very thin bedded turbidites with hemipelagic mudstone, although there are also minor intervals of interbedded sandstone and mudstone which display thickening upward trends.

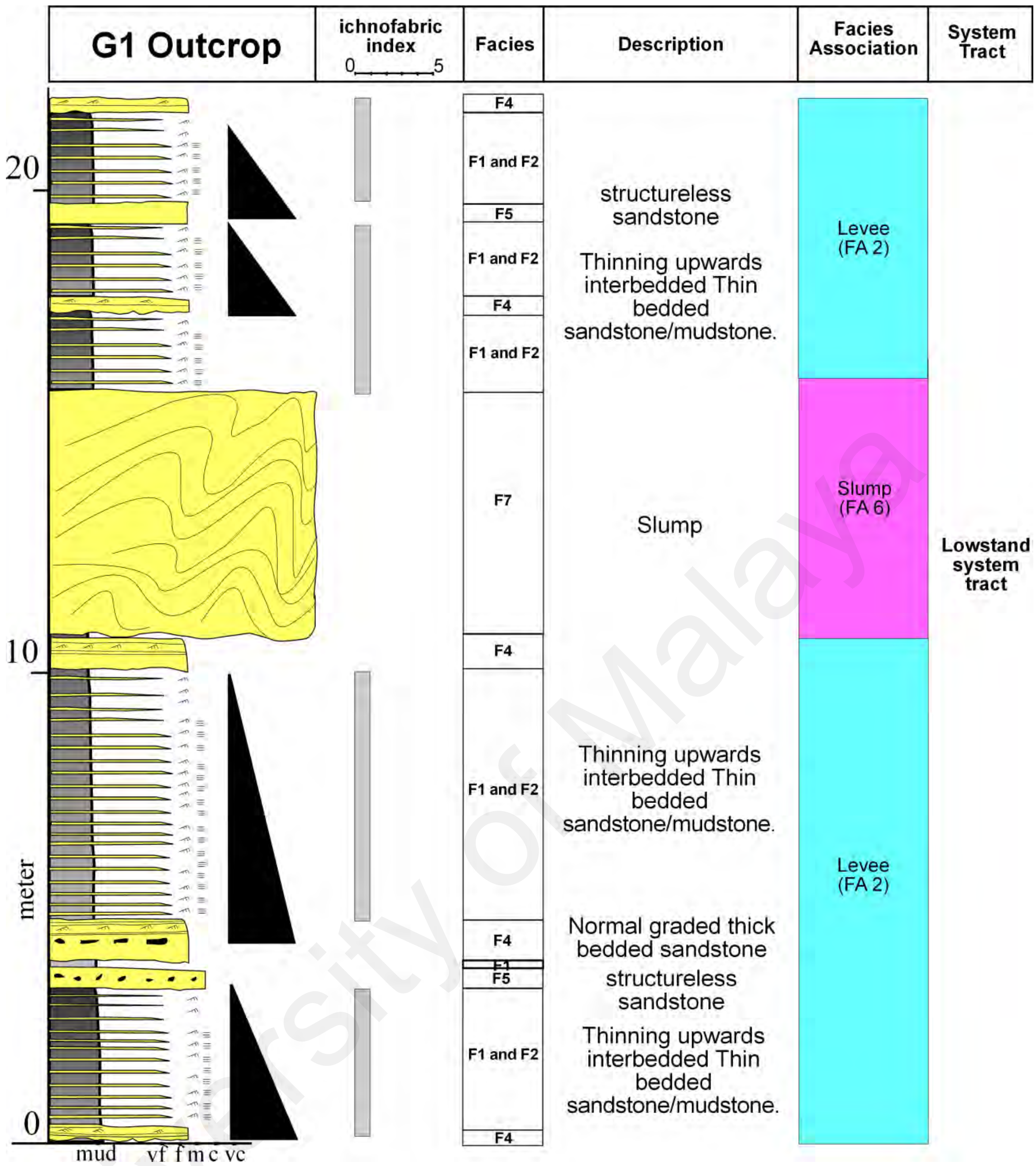


Figure 5.1. Outcrop G1 section. Refer to Fig. 1.1B for outcrop location.

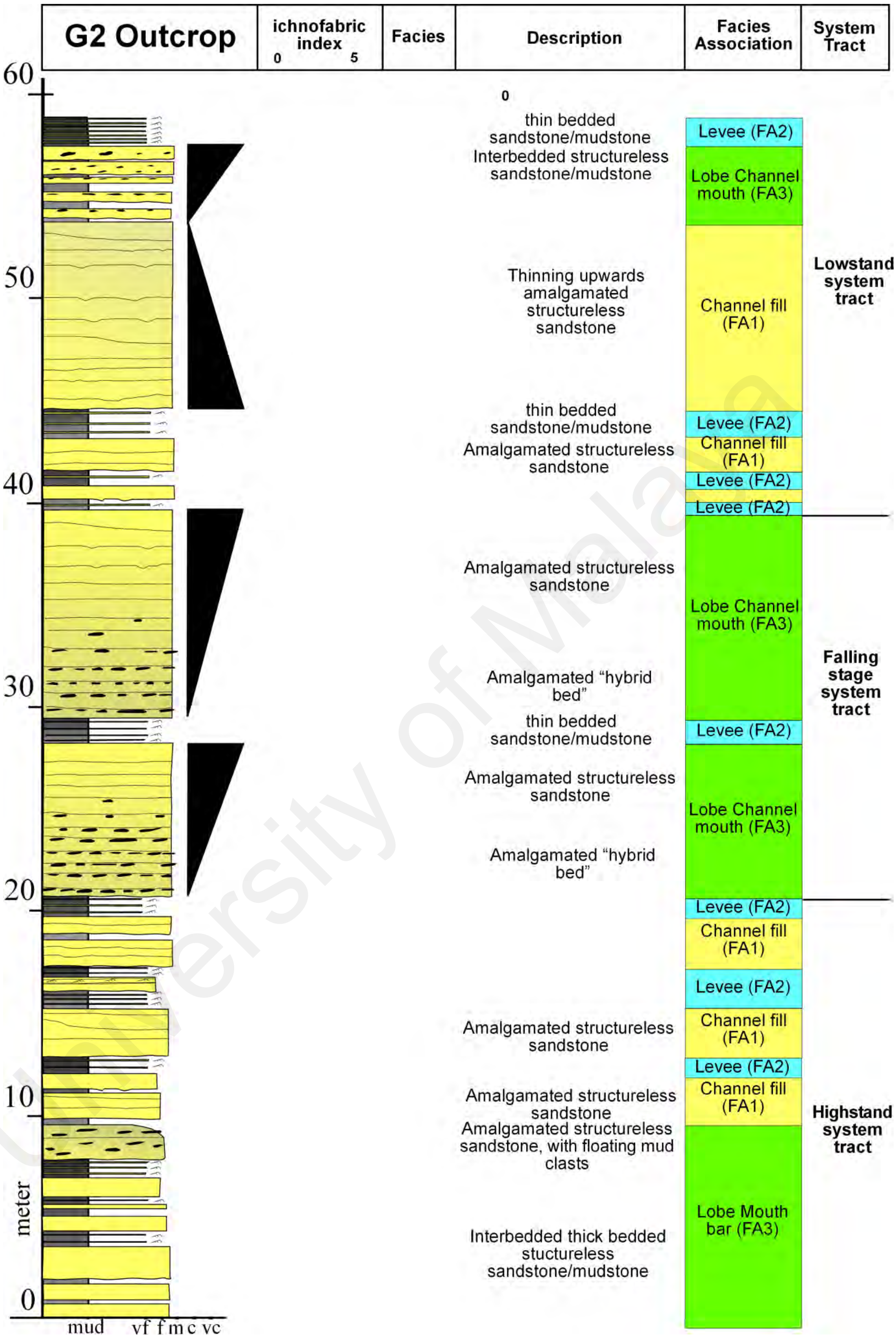


Figure 5.2. Outcrop G2 section. Refer to Fig. 1.1B for outcrop location

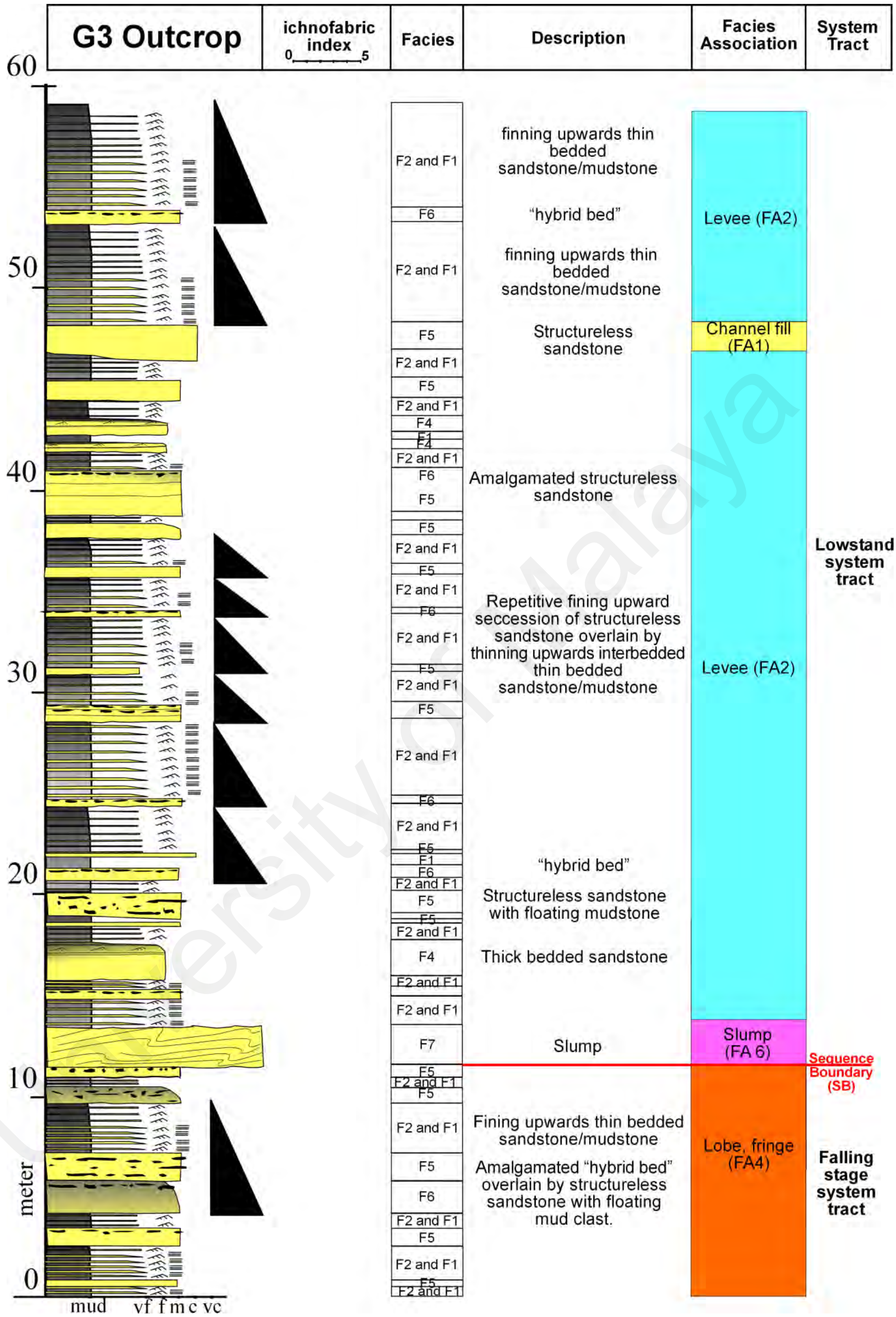


Figure 5.3. Outcrop G3 section. Refer to Fig. 1.1B for outcrop location.

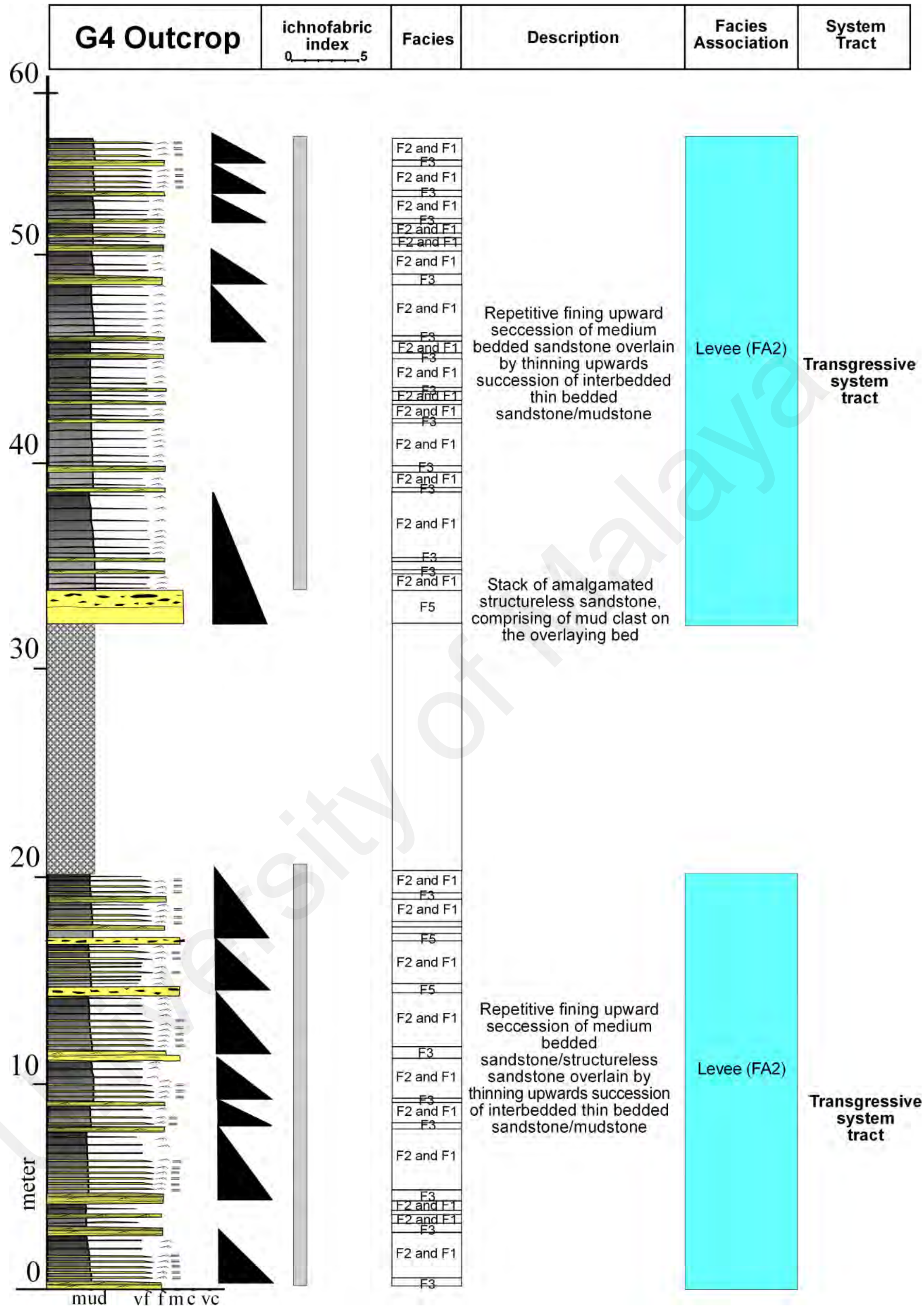


Figure 5.4. Outcrop G4 section. Refer to Fig. 1.1B for outcrop location.

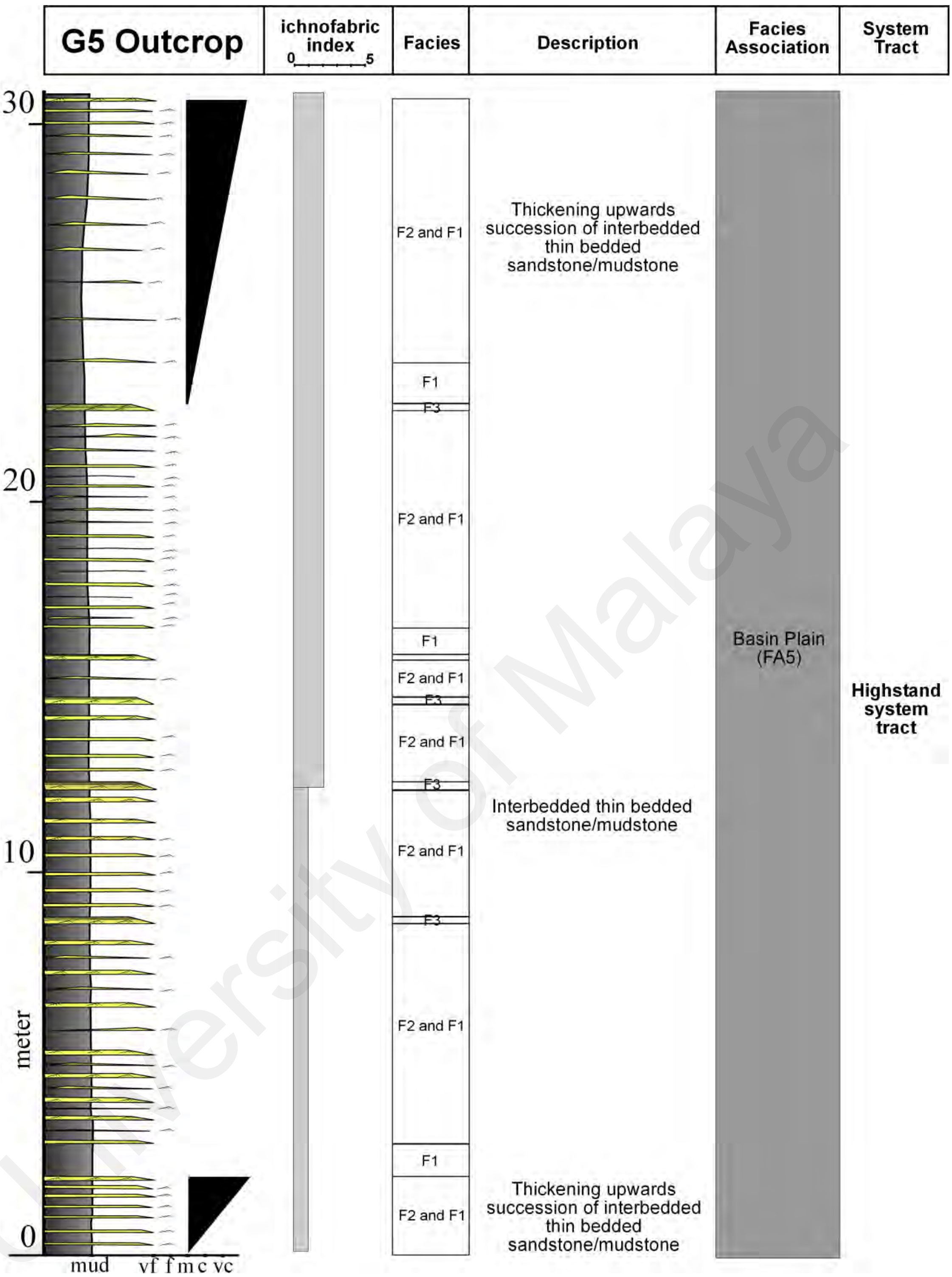


Figure 5.5. Outcrop G5 section. Refer to Fig. 1.1B for outcrop location.

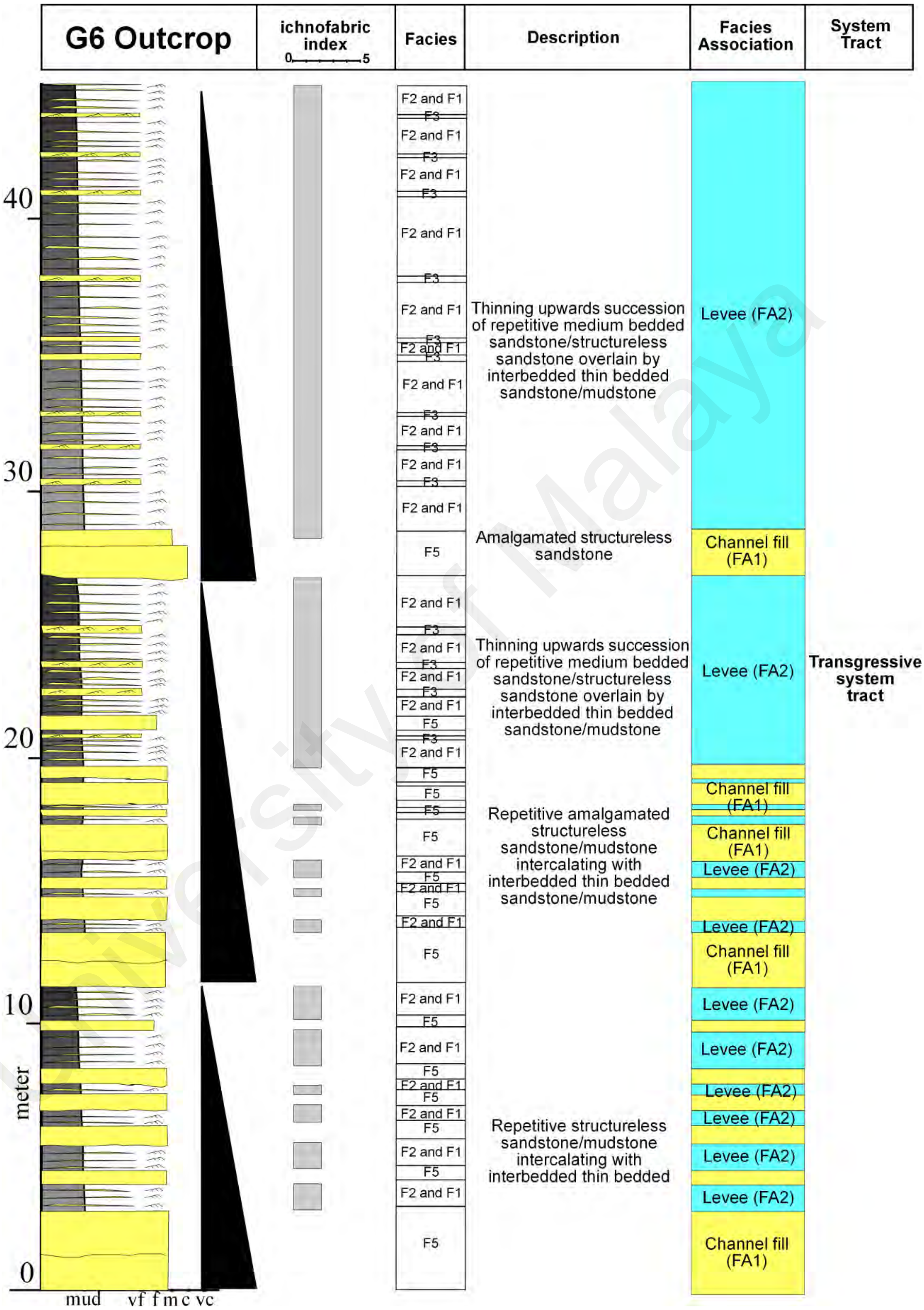


Figure 5.6. Outcrop G6 section. Refer to Fig. 1.1B for outcrop location.

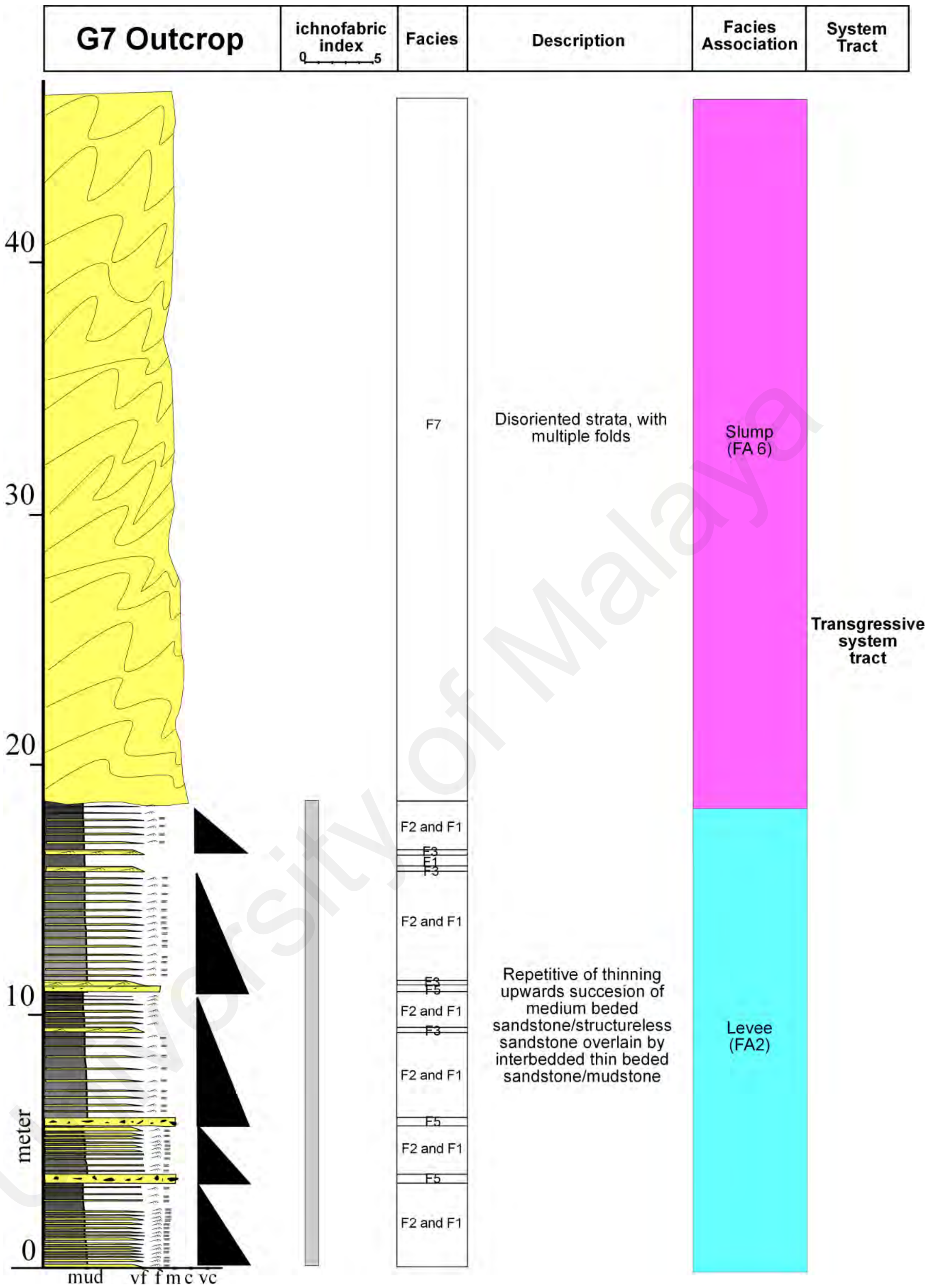


Figure 5.7. Outcrop G7 section. Refer to Fig. 1.1B for outcrop location.

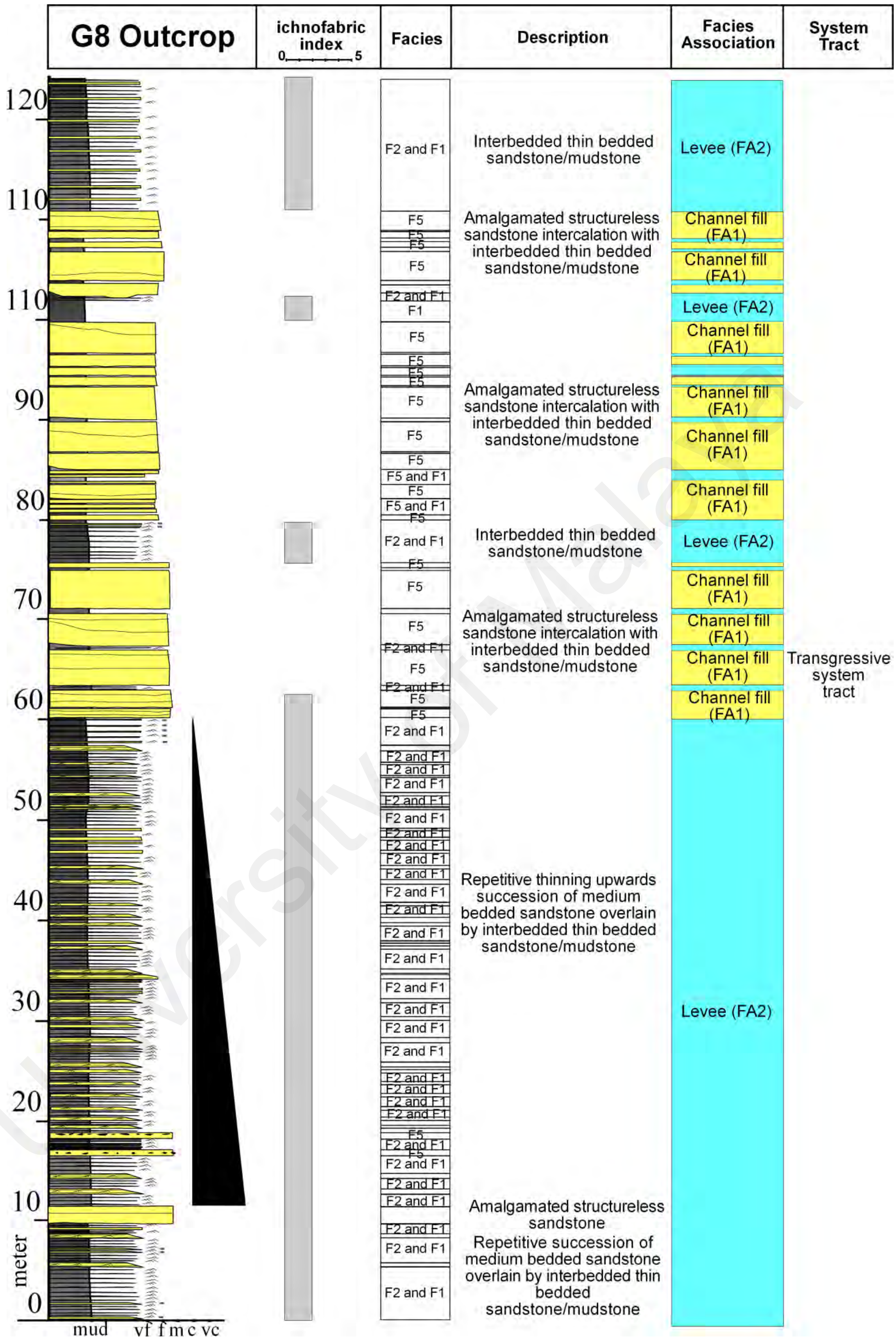


Figure 5.8. Outcrop G8 section. Refer to Fig. 1.1B for outcrop location.

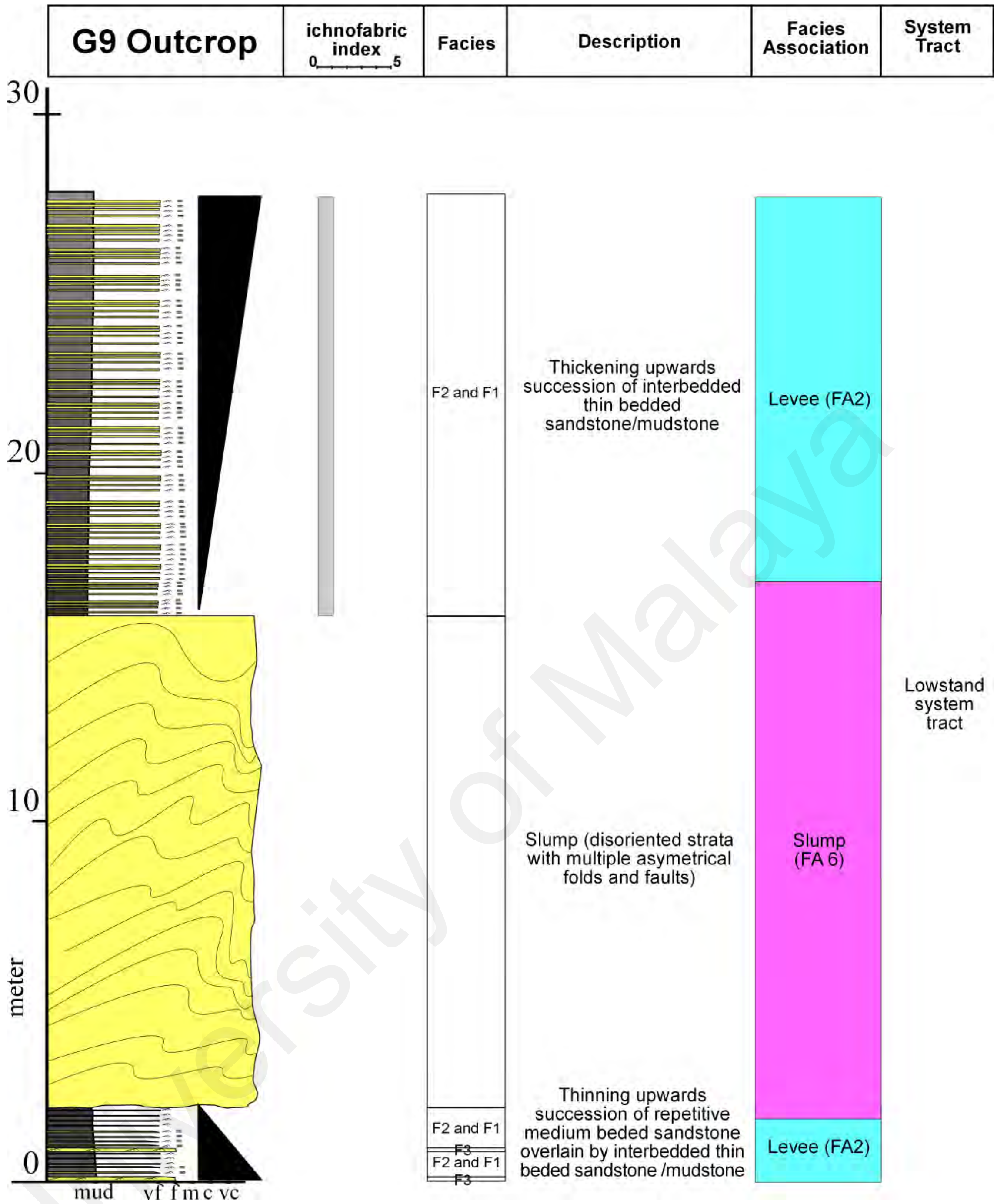


Figure 5.9. Outcrop G9 section. Refer to Fig. 1.1B for outcrop location.

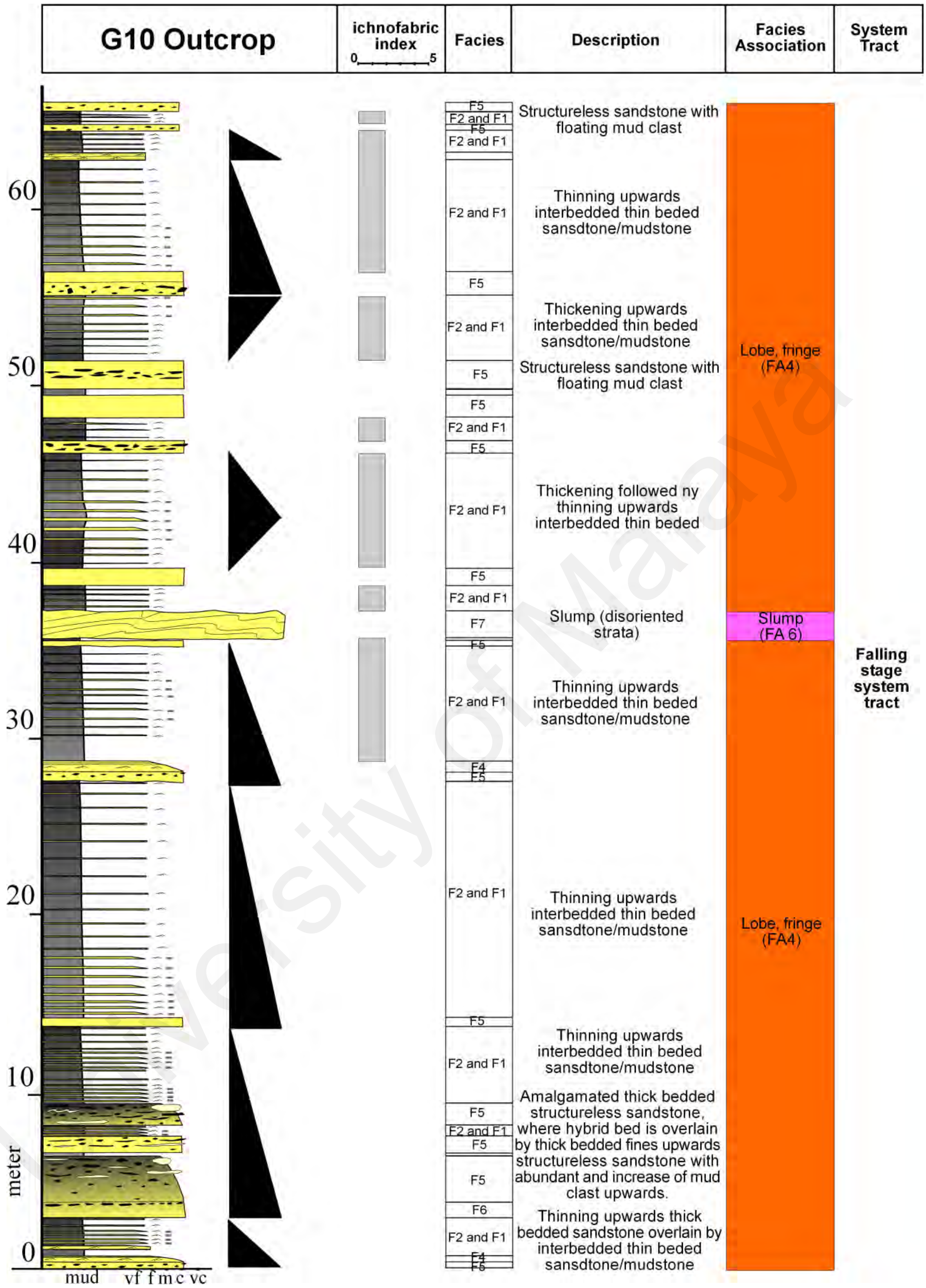


Figure 5.10. Outcrop G10 section. Refer to Fig. 1.1B for outcrop location.

5.2. Progradational Frontal Splays of the Pelagus Member

Thickening upward successions 8-10 m thick, consisting of amalgamated, thick-bedded structureless sandstone of the lobe (channel mouth) facies association (FA3), intercalated with thinner levee deposits (FA2) in the Pelagus Member are interpreted as the deposits of prograding, fan-shaped lobes (e.g. Mutti, 1977).

Lobe (fringe) deposits (FA4) present in some studied sections of the Pelagus Member **(Fig. 5.3 and 5.10)** are interpreted as representing the distal part of a frontal splay system or fringe. The sheet-like and tabular geometry of the beds are consistent with an unconfined environment (Mattern, 2002). Thick levee deposits (FA2) overlying the Lobe (fringe) facies association (FA4) indicates progradation with associated basinward shift in depositional loci.

Although the composition within lobes is sand rich, mud can be deposited during the absent of flow or transported as suspension within turbidity flows. This is indicated by the presence of poorly developed levee deposits intercalated between the thickening upwards lobe deposits**(Fig. 5.2)**. The intercalation of channel fill and levee deposits overlying lobe deposits is interpreted as indicating either lateral migration of the depositional lobes or the presence of multiple channels (braided type of channel) (Posamentier and Kolla, 2003; Posamentier and Walker, 2006).

Similar erosional features also appear in the turbidite facies of the Carboniferous Upper Ross sandstone in Western Ireland (Lien et al., 2003) and in the northern Apennines (Mutti and Lucchi, 1972). In the case of the Upper Ross sandstone, the thickening upward successions are in close relationship with channel fill deposits, which they interpreted as not being frontal lobe deposits, but spill-over deposits instead.

However, Mutti and Lucchi (1972) interpreted the thickening upward succession in the northern Apennines as the deposits of progradational depositional lobes. Similar stacking pattern and amalgamation is also described in the Karoo Basin, South Africa by Hodgson et al., (2006), exposing progradational and aggradational phases. Here, isolated channels cutting into thin bedded turbidites are commonly associated with retrogradational successions. Similar isolated channels are also present in the Pelagus Member, where they underlie thickening upward successions **(Fig. 5.2)**. This is consistent with their interpretation as part of the retrogradational phase of deposition, with the overlying thickening upward succession representing the subsequent progradational phase.

It is difficult to differentiate between high density turbidites and debrites, as both processes form structureless sandstone (Saito and Ito, 2002; Mutti et al., 2003; Shanmugam, 2000), and the beds in the Pelagus Member are not significantly exposed laterally. However, structureless sandstone beds in the lobe deposits of the Pelagus Member are commonly associated with floating mud clasts **(Fig. 3.4C-F)** and resemble the D3-type hybrid beds of Hodgson (2009). This suggests deposition by high density turbidity flow. D3 hybrid beds (Hodgson, 2009) are typical of lobe fringe deposits, which are characteristic of the distal part of splays/lobes. The initial flow carrying the sediments within the hybrid beds was probably a long-lived high density turbidity flow, favouring long distance sediments entrainment, and enabling the deposition of structureless sandstone at the distal part of lobes. Amalgamation of beds suggests that the duration between flows was short (Khan and Arnott 2011).

The lobes are interpreted to have been formed during the falling stage- lowstand system tracts. Sand-rich sedimentation is usually restricted landward of the shelf-edge during the transgressive and highstand systems tracts. Large amounts of coarse grained sediments are usually transported basinward of the shelf-edges during the falling stage-lowstand phase of sea level, when sediments are reworked basinwards due to the rejuvenation of extended fluvial channels (Posamentier and Walker, 2006), slope instability or tremors. Thickening upward deepwater successions such as those observed in the Belaga Formaiton, are characteristic of progradational phases associated with the FSST and LST. The beds within the thickening upwards successions display erosional features. Similar erosional features also appear in progradational successions of the Upper Ross (Lien et al., 2003; Mutti and Ricchi Lucchi., 1972). In the case of the Upper Ross, the thickening upward successions are in close relationship with channel fills. These deposits have been interpreted by (Lien et al., 2003) as being spillover deposits, rather than frontal lobes. However, Mutti and Lucchi (1972) interpreted the thickening upwards successions as the deposits of progradational depositional lobes. The channel-fill deposits overlying the thickening upward successions in the Belaga Formation are interpred as indicating continued basinward progradation and associated drifting of sediment loci during relative sea level fall **(Fig. 6.4A-C).** Increasing sediment supply delivered by channels resulted in erosion and reworking of preceding strata, including channels and antecedent lobes, during relative sea level fall. The channels extended further basinward, eventually extending the transition point **(Fig. 6.2).** The transition point marks the point where the levee height is no longer able to fully confine the sand prone, high concentration part of the turbidity flow (Posamentier and Walker, 2006). This results in overflow and deposition of a sheet-like fan further basinward. Low relief and shallow leveed channels can be associated with frontal splays. The intercalation of thin levee deposits indicates lateral shifting of channels (Posamentier and Walker, 2006).

The occurrence of thick levee deposits (10-30 m) in the Pelagus Member **(Fig. 5.1, 5.2, 5.3)** indicates the presence of a leveed channel system. However, channel and levee deposits are only observed in vertical contact with other. No outcrops expose the lateral change between the 2 facies associations. Thick levee deposits overlying the Lobe (fringe) facies association (FA4) at the G3 **(Fig. 5.3)** section indicates a basinward shift in depositional loci. The levee deposits are believed to be parts of a single leveed channel system, because they display consistent parallel bedding orientation and also single direction of ripple cross-lamination. The levee deposits of the Pelagus Member display relatively thick beds (1-150 cm). Slump intervals are also common, as well as deposits interpreted as crevasse splays. Slumps due to levee failure are a common occurrence, particularly along the outer bend since it is steeper compared to the inner bend (Posamentier and Walker, 2006), implying a meander leveed channel system.

5.3. Leveed Channel System of the Kapit Member

Leveed channel successions are up to 120 m thick and are composed of tens of metres thick intervals of levee deposits (FA2) in close vertical relationship with equally thick channelfill deposits (FA1)(Fig. 5.8). The levee facies association (FA2) is interpreted as overbank deposits from a single sourced, confined channel system. The channel fills are characterised by thick bedded amalgamated sandstones, with an erosional base displaying a concave upward profile. Intercalated levee deposits are interpreted as reflecting channel lateral migration in a meandering system **(Fig.**

5.11). This interpretation involves a single flow event with lateral accretion, similar to point bar development in fluvial systems (Posamentier and Walker, 2006). The development of sinuous leveed channels is most likely due to the low sand:mud ratio, which aids in enhancement of the highly sinuous leveed channel system (Possamentier and Kolla, 2003; Pirmez et al., 2000). Leveed channels typically develop during the late lowstand to transgressive phase, during which the system is more mud rich (Posamentier and Kolla, 2003). As the phases progress, the channel becomes more sheet-like and intercalation with levee becomes more frequent, indicating the development of channels having low width:depth ratio and greater channel relief (Pirmez et al., 2000).

The levee deposits of the Belaga Formation are very similar to those described from the Late Miocene Mount Messenger Formation of the Taranaki Basin (Rotzien et al., 2014) and the Neoproterozoic Isaac Formation of southern Canada (Khan and Arnott, 2011) and mudstone and thin-bedded turbidites associated with Upper Cretaceous Wheeler Gorge Conglomerate, California (Walker, 1985), displaying thinning upwards of medium to thin 595 bedded turbidites packages, similar facies to F3 and F2.

Variations in levee bed thickness observed in the Belaga Formation possibly reflect relative distance from the channel levee crest. Thicker beds comprising of 3-10 cm thick F2, with 30 cm thick F3, common occurrence of 30-40cm thick F5, with individual levee packages 2-7 m thick are interpreted as proximal levee deposits **(Fig. 5.1, 5.3)**, while thinner beds of F2 (less than 3 cm) and F3 (15-25cm thick), rare occurrence of dm thick F5, with individual levee packages 50-200cm thick, represent distal expressions of levee deposits **(Fig. 5.4, 5.6, 5.7, 5.9)**.

The presence of inferred crevasse channel deposits associated with the levee deposits in the Belaga Formation **(Fig. 5.3)** suggests that the levee deposits were developed along the outer bend of channels (Posamentier and Kolla, 2003).

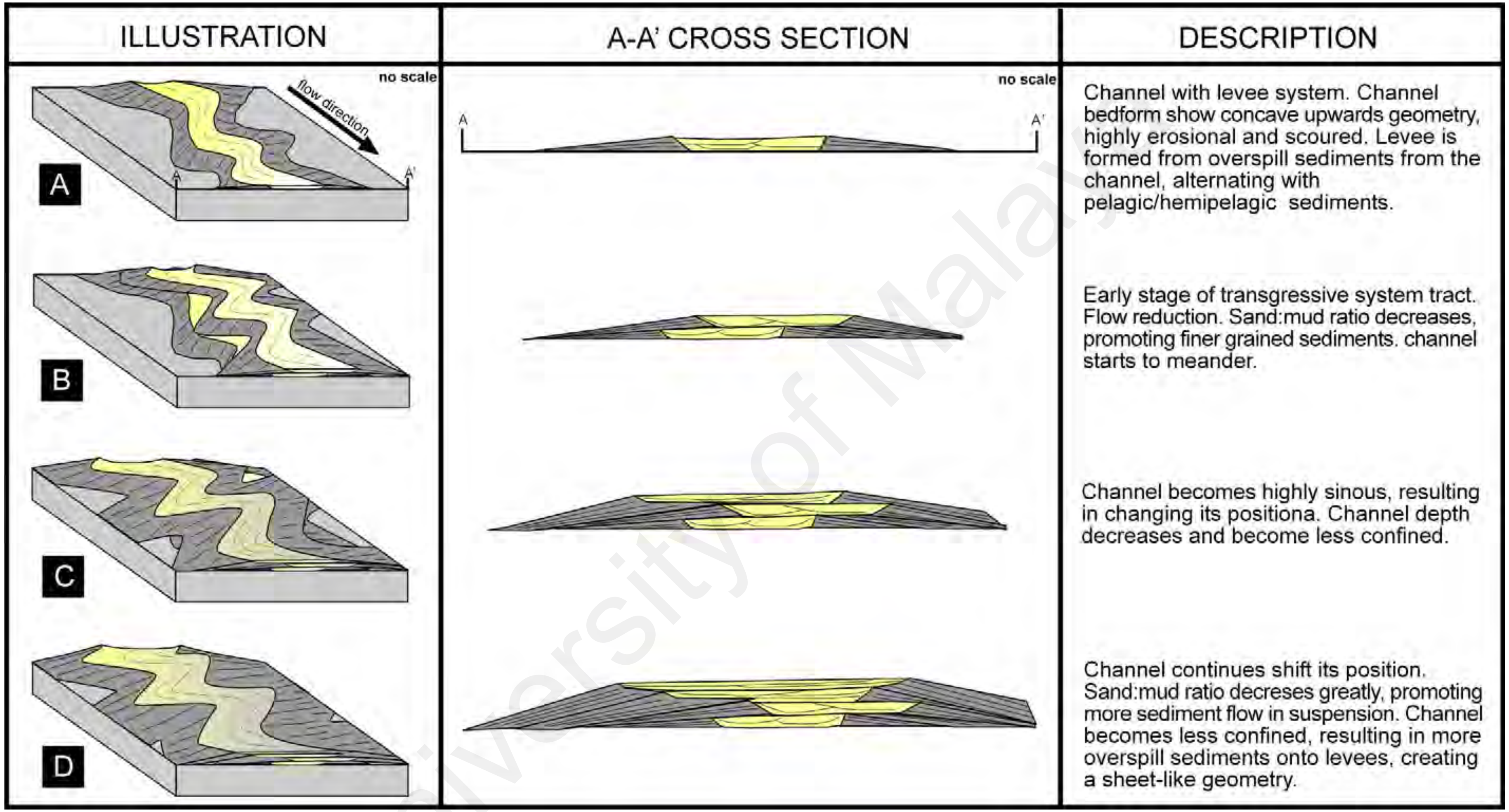


Figure 5.11. Illustration of a migrating leveed channel deposits in a single flow leveed channel. A-D show decrease in sand:mud ratio, resulting in the formation of leveed channels with meander and shallowing the channel's depth while widening at the same time, changing the bed from concave upward/spoon like to sheet-like geometry.

CHAPTER 6: DISCUSSION

6.1. Effect of sea level fluctuation on fan architecture

A depositional model for the Paleocene to middle Eocene Belaga Formation is constructed based on the vertical and lateral arrangement of facies associations in outcrops around Sibu, Sarawak **(Fig. 6.3)**. However, there are limitations to the interpretations that can be made, due to the limited lateral extent of outcrops and the steep dip of strata. This also makes high resolution correlation between outcrops impossible. Thus, the interpretation is mainly made based on the similarity of facies and facies association in outcrops, comparing with other outcrops from the other parts of the world. In this study, the writer attempts to interpret the larger scale depositional history of the Kapit and Pelagus members in terms of relative sea level change **(Fig. 6.1)**

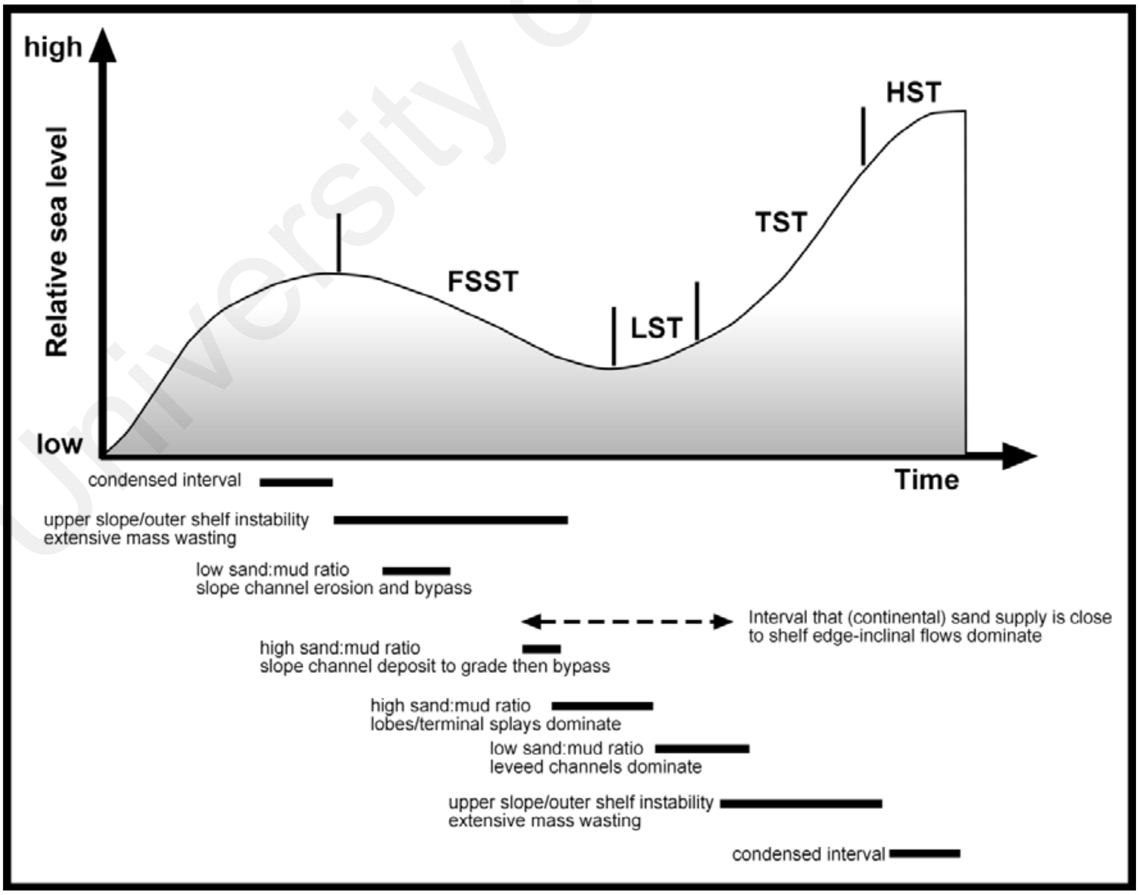


Figure 6.1. Sediment transport depiction related to the sea level changes (modified from Arnott, 2010; originally from Posamentier and Walker, 2006)

6.1.1. Falling Stage Systems Tract (FSST)

The Falling stage systems tract (FSST) (Plint and Nummedal, 2000) is also termed the early lowstand system tract by Posamentier and Allen (1999).In the beginning of the falling stage, shelf deposition moves further basinwards, which reinitiates deep water system deposition beyond the shelf-edge, characterised by thick, mud-rich, high density flows. During this stage, the slope is incised, with new sediment input, reactivated from the shelf consisting of thick mud rich, high density flows and high transport efficiency. Consequently, the gravity flow character is changing and out grade with the existing slope, triggering erosion. This produces an erosional zone in the upper and mid slope, and a bypass zone near the base of the slope. Coarse sediment is transported and deposited on the basin floor, supplied from the eroded slope or shelf. Slumping is common during this stage due to the extensive exposure of the shelf, causing mass transport deposits (MTD's) (e.g. Algar et al, 2011), which can have dimensions of 50-100 m thick and cover an area of up to 10000 km2 (Piper et al., 1997).

Fining upward, lobe fringe facies associations observed in the Belaga Formation probably represent deposition during submarine fan initiation (e.g. G3 and G10 sections) (Haughton, 2009). Sediment at the lobe fringe was supplied from the shelf or slope, due to over-steepening or unstable slope, resulting in sediment failure, triggering the initial flow. The flow rushed through channels and at the same time eroded the underlying sediments from the slopes and fan comprising thin bedded sand/silt/mud from the previous highstand systems tract. The entrained sediments were then transported further basin wards by turbidity flow. The flow is supported by low friction created by the previous turbidites, allowing the flow to bypass the fan and become deposited further basinward (Mohrig et al., 1998; Haughton et al., 2003; Hodgson, 2009). As the flow reached the basin, it became unconfined and spread in a fan shaped geometry **(Fig. 6.4A-C).**

At the outcrop **(Fig. 5.3, 5.10)**, the falling stage systems tract is characterised by repetition of thick bedded (50-100 cm) amalgamated channel fill and lobes facies association, alternating with relatively thinner pelagic and hemipelagic mud or levee facies association (<100cm).

The presence of small scale to large scale slumps of typical basin sediments and characterised by interbedding of thin to very-thin bedded turbidites, maybe an indicator that the basin was experiencing relative sea level fall, resulting in massive and extensive in erosion, which eventually caused slumping to occur.

6.1.2. Lowstand System Tract (LST)

The Lowstand systems tract is considered as deposition after maximum relative sea level fall, which includes accumulation after the onset of relative sea level rise. LST lies above the FSST and is overlain by a transgressive surface. During the beginning of the lowstand systems tract, the sedimentation towards the basin is significant. As the relative sea level is dropping, the sand:mud ratio increases, resulting in decrease of transport efficiency. Submarine canyons act as erosion and bypass areas. Channels are generally poorly confined and have low sinuosity. Channels act as conduits for transporting sediments toward basin. With time, sand:mud ratio increases, causing the channel gradient to become less than that of the grade, promoting extensive and aggradation within channels. As relative sea level starts to rise during the end of the lowstand systems tract, sheet-like sandstone bodies, consisting of amalgamated channel sandstone which gradually becomes more interstratified with fine grained strata upwards, start to develop (Arnott, 2010)**(Fig. 5.2)**.

The beginning of the LST is indicated by the coarsening upward successions shown by the increase in thickness of fine to medium grained sandstone turbidites/debrites **(Fig. 6.4).** The source of sediments is interpreted to have come from the shelf and eroded submarine canyons. Erosion was probably initiated by over-steepening. Sandstone within channels shows absent to very weak grading and are amalgamated, which is interpreted as products of turbidity flows. This is supported by the presence of deep scours (>2 cm), indicating turbulent flows, and aggrading deposits can occur during steady and uniform turbidity flows (Kneller and Branney, 1995). This thick amalgamated bed succession is overlain by what is interpreted as channel fill, comprising onlapping/offlapping beds, indicating complex erosion and filling, happened during the end of LST. The occurrence of structureless sandstone beds indicate high deposition rates, which occur at channel proximal localities (Kane et al., 2007). The reason for channel fill deposition is that during the end LST, relative sea level starts to rise, resulting in the flow with low energy. The sediment supply was relatively high preceding the FSST and beginning of LST. This promotes the development of an extensive submarine fan, as a result of the change in position of the depositional loci farther basin ward. The channels that previously act as conduits become depositional site of the relatively coarser sediments within the flow. This reduces the sand:mud ratio within the flow, improving transport efficiency, enabling the entrainment of finer grained sediment, extending the channel and lobe formation distalwards **(Fig. 6.4D).**

6.1.3. Transgressive Systems Tract (TST)

The transgressive system tract develops as the rate of relative sea level increases as well, resulting in starve of the sediment supply. This leads to retrogradational stacking of fans (Vail, 1987). At the early TST sediment supply towards the basin is gradually diminishing, as the sediments are retained on the shelf during relative sea level rise (Posamentier and Walker, 2006), resulting in decrease in sand:mud ratio. Consequently, flows that bypass the canyon will create channels characterised by high sinuosity and confined at the middle and lower parts of the slope. The high sinuosity results in the development of lower depth:width ratio and greater channel relief (Primez et al., 2000). These conditions produce a well-developed fine grained levee bordering erosive based, lateral-migrating confined channels. With time, sediment supply become more diminished further downflow, reducing the sand:mud ratio. Enhanced levee development causes the leveed-channel-to- lobe transition to prograde basinwards (Saller et al., 2008). Eventually, due to the comparatively low sediment influx, depositional-lobe complexes will begin to retrograde **(Fig. 6.4E).** However, Posamentier and Walker (Fig. 29, 2006) suggested that during sea level rise, flows of sediments are relatively muddier due to the relatively low sand:mud ratio. This results in the extension of leveed channel formation farther basinward. In the outcrops **(Fig. 5.6, 5.8)**, this TST is characterised by thick succession (up to 15 m) of alternating thin-very thin bedded sandstone and mudstone levees facies association, overlain by succession of channel deposits comprise of thick bedded (1-3 m) fine-medium grained sandstone where amalgamations are common. Levees are commonly conformably overlain by channel deposits. The presence of levees intercalating with channel fill deposits and incision of levees by channel fills are most likely because of the temporal shifting of channel position due to lateral migration as an consequence of TST where channels become highly sinuous.

Levee deposits are characterised by thinning upwards cycles (Walker, 1985; Normark et al., 1997; Skene et al., 2002; Beaubouef, 2004), which are exclusively interpreted as deposits from the channelised turbidity currents, where levee height is connected to through-channel and overbank flow character (Haughton, 2009; Hiscott et al., 1997). No sand- sized particles overtop the levee crest (Chough and Hesse, 1980) from overspiling turbidity currents, which explain that during the channel migrate/shift its position laterally in highly confined and sinuous channel, the deposition of levee is shifted as well and might be one of the reasons for mudstone deposition between channel fill deposits in one of studied outcrops **(Fig. 5.8)**.

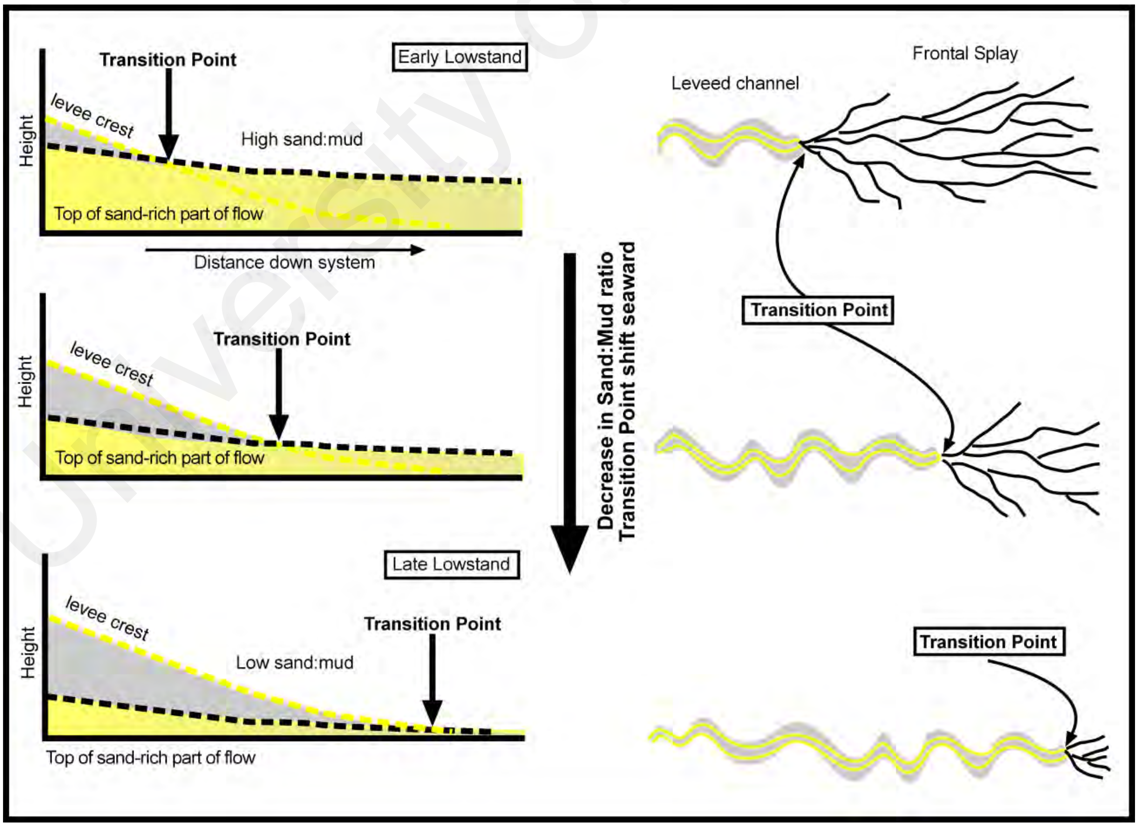


Figure 6.2. Illustration on "transition point" extension during the Early Lowstand - Late Lowstand system tracts (modified from Posamentier and Walker, 2006).

Deep water channels are similar to fluvial channels in always developing a longitudinal profile graded to a base level, in this case the gravity base (Posamentier and Walker, 2006). Abreu et al. (2003) described lateral accretion sets in a deep-water meandering channel in off-shore West Africa, which mimics the stratigraphic architecture of fluvial point bars. Deep marine channels become poorly confined at the downflow end of the lobe (Pirmez et al. 2000) and rather than forming a single channelised system, the deposition displays a sheet-like bedding geometry, characterised by shallow channels and being sand-prone. Possamentier and Walker (2006) described a point called the "transition point", which is the point where levees no longer effectively confine the basal sand-rich channel deposits **(Fig. 6.2).** Poorly confined channels are commonly display a steeper margin on the outer-bend (Arnott, 2010) and a step-flat base morphology, indicating lateral migration of the channel system (Eschard et al., 2003; Navarro et al., 2007). Along the steep margin, the levee strata are either in erosional contact with channel strata or separated by at thin, fine-grained bypass unit (Beaubouef et al., 1999), whereas along the inner side, the channel-fill strata either onlap or grade laterally into levee deposits (Arnott 2010). This may explain the vertical stacking pattern of alternating channel fill and levee deposits in some of the studied outcrops (outcrop G6 and G8).

6.1.4. Highstand systems tract (HST)

The highstand systems tract occus as relative sea level rise slows down and the rate of sediment supply outpaces it, resulting in the progradational stacking. During the highstand system tract, submarine fan deposition ceases and deposition retreats shelfward. Consequently, channels also retreat towards onshore due to the shortage of sediment supply from the shelfs. Thus, deposition in the deep marine basin is dominated by fine grained siliciclastics, biogenic materials and pelagic sediments (Arnott, 2010). However, Posamentier and Walker (2006) argue that sediment supply towards the basin is not totally stopped although it is much reduced, and minimal amount of sediments are deposited at the shelf edge. In fact, during the highstand system tract, finer grained channels and levees could still be formed, with little or no lobe formation **(Fig. 6.4F).**

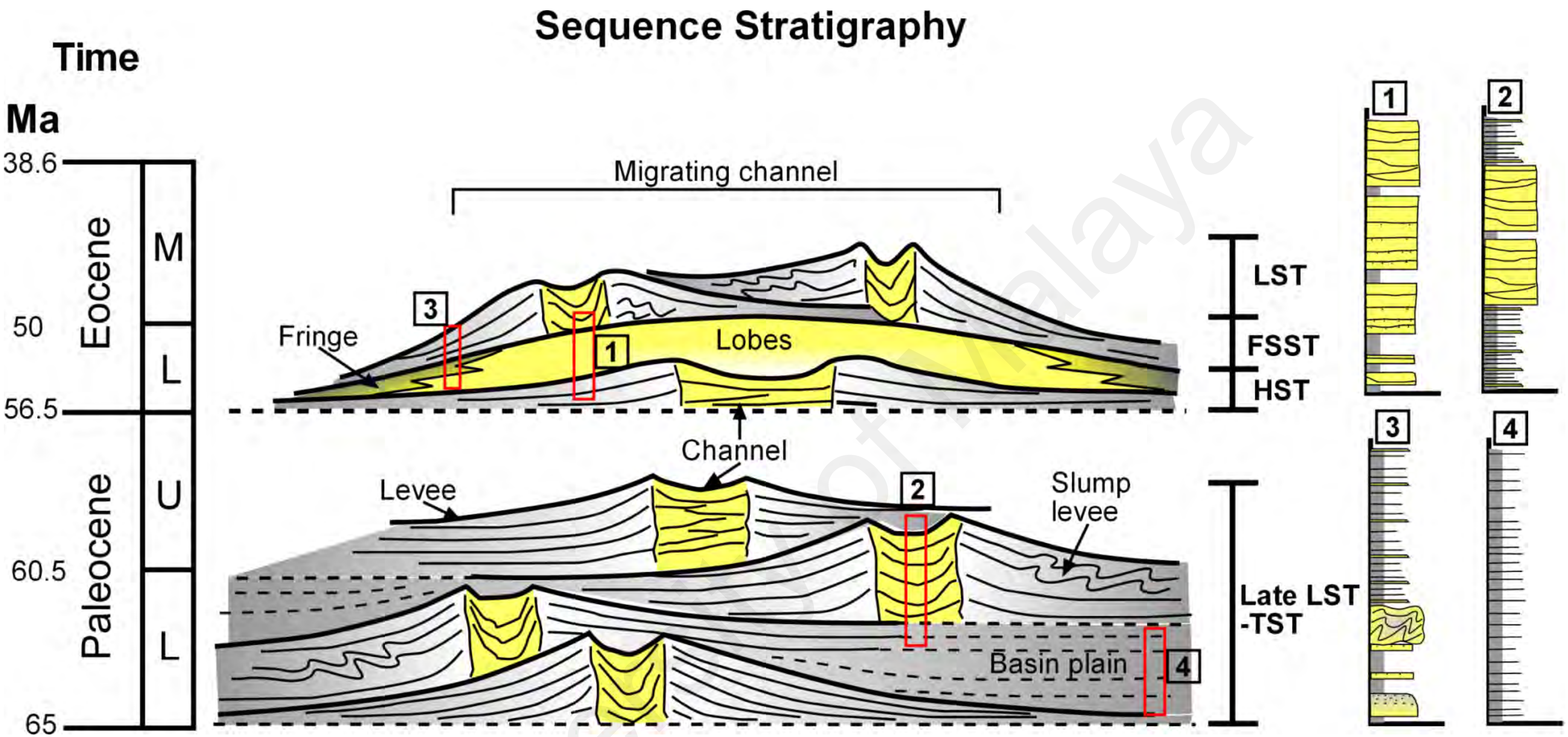


Figure 6.3. Sequence stratigraphy of Paleocene-Eocene Belaga Formation. Paleocene sediments (Kapit member) show Late Lowstand system tract (LST) to Transgressive system tract (TST) characteristic, displaying well developed channel and multiple channel migration. Eocene sediments (Pelagus member) show Highstand system tract (HST) characterised by small leveed channel overlain by Lobe deposits which mark the initiation of Falling stage system tract (FSST) and being overlain by lobes are overlain by leveed channel, marking the period of Lowstand system tract (LST)

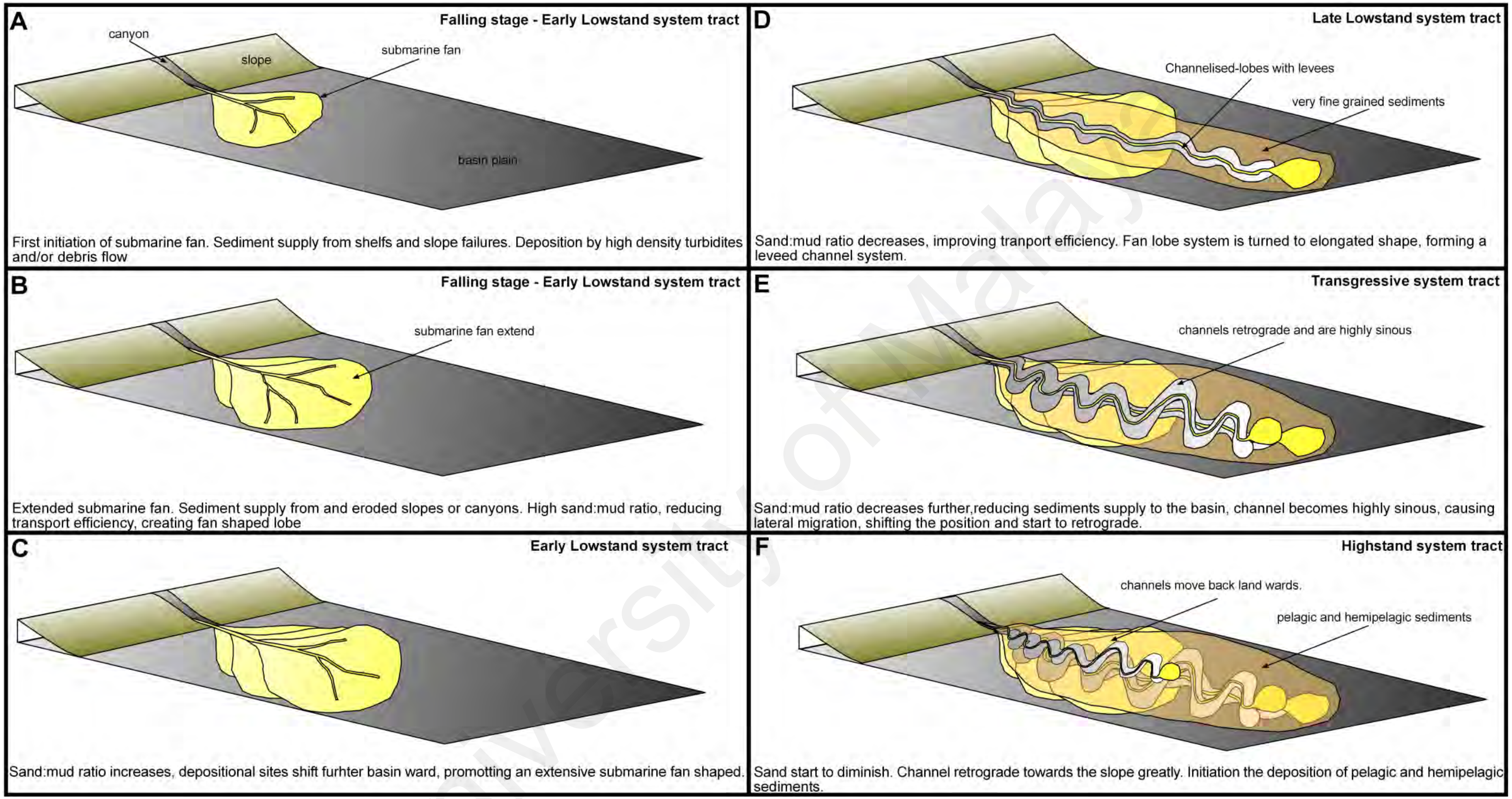

A
Falling stage - Early Lowstand system tract
canyon
slope
submarine fan
basin plain
First initiation of submarine fan. Sediment supply from shelfs and slope failures. Deposition by high density turbidites and/or debris flow
D
Late Lowstand system tract
Channelised-lobes with levees
very fine grained sediments
Sand:mud ratio decreases, improving tranport efficiency. Fan lobe system is turned to elongated shape, forming a leveed channel system.
B
Falling stage - Early Lowstand system tract
submarine fan extend
Extended submarine fan. Sediment supply from and eroded slopes or canyons. High sand:mud ratio, reducing transport efficiency, creating fan shaped lobe
E
Transgressive system tract
channels retrograde and are highly sinous
Sand:mud ratio decreases further,reducing sediments supply to the basin, channel becomes highly sinous, causing lateral migration, shifting the position and start to retrograde.
C
Early Lowstand system tract
Sand:mud ratio increases, depositional sites shift furhter basin ward, promotting an extensive submarine fan shaped.
F
Highstand system tract
channels move back land wards.
pelagic and hemipelagic sediments
Sand start to diminish. Channel retrograde towards the slope greatly. Initiation the deposition of pelagic and hemipelagic sediments.

Figure 6.4. Illustration of deep marine depositional evolution due to the effect of sea level fluctuation. A) Shows the early formation of submarine fan as the entrained sediments from the confined canyon reaches the unconfined basin, changing the depositional shape from channelised to sheet geometry. B and C) As relative sea level falls, more sediments is transported basinward, sand:mud ratio increases, promoting more sediment from onshore or/and reworked sediments from the shelf and slope, even reworking antecedent strata from the channel, extending the sediment loci farther basinward, consequently extending the transition point, resulting in the formation of fan farther basinward. D) During the late lowstand stage. sand:mud ratio is relatively low compared to early lowstand, consequently extending the transitional point farther basinward and promoting leveed channel formation, with a frontal splay formation at the end of the channel. E) During the transgressive stage, sand:mud ratio decreases greatly, leveed channels become highly sinous, transitional point retreats onshore, resulting in back-stepping of lobes. F) During the highstand stage, channel retreat due to the diminishing sediment input, resulting in the deposition of very-fine grained, hemipelagic and pelagic deposits.

During this period, mud rich sediments are deposited. Therefore, when the subsequent lowstand occurs, more extensive channelised systems will be formed due to the low sand:mud ratio, overlying the preceding strata. In the outcrops **(Fig. 5.5)**, the HST is characterised by repetition of thin-very thin (1-10 cm) bedded turbidite sandstone with thin-medium (1-30cm) bedded pelagic/hemipelagic mudstone, which can be up to 15 meters thick. This interbedded sequence does not show any cyclicity (fining/thinning or coarsening/thickening). However, there are occurrences of slight changes in bed thickness vertically, but still under the same thickness category.

6.2. Fan architecture of Palaeocene to middle Eocene Belaga Formation

Based on the facies composition, stratigraphic relationships of facies association and the palaeocurrent trends, the writer infers that both the Kapit and Pelagus members of the Belaga Formation, were deposited within a submarine fan system comprising elements such as leveed channel and frontal splays, with the facies formed by variety of gravity flow processes (turbidity flows, debris flows and associated slumps)**(Fig. 7.2).**

There are two variants of levee deposits observed in the depositional system: (1) thicker bedded levees (e.g. G1, G3 and G7 sections), having relatively thicker beds (1-150 cm), and; (2) thinner bedded levees (e.g. G4, G6, and G9 sections) with bed thicknesses of 1-60 cm. Both variants display the same alternation of facies (F1, F2, F3, F4 and F5). Thicker bedded levees probably indicate a proximal setting, near the levee crest of the outer bend of a sinuous channel. This is supported by the associated occurrence of crevasse channel deposits.

Crevasse channel deposits cut into levee deposits at section G3 **(Fig. 5.3).** However, not all outcrops with thicker bedded levee deposits are associated with crevasse channel deposits. At section G3, thick levee deposits overlie a lobe fringe facies association, implying the presence of adjacent channel fill deposits. Crevasse channel deposits indicate the presence of crevasse splays, and implies channel development in a basin floor environment rather than a slope setting (Posamentier and Walker, 2006).This is because channels at the slope tend to flow parallel to the slope, creating relatively straight channels, instead of meandering ones (Posamentier and Walker, 2006). Channels in basin settings have the tendency to become sinuous, depending on the sand:mud ratio (Reading and Richards, 1994). Crevasse channels form during periods of sudden high sediment input, with sediment entrained within the high velocity channel flow breaching the levee crest and resulting in the levee strata being cut by the overflow sediments. The sediments were probably transported basinward as a consequence of relative sea level fall and/or from single to multiple seismic events related to the active tectonic of the Sarawak Orogeny. Another evidence of outer bend levee deposition is the common presence of slumps. Slumps are more likely to happen during the falling stage and beginning of lowstand phase, when there was significant basinward shift of depositional loci, producing mass transport deposits (MTD's) which can have dimensions of 50-100 m thick and 10000 km^2 (Piper et. al., 1997; Arnott, 2010).

Stratigraphically, the slump deposits are situated at the boundary between lobe (fringe) (FA4) and levee facies associations (FA2). Slumps are commonly associated with levee deposits, especially at the outer bend since it has steeper slopes compared to the inner bend (Posamentier and Walker, 2006). Channel deposits in the Kapit Member sharply overlie thick levee successions **(Fig.5.8).** The levee deposits are relatively thinner-bedded (1-30 cm) compared to the levee deposits formed at the inner bend of channels **(Fig. 5.4, 5.6 and 5.9)**. It is possible that these levee deposits might not have been deposited particularly within the inner bend, but are the distal parts of either inner- or outer- bend levees, which display the same thinning upward characteristics of distal outer bend levee from DOB1 Isaac Formation, Southern Canada (Khan and Arnott, 2011). This relationship between channel and levee is similar to the model from graded conglomerates of the Upper Cretaceous Wheeler Gorge Formation (Walker, 1985), where part of the distal levee is overlain by thick channel succession as well as channel being overlain by thick levee succession **(Fig. 6.5).**

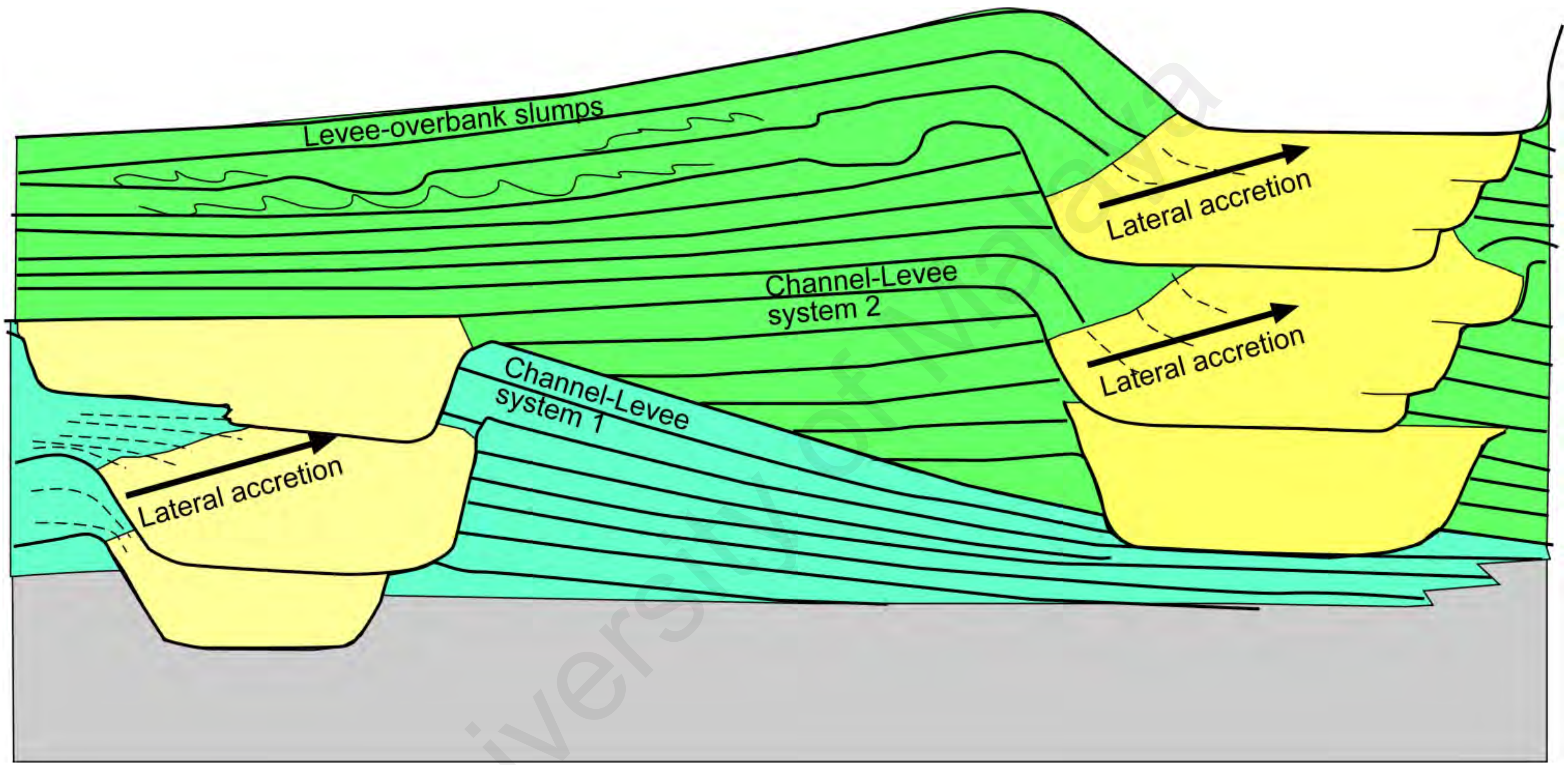
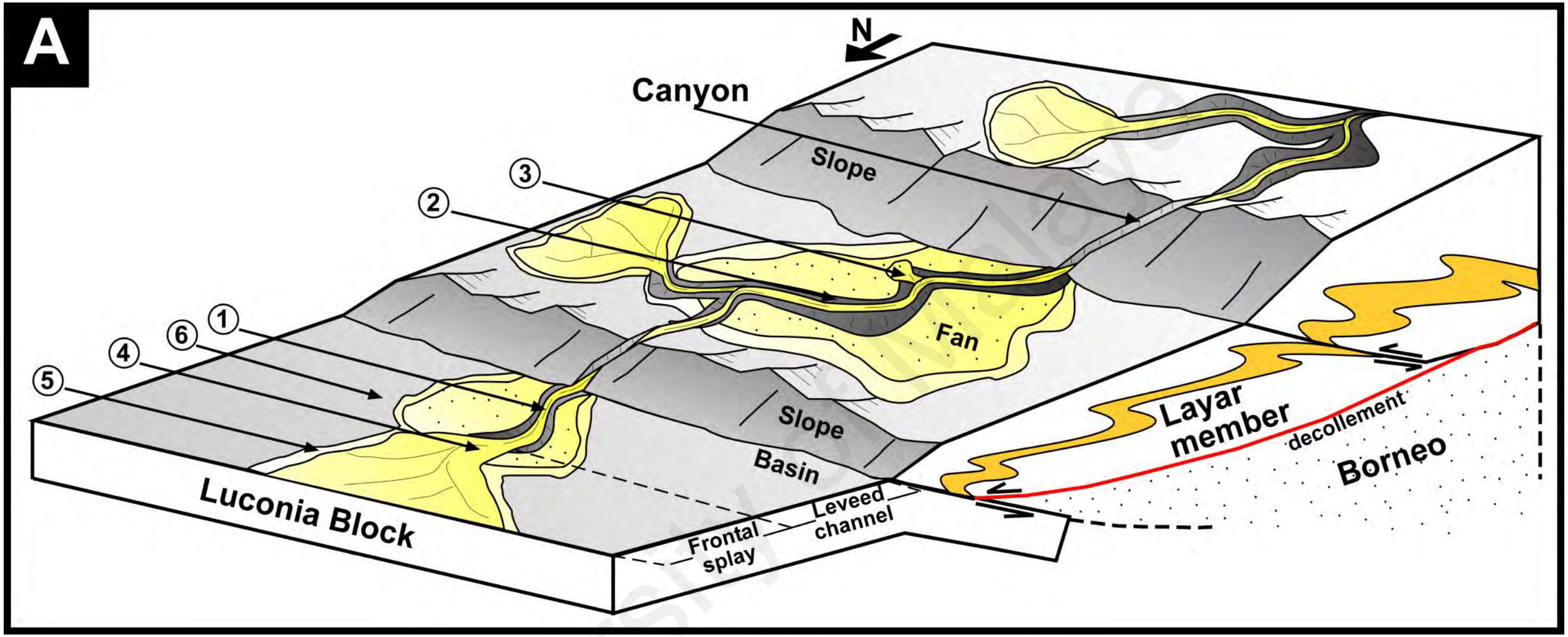


Figure 6.5. Illustration of interpreted leveed channel system of the Kapit Member. Modified after Walker, 1985.


A
N
Canyon
Slope
3
2
1
6
4
5
Fan
Slope
Basin
Layar
member
decollement
Borneo
Luconia Block
Leveed
channel
Frontal
splay

continued

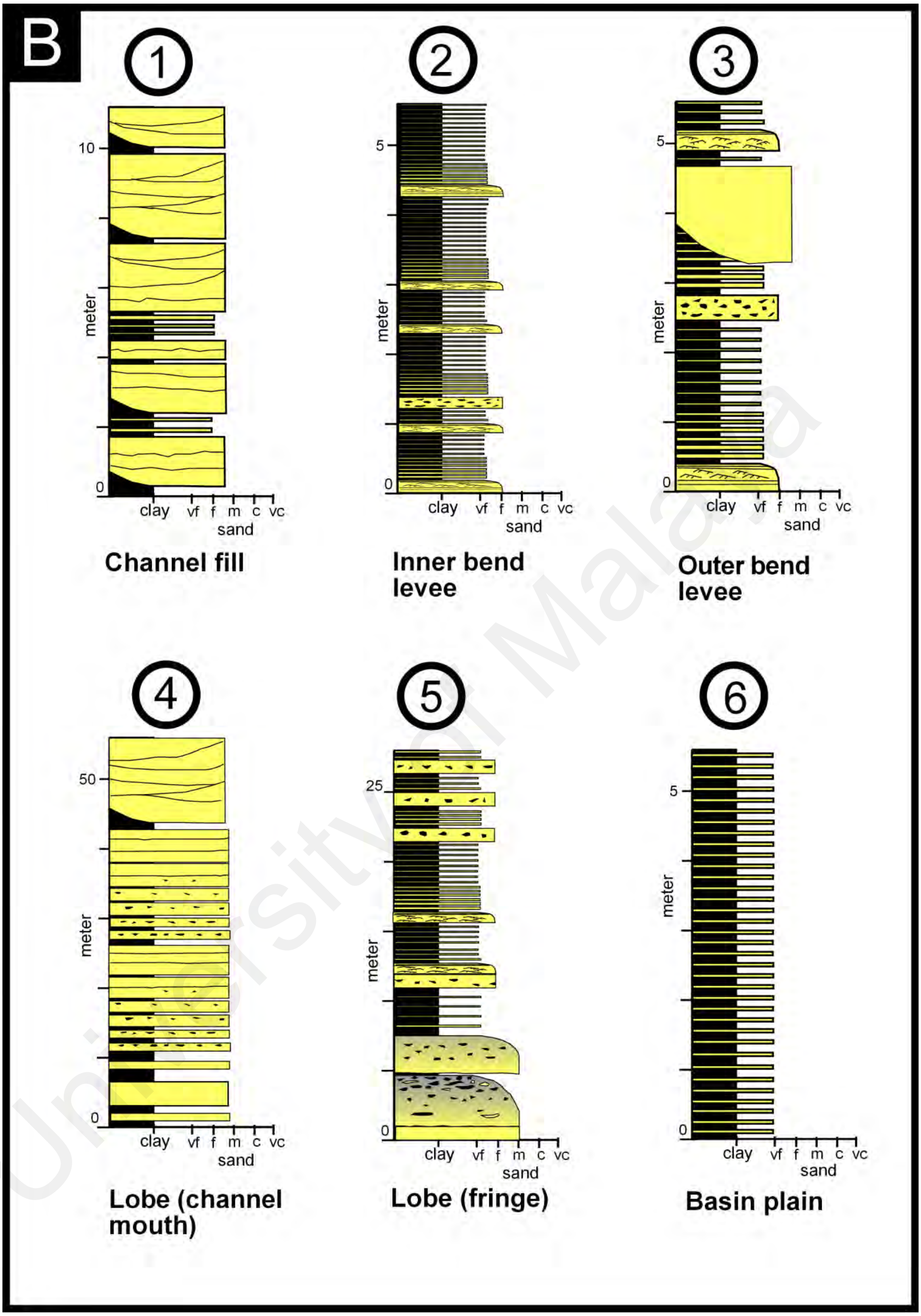


Figure 6.6. A) Schematic simplified depositional model for the Paleocene to middle Eocene Belaga Formation (modified after Reading and Richards, 1994). B) Idealized facies associations of the Belaga Formation. Location marked in Fig. A.

The channel fill deposits of the Kapit Member are intercalated with thin (<2m) levee deposits, interpreted as indicating the presence of low relief levees overlain by highly erosive channels. The low relief of levees may indicate a distal channel setting, where channels are less confined and channel deposits are more sheet-like. The intercalation of levee and channel fill deposits suggests development of either single channel, experiencing multiple lateral migration/shifting or there were several channels forming a braided channel as common in frontal splays. However, the data displayed in this paper is more supportive of a single channel rather than braided channel system, based on the presence of off-lap and on-lap bed relationships, thick succession of levee successions and levee deposits overlain by thick channel fill deposits **(Fig. 5.8)**, which is similar to channel fill deposits in the Ross Formation (Posamentier and Walker, 2006) and Upper Cretaceous Wheeler Gorge, (Walker 1985). However, it should be noted that channels of the Upper Cretaceous Wheeler Gorge are composed of graded conglomerate.

The thickening upward frontal splay successions of the Belaga Formation are very similar to the frontal splay succession in the Carboniferous Lower Ross Formation (Lien et al., 2003). The bottom part of the thickening upward succession contains D3-type ―hybrid beds” (Hodgson, 2009) indicating submarine fan initiation. Upward decrease in mud clast content suggests basinward shift of the depositional loci. Two thickening upwards successions are observed, overlain by fining upward channel fill deposits, separated by thick (1-2m) successions of low relief levee deposits. The levees suggest the presence of multiple channels occurring within frontal splays **(Fig. 6.6)**. This intercalation of thickening upwards amalgamated sandstone with the levees probably indicate that the frontal splays comprised multiple channels which produced the intercalated succession of channel fill and levee deposits. Within frontal splays, channels are less confined, producing shallow channels and low angle levee crests. The frontal splays were formed in a basinal setting, as sediments within a single flow channel spread in a splay manner, as the channel becomes less confined and reach the stage of "transitional point" where levee height was no longer able to confine the sand prone and high concentration sediments within the channel.

Referring to geological map of Wolfenden (1960), the thicker bedded levee deposits are restricted to the Pelagus Member **(Fig.5.1 and 5.3),** where multiple small slumps are common (3-30m thick). Thinner bedded levee deposits are present in both theKapit (**Fig. 5.4, 5.6, 5.8 and 5.9)** and Pelagus **(Fig. 5.7)** members. The G9 section **(Fig 5.9),** which is part of the Kapit Member,contains massive slumps (up to 15 m thick), which are similar to the levee failure deposits described from the Waitemata Group in the Takapuna section, Auckland, New Zealand (Ballance, 1964; Gregory, 1969), comprising interbedded thin sandstone-mudstone packages. Frontal splay succcessions, comprising lobe with associated channel and levee deposits, are only present in the Pelagus Member, while the Kapit Member is only composed of facies indicative of a leveed channel system.

In terms of depositional architecture, the submarine fan system is similar to the Joshua channel in the Gulf of Mexico (Posamentier, 2003; Posamentier and Kolla, 2003), offshore Makassar Strait, Indonesia (Posamentier et al, 2000), frontal splays of the Carboniferous Ross sandstone of Western Ireland (Lien, et al, 2003), Lingan Fan and Tembungo deep water sand of NW Sabah (Grant 2003; Salahudin et al., 1996) and probably the Upper Cretaceous Wheeler Gorge (Walker, 1985).

The Rajang Group fold and thrust belt has been interpreted as an accretionary prism associated with southward subduction of the proto-South China Sea beneath

Borneo (Hamilton, 1979; Rangin et al., 1990; Tongkul, 1997). However, subduction ceased in the Paleocene (Hutchison, 1996, 2005, 2010) or even earlier in the Late Cretaceous (Moss, 1998). Thus the Kapit and Pelagus members of the Belaga Formation were deposited post-subduction. This has led to the interpretation of a post-collisional foreland remnant ocean basin by Moss (1998).

The facies characteristics and stratal architecture of the Kapit and Pelagus members, Belaga Formation, is dominated by thin graded turbidites interbeddeding with mudstone facies of the levee facies association. A meandering submarine channel system is interpreted based on the low sand:mud ratio within flows and the occurrence of massive slumps, indicating a mud-rich depositional setting (Reading and Richards, 1994). The Kapit and Pelagus members probably formed an elongated, mud-rich, single point-fed source, submarine fan depositional system. Two types of depositional environments/elements are identified in the Belaga Formation deposits around Sibu: (1) leveed channel, and; (2) frontal splay. Both are interpreted as elements of the same depositional system, with unconfined turbidity currents forming frontal splays basinward of leveed channels **(Fig. 6.6)**.

Slumps and crevasse channels were most likely common along the outer bend of channel levees. Two packages of thickening upward successions, comprising amalgamated thick bedded (30- 100cm) of F6 and F5 facies, separated by 1m of interbedded F1 and F2 facies in G2 outcrop **(Fig 5.2)** represent a progradational succession (Mutti, 1977; Normark, 1978; Ricci-Lucchi and Valmori, 1980; Walker, 1992) that developed during periods of lowstand (Posamentier and Walker, 2006). Generally, feeder systems type can be an important factor controlling the complexity of geometry and morphology of a submarine fan system (Reading and Richards, 1994).

The low sand:mud ratio of the Belaga Formation submarine fan system suggests an elongate geometry. Higher sand:mud ratios produce a more spread out, fan-shaped fan geometry.

Palaeocurrent readings from older strata of the Rajang Group have been used to interpret a generally SW-NE oriented axis for the Rajang Sea remnant ocean basin and the development of an axial submarine fan system. The Late Cretaceous Lupar Formation has a dominant palaeocurrent direction towards the NE (Tan, 1979, 1982), while the Embaluh Group (correlative of the Rajang Group in Kalimantan) displays a dominant eastward palaeocurrent direction (Moss, 1998).

The general northward palaeocurrent direction for the Pelagus and Kapit member channel fills in the Belaga Formation is a little problematic. However, this suggests a more complex depositional system, possibly with smaller ‗mini-basins' and different sediment sources. It is possible that the change resulted from the change from subduction to post-collision.

6.3. The correlation between the observed sedimentation and the Rajang Group tectonics

The Rajang Group fold and thrust belt has been interpreted as an accretionary prism associated with southward subduction of the proto-South China Sea beneath Borneo **(Fig 6.6A) (**Hamilton, 1979; Rangin et al., 1990; Tongkul, 1997). In contrast, available tectonic scenarios assume that subduction ceased in the Late Cretaceous or Paleocene, inferring a syn- to post- subduction (or collisional) foreland genesis for the Paleocene - Middle Eocene Kapit and Pelagus members of the Belaga Formation

(Hutchison, 1996, 2005, 2010; Moss, 1998). In fact, most of researchers demonstrate that the deformation in the Rajang Group was associated with the so-called Sarawak Orogeny that ended somewhere towards the end of the Eocene by the development of a regional unconformity (e.g., Hutchison, 1996; Cullen, 2010). Our observations and all previous studies have shown that the Belaga Formation is dominated by the deposition of turbidites that was syn-kinematic with the formation of many thrusts and transpressive structures, which brought most of the strata to sub-vertical positions (e.g., Honza et al., 2000). Given the Late Cretaceous - Paleocene structuration of the Lupar Fault zone and Lubok Antu Melange with the associated ophiolitic suite and their interpretation as a suture zone (Tan, 1979; Hutchison, 1975, 2005), it is rather clear that the convergence between the Luconia Block and NW Borneo was more complex than simple docking of these continental units. The Late Cretaceous - Paleocene formation of the suture zone recorded the onset collision, which was a longer process where shortening and continental subduction continued throughout the Eocene times, when the ―Sarawak Orogeny” was recorded as the final deformation in a remnant, likely continental floored basin, and was followed by the observed Oligocene regressional stage (Fyhn et al., 2010; Cullen, 2010; Hall, 2012). A long-lived Eocene collisional process associated with significant transpression explains the observed syn-kinematic character of the Belaga Formation deposition and the development of sub-basins capturing along-strike sedimentary influx. The dominantly turbiditic facies is otherwise common for the collisional shortening observed in many Mediterranean orogens, where such ―flysch” deposition stages continued during the subduction of thinned continental domains, and are followed by regressive stages of molasse deposition, as observed in the Carpathians, Apennines or the Dinarides-Hellenides systems (e.g., Matenco et al., 2016; Oszczypko, 2006; Picotti and Pazzaglia, 2008; Săndulescu, 1988; Schmid et al., 2008)

6.4. Comparison between the Belaga Formation and the onshore and offshore submarine fan systems of NW Sabah

Oligocene-Miocene deepwater deposits of the West Crocker Formation (WCF), exposed onshore Sabah, share many similar facies characteristics with the Belaga Formation (Jackson et al., 2013; Zakaria et al., 2013; William et al., 2003). The WCF of Sabah also comprises gravity flow deposits (high-low density turbidites, debrites and associated slumps). The WCF strata have also been interpreted as channel-levee and lobe/splay elements of a large submarine fan depositional system. However, the WCF is sand-dominated, with relatively thicker beds for both thick and thin beds, having less and thinner mudstone interbeds, compared to the mud-dominated and thin-bedded Belaga Formation. The WCF deposits are characteristic of a sand rich, deep marine system (Readings and Richard, 1994), rather than the mud- rich system of the Belaga Formation. Current-reworked facies, which are present in the Belaga Formation, have not been reported from the WCF. This indicates the presence of bottom currents strong enough to transport sediment in the Belaga Formation basin.

The WCF has been used as a suitable outcrop analogue for Neogene, hydrocarbon-bearing deepwater reservoirs offshore Sabah (Zakaria et al., 2013). However, it cannot be considered a direct analogue, due to major differences in tectonic setting, with the WCF being syn-tectonic and associated with subduction and the development of an accretionary prism (Zakaria et al., 2013; Jackson et al., 2009; William et al., 2003), while the Neogene turbidites of offshore Sabah are post-tectonic, post-collision and influenced by toe-thrust faulting

A comparison between the Belaga Formation and the Neogene, offshore Sabah turbidite systems is presented here. The post-subduction interpretation for the Kapit and Pelagus members, Belaga Formation, is probably more similar to the tectonic setting of offshore Sabah, which is associated with a foreland basin, with small, unconnected fan systems deposited in small, ponded mini basins (Casson et al., 1999; Lambiase and Cullen, 2013, Jones et al., 2016). The Paleocene-mid Eocene Belaga Formation and the Neogene offshore Sabah turbidites systems also both share the same depositional trend, with deposition generally perpendicular to regional, SW-NE structural trend of Borneo (e.g. Salahuddin et al., 1996; Jones, 2016). The perpendicular trend is probably more typical of post-subduction submarine fan systems, due to the deposition of thick, foreland basin molasse deposits, which supplied sediment to the deep basin.

Published descriptions of the facies in the Neogene, offshore Sabah fan systems are sparse. Jones et al. (2016) described up to 200 m thick successions characterised by basal, massive bedded turbidites overlain by thin-bedded turbidites (cm-thick) and interbedded mudstone in the Kikeh Field, resembling the fining upward channel fill successions of the Belaga Formation. Similar channel fill successions are also observed in the Late Miocene Pink Fan of offshore Sabah. However, the Pink Fan amalgamated channel deposits are interpreted as representing a highstand, braided system associated with a slope canyon channel system (Grant, 2003). Seismic images indicate highly sinuous individual channels and the presence of lateral accretion, but with no levee development.

The Lingan Fan is younger and overlies the Pink Fan of offshore NW Sabah. It is also a lowstand, lower slope submarine fan developed in a ponded, structurally controlled _mini-basin‘ (Salahuddin, 1996). Seismic images indicate presence of _gull wing' seismic facies interpreted as leveed channels, associated with interpreted lowstand channel and channel lobe facies. The lobe facies are overlain by the leveed channels.

The Late Miocene Tembungo deepwater sands of offshore NW Sabah have also been interpreted as representing a lowstand submarine fan system. Integrated core, wireline and seismic data indicate the presence of facies architecture very similar to the Paleogene Belaga Formation of Sibu (Ibrahim, 2003). Muddy deposits form basin plain and slope facies associations. Heterolithic successions are interpreted as proximal overbank. Scoure-based, fining upward successions up to 25 m thick, with basal amalgamated massive sandstone overlain by thin bedded Tabc and Tbc turbidites are interpreted as channel fill. Coarsening and thickening upward successions up to 5 m thick and composed of thin to medium bedded turbidites interbedded with mudstone have been interpreted as prograding lobe deposits. The Tembungo system is also resembles the Belaga Formation in being fine grained and with a low sand:shale ratio (Ibrahim, 2003). The prograding lobe and channel fill deposits were interpreted as representing basin floor fans progradation during the early lowstand phase, while the muddier, leveed channel system developed during the late lowstand, when relative sea level started to rise.

In summary, the facies architecture and composition of the Paleogene Belaga Formation exposed around Sibu is very similar to that observed in the Neogene submarine fan deposits of offshore NW Sabah. All these systems comprise prograding sand-dominated lobe/frontal splay deposits being overlain by channel fill and overbank deposits. The Pink Fan displays an amalgamated channel system with a braided pattern and no levees. However, many other Neogene submarine fans of offshore Sabah have leveed channel strata overlying prograding lobes.

Despite these similarities, there are also significant differences between the Belaga and Neogene Sabah fans. Some of the offshore Sabah systems have been interpreted as being sand-dominated, in contrast to the mud-dominated Belaga Formation. Another major difference is the close association of thick mass transport deposit (MTD) intervals with the leveed channel systems of the Neogene fans (e.g. Algar et al., 2011). The mud-dominated MTDs are really extensive (tens of km) and possibly represent downslope transported deposits of updip slope failures. Thick MTDs have not been observed in the Belaga Formation exposed in Sibu. Debrites in the Belaga Formation are relatively thin and probably represent only local remobilization. The slumped levee deposits also most probably only represent local overbank collapse. It is possible that the observed absence of MTDs is due to preservational bias, given the limited extent and complex tectonic deformation of the Sibu outcrops. However, it is equally likely that MTDs rarely developed during Belaga Formation deposition, implying less frequent tectonic disturbance and/or lower sediment supply.

CHAPTER 7: CONCLUSION

This study is concluded into 6 points, they are;

1. A detailed facies analysis of the Paleocene to middle Eocene Kapit and Pelagus members of the Belaga Formation exposed around Sibu, Sarawak indicates the presence of facies representing a mixture of pelagic/hemipelagic, turbidity current and debris flow deposits, with evidence of bottom current reworking.

2. Strata of the Kapit and Pelagus members form 6 types of facies associations: (i) Channel Fill (FA1) characterised by tens of metres thick, sharp-based succession of interbded debrites and turbidites displaying internal channel scours; (ii) Levee (FA2) characterised by dm to m thick, thinning and fining upward cycles of medium to thin bedded turbidites displaying palaeocurrent orientations perpendicular to main flow orientation; (iii) Lobe Channel Mouth (FA3) characterised by m-thick, thickening upward successions of interbedded debrites and high density turbidites and mudstone; (iv) Lobe Fringe (FA 4) characterised by up to 65 m thick, thinning upward successions of interbedded debrites, mudstone and thin bedded turbidites; (v) Basinal Plain (FA5) characterised by up to 20 m thick successions of interbedded mudstone and thin bedded turbidites, and; (vi) Slump (FA6) characterised by folded intervals of interbedded turbidites and mudstone with associated rotated sandstone blocks.

3. A mud-rich, basin floor submarine fan depositional system is interpreted for the Paleocene to middle Eocene Belaga Formation. Tens of metres thick successions that grade upward from lobe (channel mouth) deposit into channel fill deposits represent the deposits of prograding, frontal splays. Successions of interbedded levee and channel fill deposits up to 120 m thick are interpreted as representing the deposits of leveed

channels. The channel system was sinuous and meandering, with a low sand:mud ratio. The low sand:mud ratio suggests anelongate fan morphology.

4. The Paleocene to middle Eocene Belaga Formation is compared to submarine fan deposits of the Oligocene-Miocene West Crocker Formation of NW Sabah. Both systems comprise leveed channels and splay depositional elements. However, the WCF is a sand-rich submarinefan.Also, the tectonic settings are different, with the WCF being syn-tectonic and associated with subduction and the development of an accretionary prism, while the Neogene turbidites of offshore Sabah are post-collision and influenced by toe-thrust faulting.

5. The Belaga Formation is also compared to the Neogene, hydrocarbon-bearing submarine fans of offshore NW Sabah. The Neogene fan systems also display depositional trends which are perpendicular to the regional (SW-NE) structural trend. They also display similar leveed channel and frontal splay elements, with some examples being fine grained systems, while others are more sand-dominated. Tha major difference is the absence of laterally extensive MTDs in the Belaga Formation, which suggests a less tectonically influenced setting and/or a lower sediment supply.

6. The Paleocene-mid Eocene Belaga Formation and the Neogene offshore Sabah turbidites systems both share the same depositional trend, with deposition generally perpendicular to regional, SW-NE structural trend of Borneo, and also very similar tectonic settings, i.e. post-subduction foreland basin.

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