Growth of folds in a deep-water setting

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ABSTRACT

Regional 3D seismic data of the deep-water area offshore NW Borneo provide a detailed picture of the interaction between sedimentary processes on a continental slope and the growth of major folds over a time period of ca. 3.5–5 Ma. In the deep-water area, the estimated rates of fold propagation of tens of mm/yr\(^{-1}\), shortening rates of mm/yr\(^{-1}\), and fold segment lengths of tens of kilometers indicate the studied folds are similar in scale and deformation rate to folds in orogenic belts such as the Zagros Mountains and Himalayas. Feedbacks between sediment dispersal patterns, antiline growth, and structural style are manifested in many ways, and are enhanced by the presence of weak, poorly lithified, synkinematic sediments at fold crests that undergo mass wasting as the fold grows. As folds tighten they range in geometry through simple folds—folds affected by crestal normal faults, folds with crestal normal faults and rotational slides, and folds with forelimb degradation complexes and pronounced erosional unconformities. The unconformity surfaces are either elongate parallel to the fold axes (related to local mass wasting) or perpendicular (related to flows crossing the anticlines). Mass wasting processes in the study area that are large in scale compared with folds (i.e., giant landslides) have modified antiline shape by erosion, and are little affected by antiline topography. More commonly, gravity flows are relatively small compared with the anticlines, and transport pathways are influenced by antiline surface topography. The factors influencing sediment pathway changes during antiline growth include: proximal to distal propagation of folds, lateral propagation of folds, and changing locations of fold growth with time. Assuming an overall consistency in sediment supply, local relative changes in sediment supply to individual piggyback basins are defined by relative changes in sediment supply (generally increasing with time) with respect to growth in antiline amplitude. Such changes may be due to sediment infilling of depressions farther updip and prograding downslope with time or to changes in rate of deformation with time. Initial sediment pathways were predominantly subparallel to growing faults. As folds matured, canyons and channels exploited low points (e.g., fold linkage or intersection points) or weak points (e.g., mud pipes) on folds to initiate transverse channel systems. At a late stage, transverse channels carved up the once long and linear surface ridges into isolated segments.

1. INTRODUCTION

Deep-water fold belts occur in three main settings: (1) accretionary prisms, (2) toe thrust and fold regions associated with gravitational deformation of oceanward-dipping, passive margin sediments underlain by mobile salt, and (3) passive or active margins where rapid sedimentation and development of overpressured mobile shales are associated with input from one or more major deltas. These regions are home to fold-thrust belts equivalent in size to the external regions of major orogenic belts such as the Jura Mountains of the Alps and the Zagros Mountains of Iran and Iraq (e.g., Morley and Guerin, 1996; Rowan, 1997; Trudgill et al., 1999; Ajakaiye and Bally, 2002; Hooper et al., 2002; Jackson et al., 2008). While deep-water folds have been known for a long time from 2D seismic-reflection data, the detailed geometry and evolution of deep-water folds have not been documented until recently. The advent of deep-water drilling technology has led to the commissioning of numerous 3D seismic surveys in the search for hydrocarbons. The 3D seismic data provide a wealth of information about the detailed structure of deep-water folds, and how fold growth has interacted with sedimentation and gravity (e.g., Hooper et al., 2002; McGilvery and Cook, 2003; Ingram et al., 2004; Corredor et al., 2005; Heinio and Davies, 2006; Morley, 2007a, 2007b). Deep-water fold belts form very large provinces, equal in complexity to the external zones of orogenic belts, but are much less completely described.

Models for the interaction of growing folds with synkinematic sedimentation have been determined from a range of studies including theoretical stratal geometries determined from trigonometry-based, fold-growth simulations, to the detailed interaction of sedimentation and fold growth, usually based on well-exposed outcrops or 2D seismic data (e.g., Mutti et al., 1988, 2000; Suppe et al., 1992; Shaw and Suppe, 1994; Hardy and Poblet, 1995; Hardy et al., 1996; Delcaillau et al., 1998; Evans and Elliott, 1999; Lopez-Blanco, 2002). Geometric approaches describing fault-propagation folds and folds with forelimb degradation have assumed a simple buildup of synkinematic sediments onto a growing fold high that can generate a range of patterns of thinned, rotated, and onlapping growth sequences (e.g., Suppe et al., 1992; Hardy and Poblet, 1995; Hardy et al., 1996). These models appear to work well for folds growing in continental and shallow marine environments. However, the deep-water environment has different erosional and sedimentary characteristics than folds in other settings. Only in deep-water settings can poorly lithified syntectonic sediments accumulate on top of growing anticlines that have a strong topographic expression at the seafloor. The strong topography causes interplay between structure and gravity flows moving down the continental slope. The patterns of synkinematic sediment deposition, sediment removal, and sediment redistribution significantly impact fold shape (e.g., McGilvery and Cook, 2003; Nigro and Renda, 2004). Using a 3D seismic data set

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gathered across young (Pliocene-Recent) folds offshore Brunei as an example, this paper investigates the ways in which folds develop through time in a deep-water setting and interact with sediment transport down a continental slope. It is believed that the 3D data set provides a level of detail about fold-sediment interaction that has not been so comprehensively demonstrated previously, although the fundamentals of the processes are well established.

2. PREVIOUS WORK

The NW margin of Borneo has been extensively explored for hydrocarbons, and as a result, the geology of the shelf and upper slope areas of Brunei, Sabah, and Sarawak are well known from drilling and seismic-reflection data (James, 1984; Sandal, 1996; Petronas, 1999). The shelf is dominated by prograding shallow marine sequences of middle Miocene–Recent age that attain thicknesses in excess of 10 km in places. The sequences are deformed by growth faults, shale diapirs, and inversion-related anticlines (Morley et al., 2003).

The NW Borneo fold-thrust belt is developed on the slope in the deep-water equivalents of the extensively drilled middle Miocene–Recent shelf sequences (Fig. 1). The slope dips between one and three degrees into the deepest region of the South China Sea—the Northwest Borneo trough. Until recently the fold-thrust belt was only known through sparse industrial and academic 2D seismic lines (James, 1984; Hinz and Schluter, 1985; Hinz et al., 1989; Sandal, 1996; Schluter et al., 1996; Ping and Hailing, 2004). The 2D seismic data show an extensive train of offshore-verging folds, spaced between 5 and 15 km apart, affects the slope. They appear to be mostly fault propagation folds related to imbricate faults at depth that sole out into one or more detachments. The thrust belt has a classic taper geometry, with the basal décollement dipping shelfward; at the shelf-slope break the detachment probably lies at a depth of ~10 km (Fig. 1; Morley, 2007a). The 2D seismic data demonstrate that the NW Borneo trough is the site of inactive subduction where the oceanic crust of the proto–South China Sea was subducted during the Paleogene but became jammed in the latest Early Miocene when thinned continental crust of the Dangerous Grounds block entered the subduction zone (Levell, 1987; Hall, 1996). The Miocene uplift and erosion history of NW Borneo has at least in part been attributed to oceanic slab breakoff during the middle–late Miocene and buoyant rise of the leading edge of the Dangerous Grounds continental crust (e.g., Hutchison et al., 2000; Morley and Back, 2008). Despite the termination of subduction, global positioning system (GPS) data indicate differential motion continues to occur between NW Borneo and Sundaland driven by collisional events to the south and east of Borneo (Simons et al., 2007).

The present-day fold-thrust belt is largely of latest Miocene–Holocene age (Sandal, 1996; Ingram et al. 2004). Hence the deformation is unrelated to the earlier period of active

![Figure 1. Regional map of NW Borneo showing location of the NW Borneo trough and study area, offshore Brunei Darussalam.](image-url)
subduction of oceanic crust. The active folds trend NE-SW, parallel to the old trench (Fig. 2), indicating the suture zone between the Dangerous Grounds block, and NW Borneo is a major zone of weakness that reactivated from the latest Miocene–Recent (Hall and Morley, 2004).

GPS data (Simons et al., 2007) show that northern Borneo is moving west at ~4–6 mm/yr with respect to Sundaland (i.e., South Borneo, Peninsula Malaysia, and Indochina). This motion is oblique to the NE-SW–striking folds of the continental slope and oblique to the modern maximum horizontal stress direction (NW-SE) determined for the shelf immediately adjacent to the GPS points (Tingay et al., 2005). While the stress tensor for the deep-water area may have a component related to gravity-driven deformation, another significant component

Figure 2. Edge map of the water-bottom reflection mapped from 3D seismic-reflection data, for the outer shelf and slope region of offshore Brunei Darussalam (the seafloor morphology shown here has also been illustrated by Demyttenacre et al. [2000] and McGilvery and Cook [2003]). See Figure 1 for location. The darker gray shading indicates greater slope dips on the external anticlines compared with the upper slope. Black arrows mark some examples only of the terminations of headless channels along the upper slope. I, II, III, IV, V, VI, VII, VIII, IX, and X denote anticlines. Letters A to M denote locations along important sediment pathways from the upper slope to the lower slope. Incised canyons through anticlines occur at locations A, E, F, I, and M. Inset shows a perspective view (looking SE) of the edge map.
of the stress tensor is regional stress related to Australia-Sundaland collision. The gravity-driven component of stress may help explain the difference between the modern stress orientation and GPS-defined block motions. Hence the deep-water folds probably arose due to the combined effects of gravity-driven deformation and from regional stresses (Morley et al., 2008).

Recently the ability to explore for hydrocarbons in deep water has opened up the margin to extensive 3D seismic surveys (Demytenacre et al., 2000; Ingram et al., 2004). Publications using the 3D data have described the petroleum system of the fold-thrust belt in Sabah and the modern sedimentary pathways from the shelf to deep-water offshore Brunei (Demytenacre et al., 2000; McGilvery and Cook, 2003; Ferguson et al., 2004; Gee et al., 2007; Morley, 2007a). The 3D data enabled detailed mapping of the seafloor, which provides highly detailed images of modern depositional systems (Fig. 3).

3. DATA

Ten thousand square kilometers of 3D seismic data were acquired by Petroleum Geo-Services (PGS) in 2000 and 2001 across the deep-water area of Brunei. The data were originally processed by PGS in 2001. This paper is part of a study at the University of Brunei Darussalam utilizing the deep-water 3D survey provided by an agreement between the Petroleum Unit, Total, and Brunei Shell. The data set used in this study was clipped at one second below the seafloor; the deeper seismic data were not available for study. Calibration of reflections from the 3D survey with the well and seismic data shown in Ingram et al. (2004) and Sandal (1996) from the shelf and adjacent deep-water areas indicates the upper one second of data (~0.9–1 km thickness of section) is of predominantly Pliocene-Holocene age. The upper one second of data provides a wealth of detailed information about how sedimentation and folding have interacted within the synkinematic section.

4. METHODOLOGY

Seismic interpretation was conducted using Landmark software. Five horizons were picked for the study: the seafloor (Fig. 4); a shallow horizon (horizon A) that coincides with an unconformity on the crests of several folds; the top (horizon B) and bottom (horizon C) of a distinctive low reflectivity package; and the deepest possible horizon (D) that could be mapped around most of the study area (Figs. 4 and 5). It is not possible to generalize too much about the characteristics of the different horizons because there are important lateral changes within reflection-package characteristics due to changes in sediment type, structural location, and timing of deformation. All the horizons were picked within reflection packages that in places thin onto anticlines and thicken into synclines, and hence are synkinematic with respect to folding.

For high-resolution interpretation across the area, horizons B, C, and D were picked every 25 lines and traces, and horizon A was picked every 50 lines and traces. For 3D seismic data, the location of vertical sections through the data that lie parallel to the acquisition direction are called lines or inline, those orthogonal to the acquisition direction are called traces, while sections selected in any other orientation through the data are called arbitrary lines. In some areas of high-density faulting, inlines and traces were picked every 2–10 lines. In areas where there was high reflection continuity and structural disruption was minimal, horizons were picked every 50 inlines and traces. Line spacing is every 12.5 m, hence the spacing of the 25 × 25 grid is 312.5 m. The main mapped area is ~50 km long in a strike direction (NE-SW) and up to 60 km long in a dip direction (NW-SE). The final infill of the picked grid to produce a continuous horizon was conducted with Landmark’s zoned auto-picker (ZAP). Automatic infill of the picked grid uses seismic waveform characteristics constrained within a user-defined interpretation window, and the picked seismic data as seed points, to produce a continuous horizon within 3D seismic data.

Derivative images of time-structure map dip-and-strike characteristics based on changes in curvature can be generated by a number of different algorithms (e.g., Brown, 1996). In this study, edge maps were found to give the best results for detecting subtle structural and stratigraphic signatures of comparatively minor features such as channels, scarps, craters, and small faults, as well as showing large-scale folds and fans (e.g., Fig. 2).

To determine the sedimentation patterns and fault patterns in map view from contrasts in seismic amplitude, two methods were used: (1) the amplitudes along the auto-picked horizon, and (2) the root mean square (RMS) of amplitude values for a 10 ms window of data sampled above or below a smoothed version of the picked horizon using Landmark StratAmp software. Contrasts in amplitude values in map view can show distinctive sedimentary forms such as channel or fan-shaped bodies of lower or higher amplitude relative to the adjacent areas (e.g., Fig. 3).

To understand how folds have interacted with deep-water sediments with time, the following aspects of the seismic interpretation were used most extensively (see section 8): (1) interpretation of vertical sections; (2) isochron maps of intervals between horizons C and D, B and C, A and B, and seafloor and A; (3) edge maps of the five mapped horizons; and (4) RMS amplitude maps for the five horizons.

The vertical cross sections show the detailed interaction between reflections and fold growth, as shown by thinning-thickening patterns, reflector termination patterns (e.g., onlap, erosional truncation, and downlap), sedimentation style (e.g., presence of degradation complexes, slumps, and landslides), and unconformities. To understand how folds and sedimentation patterns have evolved during fold growth in map view, the isochron maps are used in conjunction with the RMS amplitude, edge, and difference maps. The isochron maps show thickness changes for a particular interval, which tend to be related to fold growth, erosion, and deposition in fans and channels. The geometry of thickness changes cross-checked with reflection characteristics on vertical cross sections makes for a clear interpretation of the isochron maps. Typically, sedimentary patterns, such as sinuous channel, or fan-shaped features visible on the isochron maps, also show on the RMS amplitude maps. The RMS amplitude, difference, and edge maps are derived for a particular horizon, or narrow time window around the horizon, where the best resolution may be ~15–30 m. The isochron maps represent a thicker interval of investigation with the isochron interval ranging from zero thickness up to around 500 ms (~450–500 m). This means the isochron map is less detailed in resolving specific sedimentary features than the other maps. However, in general terms there is a good correspondence between the features imaged on the different map types.

5. MODERN SETTING

Figures 2, 3, and 4 show the regional context of the area in map view and in cross section. Figure 2 is an edge map of the water-bottom time-depth across the outer shelf, slope, and Borneo trench area. The outer shelf is relatively smooth, and is cut by a few active growth faults. The continental slope is strikingly different. It is dominated by an array of terraces (underlain by synclines) that step down the slope via a series of prominent, relatively steeply dipping, NE-SW-trending ridges related to the crests and forelimbs of growing anticlines. The synclines form sedimentary depocenters with a subhorizontal seafloor (Fig. 4). Anticlines tend to have a strong surface expression and locally steeper slopes on the fold limbs than the regional dip of the continental slope, but this expression changes significantly passing down the continental slope. In Figure 2 the steeper slopes are associated with the more distal 4–5 anticlines. Anticlines more
Figure 3. (A) Root mean square (RMS) amplitude map of the seafloor, illustrating the main structural and sedimentary features active today (also see McGilvery and Cook [2003] for a similar water-bottom map). See Figure 1 for location. (B) Interpretation of the main structural and sedimentary features seen on the amplitude map in Figure 3A. Black arrows mark some examples only of the terminations of headless channels along the upper slope. I, II, III, IV, V, VI, VII, VIII, IX, and X denote anticlines. Letters A to M denote locations along important sediment pathways from the upper slope to the lower slope. Incised canyons through anticlines occur at locations A, E, F, I, and M. Inset shows the location of the figure with respect to the edge map in Figure 2. Red—high amplitude, black—low amplitude.
Figure 4. Regional seismic lines showing the mapped horizons; two (A and B) are inlines, while (C) is an arbitrary line. The arbitrary line follows the modern cross-fold sediment transport path down the slope (see Fig. 3 for location). TWT—Two-way time.
proximal to the shelf are strongly dissected by NW-SE-trending linear to anastamosing channels and are lower relief features than the distal anticlines (Figs. 2 and 4). In Figure 2, the smooth area of the outer shelf is not cut by any channels; all the channels arise along the upper slope or at anticlines and do not connect with any channel system on the shelf. The southwestern part of the shelf in Figure 2 is cut by some active growth faults, and they are not disturbed by any channels. Hence it is clear the channels crossing the slope are headless.

A region lacking clear ridges and channels, with a speckled, blocky appearance, dominates the western half of Figure 2. As described in detail by Gee et al. (2007), this region is covered by a giant, young landslide, called the Brunei Slide. Figure 3 shows an amplitude map with an interpretation of the main structural and sedimentary features of the study area. Two fans (K and J; Fig. 3), fed by channel systems oriented transverse to the fold axes, are prominent features of piggyback basin development (McGivern and Cook, 2003). The amplitude map shows that the areas of higher-energy sedimentation (channels, landslides, and fans) tend to be regions of relatively high amplitude (yellow-red colors); the areas of more aggradational deposition (anticlinal crests and more isolated parts of synclinal basins) are lower-amplitude areas (shades of gray). The more landward folds are commonly affected by shallow mass movement processes and mud volcanoes (Fig. 3), while the western part of the area covers the eastern margin of the Brunei Slide.

### 6. VERTICAL SEISMIC PROFILES

In many figures, vertical seismic lines are presented; their vertical scale is given in two-way time (TWT), in units of seconds (s) or milliseconds (ms). An approximate depth conversion can be made assuming an average seismic velocity of 1900–2000 m/sec for the first second (TWT) of data below the seafloor, which means 100 ms are equal to ~95–100 m. The sections below describe changes in fold geometry passing upslope from the most external or distal to the more proximal folds.

#### 6.1. Simple Folds

In the most distal anticline, the internal folded reflections lie subparallel to the seafloor. The seafloor is folded and has a convex-up geometry. Much of the section maintains its thickness around the fold, and only the uppermost reflectors show thinning toward the fold crest. The fold is unaffected by normal faulting of the crest (Fig. 4B, anticline IX).
6.2. Simple Fold with Normal Faults

In many folds, including the most external fold in some places, planar to slightly curved faults, dipping ~40°–65° affect the crest (Figs. 6 and 7). These faults tend to penetrate deepest (500–700 ms, ~460–650 m) at the fault crest, and die out at shallower levels passing toward the fold limbs. In map view the faults can be strongly curved (Fig. 8). Faulting is most intense at the fold crest and dies out laterally (down plunge) as fold amplitude decreases. In cross section the planar nature of the faults and different depths at which they die out downward indicate that despite their curved map patterns, which suggest rotational detachment type faults, the normal faults actually do not link up with a detachment. Hence they cannot be viewed as classic rotational slides. The faults are most intense on the steeper anticline slopes, and fault orientation changes as the strike of the slope changes. These observations suggest the faults are gravity-driven features, related to slope instability created by growing anticlines (Morley, 2007b).

6.3. Simple Fold with Normal Faults and Mass Movement

Overall the convex-up fold geometry is still preserved, and the fold resembles simple folds with normal faults as described above, but additionally the fold forelimb is affected by rotational slides, which tend to detach parallel to the forelimb slope at depths of ~100 m or less (Figs. 7B, 8A [locations a, b, c, d], and 9C). These slides are linked with listric extensional faults updip and folds and thrusts at the downdip termination of the slide. Figure 8A shows an edge map of the seafloor where slump scars break up a forelimb that is also affected by planar normal faults. An amplitude map of horizon B (Fig. 8B) shows the deeper fault pattern within the anticlines. The result of the interaction between slumps and normal faults produces an angular discordance between the seafloor reflection and the internal reflections of the fold, here called forelimb unconformities (Fig. 7). Multiple unconformity surfaces are frequently present, indicating repetition of the faulting and slumping process during fold growth (Fig. 9). Landslides produce depressions, benches, and bulges in the slope that modify the overall convex-up fold geometry (Fig. 8A, locations a and c).

6.4. Mature Folds with Angular Unconformities and Degradation Complexes

In the folds described above it is generally possible to trace horizons around from the backlimb to the forelimb, particularly deeper in the fold, except where slumps have locally removed material or have disorganized layering. In mature folds it is not possible to correlate horizons directly across the fold on dip lines from the forelimb to the backlimb. The thickness of the backlimb section is much greater than the forelimb due to the fold asymmetry and its effects on where sediment accumulates (Figs. 10 and 11). The more gently dipping backlimb areas, while displaying thinning onto the fold crest, are generally areas of sediment accumulation, little disturbed by post-depositional gravity processes. Conversely, mass wasting or erosion by currents or flows to produce degradation complexes episodically affects the fold crests and more steeply dipping anticline forelimbs (Morley, 2007b).

In the relatively steeply dipping forelimbs of some folds the continuity of reflections is completely lost. These areas are characterized by chaotic or seismically transparent zones with disorganized wedge and lens-shaped packets of high-amplitude reflections (labeled degradation complexes in Fig. 9 and at location a in Fig. 10). Some weak internal reflections tend to show lower dips than the overall dip of the forelimb (Fig. 10). In places the wedge-shaped packets (Fig. 10, location a) terminate against reflective units within the forelimb syncline. These very distinctive forelimb packages are interpreted to represent degradation complexes, where material removed by mass movement from the anticline crest and upper forelimb is redeposited at the foot of the forelimb (Heinio and Davis, 2006; Morley, 2007b).

On dip lines, the presence of a crestal unconformity can be difficult to define due to complications with gas effects causing strong amplitudes below a bottom simulating reflection (BSR) associated with gas hydrates. However, on strike lines, one or more angular unconformities tend to stand out clearly (Fig. 11). The deepest unconformity tends to cut through units tilted and rotated by normal faults, and dip in the same direction as the forelimb. Some folds are at this stage today (Figs. 7B and 7C). Usually sediments are deposited above the unconformity slope. In the simplest case, these units are little disturbed by mass movement, and their base may be a relatively uncremented surface that is also affected by planar normal faults. The presence of an angular unconformity in a mature fold can be used to estimate the 2007b).

Figure 6. Seismic line across the most external anticline (IX), which shows the seafloor is folded quasi-conformably with deeper reflections. HA, HB, and HC—seismic horizons A, B, and C, respectively. See Figure 3 for location. TWT—Two-way time.
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6.4. Mature Folds with Mud Pipes and Gas Chimneys

Folds at the mature stage of development display angular unconformities and degradation complexes; they are also commonly associated with mud pipes and gas chimneys. Pipes and chimneys are readily identifiable in cross section as narrow, vertical zones of transparent to chaotic zones of seismic character. Mud pipes typically terminate at a local depression in the seafloor. In map view the depressions are circular to oval features a few hundred meters across. For example, the seafloor amplitude map (Fig. 3A) shows numerous circular features, interpreted to be mud volcanoes that are aligned along the topographic ridges (Fig. 3B).

In general the external folds are little affected by any large pipes, and only one small gas chimney affects external fold IX. The large mud pipes are the only type of fluid escape feature that, based on geochemical analysis of water-bottom grab samples, is associated with thermogenic hydrocarbons (Warren et al., 2009). This suggests a deep origin (probably >3 km) for the associated mud and fluids (Fig. 10D). Smaller pipe-like features appear to be gas-rich fluid escape features. The seismic reflections of multiple horizons are locally disturbed (pulled-up or pushed-down) in a narrow vertical band (e.g., Gay et al., 2003). These changes are attributed to a lateral change in velocity (velocity pull-up or push-down) and are known as velocity-amplitude anomaly structures (VAMPS). In the case of dewatering features, the pull-downs may in part be a geological feature, not a velocity feature, which arose due to volume loss from the fluid source layer. Figure 10E shows two examples of vertical pipe-shaped anomalies. The velocity pull-up anomaly suggests the presence of gas hydrates; the pull-up effect dies out downward. The chimney with pull-down geometry (Fig. 10E) indicates the presence of free gas in pore spaces and/or mechanical collapse of layers due to dewatering. The vertical pipe anomalies do not extend deeper than 600–700 ms (~550–650 m); hence they are shallow features and appear to rise off a seismically transparent (white) layer that, from drilling upslope, is known to be very shale rich. Clusters of these pipes are found only on the backlimb side of anticlines (Fig. 10D). The pipes are interpreted to represent lateral migration of fluids out of the synclinal depocenter, updip to a level where fluid pressure is large.

Figure 7. Seismic lines a, b, and c across anticline VIII showing rotational slide affecting forelimb of anticline, also affected by deeper penetrating and more planar normal faults. The sections are dip lines that are higher toward the fold crest passing from a to c; HA, HB, and HC—seismic horizons A, B, and C, respectively. See Figure 3 for location. TWT—Two-way time.
enough to exceed the minimum principal horizontal stress, and permit hydraulic fracturing up to the surface.

Fold height and maturity appear to control mud-pipe and gas-chimney spatial density, suggesting a strong link between shortening, pore space reduction, and overpressuring. While shortening is clearly manifest at a large scale as folding and thrusting, what is less well known is the degree to which bulk shortening of the young, and porous, deep-water section is achieved by tectonic compaction. It has been observed in fold-thrust belts that bulk shortening by pressure-solution cleavage precedes folding (e.g., Morley, 1986a, 1986b; Whitaker and Bartholomew, 1999; Tavani et al., 2006), and bulk shortening by loss of pore space in the shallower, less compacted deep-water sediments is likely to occur as well. This would result in a component of the overpressure being related to shortening, not just disequilibrium compaction associated with rapid burial of fine-grained sediments (e.g., Osborne and Swarbrick, 1997). Uplift of overpressured units in the cores of anticlines raises them higher relative to the same units on the anticline flanks and synclines and thus brings them closer to a stress state where natural hydraulic fracturing will occur. The effects of mass movement and erosion removing some of the overburden at the crest will further enhance the propensity of anticlinal fold cores to preferentially spawn mud pipes. In some folds, mud pipes are associated with forelimb degradation complexes (Fig. 10), and hence in addition to erosion and slumping of material, the degradation complexes also contain mudflows resulting from eruptions of overpressured mud from the pipes.

Figure 8. (A) Edge map of the seafloor reflection showing normal faults and slumped area associated with the external anticlines (IX and VIII); see Figure 3 for location. Location a—upper ridge associated with extensional rotation in the upper part of a major rotational slide; location b—most intense region of mass movement cutting back high into the fold crest at the apex of the fold; location c—ridge associated with toe thrust and fold region of slide; d—anticlinal ridge at tip of imbricate thrust fault (deep penetrating). M—incised canyon on anticline crest. (B) Root mean square (RMS) amplitude map of horizon B. The rotational slide region of location a disturbs the regular pattern of deep-penetrating normal faults seen beyond location a.
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Figure 9. Examples of changes in unconformity geometry passing along the strike of anticline V. (A) Anticline morphology in saddle area, where flows can traverse an anticline, passing toward the anticline; crest sections (B) and (C) show increasing erosion, the effects of normal faults, and mass wasting. HA, HB, HC, and HD—mapped seismic horizons A to D, respectively. See Figure 3 for location. BSR—bottom simulating reflection; TWT—Two-way time.

Decreasing fold interlimb angle

Thinning of upper non-kinematic layers in forelimb by slumping during early stages of fold growth?

No pronounced thinning onto fold crest

V-shaped lateral terminations to disorganized reflection packages wedge-shaped packages = mass movement complexes on forelimb

Minor erosional bench

Transport path unconformity merging with forelimb unconformity

Degradation complex

Successive forelimb unconformities

Degradation complex

TWT (s.)

1 km

A

B

C

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Figure 10. Example of two anticlines (VIII and V), the geometry of which at the seafloor is strongly modified by mass-wasting and mass-movement processes. The northwestern fold (VIII) is a simple fold with normal faults; the southeastern fold (V) is affected by erosion, forelimb degradation complexes (labeled “a”), and a mud pipe. HA, HB, and HC—mapped seismic horizons A to C, respectively. BSR—bottom simulating reflection interpreted to indicate the presence of gas hydrates. (A) Sketch illustrating the geological features of the seismic line; (B) seismic line at approximately vertical = horizontal scale; (C) seismic line at more typical display scale of 2× vertical exaggeration (better for seeing details of seismic character); (D) geological cross section based on (A) above and deeper 2D seismic-reflection data illustrating likely overpressured fluid-distribution and fluid-migration pathways; (E) examples of shallow fluid and gas pipes developed on the backlimb of the anticline, from lines along strike from the section shown in (B) and (C) above. See Figure 3 for location. TWT—Two-way time.
7. TIME-STRUCTURE MAPS

The regional horizons A, B, C, and D were picked within reflection packages that in places thin onto anticlines and thicken into synclines, and hence are synkinematic with respect to folding (Fig. 4). The deepest horizons (D and C) generally lie within lower-amplitude, more continuous reflection packages, and represent the early stages of fold development. In general, reflection continuity is good, and reflectors tend to be sub-parallel; however, in places, some highly discontinuous to chaotic units are present.

Much of the synclinal basin fill is characterized by a relatively high-amplitude package of reflections with an irregular internal geometry that overlies horizon C; they also thin and onlap onto anticlines. Horizon B is located at the base of the high-amplitude reflection package in the central slope area, and horizon A is a shallow horizon within the higher amplitude basin fill. The synclinal basin fill contains a wide range of reflection-pattern types, ranging from parallel, continuous reflections, to discontinuous mounded or wavy forms. Erosional truncations are frequent. Extensive tabular to highly discontinuous seismically transparent, internally chaotic to partially organized reflections frequently occur in the piggyback basins. These transparent units represent landslides, debris flows, turbidites, and other products of mass movement and mass wasting (e.g., McGilvery and Cook, 2003; Gee et al., 2007).

Time-structure maps of horizons A, B, C, and D are shown in Figure 5. The deeper horizons (c and d in Fig. 5) define well-developed, simple fold shapes. The upper slope anticlines (I, II, and III) are relatively low amplitude compared with anticlines IV, V, VI, and VIII, which have deep synclinal depocenters. The fold hinges are curvilinear with prominent saddles and crests developed along their length. The anticline trend VI and VII has three distinct crests and saddles that probably reflect the connection of three isolated folds. Horizons A and B (Figs. 5A and 5B) define very subdued fold shapes compared with horizons D and C. Synclines form flat-bottomed lows, and the anticline ridges are mound features dissected by NE-SW–trending low linear features. It is apparent from vertical seismic sections that horizons A–D infilled fold-created bathymetry and that the fold geometry was smoothed out during deposition of units represented by horizons A and B. The time-structure maps (Figs. 5A and 5B) also illustrate the erosional degradation of the anticline crests and folds. However, the details of this erosional degradation are much better imaged on isochron, amplitude, and edge maps, and are discussed below.

8. LARGE-SCALE SEDIMENTATION PATTERNS ASSOCIATED WITH FOLD GROWTH

The large-scale patterns of fold evolution were discerned from the isochron maps derived from time-structure maps, coupled with amplitude extractions of the mapped horizons. For this study there were no wells within the area of 3D seismic data with which to calibrate rock data with amplitudes; however, Ingram et al. (2004) provide some indication about lithologies and stratigraphy based on well and seismic data from adjacent areas of Sabah. The amplitude map of the seafloor (Fig. 3) shows that high amplitudes are associated with clear channel, fan, and lobe complexes (also see McGilvery and Cook, 2003). These higher-amplitude areas...
may be associated with increased sand content within a predominantly shale-prone section. Similarly on amplitude maps of deeper horizons, high amplitudes tend to be associated with the fill of synclines, and also commonly define channel and fan geometries.

Isochron maps of the early synkinematic stage (between horizons C–D and B–C) show that the anticlines tend to be long (25–30 km), continuous features that are infrequently breached by channels and canyons (Fig. 12). Most sediment transport was axial, along the synclinal piggyback basins, as indicated by higher amplitude areas in the synclines compared with the anticlines (Figs. 13, 14, and 15). Areas H and B (Figs. 13 and 14) are the main regions of early transport across folds. Area H occurs at the down-plunge termination of fold V against fold IV (Figs. 13 and 14), while area B is a persistent area of SE- to NW-directed transport on all amplitude maps except horizon D (Fig. 15). The persistence of area H to cross-fold transport is due to the relatively low wavelength and amplitude of the folds (Fig. 4), whose growth appears to have kept pace with sedimentation; hence the folds created little surface topography. Passing to the SW the same folds (II and III) increase in amplitude and interrupt NW-SE transport during horizon C and D times (Figs. 13 and 14). For horizon C, at location C (Fig. 14), the high-amplitude reflections curve around the plunging nose of fold VI, indicating the fold was a significant feature at the seafloor. The situation for horizon D is different—high-amplitude reflections cover the region of the plunging nose, and SW of location D (Fig. 13) a channel-like feature trends perpendicular to the anticline axis, suggesting there was no fold topography during horizon D times. The isochron maps for horizons D and C also support post–horizon D propagation of anticline IV into the area SW of location C (Figs. 12C and 12D), and establishment of an anticlinal ridge during the interval between horizons C and B.

The amplitude map of horizon C (Fig. 14) shows that a straight, NW-trending, narrow, linear feature runs across anticline VIII west of area M, unaffected by the fold, indicating the fold was not present at that time. The piggyback basin between anticlines VIII and VI contains small, patchy, high-amplitude areas (Fig. 14 locations O and N) that may be poorly developed, or partially eroded fans fed from the east (area Q) through channels in the vicinity of J. Conversely, area P (Fig. 14) appears to be supplied from the west. In the cases of fans O, N, and P, the fans were fed by flows that followed synclinal axes, and the sediment source areas lay outside of the study area (Fig. 15C). The thickness difference between the syncline and the anticlines VII and VIII on the isochron map (Fig. 12C) suggests that the maximum height of the anticlines above the synclinal depocenter was ~350 m.

The isochron map for horizon B–A (Fig. 12B) shows a significant change in geometry from the deeper isopachs. Instead of the dominant NE-SW trends, the contours display a more rectilinear pattern; the anticlinal ridges are no longer so continuous. This pattern reflects the breaching of anticlines by NW-trending channels and canyons (Fig. 15A). The amplitude map of horizon B (Fig. 16) shows some differences with the deeper horizons. The channel systems from area B fed sediment into the piggyback basins between folds II, III, and IV to the west. Small transverse canyons are also highlighted crossing folds III and VI. Area E to K marks an area of high amplitudes in a low area between folds VII and VI; it does not show the organization and well-developed transverse feeder systems and fans that characterize the subsequent infill of the piggyback basin (Figs. 3, 15A, and 17).

The amplitude map of horizon A clearly shows that between horizons B and A well-developed transverse canyons and channels started to dissect folds I to VI (Figs. 15A and 17). Elongate, sinuous, high-amplitude packages both cut across anticlines and run parallel to synclines. The amplitude geometries are interpreted to represent channel and fan patterns very similar to those seen on the seafloor amplitude map (Fig. 3). The horizon A–seafloor isochron map also shows further development toward a strongly rectilinear contour pattern where transverse (i.e., NW-SE trends) features are almost as strong as the fold trends (Fig. 12A).

While many areas have evolved considerably with time, area B on amplitude maps of horizons C, B, A, and the seafloor (Figs. 3, 14, 15, 16, and 17) has remained an area of persistent SE- to NW-directed sediment transport. Only for horizon D (Fig. 15) is the pattern not well developed. However, the isopach maps show that for the horizon D to C interval the area was a low, and hence a likely trap for sediments. Subsequent lateral propagation of anticlines I and II into area B eliminated the piggyback basin, but anticlinal uplift was not strong enough to be a barrier to sedimentation, and deposition largely kept pace with, or exceeded the anticline growth rate. Thus the SE to NW sediment transport pathway was kept open from horizon C to the present.

The piggyback basin between anticlines VI and IV changes geometry with time. For horizon C the basin is deepest adjacent to anticline IV (Fig. 5); this is also shown in the isochron map (Fig. 12C, location Q). The deep syncline is bounded by two anticlines (V and IV), and at the oblique intersection of these two anticlines (area H) there was a high in the synclinal trough (Fig. 12C). The isochron map for horizons C–B shows that the syncline adjacent to anticline V continued to deepen at a relatively high rate, while adjacent to anticline IV subsidence in the syncline slowed. This resulted in an along-strike ramp dip of the synclinal floor toward the anticline V synclinal area (Fig. 5B). During its early development the piggyback basin showed patchy development of high-amplitude areas (Fig. 13). There is a possible feeder system into the piggyback basin through area H, and some poorly developed potential channels transported sediment out of the basin around area J (perhaps reflecting an episode when the basin around area Q was completely filled; see Fig. 13). Horizon B shows the highest-amplitude concentration occurs in the area between K and E (Fig. 16). This is the main isopach thick of the basin (Fig. 12B). Late in the basin fill during deposition of horizon A, extensive deposits of gravity-driven sediments affected the area, notably fans J and K (Fig. 17). The WSW-plunging piggyback basin between anticlines VI and IV displays basin oblique transport paths that link with a high-amplitude channel system that runs parallel to anticline VI. The linked system of channels then cuts across the anticline at location I (Figs. 15A and 17). This feeder system is not seen today (Fig. 3).

The isopach and amplitude maps described above show that as the folds evolve and grow the initial ridge and synclinal piggyback basin morphology becomes modified, and anticlinal ridges become breached. In assessing the impact of folds on sedimentation patterns, it is important to consider both the lateral propagation and evolution of folds, as well as the internal to external (seaward) propagation. For a mature fold, the changes in fold geometry through time show similar patterns to the external to internal changes in fold geometry discussed for the modern setting. The next sections examine how anticline geometry changed through time, and how the anticline breaching channels and canyons developed in response to fold geometry.

9. ORIGIN OF VARIATIONS IN FOLD GEOMETRY

Fold growth in the most external anticlines rotates the seafloor parallel to the dip of shallow reflections (Figs. 4 and 7). As folds become more tightly folded and higher amplitude, their geometry changes, and the seafloor slope angle can diverge from that of the shallow reflection (Figs. 4, 18, and 19). To try and quantify these changes, a number of basic fold measurements were compared with slope angle (half wavelength from the anticline crest to the footwall.
Figure 12. Isochron map for the intervals between: (A) horizons A–seafloor, (B) B–A, (C) C–B, and (D) C–D. Black areas are regions where the horizons could not be mapped due to either unconformities, the horizon being deeper than one second from the seafloor, or unmappable within degradation complexes or mud-pipe and mud-volcano provinces. I, II, III, IV, V, VI, VII, VIII, IX, and X denote anticlines. Letters A to M denote locations along important sediment pathways from the upper slope to the lower slope. Incised canyons through anticlines occur at locations A, E, F, I, and M.
Figure 13. Amplitude map for horizon D. Black areas are regions where the horizons could not be mapped due to either unconformities, the horizon being deeper than one second from the seafloor, or unmappable within degradation complexes or mud-pipe and mud-volcano provinces. I, II, III, IV, V, VI, VII, VIII, IX, and X denote anticlines. Letters B to M denote locations along important sediment pathways from the upper slope to the lower slope. Incised canyons through anticlines occur at locations A, E, F, I, and M. Inset shows location of map with respect to edge map in Figure 2. Scale is relative amplitude in arbitrary units. High-amplitude areas—red; low amplitudes—black.
Figure 14. Amplitude map for horizon C. Black areas are regions where the horizons could not be mapped due to either unconformities, the horizon being deeper than one second from the seafloor, or unmappable within degradation complexes or mud-pipe and mud-volcano provinces. I, II, III, IV, V, VI, VII, VIII, IX, and X denote anticlines. Letters A to M denote locations along important sediment pathways from the upper slope to the lower slope. Incised canyons through anticlines occur at locations A, E, F, I, and M. Fan geometries are labeled N, O, and P. A high-amplitude area associated with axial transport along a syncline is labeled Q. Inset 1—Detailed amplitude map of fan P; inset 2—detailed amplitude map of fan O showing radial shear grooves and channels. Inset 3 shows location of map with respect to edge map in Figure 2. Scale is relative amplitude in arbitrary units. High-amplitude areas—red; low amplitudes—black.
Figure 15. Summary maps of amplitudes and fold evolution for the following horizons: (A) horizon A, (B) horizon B, (C) horizon C, and (D) horizon D, with interpretation of sediment transport paths, to show how the geometry and location of active structures and sedimentation patterns has changed with time. I, II, III, IV, V, VI, VII, VIII, IX, and X denote anticlines. Letters A to M denote locations along important sediment pathways from the upper slope to the lower slope. Incised canyons through anticlines occur at locations A, E, F, I, and M. High-amplitude areas—red; low amplitudes—black.
Figure 16. Amplitude map for horizon B. Black areas are regions where the horizons could not be mapped due to either unconformities, the horizon being deeper than one second from the seafloor, or unmappable within degradation complexes or mud-pipe and mud-volcano provinces. Inset shows location of map with respect to edge map in Figure 2. I, II, III, IV, V, VI, VII, VIII, IX, and X denote anticlines. Letters B to M denote locations along important sediment pathways from the upper slope to the lower slope. Incised canyons through anticlines occur at locations A, E, F, I, and M. Scale is relative amplitude in arbitrary units. High-amplitude areas—red; low amplitudes—black.
Figure 17. Amplitude map for horizon A. Black areas are regions where the horizons could not be mapped due to either unconformities, the horizon being deeper than one second from the seafloor, or unmappable within degradation complexes or mud-pipe and mud-volcano provinces. I, II, III, IV, V, VI, VII, VIII, IX, and X denote anticlines. Letters A to M denote locations along important sediment pathways from the upper slope to the lower slope. Incised canyons through anticlines occur at locations A, E, F, I, and M. U1 and U2 denote large transport-path unconformities. Inset shows location of map with respect to edge map in Figure 2. Scale is relative amplitude in arbitrary units. High-amplitude areas—red; low amplitudes—black.
syncline, amplitude, interlimb angle, Figs. 18 and 19). Of these different measures the clearest association of changes in anticline slope angle was with the interlimb angle (Fig. 19).

Figure 19 shows a graph of fold interlimb angle plotted against average surface slope dip above the anticline, with points coded to represent the different fold styles described above. Slope dip was calculated from depth-converted seafloor profiles. Depth conversion used 1500 m/s as the velocity of seismic waves traveling through water to convert two-way traveltime to depth. The graph shows that in general the transition to more intense faulting and mass movement processes is accompanied by an increase in interlimb angle and in surface slope dip. For the category of simple folds, the interlimb angle values are quite widely scattered between 180° and 150°. This is because included in the category are the small upper slope folds in the eastern corner of the area (Fig. 4A) that have relatively small interlimb angles (165°–150°) but appear to have been almost continuously buried by sediment. Also in the category are the youngest folds in the province, with high interlimb angles (180°–170°). Hence crest uplift rate relative to sedimentation rate, as well as interlimb angle, is an important (but difficult to quantify) factor. The deep normal faults tend to develop in folds with interlimb angles between 170° and 140°, and surface slopes of 4°–12°. Large rotational slump scars plus normal faulting tend to occur where interlimb angles are smaller than 165° and slope angles are greater.

Figure 19. Fold interlimb angle versus slope dip of seafloor above anticlines; symbols indicate the type of associated fold geometry. Note: variation within the simple fold group is affected by fold amplitude and wavelength, not just interlimb angle, because where sedimentation has kept pace with the growth of relatively short-wavelength, low-amplitude folds, the fold crests of relatively tight folds can be buried and consequently are not exposed to mass-movement and mass-wasting processes.
than 6°. For folds affected by degradation complexes, surface slope dip does not increase beyond ~12°–13°, while the interlimb angle increases. Once a critical slope angle is reached, mass movement processes remove the crest of the anticline as it continues to tighten.

The data discussed above show that gravity exerts a strong modifying influence on fold geometry in deep-water environments. Slope dip is very important both as a pathway for sediment transport and dispersal and for the initiation of sediment movement due to slope instability (also see McGilvery and Cook, 2003). The shallow fold geometry strongly reflects the influence of gravity on fold shape and internal structure. The typical changes in fold geometry described above vary according to either: (1) the maturity of the fold (i.e., passing from the youngest, deepest water folds to the older folds located toward the continental shelf), or (2) lateral changes in fold style along an anticlinal ridge. In both cases it is the vertical rate of fold growth (i.e., the effects of fold geometry and rate of shortening), coupled with sedimentation rate, that dictates whether a particular segment of an anticline will act as a barrier to sediment, or be breached. Sedimentation rate is important because if sedimentation can keep up with fold growth, the fold will never develop a topographic ridge, but will be continuously covered with sediment. Hence in Figure 4A both the most external fold (IX), and the small, buried internal folds (I, II, and III) have a simple fold shape. A planar slope profile is associated with the arbitrary line that follows a well-developed sediment transport pathway down the slope (Fig. 4C), showing that locally a combination of sedimentation infilling lows, and erosion smoothing of highs can eliminate structural topography.

10. LOCAL UNCONFORMITY DEVELOPMENT AROUND FOLDS

As folds grow, the initial pattern of parallel reflectors, or reflectors that thin toward the fold crest, become modified. The development of local erosion surfaces or unconformities is widespread, and their location and geometry varies greatly around folds. The early stages of unconformity development occur where slumps and mass movement affect the forelimb (Figs. 7, 8, and 9; forelimb unconformities) and give rise to truncation of tilted fault blocks in the forelimb of the anticline. Later deposition parallel to or onlapping the slope may then be incorporated into the fault blocks and preserved. Alternatively, renewed mass movement may repeatedly thin the forelimb and erode deeper into the forelimb, crest, and even the upper backlimb of the anticline (Fig. 9C). At the foot of the forelimb,

Figure 20. Example of episodic fold growth (fold IV) based on stratatal geometries. (A) Uninterpreted seismic line. (B) Interpreted seismic line showing evidence for episodic growth of the anticline. (C–E) Earlier stages of fold growth illustrated in line drawings. (C) Final stage of fold growth prior to infill of the topography; (D) early fold topography smoothed out, period of deposition and no folding; (E) early stage folding. See Figure 3 for location. TWT—Two-way time.
sediments in the adjacent piggyback basin thin onto or onlap the forelimb. Forelimbs may display a steeply dipping unconformity surface in the forelimb that is onlapped by sediments in the adjacent syncline. The unconformity in places bounds a degradation complex and passes into a high- or low-angle unconformity at the fold crest (Fig. 9). As the forelimb grows, the synkinematic strata evolve from progressive thinning onto the crest to nondeposition on the forelimb (Figs. 7 and 9). The nondeposition stage results in the late synkinematic section resembling a large extensional tilted fault block, where the rotated fold backlimb is equivalent to the footwall block, and the forelimb is similar to the degraded normal fault scarps described from the North Sea (e.g., Stewart and Reeds, 2003; Fig. 20). On more steeply rotated back limbs, sediments may also onlap a local (backlimb) unconformity.

Passing along strike from the crest of an anticline to a lower saddle area, steeply dipping forelimb unconformities tend to pass laterally into conformable section. Then in the lowest part of the saddle, a different geometry unconformity is developed that is elongate parallel to the sediment transport direction. The transport path unconformity initially forms at a low angle (1°–3°; Figs. 4C and 21), and is elongate subparallel to the surface slope (Figs. 17 and 22). The difference in geometry between the two unconformities reflects the changes in processes that created them. Forelimb and backlimb unconformities result from local mass movements that degrade the crest and steep forelimbs and/or back limbs. Such unconformities are elongate parallel to the anticline. In the saddle areas fold growth is slower, or has ceased, and the region has become a pathway for sediment movement transverse to the anticlinal ridges. Sediment transport causes erosion of the saddle area; hence the transport path unconformities display dips similar to the present-day slope. In some folds the transport path unconformities cut into folds with older forelimb and backlimb unconformities resulting in the higher angle unconformities meeting the transport path unconformities (Figs. 9A and 23D). As seen in accretionary prisms and salt mini-basins, knickpoint systems can arise along deep-water channels that incise because the updip basins have filled to spill point (see review in Mitchell, 2006). Figure 4C shows an arbitrary seismic line following a major channel system down the NW Borneo slope; the gradient is remarkably constant, steepening only in one place where a knickpoint affects anticline VIII. Hence while transport path unconformities may partially have arisen as knickpoints, as shown in Figures 4C and 21, they are more generally tracts of constant gradient that maintain their position by erosion keeping pace with anticline uplift.

Fold growth is not necessarily smooth and progressive; there is commonly evidence for several cycles of growth and unconformity development. Particularly the forelimb and crests can show outward and upward propagation of unconformities (Figs. 4 and 21). Where low-angle unconformities have developed, the overlying sediment is gently inclined above a fold at depth. In cases where the fold starts to grow again, folding of the sediment above the unconformity produces a broader wavelength fold with much gentler limb dips than the core of the fold (Fig. 4C, fold II; Figs. 20 and 21).

West of the study area is a different type of erosion surface and unconformity related to mass movement on a much larger scale than the processes discussed in this paper. The giant Brunei slide is related to sediment input from the Baram delta and forms the uppermost depositional unit in the western part of the area (Fig. 2); it has been described in Gee et al. (2007), and so is just briefly mentioned here. This landslide is over 150 km long, 50 km wide, and tens of meters thick. It represents a catastrophic slope failure event that resulted in a widespread erosion surface at the base of the landslide that truncates the crests of several anticlines. Hence it is important to note that although anticlines generally control mass wasting location and distribution, large mass wasting events can eliminate the anticline topography and override the structural control.

Figure 21. Example of low-angle transport path unconformity affecting the low-relief saddle area of a fold. (A) The main transport area with very flat-lying unconformity. (B) Moving east of the main transport area, the top of the fold begins to develop a little topography. For more complete topographic development still on the margins of the transport path, see Figure 20. See Figure 3 for location. TWT—Two-way time.
11. CONTROLS ON LOCATION OF TRANSVERSE CHANNELS AND CANYONS

In the data set there are a number of important pathways for sediment transport transverse to anticlinal ridges (Figs. 15, locations B, C, D, E, H, I, J, and M). The structural causes for the pathways include (Fig. 22): (1) fold saddles, where two en-echelon or inline anticlines have propagated toward each other and linked to form a region of relatively low relief (Fig. 22B). Continued fold growth failed to eliminate the relatively low relief area of the linkage zone. This is a structurally low region, not just a region where erosional incision has kept pace with anticline growth (Figs. 3 and 15, locations E and M). (2) Overlapping folds whose axes are oblique result in one fold meeting partway along the length of the other fold (Fig. 22A). For folds IV and V, this geometry sets up an early pathway from one piggyback basin to another (Figs. 13, 14, location H). (3) Mud pipes may weaken and facilitate erosion of a fold crest so that late-stage transverse channels develop pathways across the former fold crest (Fig. 22C). For example, a large canyon system crosses the crest of the anticline V at location F (Figs. 3 and 17). (4) Regions where fold growth is slow relative to sediment supply can also become pathways. Such areas may be due to the initial stages of fold growth or to shortening being distributed across an unusually large number of folds, resulting in below-average uplift on the individual folds involved (Figs. 2 and 15, location B).

Figure 23 shows two seismic lines where a transverse channel cuts a fold crest with mud pipes (3 above). The gently inclined section above horizon B (Figs. 23B and 23D) lies unconformably on the eroded anticline V, and was deposited after the transverse channel cut across the anticline. This important sediment pathway overlies a tightly folded anticline. In contrast, to the east, the less tightly folded part of anticline V (with no transport path unconformity; Figs. 23A and 23C) is eroded and onlapped by the post-folding sediment, and lies at the edge of the pathway.

Slightly emergent fold crests, and low-relief portions of plunging folds have been traversed by gravity flows. Evidence for this is a series of striations (shear grooves) caused by shear at the base of one or more subaqueous landslides, possibly enlarged by turbidite flows formed along a large proportion of the fold (Figs. 3 and 24).

12. ESTIMATES OF RATES OF DEFORMATION

It is difficult to obtain sedimentation and uplift rates in the absence of detailed dating of the section. Ages can, however, be estimated by carrying seismic horizons dated by wells from the shelf and upper slope of Brunei in Sandal (1996) into the deep-water area, and from dates provided in Ingram et al. (2004) for an adjacent deep-water area in Sabah. These data show the upper one second of seismic data is largely of upper Pliocene to Recent age, with lower Pliocene to possibly late Miocene age strata present in the cores of the most uplifted anticlines. The character of the reflection packages indicates horizon D overlies the Lingan Fan II of Ingram et al. (2004), suggesting an age of ca. 3.7–3.5 Ma for horizon D. The thicknesses of piggyback basin fill between horizon D and the seafloor typically range from 550 to 900 m, yielding average sedimentation rates of ~0.16–0.25 mm yr⁻¹. For a longer time interval (9 m.y.), the age data provided by Ingram et al. (2004) yield similar deposition rates of 0.28 mm yr⁻¹ in a syncline and 0.13 mm yr⁻¹ on an anticline crest.

For the most external anticline (IX, Fig. 6) the seafloor at the fold crest lies 350 ms above the seafloor in the adjacent piggyback basin (262 m), while the fold amplitude for horizon C is 500 ms (450 m). The depth conversion from two-way time assumes a velocity of 1500 m/s for water and 1800 m/s for the poorly compacted uppermost part of the sedimentary section. The 188 m difference in height from anticline crest to syncline trough between fold amplitude and the seafloor depth reflect both thinning of section onto the anticline, and thinning by normal faulting. For anticline IX, fold growth appears to have begun just below horizon B; strata thickness to the seafloor in the piggyback basin is ~450 m. Assuming average sedimentation rates between 0.16 and 0.28 mm yr⁻¹ for the piggyback basin fill, then the age of onset of folding lies between 2.8 Ma and 1.6 Ma. The absolute ages are based

Figure 22. Schematic diagrams of maps illustrating the ways that different fold geometries can influence where gravity flows can traverse growing folds at the seafloor. (A) Transport path between en-echelon folds. (B) Transport path across saddle region in fold. (C) Transport path across fold crest.
Figure 23. Example of sediment transport pathway across the crest of a fold, affected by mud pipes, anticline V. (A) and (C) are uninterpreted and interpreted versions of the same line, and (B) and (D) are uninterpreted and interpreted versions of the same line. See Figure 3 for location. HA, HB, and HC—mapped seismic horizons A to C, respectively. TWT—Two-way time.
on the ages of seismic horizons discussed at the start of this section. For 450 m uplift, the average uplift rate of the anticline crest above the level of the seafloor in adjacent synclinal areas is between 0.16 and 0.28 mm yr\(^{-1}\), similar to the sedimentation rates of the adjacent piggyback basin. For the anticlines located higher up the continental slope, growth started earlier, and uplift has been greater (Fig. 13, anticlines II, III, and IV). For anticline V, the fold amplitude of horizon C is ~1190 m (Fig. 9), and at the crest, horizon C has been eroded off. Anticline VI (Fig. 5) shows a very similar situation with a fold amplitude for horizon C of 1200 m. Using similar sedimentation rates to those above, average uplift rates for anticlines V and VI range between 0.24 and 0.42 mm yr\(^{-1}\). Hence there is at least a hint from the uplift estimates that anticlines affected by unconformities and degradation complexes are associated with both greater total uplift, and somewhat more rapid uplift rates than simple anticlines.

The amount of shortening across the fold belt, based on line-length measurements of horizon C is small; 6.5 km across a region that today is 55 km long, or ~10%. Assuming a 3 Ma age for horizon C based on correlation of seismic horizons to wells on the shelf and the deep-water wells in Ingram et al. (2004), the approximate shortening rate is 2 mm yr\(^{-1}\). This shortening rate is close to the 3–4 mm yr\(^{-1}\) difference in westerly motion between south (Sundaland) Borneo and NW Borneo derived from GPS data (Simons et al., 2007). The differential motion is occurring at ~45° to the orientation of the folds; hence if there is strain partitioning, then the amount of shortening in the fold-thrust belt should be considerably less than the westerly motion. Horizon C lies within the synkinematic section, so the shortening estimates are not the maximum amounts. The maximum percentage shortening across the belt based on line-length measurements of deeper, pre-kinematic section from 2D seismic data is ~17%. The rates estimated above are the same order of magnitude as external folds from the subduction-related setting of Taiwan, where a study by Delcaillau et al. (1998) estimated uplift rates for the Pakuashan anticline of ~0.3–10 mm yr\(^{-1}\) over the past 500 ka, with an E-W shortening rate of ~4–8 mm yr\(^{-1}\). Zagros folds have also been estimated as shortening at similar rates (3–4 mm/yr\(^{-1}\); Oveisi et al., 2007).

From the amplitude maps, it is possible to see the distal propagation of folds evolving between the mapped horizons. Between horizons D and B the most external fold moved from fold VI to fold IX (Fig. 15), a distance of 20 km over a time period of ca. 1.2–2.0 my, based on sedimentation rates discussed above. Within this period of time, fold VIII propagated to its full length (50+ km). Indicating lateral propagation of a fold is faster than propagation of folds in the transport direction by at least 2.5 times. Chen et al. (2007) report lateral fold propagation up to 40 km/m.y. based on magnetostratigraphic timing of outcrop sections, for the Kashi anticline. This is a very similar rate to lateral propagation estimated here.

Figure 25 shows two stages of fold development interpreted as fault propagation folds, based on 3D seismic data for the upper 1 km discussed in this paper and 2D seismic for the deeper section in the adjacent area of Sabah.
Growth of folds in a deep-water setting

(Ping and Hailing, 2004). The sections show that near the fault tip the more external anticline (VIII) dips at a lower angle (20°) than the more internal anticline (35°). In a simple kink-band, fault-propagation fold model, this would require that the initial footwall ramp dips were different in the two cases. However, regional 2D seismic data show the imbricate faults have a listric shape, matched by the fold backlimbs having a smoothly curving shape (James, 1984; Ping and Hailing, 2004). As the fault continued to propagate, it steepened; consequently, for a constant strain rate, uplift associated with a fold would increase due to the increase in fault dip. Hence the higher uplift rates discussed above for the more internal anticlines (e.g., V and VI) compared with the external anticlines can be explained by changes in fault dip (e.g., IX and VIII). The listric shape of the imbricate faults and the different depths at which the blind imbricate fault terminates with respect to fold geometry supports this interpretation. The simple fold geometry with normal faults developed while the thrust fault was blind, the fault tip lying ~0.4 km or deeper below the seafloor. Degradation complex and erosional unconformities are associated with a thrust that was emergent or very close to the surface.

13. DISCUSSION

How growing folds at the surface interact with their environment varies significantly passing from continental, to shallow marine, to deep marine settings (Fig. 26A). Continental and shallow marine environments have two modes of fold-environment interaction as follows:

1. Nondeposition or erosion of the anticline crest due to its emergence either at the land surface or above wave base. Any sedimentation associated with the anticline onlaps the flanks and is deposited in the synclinal basins (Fig. 26A). The evolution of drainage and uplifted terraces can be used to infer fault development and propagation in a continental setting (e.g., Delcaillau et al., 1998). (2) The anticline is covered by sediment. If a fold develops within a subsiding basin (usually a foredeep basin or a large thrust-sheet–top basin), then it is possible for sedimentation to keep up with fold growth resulting in the anticline crest being spared from major erosional truncation (e.g., Apennines foredeep basin, Bally, 1989). These have been referred to as fill-to-the-top models (see review in Castelltort et al., 2004; Fig. 26B).

Examples of syntectonic fold growth used to support kink-band–style, fold-growth models (e.g., Suppe et al., 1992; Shaw and Suppe, 1994) typically come from fill-to-the-top settings in shallow marine and continental environments. The pre-kinematic sedimentary strata in these settings are typically well lithified, and the erosional products of the folds give rise to coarse-grained clastic sediments. As discussed by Nigro and Renda (2004) and Morley (2007b) deep-water folds (Fig. 26C) are different, for three main reasons: (1) folds are protected from subaerial and wave-related erosion; hence thicker, and more widely distributed synkinematic sedimentation can occur across anticlines, as well as within synclinal piggyback basins; (2) the synkinematic sediment deposited across anticlines during growth is poorly lithified, predominantly fine grained and weak, and hence liable to be very sensitive.

Figure 25. Relationship of shallow structure determined from 3D seismic data (Fig. 9) and the deeper fold geometry (based on regional 2D seismic lines; Ping and Hailing, 2004). The folds are modified fault-propagation folds with listric imbricate fault geometry. (A) Simple fold with crestal normal faults; (B) fold affected by degradation complexes and overpressured mud and fluid pipe; (C) illustration of fold morphology in (B) without the effects of erosion and mass wasting of the crest and forelimb.
to the effects of gravitational force during fold growth; and (3) the main erosional processes are a variety of mass-movement and mass-wasting phenomena, including rotational slides, fast and slow creep mass movement, debris flows, and turbidity flows, and currents (e.g., McGilvery and Cook, 2003; Heinio and Davies, 2006; Gee et al., 2007). Sheet-type flows are considerably more extensive in the deep-water environment than they are in shallow marine or continental settings, and the effects of their erosive power on the poorly lithified sediments forming anticlines (e.g., Fig. 24, Gee et al., 2007) will be much greater, and create different transport paths in deep water than in continental and shallow marine settings.

Propagation and linkage geometries in normal faults have been extensively documented, and the less well documented lateral propagation of folds in response to linkage of underlying thrust faults has been treated in a similar way (e.g., Dahlstrom, 1969; Nicol et al., 2002; Delcaillau et al., 2006). In both normal faults and folds, propagation and linkage patterns have been determined from synkinematic sedimentation patterns and geomorphology associated with active faults and folds that have considerable surface expression. Gravity-driven sediment transported through syntectonic topography related to different structural styles (folds and rifts) exploits variations in fold or normal fault geometry to move down the regional slope via structural lows such as fold saddles and fault displacement lows and the plunging terminations of en-echelon folds or the equivalent relay ramps in normal faults.

The folds investigated here show mature linkage geometries after only ~17% shortening, indicating, that like normal faults, in many cases linkage of folds and their underlying thrust faults occurs at a relatively early stage in their strain history (e.g., Morley, 1999). Another aspect of normal fault growth is how displacement patterns become reorganized during linkage: does the area of maximum displacement relocate to the center of the linked fault system, or do the linkage points remain persistent regions of low displacement during the life of a long fault (e.g., Cartwright et al., 1995; Morley, 1999)? In the examples of folds described here, some of the changes in syncline depocenter locations suggest shifting patterns of thrust displacement at depth; the best example is NW of anticlines V and IV where area Q is a prominent depocenter (Fig. 12C) that developed between horizons B and C and then is much less active subsequently (Fig. 12B), while areas N and P along the same fold trend remained active between horizons B and A. It is also possible for saddle areas along anticlines to remain as relative lows because they represent the areas of latest fold development, and once fold segments are joined, the saddle areas subsequently grow in amplitude at similar rates to the crestal areas. However, this commonly does not appear to be the case. At the crests of more mature folds (e.g., anticline III) there is evidence for continuous growth as shown by progressive rotation and thinning of section onto the anticline. In the saddle areas for the same age sequences, there are periods where uniform thicknesses of sediment were deposited across the anticline, followed by renewed folding (Fig. 20). In the case of area K between folds VII and VI, early fold activity (Fig. 12C) was followed by inactivity with no reactivation (Figs. 12A and 12B). Thus anticline growth in the saddle areas is commonly more episodic than in the crests, suggesting incomplete kinematic linkage of the underlying fault systems.

With some caveats, the different ways flows exploit fold transfer zones and linkage points described here are applicable in terms of fold evolution and rate of processes to a wide variety of tectonic settings (see section 11). For example, geomorphic criteria to help identify subaerial lateral fold propagation have been documented by Delcaillau et al. (2006, 2007) from the Siwalik ranges of the Himalaya. These workers have been able to distinguish between anticlines propagating in a single direction versus linkage of several inline folds. The Zagros Mountains show paleodrainage patterns that are incompatible with present-day fold geometry, but which can be interpreted in terms of lateral propagation of fault and fold segments and linkage fold segments that are typically 20–40 km long (Ramsey et al., 2008). The data suggest that for Himalayan, Zagros, and deep-water folds, shortening rates are in the order of mm/yr⁻¹, while lateral and transport direction fold-propagation rates are at least an order of magnitude faster.

Sedimentary flows have originated from a number of sources: (1) locally from growing anticlines, (2) from high up the continental slope, sourced by headless channel systems, which are active today, and (3) from the shelf. Shelf-sourced (more sand-rich) sediments appear as a source of flows primarily during times of sea-level lowstands and tectonic uplift of the shelf (Ingram et al. 2004). The present-day time of sea-level highstand is reflected by the absence of channels on the shelf connecting to the deep water, the presence of headless channel
systems along the slope and the dominance of mud and silt, and virtual absence of sand in numerous grab samples taken across all environments in the deep-water area.

The way in which deep-water fold-belt structure influences sediment pathways on a slope described here is not just applicable to overpressured shale-prone parts of large deltas and accretionary prisms (e.g., McAdoo et al., 1997). There are many parallels to be found between the development of deep-water, salt-related mini-basins and the thrust-related piggyback basins discussed here. One common feature of tectonically active continental slopes may be the triggering of landslides at higher slope angles than on tectonically quiescent margins. McAdoo and Watts (2003) note that the California, Texas and Louisiana, and New Jersey and Maryland continental margins undergo landslides on slopes less than 4°, while for the Cascadia accretionary complex, offshore Oregon, most landslides occur on slopes over 15°. In the anticlines described here, sediment accumulation and episodic mass wasting is observed to increasingly modify fold geometry as surface slope increases, up to a maximum of 13° dip of the seafloor on anticline forelimbs.

Salt-related depocenters, deltaic deep-water fold belts, and accretionary prisms create depocenters between actively uplifting ridges or mounds and force gravity-driven sediments to deflect around them, or carve a path through them, commonly exploiting structural lows (e.g., McAdoo et al., 1997; Sinclair and Tomasso, 2002). In both deformation styles, the minibasins farthest upslope fill first; then sediments spill over in basins farther downslope. In a separate paper focusing on the evolution of the minibasin between anticlines IV and VIII described here, Morley and Leong (2008) discuss basin development in terms of “fill and spill” cycles developed for salt-related basins of the Gulf of Mexico (e.g., Prather et al., 1998; Sinclair and Tomasso, 2002; Booth et al., 2003). While there are potential significant differences in strain rate, range of structural styles, sequence of deformation, ridge geometry, mini-basin area and shape, between the main deep-water deformation styles, which may lead to some significant differences in structure-sediment interaction, in general, the similarities are compelling.

14. CONCLUSIONS

The folds described in this study display dimensions, growth rates, propagation and linkage geometries that are very similar to large-scale folds seen in orogenic belts. Consequently the ability to obtain high-quality, 3D seismic images of deep-water folds and their synkinematic strata, which is lacking in many orogenic belts, offers the potential for using deep-water fold belts to better quantify the growth and behavior of fold-thrust belts in general. Despite the many similarities between different kinds of major fold-thrust belts, fold growth in a deep-water setting occurs in an environment that protects synkinematic and pre-synkinematic sediments from wave and subaerial erosional processes that affect folds in continental and shelfal environments. Consequently deep-water folds are formed in weak rocks that differ in lithology, degree of lithification, and mass movement processes compared with orogenic belts. There are strong feedbacks between sedimentation, fold growth, and fluid migration in the deep-water environment. The key points arising from this study are as follows:

1. In the deep-water environment, the structural evolution of folds is strongly linked with sedimentary processes and their modifying influence on fold morphology.

2. Shortening in the toe fold-thrust belt may result from a stress field that combines the effects of gravity-driven and far-field stress. However, estimated shortening rates (~10% shortening in 3 m.y., at a rate of ~2 mm yr⁻¹) seem appropriate for purely regional-driven strains as determined by GPS measurements (Simons et al., 2007).

3. Deep-water sedimentation rates time-averaged over millions of years for the area range up to ~0.28 mm yr⁻¹. Such rates are sufficiently slow that for most growing anticlines, the uplift of the fold crest above the seafloor in the areas of the synclines exceeds the sediment rate by ~0.16–0.42 mm yr⁻¹. Hence topography associated with growing anticlines exerts a strong influence on sediment pathways along the continental slope, particularly on sediments transported in systems linked to headless slope channels. Conversely much larger submarine landslides, such as the giant slide in the western part of the area, are little influenced by structural topography, and can actually decapitate the crests of anticlines.

4. As anticlines tighten and increase in amplitude (either along-strike, or with time as folds grow), folds show the following tendencies: (a) at high interlimb angles (180°–165°) the fold shape is simple, and the seafloor lies subparallel to the upper internal reflections of the fold. The excess seafloor dip overlying the forelimb ranges up to 4°; (b) interlimb angles between 170° and 140° are associated with a range of simple fold geometry that is either weakly or strongly modified by swarms of normal faults that affect the fold crest. Generally as interlimb angle decreases, fault density increases. The normal faults are gravity driven, and appear to be a deep-seated, mass-movement phenomenon. The normal faults die out downward at different levels and do not detach into rotational slides. Shallow rotational slides are common features affecting the forelimbs of many folds with interlimb angles between 165° and 140°. Rotational slides tend to occur where the seafloor dips above the forelimb attain angles of 6°–11°. Where folds have tightened to interlimb angles less than ~145°, the crests of folds become strongly modified by erosional unconformities and forelimb degradation complexes. Generally the seafloor slope above folds does not attain dips greater than 13°, indicating that mass-movement and mass-wasting processes keep pace with fold growth at such dips.

5. Large (hundreds of millimeters in diameter), mud-pipe intrusions and gas chimneys are mostly located in anticline fold cores, particularly the tightest, highest-amplitude folds. The presence of hydrothermally generated hydrocarbons suggests a deep (>3 km) source for the fluids. The fluids probably migrate up zones of weakness at depth (thrust planes) until they are able to propagate subvertical hydraulic fractures to the surface. The presence of shallow overpressured fluids in the crests of anticlines is also likely to aid slumping of shallow crestal and forelimb sediments. Migration of fluids arising from compaction of sediments in synclinal depocenters occurs in clusters of small (<100-m-diameter) pipes in the backlimbs of anticlines. The pipes rise from source depths of ~500–700 m.

6. Folds exert a strong influence on sediment pathways. During the early synkinematic stages of fold development, most sediment transport is along synclinal piggyback basins parallel to the fold trends. Consequently early sediment pathways from one piggyback basin to another lie along low-angle entry points in particular at stepovers between en-echelon folds. As folds grow, pathways transverse to the folds begin to be exploited. Drainage transverse to the fold direction develops at structurally low areas, such as: (a) saddles in a fold trend, where two folds have joined along-strike, (b) regions where two oblique fold trends meet, (c) areas where fold growth has been slow, perhaps due to shortening being distributed across more folds than is typically the case. However, transverse channels can also develop at structural highs weakened by mud pipes where rapid erosion can occur.

7. An overall internal to external younging of fold initiation is discerned for the deepwater offshore area; however, across a 40- to 50-km-wide belt (in the transport direction), it is apparent that five to seven folds can be active at the same time, and that could involve any fold in the belt (Fig. 15). For the duration of the...
Morley

3.5–4 m.y. activity investigated here, the active fold belt has moved northward by ~50 km. Lateral fold propagation is at least 2.5 faster than propagation in the transport direction. Fold activity also varies laterally—a segment of a fold may be active while another portion of the fold is sealed by an unfolded unconformity, and hence has remained inactive for some time. This probably reflects incomplete kinematic linkage of thrust fault systems at depth. Saddle areas also arise where the rate of fold growth changes from one part of a fold trend to another with time. This is particularly evident for folds V and IV, and the way the location of the thickest sedimentary section in the adjacent piggyback basin changes with time (Figs. 12B and 12C).

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REFERENCES CITED

Ajakaye, D.E., and Bally, A.W., 2002, Course manual and REFERENCES CITED thanked for constructive reviews that helped improve the version of the manuscript benefi ted from comments by Arnold Bouma. Trevor Elliott and Stefan Back are acknowledged. In particular, Joe Lambiase, Martin Gee, John Warren, and Angus Ferguson are thanked for discussions about the geology of the area. An early version of this manuscript benefited from comments from Arnold Bouma. Trevor Elliott and Stefan Back are thanked for constructive reviews that helped improve the final version of the manuscript.

REFERENCES CITED

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REFERENCES CITED

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REFERENCES CITED

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REFERENCES CITED

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REFERENCES CITED

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REFERENCES CITED

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REFERENCES CITED

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REFERENCES CITED

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REFERENCES CITED

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REFERENCES CITED

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Ping Yan, and Hailing Liu, 2004, Tectonic-stratigraphic divi-
sion and blind fold structures in Nansha Waters, South
Prather, B.E., Booth, J.R., Steffens, G.S., and Craig, P.A., 1998, Classification, lithologic calibration, and strati-
Barnett, D.I., and Newall, I., 2000, The geology and hy-
drocarbon resources of the Arafura Shelf, Northern Australia: Australian Petroleum Geologists and
Whitaker, A.E., and Bartholomew, M.J., 1999, Layer paral-

Tingay, M.R.P., Hillis, R.R., Morley, C.K., Swarbrick, R.E., and Drake, S.J., 2005, Present day stress orien-

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