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Tertiary basins of S.E. Asia—their disparate tectonic origins and eustatic stratigraphical similarites

CHARLES S. HUTCHISON,
Dept. of Geology,
University of Malaya, Kuala Lumpur

Abstract: Following the Indosinian Earliest Jurassic compressive orogeny, Tertiary basins of Southeast Asia developed during the Late Mesozoic and Early Cenozoic by extensional tectonics combined with wrench fault control. The major basin-forming mechanisms are: rifting of Atlantic-type miogeoclinal margins, aulacogens; fore-arc, intra-arc, and various back arc extensions. Tertiary compressive movements were important only in some basins such as those lying between India and Burma, between the South China Sea microcontinents and Borneo, and between Australia and the Banda arc. Elsewhere throughout most of the region, those collisions were transmitted as major basement wrench movements which resulted in non-compressive open cover folds on the Tertiary basin fill.

With the exception of marginal seas, sedimentation kept pace with basin dilation and subsidence, and in the absence of compressive orogenic events which might cause uplift, basin unconformities, regressions and transgressions are related to global sea level changes. The eustatic rise from 29 Ma to 13 Ma ago is evident as a widespread Mid Miocene transgression over basins which had developed earlier on continental crust. Sea level drops at 13, 9.8, and 6.6 Ma ago are well documented as widespread regressions in Indonesian and Malaysian basins.

The excellent correlation of eustatic sea levels with transgressions and regressions in S.E. Asia maybe an artifact of overdependence upon S.E. Asian basins during the compilation of the eustatic curves.

INTRODUCTION

Provided a basin is formed and developed by extensional and wrench faulting, and is free from compressional tectonics during its sedimentation history, and provided the basin has subsided enough for marine incursion, then regressions, transgressions and unconformities will be products of global sea level changes. Extensional basins begin on uplifted continental crust and their early sedimentation history is continental, fluvial and lacustrine. Eustatic sea levels have little relevance to a basin until the marine influence begins.

The folded nature of strata does not necessarily mean that the basin underwent compressive orogenic tectonics. Many folded basins are now recognized as a result of wrench faulting within the rigid basement, causing crumpling of the overlying non-rigid basin fill strata (Harding, 1974; Harding and Lowell, 1979). Such folding would continue during sedimentation, and therefore would still continue under the influence of eustatic sea levels. Folding may also be a result of normal listric faulting in an extensional environment (Bally *et al.*, 1981).

An analysis of the tectonic style will reveal if a basin has undergone orogenic compression and tectonic uplift to produce non-eustatic related unconformities.

A basin classification system should ideally incorporate a statement as to whether or not the regional stress regime changed from a dilation to a compression mode. This distinction will be useful when applying eustatic sea levels to correlate with regressions, transgressions and unconformities observed. Basins with proven or suspected compression should be excluded from a compilation of data for the eustatic sea level curve. Conversely, once the eustatic sea level curve is established, its comparison with the stratigraphy of a basin will reveal to what extent that basin has undergone compressive orogenesis and tectonic uplift, and the timing of these events.

BASIN CLASSIFICATIONS

Two attempts at classification of S.E. Asian basins appeared independently in 1975 (Murphy, 1975; Soeparjadi et al., 1975). The more detailed one by Murphy (1975) subdivided the basins into *"shelfal"*, *"continental margin"*, *"archipelagic"*, and *"marginal seas"*. Some of these categories are too broad to be useful. Soeparjadi *et al.* (1975) used such categories as *"outer arc"*, *"foreland"*, *"interior cratonic"* and *"open shelf on continental margin"*. Some of these terms are open to misinterpretation. *"Interior cratonic"* is a term more suited to such basins as the Szechwan *"Red"* basin of the Yangtze Platform and to the Khorat-Vientiane basin of Indosinia, and should not be applied to the Borneo region.

Basin classification systems should use the same terminologies as those of plate tectonics. The general tectonics-based classifications of Bally and Snelson (1980) and Kingston *et al.* (1983) are eminent in this regard, and should be followed wherever appropriate. These two classifications are broadly similar despite different terminologies. One important difference is in regard to the mechanisms of rifting. Kingston *et al.* (1983) interpret the various back arc basins of convergent plate margins as a result of divergent wrench faulting. The back arc basins of Sumatra and the Malaya Basin are referred to as "wrench or shear basins". This is an attractive concept because these back arc basins have both extensional and wrench histories (Crostella, 1981). The code system of Kingston *et al.* (1983) requires much detail and is difficult to apply without extensive knowledge of any particular basin, and hence that of Bally and Snelson (1980) is more readily used. They are, however, broadly complementary systems.

Hutchison (1984) showed that several basins peripheral to Borneo have been complicated by collison between continental fragments or microcontinents. He therefore erected an additional category *"basins on and peripheral to continental fragments"*. This category can readily be incorporated into the detailed classification of Kingston *et al.* (1983).

A fundamental problem in Southeast Asia is whether the basins which lie behind the volcanic arcs are genetically related to the convergent arc-trench system. The *"back arc"* classified should therefore be used purely in a geographic descriptive sense. Kingston *et al.* (1983) may have improved on the system of Bally and Snelson (1980) by suggesting that such basins result from a component of extension related to major continental wrench systems. They suggest that they occur where the direction of subduction is oblique to the plate margin. However it is not certain that these basins are really a result of a simple convergent tectonic system or they result from complex transform re-adjustments of platelets combined with significant extensional motion. The same problem applies to the classification *"marginal seas"*. Few, if any, of them represent *"back arc"* ocean floor spreading. Many may represent fragments of larger oceans trapped behind younger arc-trench systems. Others, like the South China Sea basin, result from rifting of a continental shelf (Taylor and Hayes, 1983). The use of a term to classify a basin does not therefore mean that we understand the origin of the basin. In many cases the terms are geographically descriptive and unrelated to tectonic evolution.

THE BASINS OF S.E. ASIA

The locations of the major basins of Southeast Asia are shown on Figure 1, and

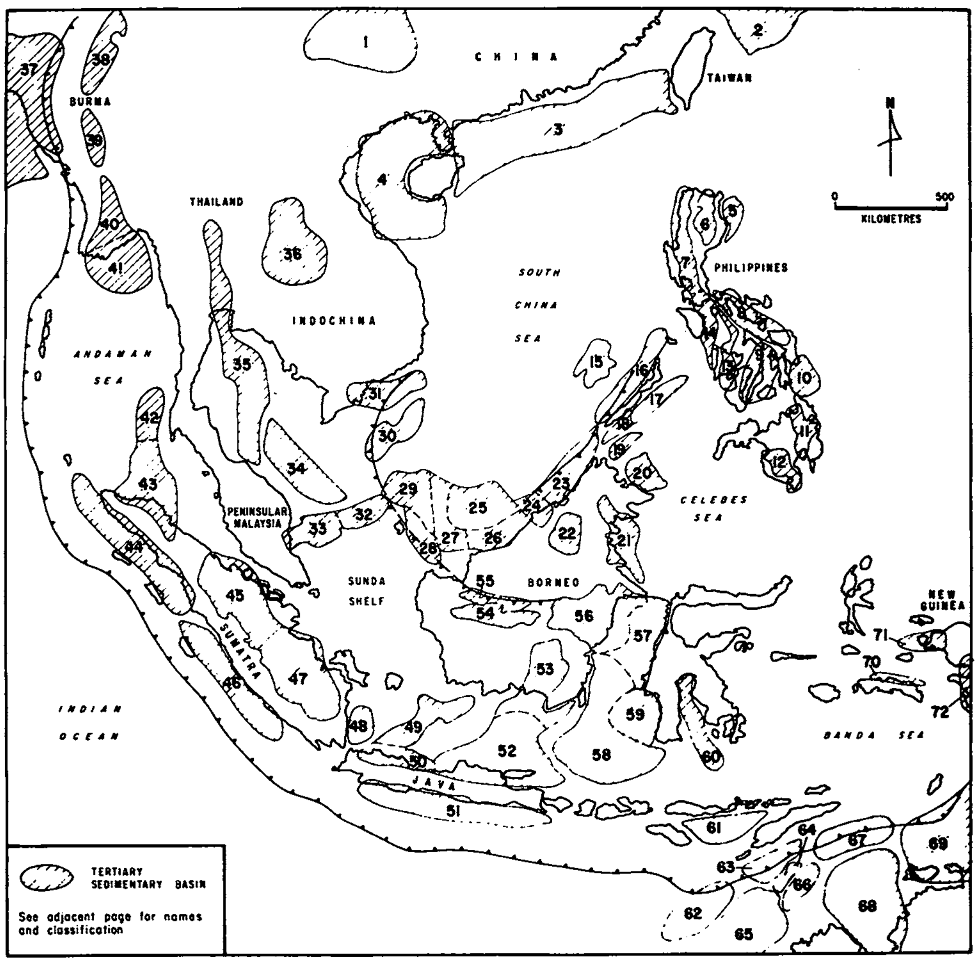

Fig. 1. The main Tertiary sedimentary basins of Southeast Asia. See Table 1 for names, classification and average geothermal gradients.

TABLE 1

KEY TO BASINS SHOWN IN FIGURE 1

No.	Basin Name	Class	Gradient.
1	Southwest China	C	27
2	East China Sea	B	
3	South China Shelf	B	
4	Gulf of Bacbo (Tonkin)	A	
5	Sierra Madre	F	
6	Cagayan Valley	G	22
7	Luzon Central Valley	F	24
8	Ragay-Samar	G	41
9	Visayan	G	31
10	Leyte	F	41
11	Davao-Agusan	G	
12	Cotabato	F	18
13	Iloilo	F	21
14	South Mindoro	G	29
15	Reed Bank	J	
16	West Palawan	D/J	27
17	East Palawan	F	22
18	Balabac	H	
19	Bancauam	H	
20	Sandakan	F	28
21	Tarakan	A	38
22	Kelabit-Long Bawan	J	
23	Sabah	D/F	28
24	Baram Delta	J	28
25	Central Luconia	J	43
26	Balingian	J	41
27	West Luconia	J	
28	Sokang sub-basin	J	56
29	North-east Natuna	J	34
30	Saigon	A	
31	Vung Tau	A	
32	West Natuna	A	38
33	Penyu	A	38
34	Malaya	K	45
35	Gulf of Thailand Chao Phraya-Phitsanulok	K	50
36	Khorat-Vientiane	C	
37	Bengal	B/E	
38	Upper Chindwin	F/E	
39	Central Basin	F/E	
40	Lower Irrawaddy	I	
41	Gulf of Martaban	I	
42	South Andaman	H	33
43	North Sumatra	I	47
44	Sibolga	F	24
45	Central Sumatra	I	61
46	Bengkulu	F	24
47	South Sumatra	I	49
48	Sunda	I	46
49	Billiton sub-basin	I	32
50	Northwest Java	I	46
51	South Java	F	
52	East Java	I	39
53	Barito	I	36
54	Melawi	I	
55	Ketungau	I	
56	Kutei	A	32
57	North Makassar	A	
58	Paternoster Platform	J	
59	South Makassar	A	25
60	Bone	A	8
61	Savu	F	
62	Scott Plateau	B	
63	Ashmore Block	B	23
64	Vulcan sub-basin	B	26
65	Browse Basin	B	27
66	Londonderry High	B	37
67	Sahul Ridge	B	47
68	Bonaparte Gulf	A	
69	Money Shoals	B	
70	Bula	F	17
71	Salawati	B	36
72	Bintuni	B	30

Average geothermal gradient in °C km^{-1}

BASIN CLASSIFICATION
A = aulacogen rifts in continental lithosphere (Ref. 1); continental interior fracture (Ref. 2): B = Atlantic-type miogeoclinal margin (Ref. 1); continental margin sag (Ref. 2): C = cratonic basin (Ref. 1); continental interior sag (Ref. 2): D = trench (Ref. 1); oceanic trench (Ref. 2): E = foredeep or collision zone (Ref. 1); continental margin sag affected on one side by a foldbelt (Ref. 2): F = fore-arc basin (Ref. 1); trench associated basin (Ref. 2): G = intra-arc basin (Ref. 3); trench associated basin (Ref. 2): H = marginal sea (Ref. 1); oceanic sag-wrench couplet basin (Ref. 2): I = back arc basin (Ref. 1); continental wrench or shear basin (Ref. 2): J = basin on peripheral to continental fragments (Ref. 3); various classifications but mainly a combination of continental margin sag and foldbelt (Ref. 2): K = Pannonian-type back arc basin (Ref. 1); continental wrench or shear basin (Ref. 2).
REFERENCES: (1) = Bally and Snelson (1980). (2) = Kingston *et al.* (1983). (3) = Hutchison (1984).

the names, classification and average geothermal gradients calculated from the data of Rutherford and Qureshi (1981), given in Table 1. The classification followed is that of Bally and Snelson (1980), Kingston *et al.* (1983) and Hutchison (1984). The classification cannot be applied with equal confidence to all basins, for the tectonic setting of some is still to be resolved. Thus, basins 54 and 55, Melawi and Ketungau, have been called back arc, but ongoing work (Williams *et al.*, 1984) has shown them to be separated by a Cretaceous melange, suggesting that they may be trench-related basins.

The majority of the basins are of a single tectonic setting, resulting from combinations of extensional and wrench faulting. Such basins are devoid of compressive tectonics and their sedimentation is expected to be predominantly influenced by eustatic sea level changes.

Two categories, foredeep or collision zone (E), and basins on and peripheral to continental fragments (J), however, have had compressive tectonics at a later stage in their development and therefore some of their regressions and unconformities should have resulted from local vertical tectonics totally unrelated to eustatic sea level changes. Deeper basins such as marginal seas (H) or trench (D) may be devoid of synsedimentation vertical tectonics but sedimentation was unrelated to subsidence and therefore eustatic sea level effects are absent. Nevertheless, Curray *et al.* (1982) have shown that turbidite sedimentation into deeper basins is more likely during periods of lowered sea level.

EUSTATIC CHANGES OF SEA LEVEL

The curve showing estimated global cycles of relative changes of sea level, deduced by Vial and Mitchum (1979) from seismic stratigraphy, is given in Figure 2. The most outstanding features are: a) the consistently high sea levels which persisted from the Mesozoic to 29 Ma ago, with a rapid fall in Late Oligocene times. This would have resulted in a spectacular Oligocene exposure of the shelf areas of the Early Tertiary continents and encouraged turbidite sedimentation into the deep oceanic basins; b) from 29 to 13 Ma ago, there was a progressive rise in sea level, which would have resulted in marine incursions into basins which developed within continental lithosphere, reaching maximum marine transgression by Mid Miocene times; c) stepwise drops of sea levels at 13, 9.8 and 6.6 Ma ago, resulting in an Uppermost Miocene low, which would have caused an important marine regression in non-compressive basins; d) throughout the Pliocene and Pleistocene there were several fluctations of sea level which would have caused regressions and transgressions in rapid succession in non-compressive basins. The Quaternary changes have been documented by Biswas (1973).

SOME ILLUSTRATIVE EXAMPLES

a. **Bengal Basin (37: Figure 1)**

This basin can be interpeted as an Atlantic-type miogeoclinal shelf, which formed as India rifted from Australia during Jurassic times, as shown by Paul and Lian (1975)(Type B of Table 1). The basement shows a transition from continental to

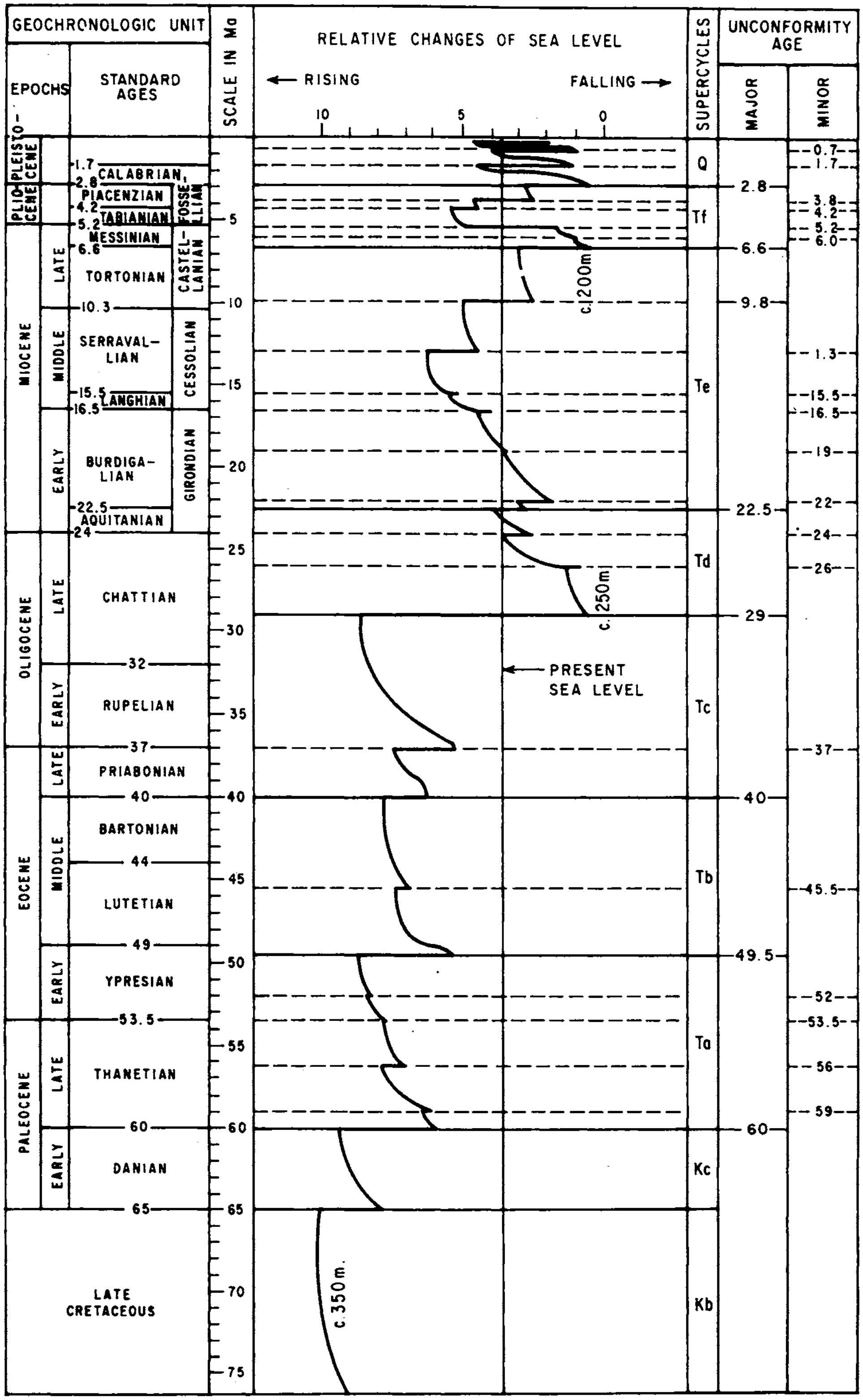

Fig. 2. The global cycles of relative changes of sea level, deduced from seismic stratigraphy, after Vail and Mitchum (1979).

oceanic crust (Fig. 3). The Miocene collision with Asia has converted the basin into a foredeep (Type E), or using the terminology of Kingston *et al.* (1983), it may be classified as a Late Jurassic-Early Miocene continental margin sag, modified during the Miocene collision with Burma by a foldbelt on the Chittagong Hills side. The collision was preceded by subduction of the oceanic lithosphere basement beneath Burma to form the Burman volcanic arc, which became extinct at the time of collision.

Although the main Indian orogenic "hard" collision is of Miocene age, there was a "soft" collision at magnetic anomaly 22 (53 Ma ago).

The abrupt change during the Late Oligocene from the deep water Disang Group to the deltaic lithofacies of the Barail Group (Fig. 3) correlates with the dramatic fall in eustatic sea level 29 Ma ago (Fig. 2). However the Late Oligocene to Pleistocene deltaic sequence does not closely follow the eustatic conditions because the change of the basin from miogeocline into a collision foredeep resulted in basin shortening and uplift of the eastern side.

b. **Back arch basins of Sumatra-Java (45, 47–50, Fig. 1)**

These basins have formed under a single tectonic regime by combined extensional-wrench faulting (Type I, Table 1). They share a common stratigraphic sequence, summarized by Soeparjadi *et al.* (1975), as illustrated in Figure 3. Such basins are therefore strongly influenced by global sea level changes.

The early stage of continental sedimentation was in rifts developed in uplifted continental crust and was therefore not synchronous in all basins. As dilation and subsistence proceeded, the sedimentation came under marine influence, but the initiation of marine incursions was not simultaneous in all basins. The rising global sea level from 29 to 13 Ma ago affected all basins and by Mid Miocene they all achieved a maximum marine character, entirely as a result of the eustatic sea level rise. The Mid Miocene to Pliocene regression (Fig. 3) is also controlled by a stepwise eustatic fall from 13 to 6.6 Ma ago. The details of Figure 3B are therefore closely comparable with Figure 2, and Category I basins have had little if any tectonic control on their sedimentation histories. Folding in these basins can be ascribed to wrench movement within the rigid basement, which does not cause basin shortening or uplift. However the wrench faulting is not always parallel to the basin length, as implied by Kingston *et al.* (1983). Crostella (1981) has documented N-S trending wrench faults incorporating significant and variable vertical movement in the NW-trending Central Sumatra Basin.

c. **Baram Delta (24: Figure 1)**

Although I have classified the Baram Delta as a basin peripheral to a continental fragment, it has had little orogenic effect from the adjacent continental fragments of Luconic Shoals and the Kelabit Highlands (Hutchison, 1984), which lie on opposite sides of the Tinjar Fault; largely because motion between these microcontinents has been strike-slip along the fault.

The Tinjar Fault is prominent on the map of Haile (1962). The river follows a NNW trend as far north as Bukit Selikah, then abruptly bends 90° towards the

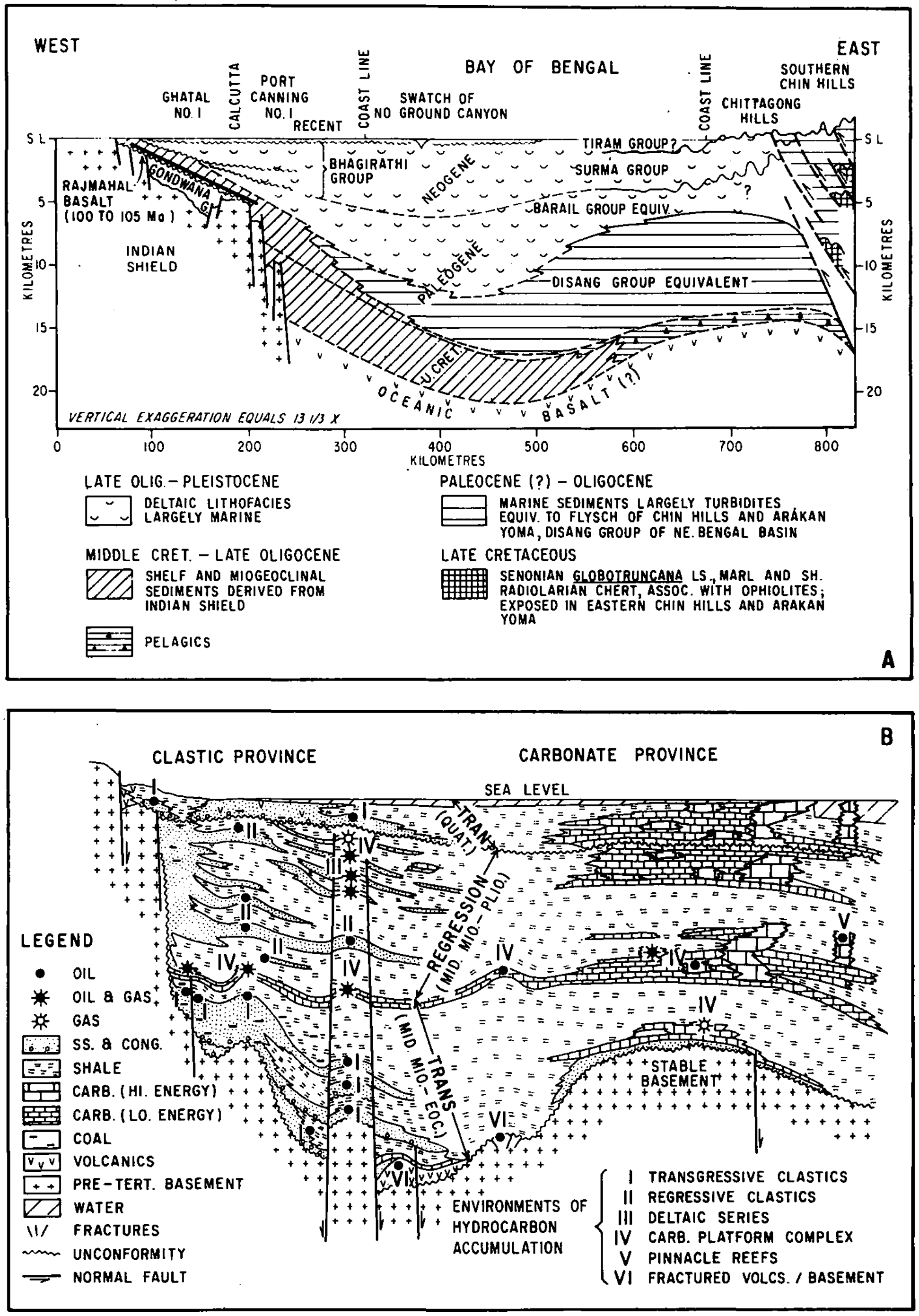

Fig. 3. A:Cross section from India to Burma across the Bengal Basin of type E (Table 1), after Paul and Lian (1975). B: Diagrammatic cross section showing the salient features of I-type Indonesian Tertiary basins, after Soeparjadi *et al.* (1975).

ENE. The NNW sector contains an undated but presumably young olivine basalt dyke parallel to the fault. Although the extrapolation of the fault from Bukit Selikah to the coast is obscure, structural complexities in the Miocene scarps, as seen on air and earth satellite photographs, suggest that the fault must have continued. Unfortunately the region is remote and no detailed studies have been conducted on this fundamental fault. Offshore, the NNW extrapolation dramatically separates the Central Luconia province with an average geothermal gradient of 43°C km^{-1} from the eastern Baram Delta with a gradient of only 28°C km^{-1} (Hutchison, 1984). Here the fault is known as the Baram Line, against which the WSW trending Palawan-Sabah or Northwest Borneo Trough abruptly terminates (Hamilton, 1979). The low geothermal gradient suggests that the Baram Delta basin is formed on oceanic crust, and the setting may have similarities to the Niger Delta, in which the river flows along a fault-bounded aulacogen (Type A), so that most of the delta is built up over oceanic crust (Hutchison, 1983, page 286). The similarity of sedimentation sytle with the Niger Delta has been documented by Rijks (1981).

The tectonic style of the Baram Delta is domianted by non-orogenic synsedimentary deltaic growth faults and associated rollover folds, which result from the superposition of deltaic sands over marine muds (Rijks, 1981), and there appears to be an absence of compressive tectonics. Therefore the delta sedimentation pattern was dominated by eustatic sea level changes.

The worldwide high at 13 Ma ago is seen as the maximum southerly transgression of Cycle IV which formed the prodelta marine basement. The three spectacular drops in sea level at 13, 9.8 and 6.6 Ma ago are seen within the overall regressive pattern of Cycle V (Figure 4). The rising, but fluctuating, sea levels since the Early Pliocene, can be seen in Cycles VI and VII (Ho, 1978). It is therefore suggested that the Baram Delta is an ideal basin on which to base the global eustatic sea level changes.

d. **The Central Luconia Province (25, Fig. 1)**

The Central Luconia Province is dominated by a Mid Miocene carbonate platform from which rise carbonate mounds and pinnacle reefs. Central Luconia shows a low degree of structural deformation. It underwent moderate faulting during the Oligocene to Lower Miocene and again between the Lower and Middle Miocene.These appear to be extensional features which resulted in basement horsts and grabens. The prolific growth of reefal carbonate build-ups was on the horsts (Epting, 1980). Such features are characteristic of a rifted miogeoclinal continental shelf. The Luconia Province has been interpreted by Taylor and Hayes (1983) as a rifted fragment of the now decimated shelf of Indochina. Therefore although I have classified the province as a basin on a continental fragment (Type J), it has all the characteristics of a small portion of an Atlantic-type miogeoclinal margin (Type B). Such an interpretation is in keeping with the tectonic stability and the strong control of eustatic sea level changes on the sedimentary history. Figure 4 (lower) is modified from Epting (1980) to emphasize this interpretation.

The pre-carbonate cycles are poorly known. The carbonate build-ups of cycles III to early V may be attributed to the rising eustatic sea level from Late Oligocene through to 13 Ma ago (Fig. 2). The stepwise fall of sea level at 13, 9.8 and 6.6 Ma ago

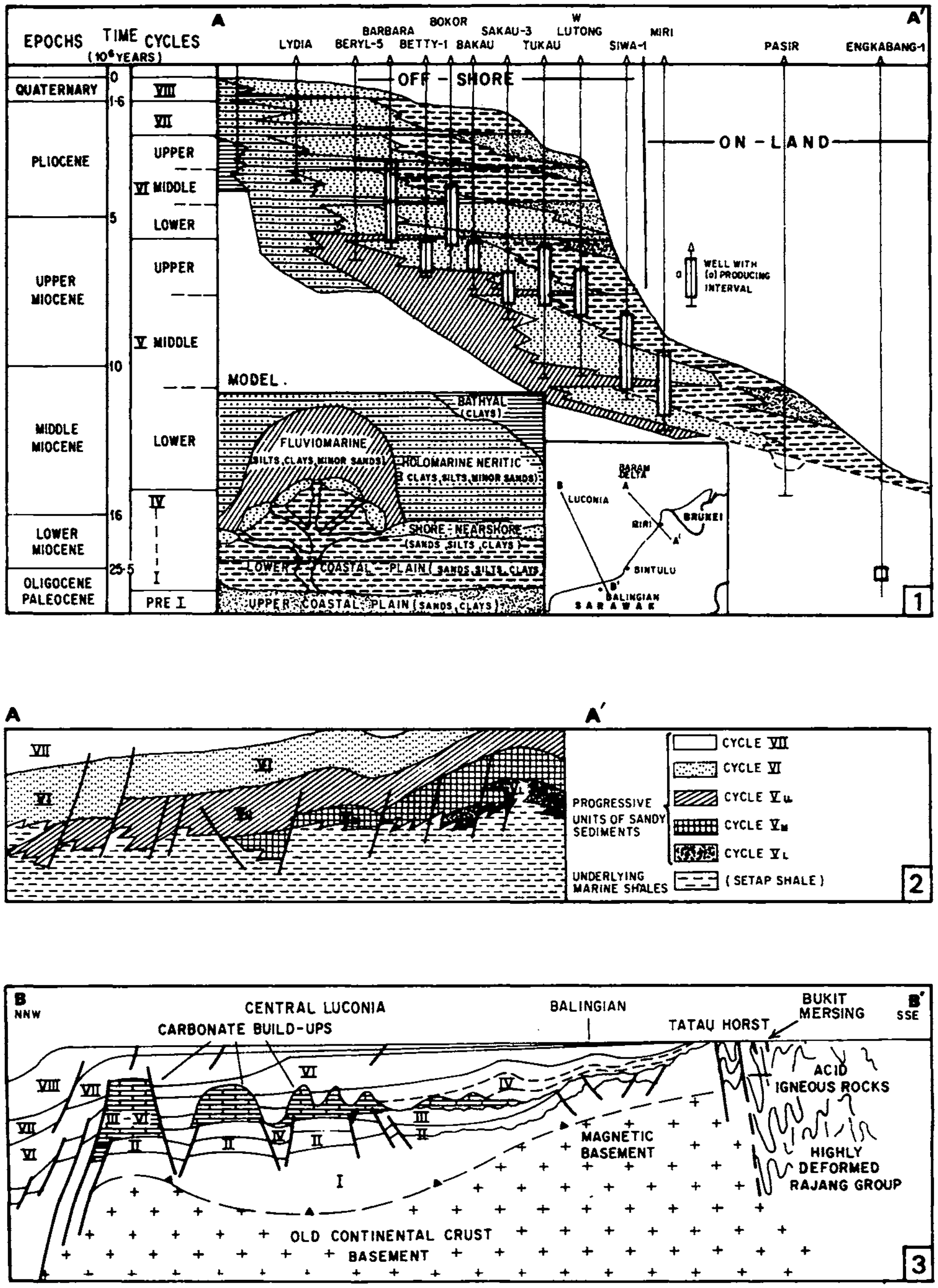

Fig. 4. Upper: Time-stratigraphic profile of the Baram Delta, after Ho (1978) and ASCOPE (1981). Middle: Structural profile of the same section (from Rijks, 1981). Lower: Structural cross section across the Tatau Horst, Balingian and Central Luconia provinces (after Epting, 1980).

resulted in emergence and extinction of the reefs during Cycles V and VI. Rising sea level from 6.6 to 5 Ma ago resulted in Cycles VI and VII burial of the dead reefs by Latest Miocene to Pliocene times.

e. **Balingian Province (26, Fig. 1)**

The Balingian Province is the southerly continuation of the Luconia Province. The Atlantic-type miogeoclinal fragment extends as far as the Tatau Horst area, with Haile (1974) showed to be the dividing line between the miogeoclinal Miri Zone, which incorporates the Central Luconia and Balingian Provinces, and the eugeoclinal Sibu Zone.

I interpret the Sibu Zone as a thick turbidite fan which was deposited upon oceanic crust. The southwards drifting of the Central Luconia-Balingian microncontinent compressed this turbidite fan against the Lupar Line accretionary wedge to form a mobile belt suture between the microcontinent and the West Borneo Basement and its Kuching Zone shelf. The mobile belt is of tightly deformed slaty and phyllitic Late Cretaceous to Late Eocene Rajang Group Belaga Formation, which has been described by Wolfenden (1960).

During this post Eocene compression, the boundary between the Miri and Sibu zones was tectonically active—the Bukit Mersing Line of Hutchison (1975).Some pillow basalts were uplifted from the oceanic floor which underlies the Belaga Formation. Some igneous activity occurred along the convergent margin. Undated but post Ecocene granodiorite was intruded at Piring Hill, closely associated with ignimbrite and minor andesite in the Upper Eocene to Oligocene Tatau Formation east of Piring Hill (Wolfenden, 1960). At Tabau Hill the anticlinally folded ignimbrite overlies andesite. The potassic granodiorite and ignimbrite indicate that the continental basement of the Miri Zone was involved in the magma genesis, while the oceanic basement of the Sibu Zone was uplifted at Bukit Mersing (Fig. 4, lower). The Balingian Province therefore represents the leading edge of the microcontinent, and it has been strongly involved in compression tectonics, whereas the posterior part—Central Luconia Province—was protected from the collision tectonics.

Accordingly the sedimentary history of the Balingian Province does not show the same strong relationship to eustatic sea level changes as does the Central Luconia Province. The strata of Balingian are predominantly of Cycles I and II, which outcrop onland as the Upper Eocene to Lower Miocene Setap Shale and Nyalau formations (Haile, 1962; Wolfenden, 1960). Structural deformation of western Balingian took place in early Miocene times (post Cycle II) which resulted in highly faulted ridges and depressions (Du Bois, 1981). Since Miocene times, western Balingian has remained relatively elevated and has been extensively eroded. During Late Miocene time (end of Cycle V) eastern Balingian was compressed and uplifted. The folds are asymmetric and reverse faulted. Transcurrent faults within the basement may have contributed to the tectonic complexity (Du Bois, 1981).

The basin classification J (Table 1) is of variable characteristics. In the terminology of Kingston *et al.* (1983) the Central Luconia-Balingian Provinces form a

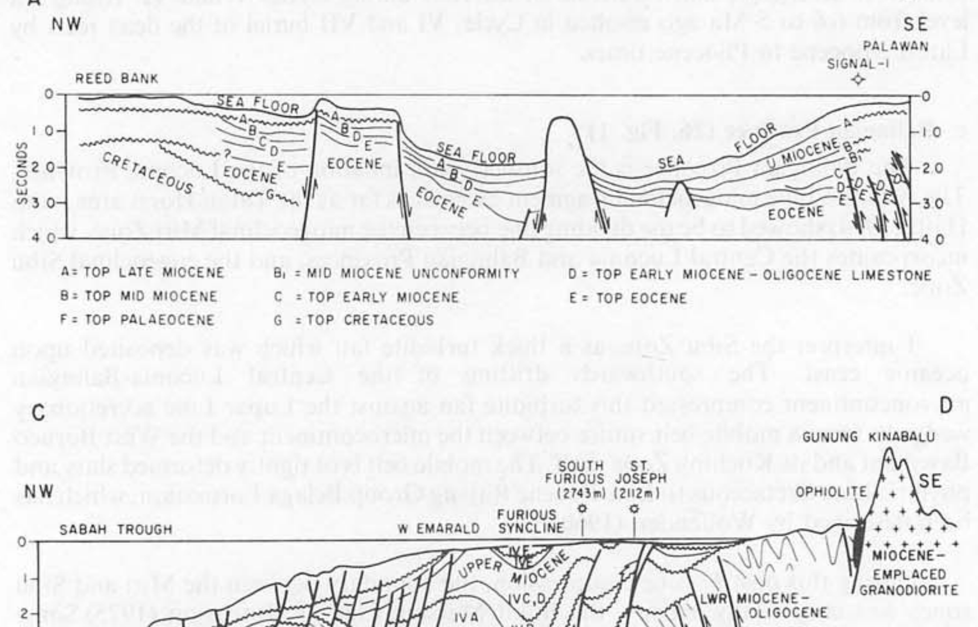

Fig. 5. Northwest-southeast cross sections across A: Reed Bank to Palawan; B: Sabah trough-Sabah basin-Mount Kinabalu. C: Time-stratigraphic section through Labuan (after Ascope, 1981; and Bol and Van Hoorn, 1980).

microcontinental fragment of a continental margin sag, deformed by collision on the Balingian side.

f. Sabah and West Palawan Basins (23, 16, Fig. 1)

Both of these basins have some former trench association and both have been tectonically complicated (Fig. 5). The complication has been caused by the southwards drift of the Reed Bank continental platform (Holloway, 1981), and Hinz and Schlüter (1983) have shown that the continental lithosphere extends southwestwards at least as far as the Dangerous Grounds. This large submerged platform also probably extends to include the Spratley Islands (Hutchison, 1984). It represents the rifted miogeocline of China, attenuated by horst and graben structures, and pushed southwards by the Mid Oligocene to Early Miocene spreading of the South China Sea Basin (Taylor and Hayes, 1983). The arrival of the microcontinent at the N.W. Borneo-Palawan Trough resulted in its southeasterly underthrusting beneath the Oligocene to Lower Miocene Crocker Formation turbidite fan, causing uplift of the fan and parts of its oceanic crustal basement as ophiolite (Fig. 5). Before this collision event, there was no landmass in North Borneo. The provenance of the Crocker Formation turbidite fan lay to the west in Sundaland, and the proto Mekong river was the only river large enough to have transported the sediment.

The Early Miocene underthrusting of the microcontinent uplifted and overturned large parts of the Crocker Formation to form land in Sabah for the first time, resulting in the Deep Regional Unconformity of Bol and Van Hoorn (1980), as shown in Figure 5.

The uplifted Crocker Formation provided the new provenance for the middle Miocene and younger shallow water deposits.

Although the region has continued to be tectonically active throughout the Miocene, some of the unconformities correlate with eustatic falls in sea level, notably the Late Miocene shallow regional unconformity. To a large extent, it must be eustatically controlled.

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