Pull-apart development at overlapping fault tips: Oblique rifting of a Cenozoic continental margin, northern Mergui Basin, Andaman Sea

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ABSTRACT

The northern Mergui Basin (Andaman Sea) contains east-northeast–west-southwest–to northeast-southwest–striking normal fault-bound basins, and north-northwest–southeast–trending strike-slip faults. The two largest strike-slip faults (Manora and Mergui) pass into extensional or transtensional basins at their tips, consistent with dextral offset. The faults provide examples of early stage pull-apart basin development at fault tips instead of the more common model for development at releasing bends. Offset of isochron markers for the Ranong Formation indicate that ~8 km of dextral offset has occurred along the Mergui fault and 4.5 km of dextral offset has occurred on the Manora fault. The strike-slip faults and associated extensional faults formed relatively late for the history of the entire Mergui Basin during the Early Miocene. The northern part of the Mergui Basin developed after a phase of west-northwest–east-southeast extension during the Oligocene in the Mergui Basin to the south, indicating a rotation in the extension direction toward the north-northwest–south-southeast with time. The basin is part of a major transtensional system involving the Sumatra, West Andaman, and Sagaing faults that accommodated the northern motion of western Myanmar as India moved north relative to Southeast Asia. Fault activity in the northern Mergui Basin decreased significantly when the broad zone of Early Miocene transtension became focused on the Alcock and Sewell Rises during the Middle Miocene, and the West Andaman and Sagaing faults began to develop and interacted in a large pull-apart geometry with the Shan Scarp Fault, and later (Late Miocene or Pliocene) with the Sagaing Fault.

INTRODUCTION

The Andaman Sea is a classic example of a Cenozoic backarc spreading center formed in a pull-apart geometry. Western Thailand and eastern Myanmar underwent transpressional crustal thickening and metamorphism during the Paleogene as a consequence of subduction (Barley et al., 2003; Morley, 2004, 2012; Searle, 2007), followed by a crustal block west of the Sagaing Fault (Burma Platelet) becoming coupled to the northwards-moving India plate between 35 and 30 Ma (see reviews in Hall, 2012; Rangin et al., 2013). Despite the overall timing and geometry of the region being well established (for a review see Curray, 2005), the structural evolution and structural style of the east Andaman passive margin, which contains the Mergui Basin (Fig. 1), has only been described on a regional scale (e.g., Polachan and Racey, 1994; Andreasen et al., 1997). The main dextral active strike-slip faults in the region controlling the pull-apart system are the Sagaing (onshore–offshore Myanmar), West Andaman (offshore), and Sumatran (onshore Sumatra, Indonesia) faults (Fig. 1; Curray, 2005).

This paper describes the structural geometry, timing, and evolution of strike-slip and extensional faults affecting the Miocene–Holocene northern Mergui Basin, in the Taiwan waters of the Andaman Sea, based on two-dimensional (2D) oil industry seismic reflection data and well data. Two strike-slip faults, the Mergui and Manora, are the focus of this study. The evolution of the northern Mergui Basin is important for understanding the development of an oblique extensional to transtensional continental margin and the Sagaing Fault. The Mergui and Manora faults are also of general structural geology interest in that they provide excellent examples of the interaction between strike-slip faults and extensional faults, and of structures formed at the tips of strike-slip faults.

GEOLOGY OF THE ANDAMAN SEA

The Andaman Sea region has long been identified as a region of plate-scale oblique motion, where backarc extension is linked with dextral shear on major north-south–trending strike-slip faults (Rodolfo, 1969; Ridd, 1971; Paul and Lian, 1975; Mitchell and McKerrow, 1975; Curray et al., 1979, 1982). According to the model of Curray (2005), ~460 km of opening has occurred, with much of the extension accommodated by dextral motion along the Sagaing–West Andaman fault system.

The Andaman Sea comprises these tectonic elements (Fig. 1): (1) Cenozoic (Oligocene–Holocene) sedimentary basins on the eastern and northern margins of the region (e.g., North Sumatra Basin, Mergui Basin, East Andaman Basin, Gulf of Martaban); (2) a deep-water trough (~3000 m deep) that extends north-south along the trend of the East Andaman Basin, and an east-northeast–west-southwest trough that branches off to the west that extends along the Central Basin; (3) oceanic crust along the Central Basin (this crust was thought to be a Middle to Late Miocene spreading center, but more recently its activity has been reinterpreted as Pliocene [Raju et al., 2004; Curray, 2005]); (4) two high, topographically irregular regions (minimum bathymetry ~1500 m) on the west side of the deep-water trough, that are split by the spreading center, called the Alcock and Sewell Rises; they may represent oceanic crust thickened by Cenozoic plateau basaltic activity, or hyperextended continental crust (Curray, 2005; Morley, 2012); (5) a number of large north-south–trending strike-slip faults that extend down the western margins of the Alcock and Sewell Rises, and down the sedimentary basins east of the rises; and (6) the accretionary complexes that have surface expression as the Andaman, Nicobar, and Coco Islands on the western margin of the Andaman Sea (see Curray, 2005, for a review).

The duration of activity of the major strike-slip faults is difficult to establish. The West Andaman fault and other faults (e.g., Diligent fault, Eastern Margin fault) on the western margin of the Andaman Sea (Fig. 1) are major geomorphological features and clearly were very important fault zones during the Pliocene (Curray, 2005; Cochran, 2010). However, there are no well-dated basins along the West Andaman fault zone that would categorically suggest a long history of displacement back to
Activity along the Sagaing fault in eastern-central Myanmar (Fig. 1) is only demonstrable for the Late Miocene–Pliocene when strike-slip motion inverted older transtensional basins forming the Begu Yoma and central basin, onshore Myanmar (Pivnik et al., 1998). Bertrand and Rangin (2003) considered a scenario where the Sagaing fault was only active during the past 4–5 m.y., when the Andaman spreading center became active. To the east of the Sagaing fault is the Shan scarp fault zone (Fig. 1). According to Bertrand et al. (1999) and Bertrand and Rangin (2003), the Shan scarp was the site of dextral transtensional activity from the Oligocene, as proposed by Curray (2005). Seismic reflection data along the southern portion of the West Andaman fault zone suggest that it was only active since the Middle Miocene or later (depending upon how unconformities on the seismic data are correlated; Berglar et al., 2010).
the Late Oligocene to the Middle Miocene. The Shan scarp fault zone could substitute for the Sagaing fault zone in the tectonic model of Curray (2005) for events prior to the Pliocene. The South Sagaing fault zone, in the East Andaman Basin (Fig. 1), is an offshore fault known from 2D and 3D seismic reflection data. It was active prior to the Middle Miocene, and is directly on the ocean-continent boundary interpreted by Curray (2005). The fault lines up directly with the active Sagaing fault to the north, but in terms of timing is similar to the Shan scarp fault zone.

GEOLOGY OF THE MERRUGI BASIN

The geology of the Mergui Basin is known in detail from seismic reflection data and drilling due to hydrocarbon exploration from the 1970s onward. The classic paper on flower structures by Harding (1985) was partially based on 2D seismic reflection data from the Mergui Basin. The basin was described as a trans-tensional basin, with north-northwest–south-southeast–trending dextral primary shear zones, and associated northwest-southeast dextral and northeast-southwest (antithetic) sinistral fault sets. Negative flower structures dominate along northeast-southwest–trending wrench faults, accompanied by north-south– to north-northwest–south-southwest–trending, oblique, en echelon normal faults.

An extensive oil industry database was utilized by Polachan (1988) to develop the cornerstone reference work documenting the general geology, stratigraphy, sedimentary evolution, and structural development of the area. The industry database was subsequently expanded; 19 exploration wells and ~20,000 km of 2D seismic data have been acquired in the Mergui Basin by various oil companies (including Exxon, Unocal, Kerr McGee, Unocal, PTTEP and the Thailand Department of Mineral Resources).

Polachan and Racey (1994) dated the main extensional phase as Late Oligocene–Early Miocene. Seismic data are suggestive of deeper, poorly imaged extensional fault–bound basins that are older (probably Early Oligocene or Late Eocene). Unpublished strontium ages (ca. 32–34 Ma) from carbonates on a basement high in an exploration well also suggest that deposition began near the Eocene–Oligocene boundary, at least in the southern part of the basin. These carbonates are underlain by red beds, which seem to correlate with the lower parts of half-graben fill moving off the high. Extension is largely characterized by boundary faults that give rise to half-grabens; maximum sediment thicknesses in the depocenters are ~3–4 km.

Fault growth continued in some places into the Middle Miocene, a few faults reactivating to the present day. However, this activity is minor compared with the earlier stage. Oligocene–Early Miocene extension is the dominant tectonic imprint on the main Mergui Basin area and the North Sumatra Basin.

Following cessation of major continental rifting during the Early Miocene, western Myanmar was translated away from the Mergui Basin along north-south–trending strike-slip faults that developed along the boundary between Sundaland continental crust and backarc oceanic crust (Curray, 2005). Consequently the Mergui Basin changed from fault-dominated subsidence to thermal, postrift subsidence. Deposits of the postrift basin onlapped the old, largely inactive structural highs. In the main depocenters the postrift section (Middle Miocene–Holocene) is as thick as ~3 km.

Offshore to the west much of the continental shelf of the Mergui Basin is in unusually deep water (~400–1000 m); although this is too deep for a typical shelf, the flat, gently dipping morphology of the seafloor that ramps up to the coast is that of a shelf. West of the shelf (including the broad basement high of the Mergui Ridge; Fig. 1) is the continental slope (~60 km wide), which descends to water depths in excess of 3000 m. Polachan (1988) interpreted the Mergui Ridge as an old, eroded Cre-taceous–early Cenozoic magmatic arc related to northeast subduction of Indian oceanic crust. Late Oligocene–Miocene rifted sequences expand westward off the high to thicknesses >4 km (East Andaman Basin; Figs. 1 and 2). The presence of two large depocenters either side of the ridge indicates that the topography is at least partly related to a flexural isostatic response to loading of the crust by sediments east and west of the ridge.

The basins east of the Mergui Ridge exhibit half-graben geometries and in the main part of the basin trend predominantly north-northeast–south-southwest (Figs. 2 and 3). Back to back half-grabens (eastern and western Mergui Basin) set up the horst block called the central high (Figs. 2 and 3). Northward into the study area the basins become smaller and trend northeast-southwest. The north-northeast–south-southwest– north-south–trending predominantly extensional West Mergui fault passes northward into the north-northeast–south-southwest–trending Mergui fault, which is one of the strike-slip faults we investigated (along with the Manora fault; Fig. 2). In the northern Mergui Basin the Mergui fault and northeast-southwest–trending half-graben basins mark significant differences with the area to the south.

STRATIGRAPHY OF THE MERRUGI BASIN

Stratigraphic data for the Mergui Basin were compiled from Polachan and Racey (1994) and Andreaen et al. (1997). Ages are mostly from unpublished micropaleontology studies, in particular of foraminifera, from the exploration wells. The following nine formations are recognized within the Mergui Basin, based on Polachan (1988), Polachan and Racey (1994), and Andreaen et al. (1997; Fig. 4).

1. The Takua Pa Formation is composed of Pliocene–Holocene calcareous and/or glauconitic shales and occasional silstones deposited in an outer shelf to lower upper slope environment.
2. The Thalang Formation comprises Late Miocene glauconitic shales interbedded with silstones and fine-grained sandstones and occasional limestones deposited in a shelf to upper slope environment.
3. The Trang Formation consists of Middle Miocene deep–marine sediments subdivided into two units. The lower unit comprises glauconitic shales, rare silstones, and sandstones deposited in a low-energy basin–plain environment. The upper unit is composed of glauconitic shales, silstones, sandstones, and calcarenites. Coarser clastic units are interpreted as being deposited in a distributary channels in a mid–lower fan setting.
4. The Surin Formation is Middle Miocene in age and consists of sandstones, shales, and calcarenites deposited predominantly in a shallow–marine environment. Progradational deltaic deposits are present in some parts of the basin (particularly the north Mergui Basin).
5. The Tai Formation this is a carbonate unit that developed during the late rift stage (Early Miocene to early Middle Miocene) on structural highs across the basin. The formation is best developed on the central high, where in the Tai–I well, it is divisible into three units. The basal unit comprises dolomite, anhydrite, mudstone, and sandstone. The middle unit consists of massive corallalgal limestones, and the upper unit comprises fore-reef calcarenites, shales, and sandstones. However, wells indicate other buildups may be deeper water mounds, not coral-algal reefs.
6. The Kantang Formation is of Early–Middle Miocene age and comprises two units. The lower unit consists of glauconitic shales and occasional thin silstones and fine-grained sandstones and was deposited in a slope to bathyal basin–plain setting. The upper unit comprises glauconitic shales interbedded with silstones, fine sandstones, and calcarenites, deposited in a middle bathyal, mid–lower fan environment.
7. The Payang Formation is composed of Early Miocene shallow-marine clastics that pass
upsequence from delta plain to progradational delta deposits.

8. The Yala Formation is composed of Late Oligocene to Early Miocene slope to bathyal shales in its lower part and lower fan turbidites in its upper part. It is a fine-grained deep-water equivalent of the Ranong Formation.

9. The Ranong Formation is predominantly of Oligocene age, but may be as old as Late Eocene and as young as Early Miocene in some basins. It is a clastic formation that was predominantly deposited in a marine shelf environments, although potential continental deposits are encountered in one well. Delta systems were well developed and fluvialite, delta plain and delta front environments have been identified. On some highs, the Oligocene clastic Ranong Formation passes laterally into carbonates that are older than the Tai Formation.

The Ranong and Yala Formations represent the synrift section, while in some areas the Kantang Formation represents the postrift stage, and in other areas it is synrift. It is more problematic that in some areas (southern Mergui Basin) the age of the Kantang Formation is Early Miocene (Ranong Formation is confined to the Oligocene), while in the northern part of the basin the Kantang Formation is of Middle Miocene age. This usage creates some terminology problems because the age of rifting does not appear to be the same everywhere in the Mergui Basin. In the south the age of rifting may be Late Eocene to Oligocene. Farther north the top of the Ranong Formation is within the Early Miocene. High-resolution biostratigraphy from the Manora-1 well indicates that what is called the Ranong Formation is exclusively of Early Miocene to possibly early-Middle Miocene age (Fig. 5). At present the existing stratigraphic scheme still refers to the diachronous sand-rich synrift section as the Ranong Formation. This paper is not focused on revising the stratigraphy, and will use the formation names identified in the Manora-1 well. However, clearly a reappraisal of the stratigraphic nomenclature is needed. Figure 4 presents the stratigraphy for the northern Mergui Basin based on the Manora-1 well.

Manora-1 was drilled in A.D. 2000 on a structural high in the footwall block of the extensional region of the Manora fault (Fig. 5). Biostratigraphy from the well (conducted by Edelman, Percival and Associates in 2000; in-house report) indicates a Late Miocene interval between 670 and 830 m (subsea level), characterized by the *Globorotalia (T.*) acostacnsis* zone (N18). From 850 to 1760 m is the Middle Miocene *Globorotalia (T.*) siakensis* (N14) Zone to *Praeorbulina glomerosa* (N8) Zone, which includes *Florschuetzia meridionalis* (late-Middle Miocene–Holocene). The occurrence of *G. (T.*) siakensis* in cuttings samples below 1550 m indicates an Early to late-Middle Miocene age for the interval from 1550 to 1760 m. *Lanagio-pollis emarginatus* (late Middle Miocene–Holocene) occurs in a sidewall core at 1052 m, and in cuttings and sidewall cores between 1263 and 1760 m. An Early to Middle Miocene age is determined for the interval between 1760 and 2310 m on the basis of *G. (T.*) siakensis* in cuttings samples below 1550 m indicates an Early to late-Middle Miocene age for the interval from 1550 to 1760 m. *Lanagio-pollis emarginatus* (late Middle Miocene–Holocene) occurs in a sidewall core at 1052 m, and in cuttings and sidewall cores between 1263 and 1760 m. An Early to Middle Miocene age is determined for the interval between 1760 and 2310 m on the basis of *G. (T.*) siakensis* (earliest stratigraphic occurrence, Early Miocene, latest appearance late-Middle Miocene), *Globorotalia (T.*) peripherorum* (Early Miocene to early-Middle Miocene), *Globigerinoides sicarius* (Early Miocene to early-Middle Miocene), *Florschuetzia semilobata* (Early to Miocene), and *Florschuetzia levipoli* (Early Miocene to Holocene). The well reached total depth in green kaolinitic hard basement associated with chlorite and feldspar; it is unknown
whether the basement is altered igneous rocks or metamorphic schists (such as those encountered in wells farther south).

**DATA AND METHODOLOGY**

This study used oil industry 2D seismic data in the northern part of the Mergui Basin where two sets (1998 and 1999) of data were available for interpretation. Both surveys were shot by Western Geophysical with a 2–3 km line spacing, and the total line length in the study area is ~4000 km. The migrated lines have a 7 s record length, with 1750 samples per trace. The study area dimensions are ~80 km north-south and 80 km east-west; 80 2D seismic lines oriented north-south, east-west, and northwest-southeast were interpreted. One well (Manora-1) has been drilled in the area, and was used for calibrating lithology and ages to the seismic data (Fig. 5).

Seven seismic horizons were interpreted; each horizon was chosen to reflect significant structural or sedimentary events in the basin (Figs. 4 and 6). Time-depth structure maps were made for each seismic horizon. Six time-depth isochron maps were generated for the intervals between each horizon by subtracting the values of the overlying time-structure map from the lower time-structure map. The characteristics of the picked horizons are as follows.

1. The base of the synrift (blue) horizon marks the base of the synrift section (Figs. 6 and 7). In places the synrift section shows clear onlap onto prerift topography. Commonly the horizon is a strong amplitude reflection that separates a strongly to weakly reflective synrift sequence from the prerift basement, which shows little coherent internal reflectivity.

2. The top of the Ranong Formation (pink) was initially picked at a significant local angular unconformity associated with the Manora fault (Fig. 8C) and then carried into areas of conformable reflections. The horizon coincides with the top of the Ranong Formation in the Manora-1 well. In the study area the Ranong Formation tends to form an interval of relatively weak, discontinuous reflectivity, with a few strong, discontinuous, high-amplitude internal reflections.

3. The top of the Kantang Formation (light green, Figs. 4 and 6) underlies a well-developed series of prograding clinoforms. The sequence boundary that underlies the packet of clinoforms was picked as marking the approximate top of the Kantang Formation.

4. A horizon (green, Figs. 4 and 6) was picked within the Surin Formation at the highest event that underlies a second set of prograding clinoforms that are above the approximate top of the Kantang Formation. The horizon was carried westward into deeper water sediments that compose the Trang Formation.

5. A Thalang Formation (purple, Figs. 4 and 6) horizon follows the top of the highest clinoforms, and is marked by a toplap surface and correlative units of Late Miocene age.

6. The top of the Thalang Formation (yellow, Figs. 4 and 6) approximately marks the top of the Miocene–base of the Pliocene, and follows a high-amplitude reflection with good continuity over the entire study area. However, in the southwestern part of the study area the horizon was not mapped due to an unconformity between the Miocene and Holocene.

7. The seafloor is shown in light blue in Figures 4 and 6.

**MANORA FAULT STRUCTURAL GEOMETRY**

The Manora fault exhibits a straight north-northeast–south-southeast central segment ~25 km long that passes into northeast-southwest–trending extensional and/or transtensional fault sets at the northern and southern ends (Figs. 7, 8, and 9A). Like the Mergui fault, the extensional basin at the southern termina-
The fault patterns in the area are illustrated by the 3D perspective views of the base of the synrift surface (Fig. 10). The remnant parts of the Mergui Ridge (Fig. 10, location c) are seen as the highs at locations e and f. These highs are cut by the Manora fault (e) and the East Manora fault (f, Fig. 10). The perspective view in Figure 10A shows the variable offset along the Mergui fault at the synrift level, and at location (b) the vertical offset is drops to near zero. Location a (Fig. 10) marks the tip of the Mergui fault with a compression-related high on the east side. The perspective view in Figure 10B shows the large extensional offset of the Manora fault at location d; the fault steepens to vertical at location e.

**MERGUI FAULT STRUCTURAL GEOMETRY**

The Mergui fault is a linear, north-northwest–south-southeast fault that is ~60 km long in the study area (Fig. 7). To the north the fault terminates into a series of northeast-southwest–trending extensional splays (Fig. 7). Figure 11 shows seismic lines across the Mergui fault, and Figure 12 presents a series of cross sections based on seismic lines that show the variations in geometry along the Mergui fault. In the north the fault dips to the southeast and is comparatively low angle (Figs. 11A, 12A, and 12B). Erosion of the footwall area adjacent to the fault occurred during deposition of the Kantang Formation (Figs. 11A and 13). Southward to the north-northwest–south-southeast–trending part of the fault segment, there is commonly a convergent fault geometry seen on the seismic lines (Figs. 12C–12K), the largest displacement occurring on a east-northeast–dipping fault that is downthrown to the east (Figs. 12B–12E, 12I and 12J). Figure 11C (fault X) and Figures 12F—12H show a change in this displacement pattern, where throw at the top basement has a small reversal of offset on the east-northeast–dipping fault. Compared with Figure 11B, the sedimentary dips and thickening directions for the Kantang and Ranong Formations are reversed.

Figure 11C marks a southward transition toward the Mergui Ridge, which is seen in other lines (Figs. 11D–11F). The Mergui fault becomes more vertical, and the secondary faults that form a flower-structure geometry become narrower and less vertically extensive as the Ranong Formation thins to zero thickness onto the Mergui Ridge (Figs. 12K–12M). This thinning resulted both from onlap and later erosion. The eastern side of the fault (Fig. 11E) also shows local uplift and erosion. The westward thickening of the Ranong Formation could be interpreted as fault controlled; however, this thickening is actually more regional in extent.
Figure 5. Seismic line showing the correlation with the Manora-1 well (see Fig. 7 for location) Fm.—formation.

Figure 6. Regional seismic line showing the two main strike-slip faults (Mergui, Manora) in the study area, and the picked seismic horizons (see Fig. 7 for location). Fm.—formation.
Figure 7. (A) Base syn-rift time-structure map of the study area from 2D seismic interpretation. Dashed white lines indicate Figure 11 sections, which are longer than Figure 12 sections. (B) Same map as A, without the figure locations. See Figure 2 for location.
Figure 8 (on this and following page). Seismic sections illustrating the change in geometry of the Manora Fault from north (A), to south (F). See Figure 7 for location.
Figure 8 (continued).
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and continues tens of kilometers eastward, into the present-day deep-water area. The apparent fault-controlled thickening of the Ranong Formation is at least partially explained by factors unrelated to the Mergui fault. The west side of the fault displays a thin section because it is located on the northward-plunging nose of the Mergui Ridge. When the Mergui fault underwent right-lateral motion, the thicker Ranong Formation on the east side of the fault was juxtaposed with thinner Ranong Formation strata on the west side of the fault. Consequently, the apparently fault-related thickening into the fault on east-west seismic lines is actually related to the plunging nose.

The most southern seismic lines across the Mergui fault reveal that the fault dips to the west (Figs. 11G, 11H, 12N, and 12O). This marks the linking of the Mergui fault with the boundary fault to the west Mergui Basin (Fig. 3). The southern link is beyond the study area discussed here; however, it is useful to compare the fault plane links and geometry of the West Mergui boundary fault with the Mergui fault (Fig. 9B) because they are rather similar to those for the Manora fault (Fig. 9A), but on a larger scale.

The Mergui fault is a very clear feature on northeast-southwest–trending seismic lines, but it is absent on the three most northern northeast-southwest–trending lines. The seismic lines at the northern end of the Mergui fault provide a very clear example of the structural style at strike-slip fault terminations. The east side of the fault shows several large northeast-southwest–trending, southeast-dipping fault strands that represent the releasing bend area of a dextral fault tip (Fig. 7). Conversely, the western side of the fault is marked by a compressional fold, bound to the south by an east-northeast–west-southwest–striking reverse fault zone (Figs. 13 and 14). In detail secondary faults affecting the folded area display a mixture of reverse and normal offset. The pattern of compression on the left stepping side, and extension on the right stepping side, of a dextral fault is similar to the ideal strike-slip fault tip geometries predicted by Chinnery (1966) and Segall and Pollard (1980). The southern transition into the transtensional west Mergui Basin is also appropriate for dextral motion.

**TIMING AND DISPLACEMENT AMOUNT OF THE MERGUI AND MANORA FAULTS**

Interpretation of isochron maps coupled with stratigraphic patterns on vertical seismic lines enables a clear understanding of the timing of fault activity. The north-northeast–south-southwest–trending axis of the Mergui Ridge high is apparent on the Ranong Formation isochron map (Fig. 15A). This high was present prior to development of faulting in the study area, and was subsequently offset by movement on the Mergui and Manora faults. Based on the Ranong Formation isochron map, the central part of the ridge high collapsed due to early extensional activity along the Manora fault (Fig. 15). The regional, westward-thickening pattern of the Ranong Formation north of the Mergui Ridge is apparent from the isochron map. Offset of the 250–500 ms isochron thickness package indicates ~8 km dextral displacement along the Mergui fault (Fig. 15A, arrows a and b). For the Manora fault, offset of the preexisting basement high indicates ~4.5 km of dextral offset (Fig. 15A, arrow c). This estimate matches well with the heave on the southern extensional part of the fault segment (Fig. 15A). Extensional and/or transtensional

![Figure 9](https://pubs.geoscienceworld.org/gsa/geosphere/article-pdf/10/1/80/3334005/80.pdf)
depo centers along northeast-southwest–trending faults and fault segments in the study area are well defined and reach a maximum thickness of ~1500 ms (~2100–2300 m).

The isochron map for the Kantang Formation (Fig. 15B) shows a significant difference from the Ranong Formation; the west-thickening trend seen in the Ranong Formation is no longer present. A high is present west of the Mergui fault, where the Ranong Formation was previously thickest. The Kantang Formation on the west side of the fault is thin to absent and thins significantly toward the northern (west side) compressional fault tip. Expansion of section into northeast-southwest–trending half-graben–type faults is well defined and fairly similar to the Ranong Formation, with maximum thicknesses of ~1200–1300 ms (~1700–1900 m).

The lower Surin Formation–Trang Formation (Fig. 15C) shows a marked change compared with the Kantang and Ranong Formations, in keeping with reduced tectonic-related subsidence due to the onset of postrift passive margin subsidence (Polachan and Racey, 1994). The maximum sediment thickness and sedimentation rate are much reduced (maximum thickness ~250 ms, or 350–370 m). The areas of maximum thickness are less clearly fault controlled, and in the northeast of the area the isopach thickness is related to input of a deltaic system. The interval within the Surin–Trang to Thalang Formation shows evidence for some renewed extensional and strike-slip activity with thickening in the southern Manora fault depocenter (Fig. 15D). Both the Surin–Trang to Thalang Formations and the Thalang Formation isochron (Fig. 15E) maps show evidence for weak fault activity, where the western side of the Mergui fault is upthrown with respect to the eastern side. The Takua Pa Formation isochron shows a gradual thickness change to the west, with the maximum sediment thickness reaching only ~250 ms (~270 m over ~5 m.y. time period; Fig. 15F). Some minor fault activity along the Manora and Mergui faults resulted in some fault strands locally cutting into the Takua Pa Formation. However, it is clear from seismic sections such as the one shown in Figure 16 that the main episode of activity on the Mergui fault ceased around the end of the Middle Miocene–earliest Late Miocene, and the number of active fault strands and their displacement amount were very much reduced after that time.

The onsets of displacement on the Mergui fault appear to have occurred during deposition of the upper part of the Ranong Formation (late Early Miocene to early Middle Miocene). Prior to that time a thick Ranong Formation sequence had been deposited, but thickness patterns were not strongly influenced by the Mergui fault; a more extensional or transtensional east-northeast–west-southwest–trending fault pattern was active. In the northern compressional fault tip area of the Mergui fault, the upper part of the Ranong Formation is folded on relatively short wavelength folds and shows internal unconformities (Fig. 13). At the end of deposition of the Ranong Formation and during deposition of the Kantang Formation, the large, broad fold and thrust structure developed at the northern tip of the Mergui fault, and folding continued into the Trang Formation (Figs. 13 and 14).

**FAULT DISPLACEMENT PATTERNS**

The dip-slip component of displacement of horizons across the Manora and Mergui faults was measured from seismic reflection data.
Figure 11 (on this and following two pages). Seismic sections illustrating the change in geometry of the Mergui Fault from north (A) to south (H). See Figure 7 for location of lines A–G, and Figure 2 for location of H. TWT—two-way traveltime.
Figure 11 (continued).
Displacement can be represented in two ways, the present-day (finite) displacement measured directly from the seismic data, and the displacement for a particular stratigraphic interval. The offset that occurred during the time period between two horizons is referred to here as interval displacement. The interval displacement values are obtained by successively subtracting the displacement of a higher interval (H1) from the next highest interval (H2) to obtain the interval displacement (ID) (H2 to H1 = ID2 to ID1), then for lower horizons the interval displacement for the higher unit is subtracted from the next horizon (H3 – ID2 – ID1 = ID3 – ID2; H4 – ID3 – ID2 = ID4 – ID3; e.g., Morley et al., 2007). If the deeper value is less than the higher value, a negative number arises, which for normal faults indicates no expansion of section, i.e., no displacement occurred during that interval, and so a value of zero is given to any negative number. For normal faults the fault displacement graphs produce relatively simple patterns (e.g., Morley et al., 2007; Dutton and Trudgill, 2009); for strike-slip faults the graphs look more complex.
Figure 12. Line drawings from seismic lines illustrating in detail the changes in fault geometry along the Mergui Fault passing from north (A) to south (O) along the faults. Lines A–N are located on Figure 7.
Figure 13. Strike line through the compressional tip of the Mergui Fault. See Figure 7 for location.
Pull-apart faults, Mergui Basin, Andaman Sea

for two reasons: changes in fault dip direction and along-strike changes in fault displacement from a normal to a reverse component. For normal faults the horizon offsets are derived from footwall values minus hanging-wall values. For strike-slip faults the values are one side of the fault minus the other side (for the Manora and Mergui faults west side minus east side was used). In the case of the Manora fault where the fault dips to the east, extensional offset would show as a positive number, while extensional offset on a west-dipping fault segment would show as a negative number. That means segments with a reverse sense of displacement can also be positive or negative numbers, depending upon fault dip direction.

In the case of the Mergui and Manora faults, a large component of strike-slip as well as dip-slip motion has occurred. The lowest units (Ranong and Kantang Formations) have undergone the largest strike-slip displacement. The maximum dextral strike-slip offset for the Mergui fault is estimated as 7 km (Fig. 15A). As a test, the east-side displacement values

Figure 14. Dip line through the compressional tip region of the Mergui Fault. See Figure 7 for location.
Figure 15. Isochron maps for the study area. (A) Ranong Formation (base syn-rift to top Ranong Formation). (B) Kantang Formation (top Ranong Formation to top Kantang Formation). (C) Top Kantang Formation to Intra Surin/Trang Formations. (D) Intra Surin/Trang Formations to Intra Thalang Formations. (F) Intra Thalang Formation to Top Thalang Formation.
Pull-apart faults, Mergui Basin, Andaman Sea

for the Ranong were moved 7 km south with respect to the west-side values and the Kantang Formation values were moved 3.5 km south to account for this motion (which is assumed to have largely finished by the start of deposition of the Trang Formation), and the dip-slip component of displacement was recalculated. The result produced only a minor difference to the overall pattern of displacement when the strike-slip component is considered; consequently, strike-slip effect was ignored. The figures illustrate the present-day vertical offsets of the footwall and hanging wall.

The Manora fault interval displacement pattern clearly shows the presence of the east- and west-side depocenters (Fig. 17A). The west-side depocenter had a large component of displacement during deposition of the Ranong Formation, which narrowed during deposition of the Kantang Formation as a result of uplift on the west side of the fault, along the strike-slip–dominated segment of the fault. This uplift is also significant in that it restricted the low displacement along the Kantang fault for the east-dipping depocenter. The most significant displacement occurred during deposition of the Ranong and Kantang Formations, with more minor displacements in the succeeding formations.

The Mergui fault pattern is significantly different from that of the Manora fault (Fig. 17B). Using the west side of the fault minus the east side convention, most of the interval displacement values are positive, indicating that the east side of the fault is predominantly the downthrown side. There are a few places where interval displacement values are zero; unlike normal faults, where such values would indicate no displacement, the values here simply indicate no vertical component of displacement. The overall impression from Figure 15 is that the fault was most active during the time of deposition of the Ranong and Kantang Formations, and activity was subsequently considerably reduced.

STRUCTURAL MODEL

Northeast-southwest–trending, predominantly northwest-dipping, half-graben, bounding normal faults are on the east side of the Mergui fault; this structural style is not seen on the west side. Northeast-southwest–trending faults are present on the west side, but they do not form well-developed half-grabens. The normal faults have an inappropriate orientation and displacement pattern to be releasing bends associated with Mergui fault. Instead, the faults formed either at a releasing bend associated with the Manora fault to the east, or simply as independent extensional and/or transtensional faults, the displacements of which decrease toward the Mergui fault.

The extensional structures described herein occur at the tips of strike-slip faults, and therefore differ from the pull-apart basins within fault segments that develop at releasing bends (as described from analogue modeling, e.g., Dooley et al., 2004; Wu et al., 2009), or as sidewall ripouts (Swanson, 1989, 2005). Deng et al. (1986) described the basic characteristics of fault tip basins, where pull-apart basin geometry was controlled by the magnitude of overlap at the ends of strike-slip faults. While analogue models have been made of overlapping and underlapping faults, there is always a pre-cut link in the baseplate that joins the fault tips (Dooley and Schreurs, 2012). This link forces the faults to develop together with similar displacements, and also seems to promote the development of Riedel (R) shears that link the two main fault trends (Figs. 18A, 18B). The Mergui is a longer and higher displacement fault system than the Manora fault system; the two might have interacted, but their displacements are not as interdependent as those in the analogue models. The
Displacement during Kantang Fm. in this area affected by uplift along fault

Figure 17. Interval displacement diagrams for (A) the Manora Fault and (B) the Mergui Fault.
Pull-apart faults, Mergui Basin, Andaman Sea

The southern segment of the Manora fault ends in an east-northeast–west-southwest extensional half-graben, the bounding fault of which loses displacement westward toward the Mergui fault. The hard linkage of overlapping faults generated in the analogue models does not match the observations from the north Mergui Basin.

The Mergui fault illustrates that during fault propagation and linkage early fault tip-related structures may be present in the middle of long, linked strike-slip fault systems (Fig. 18C, panel D, location X). In this transtensional or oblique extensional environment R' trends (east-northeast–west-southwest to northeast-southwest) are much better developed than the R shear trends, which appear to be short fault segments close to the main fault zone (Fig. 10B). Ultimately, at high displacements (Fig. 18Bii, Cii), if the Mergui and Manora faults become fully linked, the two types of pull-apart basin may start to resemble each other, and features like short-cut faults (R shears) may start to develop. However, their early development history would be different. At present the Mergui-Manora fault system is at the stage of Figure 18Di, and is unlikely to be active enough to reach stage Dii.

One of the early interpretations of transfer zones in rifts envisaged that strike-slip faults were important links between extensional faults similar to the style shown in Figure 19F (e.g., Rosendahl, 1987; Gibbs 1990). However, other models downplayed the importance of such faults and suggested that most extensional transfer zones display soft linkage geometries (Morley et al., 1990). Rift basins on passive margins, where relatively large amounts of extension have occurred, show important strike-slip faults interacting with normal faults (Jeanne d’Arc Basin; Tankard et al., 1989). One of us (Morley) has interpreted seismic reflection data in many failed rifts, and only rarely seen evidence in failed continental rifts for hard-linked transfer zones between extensional faults involving strike-slip faults. The Manora fault is a very clear example of such a transfer zone, and it is not surprising that it is located on a margin that has undergone transtension or highly oblique extension, not pure extension. There are strong similarities to the Oligocene–Miocene Fang

![Figure 18. Diagrams illustrating the basic differences between pull-apart basins formed at releasing bends in the center of strike-slip faults (A, B) and at the tips of faults (C, D).](https://pubs.geoscienceworld.org/gsa/geosphere/article-pdf/10/1/80/3334005/80.pdf)

![Figure 19. Summary of strike-slip to oblique-extension fault geometries involving strike-slip faults and fault trends with R shear and R' and T fracture orientations. The relative importance and geographic extent of the different fault orientation changes, with width of the strike-slip zone and degree of transtension or oblique extension that affects the area as illustrated in figures A–F. A–D follow models described in Dooley and Schreurs (2012). (A) Pure strike-slip along a single fault. (B) Strike-slip along a broad shear zone. (C) Broad simple shear zone. (D) Transtension. (E) R' and T fracture dominated transtensional or oblique extension zone (East Andaman Basin type). (F) North Mergui Basin type setting, either transtension or oblique extension. See discussion in text for more details.](https://pubs.geoscienceworld.org/gsa/geosphere/article-pdf/10/1/80/3334005/80.pdf)
Basin–Mae Cham fault in northern Thailand, where the hard-linked transfer zone geometry is also present in the transition region from a strike-slip system to the north, and extension-dominated basins to the south (Morley, 2007).

Dooley and Schreurs (2012) reviewed the main fault patterns found in analogue models of strike-slip fault systems ranging from single strike-slip fault zones to broad fault zones and zones of transtension. In all these systems R-shears dominate in length and density over faults of other orientations. In terms of angle to the principal displacement zone, northeast–southwest faults in the north Mergui Basin range from R'–T to T-fracture orientations (Petit, 1987; Figs. 19A–19D). The extensional splays at the terminations of the Mergui and Manora faults can be understood in terms of classic strike-slip fault geometries. However, the faults of east-northeast–west-southwest to northeast-southwest orientation that surround these faults, but are not directly linked with them, are common and well developed. The Mergui and Manora faults have developed in a region where R-shear development is strongly subordinate to extensional faults developed with R’ or T fracture orientation. The East Andaman Basin is very strongly dominated by R’ or T orientations and the north-northeast–south-southwest main strike-slip fault trend, as schematically illustrated in Figure 1. Consequently, the types of fault patterns found in the East Andaman Basin (Fig. 19E) and in the north Mergui Basin (Fig. 19F) are not matched by analogue experiments (Figs. 19A–19D). The Andaman Sea is not the only region to show this structural style; parts of the Bohai Basin also show very well-developed high-angle fault trends to the main strike-slip faults (Hsiao et al., 2004, their fig. 10), which Chen et al. (2010) referred to as T-tensional systems. The conditions that could produce R’ or T fracture dominated systems are uncertain: are they really just oblique extensional systems (i.e., $\sigma_1 = \sigma_3$, $\sigma_1 > \sigma_2 > \sigma_3$), or does a stress system where $\sigma_1 = \sigma_2 > \sigma_3$ have significance for these structures? Could traction forces arising from flow of the lower crust suborthogonal to the R’ or T direction play a role? What role might preexisting structures play? These issues cannot be addressed here; however, the structural styles described here are atypical for analogue model results, and further work to understand them is required.

DISCUSSION OF REGIONAL TECTONIC EVOLUTION

The main Mergui Basin is composed of three rifting events that took place in the Late Eocene–Early Oligocene, Oligocene, and Early Miocene. These events are not seen everywhere, and it is difficult to demonstrate the Eocene–Early Oligocene event in most areas. In some places only the Oligocene rifting in prominent, in other areas rifting is of Oligocene to early-Early Miocene age, or Miocene-age rifting is dominant (Fig. 20). Locally angular unconformities can separate the rifting stages (Fig. 21), while elsewhere the stages are conformable. The oldest rifting is in the south, and younger rifting is to the north (particularly the west Mergui Basin). The area of the E-1 well demonstrates the early rifting stage (see Fig. 2 for location). The Oligocene section expands into the boundary faults and is capped by a thin Early Miocene postrift section of fairly even thickness across the underlying fault system. Inversion formed a prominent anticline that is onlapped by later Early Miocene synkinematic strata. The East Mergui Basin shows both Oligocene and Early Miocene.

Although the Manora-1 well points to the dominance of Early and Middle Miocene strike-slip and extensional development, in some areas on the western side of the Mergui fault there is evidence that older (presumably Oligocene) rifting affected the Mergui Ridge (Fig. 21). There appears to have been a significant episode of uplift and erosion along the Mergui Ridge in the Late Oligocene to Early Miocene that removed sediment from these basins; it may be linked with the phase of inversion seen in the E-1 well. Inversion during the Middle Miocene is also seen in parts of the northern Mergui Basin.

In the northern Mergui Basin the westward thickening of the Ranong Formation, or at least its time equivalent in the East Andaman Basin west of the Mergui Ridge (Fig. 1C), indicates that during the Oligocene–Early Miocene the main area of rifting was to the west of the Mergui Ridge. The distinctive structural style of the northern Mergui Basin, marked by strong Miocene north-northwest–south-southeast strike-slip fault trends, developed later than the rift basins to the south, and appears to represent a north-northwest–south-southeast–extension direction, while the simpler normal fault patterns in the Mergui Basin to the south suggest more pure extension to weakly oblique extension (i.e., approximately west-northwest–east–southeast regional extension). Although linkage is shown between the West Mergui Fault and the Mergui Fault (Fig. 9B), this linkage came relatively late in the development of the West Mergui Fault, which shows an Oligocene depo-

![Figure 20](https://example.com/figure20.png)

**Figure 20.** Cross-sections based on seismic data illustrating the timing of rift basins in the Mergui Basin. (A) E-1 area. (B) West Mergui Basin. See Figure 2 for location.
Pull-apart faults, Mergui Basin, Andaman Sea

![Seismic section showing a remnant of an Oligocene rift basin below a Late Oligocene or Early Miocene unconformity. See Figure 2 for location.](image)

The north Mergui Basin underwent more oblique extension in a north-northwest–south-southeast direction (assuming that the extension direction is parallel to the Mergui and Manora faults). The diachronous onset of postrift subsidence in the Mergui Basin during the Early to Middle Miocene indicates the time when western Myanmar began to decouple from the Mergui Basin–North Sumatra Basin area. The timing of reduction in activity of the Mergui and Manora faults described herein is in accord with this history of reduced tectonic activity commencing in the Middle Miocene.

It is generally accepted that seafloor spreading in the central basin of the Andaman Sea began during the early Pliocene, and that activity on the Sagaing fault is probably linked with opening of the spreading center (Andaman scientific cruise reports; see http://www.geologie.ens.fr/Andaman/Pages/index_rapport.html; Raju et al., 2004; Curray, 2005; Morley, 2013; Rangin et al., 2013). However, one of us (Morley) has seen seismic data that indicate the presence, in the deep-water of offshore eastern Myanmar, of seaward-dipping reflections, which can be interpreted as related to igneous activity around the continent-oceanic crust boundary (e.g., Eldholm et al., 1989, 2000). West of these reflections the crust is covered by as much as 4 km of (probably latest Miocene to Pliocene) sediment of the East Andaman Basin. It is also apparent from data from across the axial trough (Raju et al., 2004, their fig. 10B) that in the eastern half of the region interpreted as Pliocene–Holocene oceanic crust, an even thickness of sediment (~600 ms two-way traveltime, i.e., ~600 m thick) covers the axial valley, and 24 km northwest of the valley sediment may be as thick as 1.3 km (Raju et al., 2004). Data from the 2000 Andaman cruise reports (Chamot-Rooke and Rangin, 2000) show 1 km of sediment in the axial valley region. These observations do not match the pattern of thinning and younging of sediment toward the axial valley that would be expected if oceanic crust was being continually generated at the spreading center. Unfortunately, the sediments are not dated, and typical deep-water sedimentation rates (0.001 to 0.1 mm/yr) are unlikely to be applicable in this region where sedimentation rates on the shelf during the Pliocene are high (~1–1.5 mm/yr). However, deep-water sedimentation rates should be a fraction of the shelf depositional rates and using an upper estimate of 0.5 mm/yr for the deep-water area is probably generous. Using an average sedimentation rate of 0.5 mm/yr, the oceanic crust underlying 1 km of sediment in the axial trough area should be at least 2 m.y. old (1 m.y. old if shelf depositional rates are used). Consequently, we suggest that the recent axial valley marks a renewal of activity after a hiatus of uncertain duration. Late Miocene oceanic crust is inferred to be on the eastern side of the northern part of the Sewell and Alcock Rises (covering a more limited area than that interpreted by Curray, 2005), where as much as 4 km of sediment is present (Fig. 1).

The northward movement of the coupled India–western Myanmar block requires that separation of western Myanmar from the eastern Andaman margin was accomplished by dextral translation along major strike-slip faults, and by extreme thinning, accompanied by intense igneous activity in the area of the Alcock and Sewell Rises. Curray (2005) interpreted through-going strike-slip fault systems during the Early Oligo-
Figure 22. Oligocene-Holocene structural and tectonic evolution of the Andaman Sea area, mostly based on data and maps in Curray (2005), Hall (2002), Morley (2004), and unpublished industrial seismic reflection data (PTT Exploration and Production), updated to reflect current understanding of fault timing (as discussed in “Geology of the Andaman Sea” in text). a—West Andaman Fault active; b—Sumatra Fault active; c—possible area of Late Miocene seafloor spreading. Revised and updated from Morley et al. (2011).
cene stage of basin development onward, linking the Sagaing fault with the old West Andaman fault. Figure 22 presents a modification of Curry’s (2005) model view, with only mildly oblique extension dominating the Oligocene stage of development.

The history of development of the north Mergui Basin suggests that only during the Early Miocene to early-Middle Miocene did significant strike-slip deformation commence, with rotation of the extension direction to the north-northwest–south-southwest. At that time a narrow zone of hyperthinned crust then formed in the East Andaman Basin (Fig. 22). Subsequent deformation during the Middle–Late Miocene became focused further northward in the vicinity of the Alcock and Sewell Rises, and this led to development of seafloor spreading during the Late? Miocene and Pliocene–Holocene (Fig. 22). This time period marks drowning of the shelf in the north Mergui Basin. The Middle Miocene shallow-marine prograding clinoforms, which indicate very shallow water depths for the topsets, now are under 400–600 m of water, indicating a considerable transgression in the past 10 m.y. This drowning is probably related to the creation of oceanic crust and onset of thermal subsidence following the cessation of extension and/or transtension. Flow of lower crust from the shelf area toward the Alcock and Sewell Rises area is one additional cause of subsidence that needs further investigation.

The evolution of the Andaman Sea is affected by the interplay of two major tectonic processes: the northwards movement of India relative to SE Asia, and subduction. It appears to be the case that as the collision of India has progressed, the role of subduction has become less important, so that today seismicity associated with the hyper-oblique subduction zone underlying the Indo-Burma Ranges can be interpreted as that associated with slab detachment (Rangin et al., 2013). Given that during the Eocene subduction-related plutonic activity ran from Peninsular Thailand up to the Myanmar-Yunnan border (as reviewed by Palin et al., 2013), there is a strong case to be made for westwards subduction rollback of several hundred kilometers sometime between the late Eocene and the Miocene (Pubellier and Morley, 2014). Subduction rollback (possibly coupled with gravity collapse of a thickened orogenic belt) provides a convenient mechanism for driving approximately E–W extension during the early stages of formation (Late Eocene–Oligocene) of the Mergui and the contiguous North Sumatra basins, as well as basins in the Gulf of Thailand (Morley, 2001, 2012; Pubellier and Morley, 2014).

As the influence of India-Burma Platelet coupling increased during the Late Oligo-

CONCLUSIONS

The Mergui and Manora faults provide classic examples of compressional and extensional structures at the terminations of strike-slip faults (e.g., Chinnery, 1966; Deng et al., 1986; Kim and Sanderson, 2006), and how local areas of transtension and transpression are distributed along strike-slip faults. The Mergui fault has undergone ~8 km of dextral offset; it originates in the south at the northern end of the west Mergui Basin, and terminates in the north in compressional and/or inversion structures on its west side, and extensional horsetails on its east side. Displacement appears to be linked to some degree with the Manora fault to the east via a right-stepping transfer zone. Transtensional deformation favors the development of major faults that follow the R′ or T fracture orientations (northeast-southwest to north-northwest–west-southwest trends). R′ shears are difficult to define from the 2D seismic grid, implying that if they exist, they are short, low-displacement faults that are close to the main strike-slip fault trends. Classic flower-structure geometry is dominated by faults that strike at a low angle to the principal displacement zone (i.e., R, Y, and P shears). The flower-structure geometries shown here (e.g., Fig. 9) are in part related to a narrow zone of probable R, Y, and P shears, but also forming a pseudo–flower-structure pattern are the extensional and/or transtensional faults following the R′ or T fracture orientations, which are too oblique to form a classic flower-structure association.

The main part of the Mergui Basin shows predominantly extensional, half-graben–type fault geometries. The basins began developing during the Late Eocene to Early Oligocene. The Mergui Ridge formed or was already present during this time, but became accentuated by uplift that eroded Oligocene extensional basins, on the east side of the ridge (e.g., Fig. 21). In the Early Miocene the strike-slip faults and linked half-grabens in the northern Mergui Basin collapsed the northern part of the Mergui Ridge. Their development was accompanied by a change in extension direction from west-northwest–east-southeast (Oligocene) to north-northwest–south-southeast (Early Miocene), a change that promoted transtensional deformation.

The Mergui Basin is at the southern termination of a broad belt (~400 km wide) of north-south–trending dextral strike-slip deformation that affected the areas that are now western Thailand and eastern Myanmar during the Late Oligocene–Early Miocene (Bertrand et al., 1999; Bertrand and Rangin, 2003; Morley, 2004). Although the strike-slip faults display reaction into the Pliocene, >90% of the fault activity occurred during the Early and middle-Middle Miocene. The end of transtension and/or oblique extension in the north Mergui Basin appears to have been caused by northwest migration of the zone of stretching, and development of the linked system of the Sagaing, West Andaman, and Sumatra faults west of the Mergui Basin during the later Miocene and Pliocene. This time was also accompanied by a drowning of the shelf, probably related to the creation of oceanic crust and onset of thermal subsidence following the cessation of extension and/or transtension.

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