Systematic spacing and topological variations in layer-bound fault systems

Mark T. Ireland1 | Chris K. Morley2 | Richard J. Davies1

1School of Natural and Environmental Sciences, Newcastle University, Newcastle upon Tyne, UK
2PTTEP, Enco, Bangkok, Thailand

Abstract
Polygonal fault systems, sometimes termed layer-bound faults, are extraordinary features of many fine-grained sedimentary successions and have been described in a significant number of sedimentary basins over the last two decades. Their formation represents an important mechanism by which fine-grained sediments compact often resulting in a variety of complex patterns for which several controlling factors have been proposed. Here, three-dimensional seismic data from the North West Shelf of Australia are used to interpret previously undescribed characteristics of layer-bound fault systems where systematic horst and graben structures are the dominant structural style. Conjugate fault pairs, which form the horsts and grabens, frequently have a systematic spacing with graben-bounding faults exhibiting a spacing of half that of the horst-bounding faults. This systematic spacing of fault pairs indicates, (a) the presence of a mechanically weaker layer at the base of the fault system and (b) that the horizontal shortening required by the volume loss due to compaction can be accommodated without reaching saturation with respect to fault intensity. Furthermore, topological analysis indicates that areas with different patterns also have different intersection and branch characteristics, and these differences suggest that the growth of layer-bound faults is not explained by a single model. The findings have implications for the genesis and growth of layer-bound fault systems and the potential for cross-stratal fluid flow.

Keywords
fault spacing, layer bound faults, polygonal faults, topology

1 | INTRODUCTION
Polygonal faults, sometimes termed layer-bound faults, are a ubiquitous feature in many sedimentary basins on Earth (Cartwright, 2007). These fault systems are commonly found in fine-grained sedimentary successions and comprise networks of normal faults with orientations that typically form polygonal geometries on bedding planes (Lonergan et al., 1998). The bedding plane geometries of the fault systems may depart from these polygonal forms due to the influence of external stresses such as basin floor slope (Higgs & McClay, 1993), tectonic faults (Hansen et al., 2004) and the influence of stratigraphic features (Ireland et al., 2011). They are an important class of fault because they typically deform fine-grained, low permeability sediments that are
Highlights

- Horsts and grabens within a layer bound fault system exhibit a systematic spacing.
- Graben bounding faults exhibit a spacing half that of the horst bounding faults.
- Areas with a preferred fault orientation have more I nodes compared with polygonal patterns.
- Variations in topology indicate layer bound faults may not be explained by a single model of growth.
- Fault spacing, and topology are inherently linked to the local geological setting.

Frequently sealing sequences for fluids in sedimentary basins (Cartwright et al., 2007) and represent a mechanism for sediment dewatering and compaction. They have been one of the most debated and enigmatic geological structures discovered using 3D seismic data. The characteristics of complex fault networks and fault interactions are important for understanding how mechanical layers cause variations in the displacement and scaling relationships of faults (Peacock, 2002).

Over the past two decades, the increasing coverage, resolution and availability of three-dimensional (3D) seismic reflection data from sedimentary basins on Earth have led to the widespread recognition of layer-bound fault systems and numerous variations in their geometries (e.g. Ghalayini et al., 2017; Morgan et al., 2015). Increasingly analysis has shifted from qualitative descriptions to quantified structural analysis, including, but not limited to, throw and displacement, orientation and topological analysis (e.g. Morley & Binazirnejad, 2020; Stuevold et al., 2003; Wrona et al., 2017). Here, observations and interpretations from 3D seismic data from the North West Shelf, Australia are used to describe the spacing and topology of a layer-bound fault system and the implications for its genesis and growth are examined.

2 | LAYER-BOUND FAULT SYSTEMS

The origin of layer-bound fault systems has received a wide variety of interpretations, but they are generally accepted to be the result of volumetric reduction, by bed-parallel compaction, which complements the heaves on the faults, in addition to vertical compaction (Cartwright & Lonergan, 1996). Laboratory measurements (Bishop et al., 1971) and field data (Goulty & Swarbrick, 2005) suggest that low coefficients of friction on fault surfaces may be an important factor that allows these fault systems to develop (Goulty, 2008; Goulty & Swarbrick, 2005). Once faults have nucleated in the fine-grained host sediment, they can continue to grow with increasing overburden stress under laterally confined conditions, provided that the coefficient of residual friction on the fault surfaces is sufficiently low (Goulty, 2008). Increasingly, layer-bound faults have been identified in a wide range of fine-grained lithologies, including chalk (Hansen et al., 2004) and siliceous sediments (Davies et al., 2009; Seebeck et al., 2015).

Geometric and topological analyses of fault systems are important to investigate their evolution (e.g. Duffy et al., 2017; Nixon et al., 2020) and can support examining the influence of factors such as lithology variations, depositional and stratigraphic setting, gravitational instability and post-faulting compaction. Fault systems are commonly characterized according to fault geometry (e.g. Barnett et al., 1987) and fault network properties (e.g. Bour & Davy, 1998), which include fault spacing (Barnett et al., 1987; Soliva & Benedicto, 2005) and connectivity (Sanderson & Nixon, 2015, 2018). The mechanics that control and influence fault spacing in normal fault populations are applicable to both polygonal (or layer bound) fault systems, as well as other fault populations which are confined to discrete mechanical layers (e.g. Benedicto et al., 2003). Regular patterns of fault spacing can occur at different scales; and are often attributed to the thickness of the mechanical layer (Soliva et al., 2006). The role played by layer thickness in controlling vertically restricted fault systems has been linked to the differences in mechanical strength within layered sequences (e.g. Benedicto et al., 2003). There have been numerous analogue studies that have investigated the role of basal detachments (Axen, 1988) and deformable substrates (Li & Mitra, 2017), in extensional regimes. These studies have highlighted the role that rheological variations play in the formation of different normal fault patterns with both hard and soft linkages (Bahroudi et al., 2003). Recently, the application of topological analysis has provided quantitative descriptions of fault and fracture networks, and combined with geometric analyses has helped to investigate their genesis (Morley & Nixon, 2016). Topology describes the arrangement of the faults and fractures within a network focusing on the geometrical relationships between them (Jing & Stephansson, 1997). The topology of fault and fracture networks can be characterized by branches (Sanderson & Nixon, 2015) and nodes (Manzocchi, 2002). There are three types of node: isolated tips (I); crossing fractures (X) and abutments or splays (Y nodes or T nodes). Commonly faults terminate against (or abut) or cross-cut pre-existing faults, producing many Y nodes (Duffy et al., 2017; Morley & Nixon, 2016; Nixon et al., 2012). Morley and Binazirnejad (2020) described the detailed topology of polygonal fault sets to investigate
variations in connectivity and identified that the polygonal fault sets share similarities in node and branch topology with complex tectonic fault patterns in rifts.

3 | GEOLOGICAL SETTING

The Exmouth Plateau is bounded by the continental shelf to the southeast, and the Argo, Gascoyne and Cuvier abyssal plains to the northeast, northwest and southwest respectively (Longley et al., 2002). The Exmouth Plateau is part of the North Carnarvon Basin, which experienced several rifting events between the Late Carboniferous and Early Cretaceous, with seafloor spreading commencing across the Argo Abyssal Plain in the Late Jurassic and across the Gascoyne and Cuvier abyssal plains in the Early Cretaceous (Longley et al., 2002; Tindale et al., 1998). Broadly distributed intracratonic basin formation across the North Carnarvon Basin during the Permain was followed by the deposition of thick Triassic sequences across the margin (Driscoll & Karner, 1998). Following this, further extension in the late Triassic affected the margin and formed a network of basins, including in the Exmouth Plateau area (Driscoll & Karner, 1998). The latest Triassic–earliest Late Jurassic extension in the central Exmouth Plateau area is characterized by NNE-SSW trending domino-style extensional faults (Bilal et al., 2020). Post-Valanginian thermal subsidence across the Northern Carnarvon Basin was associated with the deposition of deep marine sediments which typify the Cretaceous to recent deposition (Bilal et al., 2020; Driscoll & Karner, 1998). This period is commonly associated with the onset of current-controlled deposition (e.g. Driscoll & Karner, 1998), and changes in the depositional patterns during this time have been linked to regional tectonic processes on the Exmouth plateau where the stratigraphy records plate reorganization and adjustment (Nugraha et al., 2019). The 3D seismic data are located on the Exmouth Plateau and to the west of the Kangaroo Syncline (see Figure 1).

The faulted interval comprises recent to Campanian age stratigraphy, which is constrained by the Moyet-1 well (Figure 1b). The Recent to Pliocene (Delambre Formation) interval extends down from the seafloor to the T40 sequence boundary, which is a prominent peak event in the area, tied to the nearby Moyet-1 well. The Miocene and Eocene (Bare to Upper Walcott Formations) is marked at the top by the T40 sequence boundary and the base is marked by the T27 horizon. The Eocene to Palaeocene (Lower Walcott and Dockrell Formations) is marked at the top by the T27 horizon, and at the base by the T10 horizon. A description of the interval of interest is provided in the Moyet-1 well report (Woodside, 2011). The Dockrell Formation commonly comprises marine marls and clays, and the Walcott Formation is typically composed of calcareous foraminiferal marls and clays. The Maastrichtian to Campanian (Mira and Withnell Formations) is marked at the top by the T10 horizon and the base at the K60 horizon. The top and base of the Campanian to Aptian (Toolonga Calcilutite and Haycock Marl) interval are the K60 and K40 markers respectively. The Miria Formation is commonly composed of marine clays, marls and oozes, and the Withnell Formation of calcareous clays. The primary interval of interest is the Miocene to Maastrichtian section which comprises a succession of dominantly fine-grained calcareous sediments. There are no core across the faulted interval in the study area. Nearby Ocean Drilling Program sites 762, 763 and 766 (Brereton, 1992; O’Brien & Manghnani, 1992) provide the most comprehensive data on stratigraphically equivalent lithologies and the associated physical properties for the Cretaceous to recent post-rift sequence. The porosities of stratigraphically equivalent sediments at Site 766 are between 50% and 75%, at shallower burial depths down to a maximum of 90 meters below seafloor (mbsf; Brereton, 1992). Similar lithologies are encountered in stratigraphically older sequences, and at depth ranges of 200 to 500 mbsf, porosities vary between 30% and 60% (Brereton, 1992).

4 | DATABASE AND METHODOLOGY

4.1 | Seismic data and interpretation

The 3D seismic survey interpreted in this study is located on the Exmouth Plateau on the North West Shelf of Australia (Figure 1). The survey was acquired in 2002 and provided by Geoscience Australia. The data are post-stack time migrated with a bin spacing of 12.5 x 12.5 m and a vertical resolution of ca.10 m. The data are zero-phased, with SEG reversed polarity (SEG, 2019), where a positive reflection coefficient, representative of an increase in acoustic impedance, is plotted as a black–red–black reflection. The interpreted seismic horizons were tied to the nearby exploration well, Moyet-1 (Figure 1), which is also used to describe the lithology of the interval studied (Figure 2). A series of horizons from stratigraphic reflections were interpreted and the subsequent geometrical properties (e.g. Chopra & Marfurt, 2012) and seismic attributes (Bacon et al., 2007) were used to describe the planform geometries of the observed layer-bound faults. To quantify fault orientation, we interpreted fault traces from horizon attributes (Figure 3). A variance attribute (Figure 3b) was used to enhance structural features (see Chopra & Marfurt, 2008). This attribute applies a windowed analysis across the seismic volume to calculate the degree of trace to trace similarity along the selected dip within the defined 3D window (Barnes, 2016). In addition to variance, an ant-tracking attribute was used (Figure 3c) to refine the interpretation of faults and their terminations, as ant-tracking attributes...
specifically try to eliminate horizontal variations in coherence associated with stratigraphy (see Chopra & Marfurt, 2008). Figure 4 shows key horizon maps used for the interpretation here. Detailed interpretations based on the stratigraphic horizons and the associated attribute maps allowed for the planform geometries and intersections to be mapped out in detail.
4.2 | Fault analysis

To analyse the spacing of faults, a scanline approach was adopted (e.g., Soliva et al., 2006). Fault spacing was measured along four scanlines, two inlines (IL 3400 and 3600) and two crosslines (XL 4400 and 4600), each ca. 30 km long (Figure 1d). The spacing distance was measured on the T10 horizon between bounding faults of horsts, bounding faults of grabens and faults that have the same dip direction (see Figure 5c). Across all four scanlines (IL 3400, 3600, XL 4400 and 4600), a total 264 fault spacings and segment types were measured. The T10 horizon was chosen because (a) the faults have b-type displacement depth profiles (e.g., Wrona et al., 2017) and therefore exhibit displacement maxima located towards the lower tip, (b) it is a continuous reflection that can be mapped with high confidence through the faulted interval and (c) it exhibits the most structural complexity of any single level within the fault system. The results of the fault spacing are presented as statistical data distributions, including cumulative frequency, probability distributions and histograms grouped by the different fault block types, horsts, grabens and same dip direction. The coefficient of variation, defined as $\sigma / \mu$, where $\sigma$ is the standard deviation and $\mu$ is the mean and is used to compare the degree of variation between the spacing of conjugate fault pairs. In addition, we analyse the topology of the layer-bound fault system using five sample areas (labelled Areas A-E in Figure 1d), with a total of 2161 individual fault traces interpreted from the attribute maps presented in Figure 3. NetworkGT (Nyberg et al., 2018) was used for the topological analysis of mapped fault traces that were interpreted from the seismic horizon. NetworkGT is an open source toolbox for ArcGIS which can be used to sample, analyse and spatially map the geometric and topological attributes of two-dimensional fracture networks (Nyberg et al., 2018).
The analysis characterizes the branches and nodes of the fault network and the results are presented in the form of maps, rose diagrams and histograms. Following Sanderson and Nixon (2015), we used the number of connections per branch ($C_B$) to provide a measure of connectivity of the faults systems, which is given by Equation (1) where $N_Y$ is the number of $Y$ nodes, $N_X$ is the number of $X$ nodes and $N_B$ is the number of branches:

$$C_B = \frac{3N_Y + 4N_X}{N_B}$$

The results from the topological analysis of the fault systems are presented as maps, with the proportions of different nodes and branches presented in triangle plots. The orientation of the faults is presented as rose diagrams.

5 | SEISMIC OBSERVATIONS AND INTERPRETATIONS

5.1 | Faulted interval

The primary interval of interest is the Miocene to Maastrichtian section, between the T40 and K60 seismic markers (Figure 2). The succession comprises dominantly fine-grained calcareous sediments and has a network of
normal faults developed across the survey area throughout this interval. Based on offset data from Ocean Drilling Program sites 762, 763 and 766, the faulted interval mostly likely comprised fine-grained calcareous oozes and claystones. These lithologies have high porosities (>40%) even at depths of up to 500 m. The data indicate that these lithologies undergo significant porosity reduction during shallow burial, where similar lithologies have porosities >60% in the first 100 mbsf. Individual faults may separate polygonal fault blocks of un-faulted sediment. Across much of the area, the polygonal fault blocks are often bounded not by a single fault, but by conjugate, graben forming pairs (see Figures 2 and 4b). The faults predominantly form a single tiered network, although there are occasional faults that are antithetic to large faults and which only offset the upper part of the interval (see Figure 2). There are a range of fault strike orientations evident from the T10 time structure maps (Figure 3a) and the seismic attribute maps (Figure 3b,c). Thickness variations are apparent across the area, with measurable differences in the gross interval thickness of the T40 to T10 interval (Figure 5a), with subtle thickness variations across individual horst and graben.

**Figure 5** Isochron maps of (a) K60 to T10 and (b) T10 to T27. Zooms shown in parts (c) and (d), with the position of a polygonal horst surrounded by a network of narrow grabens annotated in each case. See Figure 1d for location.
fault blocks (Figure 5). The mean isochron thicknesses for the T40 to T10 interval of each area vary between 474 ms (Area C) up to 521 ms (Area E). Table 2 shows the mean isochron thicknesses for each of the sample areas. The mean isochron thicknesses between each area differ by <50 ms. Within the areas, there are isochron thickness variations between 125 ms (Area D) and 218 ms (Area E). The isochron thickness maps for Areas A-E are shown in Figure 5a. From the isochrons, a characteristic increase in thickness within the grabens can be observed, relative to compared with the polygonal horsts (Figure 6c). The T10 to K60 interval, which is the interval where most of the layer-bound faults tip out, shows systematic thickness variations that are clearly linked to the development of the horst and graben structures (Figure 5b,d). Beneath the downthrown grabens, the interval is thinner than beneath the adjacent horst blocks (Figure 5d). Within this interval, the conjugate graben pairs commonly have lower tips which converge or intersection near the top of this section (see Figures 3 and 4).

5.2 | Fault geometry and spacing

The geometries of individual faults forming the fault network varies from planar to listric. In general, faults with a greater vertical extent (and typically with a greater displacement) have a listric geometry, whereas faults with a limited vertical extent and smaller displacements are closer to planar. Faults with their lower tips below the T10 stratigraphic horizon are more commonly listric, whereas those restricted to above T10 are more planar. There are some faults that extend, and have their lower tips co-located with the upper tips of deeper faults (see Figure 2). Figure 6 shows IL3400 and 3600 with an interpretation of horsts, grabens and fault blocks bounded by the same dip direction faults. This interpretation is based on the geometry at the T10 sequence boundary (Figure 6c), which is located towards the lower tips of the faults in the system. Table 1 summarizes the fault spacing statistics for all the measurements from the scanlines. As an example of the distribution of fault block structures, along IL3400, there are 18 conjugate faults which bound horsts, 17 which bound

TABLE 1 Summary statistics for fault spacing from the four scanlines (IL 3400, 3600 and XL 4400 and 4600)

| Fault block type | N   | Mean  | SD   | Coefficient of variance | Min  | Max   | Median |
|------------------|-----|-------|------|-------------------------|------|-------|--------|
| Same dip         | 113 | 343.6 | 228.4| 0.66                    | 72.8 | 1085.2| 257.5  |
| Horst            | 76  | 543.4 | 341.8| 0.63                    | 72.6 | 1631.6| 448.8  |
| Graben           | 75  | 236.2 | 95.1 | 0.40                    | 64.4 | 448.9 | 222.5  |
grabens and 23 fault blocks bounded by faults which have the same dip direction. At the base of the faulted interval, which is taken to be the K60 horizon, the normal faults which bound horsts and grabens rarely intersect or displace adjacent faults, which is interpreted as indicating that fault initiation and propagation are contemporaneous across the scanlines. In total, 29% of the faults form conjugate sets which bound horsts, and 28% form conjugate sets which bound grabens. The average spacing of conjugate graben-bounding faults is 236 m compared with 543 m for conjugate horst-bounding faults (see Table 1). Figure 7 shows the distribution of fault spacing for the different fault segment types and shows the spacing of all fault block types. The statistics for fault spacing are summarized in Table 1. The conjugate horst pairs exhibit a log-normal distribution (p-value = .009) (Figure 7) and 64% of this population have a spacing greater than the average for the whole sample population. The conjugate pairs that form graben pairs fit a log-normal distribution (p-value = .145) (Figure 7) and of these faults 91% have a spacing less than the average for whole sample population. Where a fault block is bound by the same dip direction faults, the segments broadly fit a log-normal distribution (p-value < .005) (Figure 7). Of these faults, 25% have a spacing greater than the average of the whole population. The standard deviation for fault spacing of grabens is 94 m, compared with 227 m for faults blocks bounded by the same dip faults and 340 m for horsts. The coefficient of variance for the spacing of faults which bound grabens is 0.40, compared with 0.66 for fault blocks bounded by faults with the same dip and 0.63 for faults which bound horsts. In each case, the coefficient of variance is < 0.7 indicating a regularly spaced network rather than random. The lower coefficient of variance of grabens (0.40) compared with horsts (0.63) and same dip fault pairs (0.66) indicates a greater degree of clustering in their development. The coefficient of variance for the spacing of both graben-bounding faults and horst-bounding faults, together with the log-normal spacing distributions (Figure 7), indicates a regular spacing that is anticlustered for both. The pronounced systematic differences in fault spacing between horst and graben conjugates suggest that fault spacing, and dip direction, is controlled not by random nucleation and growth, but a systematic control at the kilometre scale.

5.3 | Fault orientation and topology

In map view, the overall geometry of the fault system is polygonal; however, there are clearly identifiable variations.
across the survey. Figure 8 shows rose diagrams for the faults within each of the sample areas A-E. Overall, there is a wide variance in fault traces, with all sample areas exhibiting a circular variance between 0.65 and 0.75, where an orthogonal network with two distinct trends has a variance of 0.5 and an area with no obvious trend would have a variance of >0.85. The circular variances across all areas indicate that there are some prevalent trends but there are numerous faults with other orientations. In sample areas A, B and E, the map patterns qualitatively exhibit a polygonal fault, and the rose diagrams show a directional spread of fault trace orientations, commonly with two dominate orientations which are approximately orthogonal to one another (Figure 8f,g,j). For area C, the map patterns exhibit a strong preferential orientation, approximately NNE–SSW (Figure 8c) and this is reflected in the rose diagram (Figure 8h). Area D has two dominant orientations, one ENE–WSW and the other NW–SE (Figure 8d,i). Despite sample areas C and E exhibiting dominate directional trends, the overall circular variance is like that of other sample areas.

Across the survey area, there are 182 tectonic faults interpreted below the K60 horizon. These faults have a dominant orientation of NNE–SSW and are interpreted to have developed in response to the widespread Triassic–earliest Late Jurassic extension. The observed linkage with shallower layer-bound fault systems may be related to the periods of inversion during Cretaceous to recent period (Figure 2). Above these NNE–SSW trending tectonic faults, the orientation of the layer-bound faults is clearly strongly influenced by them. Where layer-bound faults in the T40 to T10 interval directly overly tectonic faults, they exhibit a dominant E–W trend (Figure 9b) which contrasts with the range of orientations for all layer-bound faults (Figure 9c). In general, these orientations are oblique to the trend of the tectonic faults (Figure 9a). This indicates a correlation between proximity to tectonic faults and orientation which is interpreted as the result of localized stress perturbation above the tectonic faults.

The topological analysis across the study area is summarized in Table 2 and in the ternary plots in Figure 10. The topology of the individual sample areas and their detailed planform geometry of the branches and node types is shown in Figure 8a–e. Overall, across all the sample areas, there is a wide range of fault orientations (Figure 8c), and an overall dominance of C–I branches (Figure 10a and Table 2). The topology of areas C and D which both exhibit strong preferential orientations (Figure 9h,i) shows very different topologies, with area C having a higher proportion of I nodes and lower proportion of Y nodes compared with area D (see Table 2). Areas A, B and E exhibit a similar planform geometry and topologically are generally more complex than Area C. Faults in Areas A, B, D and E have a higher proportion of C–C branches compared with Area C, indicating a higher extent of connectedness. In Area C, there is a greater proportion of I–I branches compared with the other sample areas. In Area C, 25% of the sampled fault branches are I–I. Using Equation (1) to determine the average number of connections per branch,
Area E has the most connections per branch (1.37), and Area C (1.01) the fewest (see Table 2).

Across the five sample areas, there are differences in the mean length of branches (Table 2). A common characteristic across all areas is that the average length of both C–C and C–I branches is lower than that of I–I branches. Areas C and E, which exhibit a strong preferential orientation have the longest average length of C–C branches, which is interpreted to be due to the dominant resultant orientations, which has reduced the probability of a fault intersecting another fault. For Areas A and B where the observed fault networks do not have a single dominant orientation or even a bimodal distribution, then the average branch lengths are shorter. In these areas, the absence of a single dominant orientation increases the likelihood of a fault intersecting another fault and as result increases the number of connections per branch. There is a strong correlation ($R^2$ of 85.85%) between average branch length and 2D fault intensity (Figure 11). Area E has the lowest fault intensity and the longest average branch length, whereas Areas A and B have the highest fault intensity and the shortest average branch length. As described earlier, the faulted interval between T40 and T10 exhibits thickness variations across each of the Areas A–E (Figure 5a). The mean thicknesses across Areas A–E vary by ±10%. Despite the limited variability in interval thickness, there are considerable variations in both fault intensity and average branch length across Areas A–E (Figure 11). For the fault network analysed here, there is significant variability in fault intensity without significant variation in the thickness of the faulted interval. It is interpreted that this variability indicates there are differences in the fault network evolution for each area. The relatively minor variations in thickness between the Areas A–E indicate that fault network properties do not fundamentally vary as a result of minor thickness variations. It is interpreted that the dominant control on the connectivity of the fault system is the prevalence of single or bimodal fault orientations, which has reduced the probability of connections.

6 | DISCUSSION

6.1 | Controls on fault spacing and orientation

The layer-bound faults interpreted in this study show a system of normal faults dominated by conjugate pure shear
The orderly variations in faults spacing and dip direction suggest that fault growth is not random but related to a systematic control at the kilometre scale. The differences in coefficient of variance could be indicative of contemporaneous development of the faults. Though we do not have another dataset for comparison, qualitative assessment of other published polygonal fault systems indicates that this is not common. For example, Soliva et al. (2006) observed similar horst–graben configurations within an overall normal distribution, but these geometries only accounted for a very small proportion of the whole population. Here, the analysis indicates that horst and graben fault blocks account for up to ca. 50% of the total population. The analogue models of Victor and Moretti (2006) also exhibited similar horst–graben fault geometries where they identified that normal faults-bounding polygonal rafts appeared regularly spaced. Vétel et al. (2005) suggested that the systematic spacing of horst–graben pairs and the exponential spacing distributions in the Turkana Rift indicate the presence of a dominating kilometre scale structural control complicated by randomly distributed smaller faults.

Outcrop and analogue models of extensional faulting in layered systems have shown that typically a single growing fault reaches a point where it cannot accommodate the progressive increase of strain without further generation of faults (Bahroudi et al., 2003; Benedicto et al., 2003; Soliva et al., 2005). In the fault system described here, there are areas with a dominance of large, unfaul ted, horst blocks surrounded by narrow grabens. It is postulated that in these areas, horizontal shortening due to volume loss related to compaction of the Miocene to Maastrichtian interval was accommodated without reaching saturation with respect to fault intensity (e.g. Ackermann et al., 2001). The style of deformation observed is consistent with that of a frictional overburden uniformly extended above a frictional detachment, as described in analogue models by Bahroudi et al. (2003). Their analogue models identified that grabens were wider above a ductile detachment than above frictional detachments. The properties of substrates are discussed later. The geometry of the areas dominated by conjugate faults is akin to the geometry of faults identified in gravity spreading systems (e.g. Schultz-Ela, 2001), which suggests that the lower most part of the Maastrichtian to Campanian (Mira and Withnell formations) interval marks a transition from brittle to ductile behaviour. It is interpreted that during early fault growth, the polygonal rafts were separated by narrow graben structures, and the increase in isochron thickness within the grabens could indicate syn-depositional faulting. For the conjugate pairs of faults forming horsts and grabens, it is likely the faults formed coevaly and intersect at their tips to form a polygonal pattern. Since that the network of polygonal grabens in Areas A, B and E has no preferred orientation, it is reasonable to interpret that the apparent extension, and ultimately strain also has no preferred orientation.

### TABLE 2

Summary statistics for the topological analysis for the five sample areas A–E

| Sample area | X% | Y% | Z% | X–Y | Y–Z | Z–X | Mean thickness (TWT) | Average branch length | C–C | C–I | I–C | I–I | C–I |
|-------------|----|----|----|-----|-----|-----|----------------------|----------------------|-----|-----|-----|-----|-----|
| Area A      | 70.14 | 2.30 | 27.57 | 32.23 | 19.38 | 11.53 | 1.13422               | 1.57                 | 451 | 471 | 443 | 455 | 485 |
| Area B      | 63.57 | 1.09 | 35.33 | 37.77 | 11.53 | 25.49 | 1.26902               | 1.21                 | 495 | 513 | 471 | 465 | 485 |
| Area C      | 74.93 | 1.18 | 23.98 | 37.73 | 11.53 | 25.49 | 1.00975               | 1.70                 | 474 | 486 | 539 | 534 | 537 |
| Area D      | 60.68 | 1.55 | 37.77 | 37.73 | 11.53 | 25.49 | 1.32646               | 1.98                 | 486 | 524 | 524 | 524 | 524 |
| Area E      | 58.33 | 2.50 | 39.17 | 39.17 | 11.53 | 25.49 | 1.3722                | 1.73                 | 521 | 541 | 521 | 521 | 521 |
In the areas with a dominant orientation, the volumetric compaction strain must also have a preferred orientation. Since layer-bound fault systems are typically restricted to a stratigraphic interval, it can be hypothesized that the mechanical thickness of the interval exhibits an important control on fault intensity. The analysis here indicates a significant variability in fault intensity for an approximately uniform thickness. As has been suggested previously (e.g. Ackerman et al., 2001), areas where faults are regularly spaced could represent a saturated state, where the fault spacing stopped evolving.

6.2 Controls on connectivity within layer-bound fault systems

The layer-bound fault system is pervasive across the survey area in the T40 to K60 interval. The topological analysis and the variations in fault orientation indicate that the development of the fault system and interactions between neighbouring faults vary considerably. Across the five sample areas, the observed differences in fault system topology suggest that their connectivity is influenced by the local stress orientations affecting the development of preferential fault trends which can impact the probability of faults intersecting one another. For example, Area C shows the fewest connections per branch interpreted as the result of the dominance of a single fault orientation. This dominance indicates that there is likely a significant local variation in the stress field from isotropic to anisotropic. In contrast, Areas A and B exhibit a more typical polygonal pattern, indicating an isotropic stress field. They also have fewer I–I branches and a shorter average branch length than Area C. Area E can be considered a hybrid area, with the south-west corner showing a strong alignment, while the north east corner shows greater variance in strike. Both fault orientation and fault intensity are factors in the connectivity fault networks, since increases in fault density and/or the range of orientations increase the chances of faults connecting (e.g. Sanderson & Nixon, 2018).

Morley and Binazirnejad (2020) described that topology offers a way to describe polygonal fault network complexity, and when combined with other observations may also help discriminate between the origins of different complex fault types. Detailed displacement and throw distributions by Stuevold et al. (2003) of a fault system from offshore Norway showed that the faults are highly segmented with irregular and rapid variations in displacement. They interpreted that an important control on these variations was the intersection geometry in the fault system. Future work should look to carry out an integrated topological and displacement analysis for layer-bound fault systems, as has been done for tectonic faults.
(e.g., Duffy et al., 2017; Morley & Nixon, 2016) to attempt to characterize the growth history of both individual and intersecting branches. Advances in automated fault extraction and subsequent methods to quantify displacement (e.g., Hale, 2013) could further the understanding of displacement–intersection relationships in layer-bound fault systems. While the analysis presented here does not definitively confirm that different sample areas have a different genesis, it is difficult to reconcile that the contrasting topologies, orientations and geometries interpreted across the fault system can be explained by a single evolution. Polygonal areas typified by Area A, which shows no preferred orientation, are topologically the most complex due to the high proportion of C–I and C–C branches, also have a high proportion of conjugate pairs forming horsts and grabens. These characteristics probably represent systematic development of a polygonal fault system with very limited external forcing, such as pre-existing structures (e.g., Hansen et al., 2004) and stratigraphic dip (e.g., Ireland et al., 2011) or the influence of regional stresses. In Area C, where the same T40 to K60 interval is faulted, there is an absence of conjugate pairs, a dominant orientation and a high proportion of I–I branches. This area has most likely been influenced by stratigraphic dip and regional stress. As would be expected, where a single orientation of fault dominates the likelihood of C–C branches and abutments or splays (Y nodes or T nodes) decreases. The variation in branch-type length may also reveal details of the genesis and growth of the fault system. The mean fault length of I–I branches is always greater than C–C branches, and this could indicate that C–C fault branches are the last fault segments to grow and their trace length is restricted by faults which have already initiated. The very low proportion of cross-cutting faults (X branches) suggests that in the case of different evolutionary stages, adjacent faults act as boundaries to fault propagation in the layer-bound fault systems described here.

Layer-bound fault systems have been proposed as pathways for fluid migration within sedimentary basins (e.g., Berndt et al., 2003; Hoffmann et al., 2019). It has been suggested previously (Stuevold et al., 2003) that intersections in polygonal fault systems are likely to be sites for preferential fluid communication. The factors that determine the role played in fluid migration pathways by these fault systems do not differ fundamentally from the processes that control the likelihood of other faults acting as migration pathways. Fault intersections have been recognized as important conduits for focused fluid flow (e.g., Sibson, 1996). In the case of fault networks with a single dominant orientation, for example, Area C, there are fewer fault intersections, and therefore the likelihood of zones of increased permeability for cross-stratal flow is reduced. In contrast, all areas, which has more than single dominant fault orientation, have a greater number of fault intersections, and therefore the likelihood of zones of increased permeability for cross-stratal flow is increased. For flow parallel to the fault plane, then in the case of anisotropic fault orientations, the resultant permeability could be increased in the same orientation. In areas with conjugate faults and more frequent X, Y or T nodes, extensional shear fractures may develop in pipe-like conduits and act as preferential pathways for cross-stratal fluid flow (Sibson, 2000). Cartwright et al. (2007) identify that polygonal fault systems, and more generally, faults embedded in a sealing sequence, are important geological features which enable seal bypass. The connectivity of a polygonal fault system is likely important in understanding the potential for cross-stratal fluid flow. Observed variations in the topology of layer-bound fault systems may strongly influence fluid flow and may be expressed in the resulting patterns of fluid expulsion features through otherwise low permeability sequences (Waghorn et al., 2018). Future studies focusing on areas where both layer-bound fault systems and vertical fluid escape features are present could look to quantify the relationship between topology, fault connectivity and fluid flow and escape features.

6.3 | Role of mechanically weak layers and detachments

In analogue models with a mobile substrate or detachment at the base (akin to evaporites or shales), the formation of polygonal fault patterns has been attributed to brittle 3D extension imposed by lateral flow of the ductile substrate (Victor & Moretti, 2006). The succession of sediment here is dominated by the fine-grained calcareous and clay-rich formations. Ductile deformation of clay-rich sediments has been observed in different geological settings and less frequently ductile deformation within these intervals has been indicated by S–C foliations reported in core (Takizawa & Ogawa, 1999). In the models of Victor and Moretti (2006), normal faults-bounding polygonal fault blocks appear regularly spaced in all of the experiments and are localized on top of silicone ridges rising from the basal silicone layer. Similar structures geometrically have been attributed to the rise of salt diapirs or shale diapirs are described by various authors (Cohen & McClay, 1996; Morley et al., 1998; Rowan et al., 1999; Vendeville & Jackson, 1992). The convergence of conjugate faults in a polygonal fault system in Lake Hope is described by Wattrson et al. (2000), who focus on the mechanisms which account for the observed folding and faulting geometries. The geometries were best explained by the presence of an underlying décollement (or detachment).

In the study area, the available well information and 3D seismic data inform us of the precise lithology of the T10 stratigraphic horizon. However, the pattern of deformation, and in particular the horst–graben geometries, indicates that
this interval is mechanically weaker and deforms differently from the T40 to T10 interval. The Dockrell and Walcott formations that make up the T40 to T10 interval are described as having a greater proportion of calcareous and biogenic components compared with the underlying (T10 to K60) Mira and Withnell formations, which has a greater clay component. The increased proportion of clay in the interval below the T10 horizon could make the formation weaker and thus prone to faults detaching into it. During shallow burial, the lithologies that are commonly affected by layer-bound fault systems are likely to be capable of deforming in a ductile manner. Furthermore, argillaceous sediments may exhibit transitional brittle–ductile behaviour which favours shear failure under high differential stresses and ductile shear under small differential stresses (Dehandschutter et al., 2005). Variations and contrasts in the rheology and mechanical properties of fine-grained sequences, as well as their stress state sequence, may favour particular layers behaving in a more ductile manner and therefore acting as a mobile substrate or detachment (e.g. Ireland et al., 2011). Additionally, the presence of listric faults in the system may qualitatively indicate that the faults initiate in the layer exhibiting more brittle behaviour and subsequently flatten down towards ductile substrate after further extension (e.g. Ellis & McClay, 1988). The interpretation that the listric geometry is the result of fault growth and detachment into a mechanically weaker unit is favoured, as it is consistent with the previous explanation of fault spacing. The interval into which the faults detach could be either a narrow zone a few metres thick, or a diffuse zone perhaps several hundred metres thick; however, this is not resolved by the current data. Further work should investigate how thick a mechanically weaker interval needs to be to act as a detachment and accommodate the individual fault displacements for layer-bound fault systems. Though subsequent compaction of the already faulted interval cannot be ruled out (e.g. Neagu et al., 2010), given that there is no systematic or pervasive fault plane flattening, it is not considered a dominant control on the observed fault geometries, and it is worth noting that Neagu et al. (2010) were unable to rule out the potential role of detachments at the lower tips in the fault plane flattening they observed.

7 CONCLUSIONS

This is the first time that a systematic spacing and linked variations in topology have been described for layer-bound fault systems. The systematic spacing of conjugate faults suggests that the volume loss due to compaction can be accommodated without reaching saturation with respect to fault intensity. Variations in orientation and topology indicate that local stress perturbations can directly influence the development and resulting topology of layer-bound faults systems. These findings are consistent with normal faults in other vertically restricted systems. The findings have implications for the mechanics of not only polygonal faults, but vertically restricted normal faults more broadly. Variations in the topology and geometry of layer-bound fault systems may strongly influence fluid flow through otherwise low permeability sequences and are therefore important for understanding the sealing integrity of overburden sequences for the geological storage of carbon dioxide and hydrogen. The observations and characteristics of the fault systems described provide valuable insights for the modelling and prediction of sub-seismic faults and fractures. To date, existing studies have looked to ascribe a single genesis to explain the formation of layer-bound or polygonal faults systems (see Cartwright et al., 2003; Goulty, 2008). However, the observations and analysis here demonstrate that the fault spacing and topology are inherently linked to the local geological setting (e.g. Davies et al., 2009), suggesting that the growth and potentially the genesis of layer-bound fault systems may be location specific. Further investigations from a wider sample of fault systems in different sedimentary basins are needed to examine whether the spacing, geometry and topology of layer-bound fault systems can help distinguish faults with different genesis and growth histories.

ACKNOWLEDGEMENTS

Seismic data were provided by Geoscience Australia and used under Open Access license. Data were interpreted using Schlumberger Petrel software provided under academic license. The interpretations used in this study are made available through data.ncl.ac.uk, this includes horizon surfaces and fault traces (https://doi.org/10.25405/data.ncl.14195204). We thank Thilo Wrona, Casey Nixon and Joe Cartwright for their thorough reviews which improved the manuscript. Early on this work benefitted from discussions with Chris Jackson and Craig Magee.

CONFLICT OF INTEREST

There are no conflicts of interest.

DATA AVAILABILITY STATEMENT

The 3D seismic data that this study is based on are openly available from Geoscience Australia at https://www.ga.gov.au/nopims. The interpretations used for the topological analysis are available from data.ncl.ac.uk.

PEER REVIEW

The peer review history for this article is available at https://publons.com/publon/10.1111/bre.12582.

ORCID

Mark T. Ireland https://orcid.org/0000-0001-9777-0447
REFERENCES

Ackermann, R. V., Schlische, R. W., & Withjack, M. O. (2001). The geometric and statistical evolution of normal fault systems: An experimental study of the effects of mechanical layer thickness on scaling laws. Journal of Structural Geology, 23, 1803–1819. https://doi.org/10.1016/S0191-8714(01)00028-1

Axen, G. J. (1988). The geometry of planar domino-style normal faults above a dipping basal detachment. Journal of Structural Geology, 10, 405–411. https://doi.org/10.1016/0191-8714(88)90018-1

Bacon, M., Simm, R., & Redshaw, T. (2007). 3-D seismic interpretation. Cambridge University Press.

Bahroudi, A., Koyi, H. A., & Talbot, C. J. (2003). Effect of ductile and frictional decoulements on style of extension. Journal of Structural Geology, 25, 1401–1423. https://doi.org/10.1016/S0191-8141(02)00201-8

Barnes, A. E. (2016). Handbook of poststack seismic attributes. Society of Exploration Geophysicists.

Barnett, J. A., Mortimer, J., Rippon, J. H., Walsh, J. J., & Watterson, J. (1987). Displacement geometry in the volume containing a single normal fault. AAPG Bulletin, 71, 925–937.

Benedicto, A., Schultz, R. A., & Soliva, R. (2003). Layer thickness and the shape of faults. Geophysical Research Letters, 30, 2076–2079. https://doi.org/10.1029/2003GL018237

Berndt, C., Bünz, S., & Mienert, J. (2003). Polygonal fault systems on the mid-Norwegian margin: A long-term source for fluid flow. Geological Society, London, Special Publications, 216(1), 283–290. https://doi.org/10.1144/GSL.SP.2003.216.01.18

Berndt, C., Bünz, S., & Mienert, J. (2003). Polygonal fault systems on the mid-Norwegian margin: A long-term source for fluid flow. Geological Society, London, Special Publications, 216(1), 283–290. https://doi.org/10.1144/GSL.SP.2003.216.01.18

Bilal, A., McClay, K., & Scarselli, N. (2020). Fault-scarp degradation in the central Exmouth Plateau, North West Shelf, Australia. Geological Society, London, Special Publications, 476(1), 231–257.

Bishop, A. W., Green, G., Garga, V. K., Andresen, A., & Brown, J. (1971). A new ring shear apparatus and its application to the measurement of residual strength. Geotechnique, 21, 273–282. https://doi.org/10.1680/geot.1971.21.4.273

Bour, O., & Davy, P. (1998). On the connectivity of three-dimensional fault networks. Water Resources Research, 34, 2611–2622. https://doi.org/10.1029/98WR01861

Breteron, N. R. (1992). Physical property relationships from Sites 765 and 766. In F. M. Gradstein, J. N. Ludden & Shipboard Scientific Party (Eds.), Proc. ODP, Sci. Results, 123, Ocean Drilling Program.

Cartwright, J. (2007). The impact of 3D seismic data on the understanding of compaction, fluid flow and diagenesis in sedimentary basins. Journal of the Geological Society, 164, 881–893. https://doi.org/10.1144/0016-764906-143

Cartwright, J., Huuse, M., & Aplin, A. (2007). Seal bypass systems. AAPG Bulletin, 91, 1141–1166. https://doi.org/10.1306/04090705181

Cartwright, J., James, D., & Bolton, A. (2003). The genesis of polygonal fault systems: A review. Geological Society, London, Special Publications, 216, 223–243. https://doi.org/10.1144/GSL.SP.2003.216.01.15

Cartwright, J. A., & Lonergan, L. (1996). Volumetric contraction during the compaction of mudrocks: A mechanism for the development of regional-scale polygonal fault systems. Basin Research, 8, 183–193. https://doi.org/10.1046/j.1365-2117.1996.01536.x

Chopra, S., & Marfurt, K. J. (2008). Emerging and future trends in seismic attributes. The Leading Edge, 27, 298–318. https://doi.org/10.1190/1.2896620

Chopra, S., & Marfurt, K. (2012). Seismic attribute expression of differential compaction. CSEG Recorder 37.

Cohen, H. A., & Mcclay, K. (1996). Sedimentation and shale tectonics of the northwestern Niger Delta front. Marine and Petroleum Geology, 13, 313–328. https://doi.org/10.1016/0264-8172(95)00067-4

Davies, R. J., Ireland, M. T., & Cartwright, J. A. (2009). Differential compaction due to the irregular topology of a diagenetic reaction boundary: A new mechanism for the formation of polygonal faults. Basin Research, 21, 354–359. https://doi.org/10.1111/j.1365-2177.2008.00389.x

Dehandschutter, B., Vandycke, S., Sintubin, M., Vandenberghe, N., & Wouters, L. (2005). Brittle fractures and ductile shear bands in argillaceous sediments: Inferences from Oligocene Boom Clay (Belgium). Journal of Structural Geology, 27, 1095–1112. https://doi.org/10.1016/j.jsg.2004.08.014

Driscoll, N. W., & Karner, G. D. (1998). Lower crustal extension across the Northern Carnarvon basin, Australia: Evidence for an eastward dipping detachment. Journal of Geophysical Research: Solid Earth, 103(B3), 4975–4991. https://doi.org/10.1029/97JB03295

Duffy, O. B., Nixon, C. W., Bell, R. E., Jackson, C.-L., Gawthorpe, R. L., Sanderson, D. J., & Whipp, P. S. (2017). The topology of evolving rift fault networks: Single-phase vs multi-phase riffs. Journal of Structural Geology, 96, 192–202. https://doi.org/10.1016/j.jsg.2017.02.001

Ellis, P., & Mcclay, K. (1988). Listric extensional fault systems—results of analogue model experiments. Basin Research, 1, 55–70. https://doi.org/10.1111/j.1365-2117.1988.tb00005.x

Ghalayini, R., Homberg, C., Daniel, J. M., & Nader, F. H. (2017). Growth of layer-bound normal faults under a regional anisotropic stress field. Geological Society, London, Special Publications, 439, 57–78. https://doi.org/10.1144/SP439.13

Golty, N. R. (2008). Geomechanics of polygonal fault systems: A review. Petroleum Geoscience, 14, 389–397. https://doi.org/10.1144/1354-079038-781

Golty, N. R., & Swarbrick, R. E. (2005). Development of polygonal fault systems: A test of hypotheses. Journal of the Geological Society, 162, 587–590. https://doi.org/10.1144/0016-764905-004

Hale, D. (2013). Methods to compute fault images, extract fault surfaces, and estimate fault throws from 3D seismic images. Geophysics, 78, O33–O43. https://doi.org/10.1190/geo2012-0331.1

Hansen, D. M., Shimmel, J. W., Williamson, M. A., & Lykke-Andersen, H. (2004). Development of a major polygonal fault system in Upper Cretaceous chalk and Cenozoic mudrocks of the Sable Subbasin, Canadian Atlantic margin. Marine and Petroleum Geology, 21, 1205–1219. https://doi.org/10.1016/j.marpetgeo.2004.07.004

Higgs, W., & Mcclay, K. (1993). Analogue sandbox modelling of Miocene extensional faulting in the Outer Moray Firth. Geological Society, London, Special Publications, 71, 141–162. https://doi.org/10.1144/GSL.SP.1993.071.01.07

Hoffmann, J. J., Gorman, A. R., & Crutchley, G. J. (2019). Seismic evidence for repeated vertical fluid flow through polygonally faulted strata in the Canterbury Basin, New Zealand. Marine and Petroleum Geology, 109, 317–329. https://doi.org/10.1016/j.marpetgeo.2019.06.025

Ireland, M. T., Golty, N. R., & Davies, R. J. (2011). Influence of stratigraphic setting and simple shear on layer-bound compaction faults offshore Maurtania. Journal of Structural Geology, 33, 487–499. https://doi.org/10.1016/j.jsg.2010.11.005

Jing, L., & Stephansson, O. (1997). Network topology and homogenization of fractured rocks. In B. Jamtveit & B. W. D. Yardley (Eds.),
Fluid flow and transport in rocks (pp. 191–202). Springer. https://doi.org/10.1007/978-94-009-1533-6_11

Li, J., & Mitra, S. (2017). Geometry and evolution of fold-thrust structures at the boundaries between frictional and ductile detachments. Marine and Petroleum Geology, 85, 16–34. https://doi.org/10.1016/j.marpgeo.2017.04.011

Lonergan, L., Cartwright, J., & Jolly, R. (1998). The geometry of polygonal fault systems in Tertiary mudrocks of the North Sea. Journal of Structural Geology, 20, 529–548. https://doi.org/10.1016/S0191-8141(97)00113-2

Longley, I. M., Buessenschuett, C., Clydsdale, L., Cubitt, C. J., Davis, R. C., Johnson, M. K., Marshall, N., Murray, A. P., Somerville, R., Spry, T. B., & Thompson, N. B. (2002). The North West shelf of Australia—a woodside perspective. The Sedimentary Basins of Western Australia, 3, 27–88.

Manzocchi, T. (2002). The connectivity of two-dimensional networks of spatially correlated fractures. Water Resources Research, 38, 1-1-1-20. https://doi.org/10.1029/2000WR000180

Morgan, D. A., Cartwright, J. A., & Imbert, P. (2015). Perturbation of polygonal fault propagation by buried poikomarks and the implications for the development of polygonal fault systems. Marine and Petroleum Geology, 65, 157–171. https://doi.org/10.1016/j.marpgeo.2015.03.024

Morley, C. K., & Binazirnejad, H. (2020). Investigating polygonal fault topological variability: Structural causes vs image resolution. Journal of Structural Geology, 130, 103930. https://doi.org/10.1016/j.jsg.2019.103930

Morley, C. K., Crevello, P., & Ahmad, Z. H. (1998). Shale tectonics and deformation associated with active diapirism: The Jerudong Anticline, Brunei Darussalam. Journal of the Geological Society, 155, 475–490. https://doi.org/10.1144/gsjgs.155.3.0475

Morley, C. K., & Nixon, C. W. (2016). Topological characteristics of simple and complex normal fault networks. Journal of Structural Geology, 84, 68–84. https://doi.org/10.1016/j.jsg.2016.01.005

Neagu, R. C., Cartwright, J., & Davies, R. (2010). Measurement of diagenetic compaction strain from quantitative analysis of fault plane dip. Journal of Structural Geology, 32, 641–655. https://doi.org/10.1016/j.jsg.2010.03.010

Nixon, C. W., Nierland, K., Rotevatn, A., Dimmen, V., Sanderson, D. J., & Kristensen, T. B. (2020). Connectivity and network development of carbonate-hosted fault damage zones from western Malta. Journal of Structural Geology, 141, 104212. https://doi.org/10.1016/j.jsg.2020.104212

Nixon, C. W., Sanderson, D. J., & Bull, J. M. (2012). Analysis of a strike-slip fault network using high resolution multibeam bathymetry, offshore NW Devon UK. Tectonophysics, 541, 69–80.

Nugraha, H. D., Jackson, C.-.L., Johnson, H. D., Hodgson, D. M., & Reeve, M. T. (2019). Tectonic and oceanographic process interactions archived in Late Cretaceous to Present deep-marine strataigraphy on the Exmouth Plateau, offshore NW Australia. Basin Research, 31, 405–430. https://doi.org/10.1111/bre.12328

Nyberg, B., Nixon, C. W., & Sanderson, D. J. (2018). NetworkGT: A GIS tool for geometric and topological analysis of two-dimensional fracture networks. Geosphere, 14, 1618–1634. https://doi.org/10.1130/GES01595.1

O’Brien, D. K., & Manghani, M. H. (1992). Physical properties of Site 762: A comparison of shipboard and shore-based laboratory results. In U. von Rad, B. U. Haq, et al. (Eds.), Proceedings of the Ocean Drilling Program, Scientific Results, College Station, TX (Ocean Drilling Program) (Vol. 122, pp. 349–362). https://doi.org/10.2973/odp.proc.sr.122.115.1992

Peacock, D. C. P. (2002). Propagation, interaction and linkage in normal fault systems. Earth-Science Reviews, 58, 121–142. https://doi.org/10.1016/S0012-8252(01)00085-X

Rowan, M. G., Jackson, M. P., & Trudgill, B. D. (1999). Salt-related fault families and fault welds in the northern Gulf of Mexico. AAPG Bulletin, 83, 1454–1484.

Sanderson, D. J., & Nixon, C. W. (2015). The use of topology in fracture network characterization. Journal of Structural Geology, 72, 55–66. https://doi.org/10.1016/j.jsg.2015.01.005

Sanderson, D. J., & Nixon, C. W. (2018). Topology, connectivity and percolation in fracture networks. Journal of Structural Geology, 115, 167–177. https://doi.org/10.1016/j.jsg.2018.07.011

Schultz-Ela, D. D. (2001). Excursus on gravity gliding and gravity spreading. Journal of Structural Geology, 23, 725–731. https://doi.org/10.1016/S0191-8141(01)00004-9

Seebeck, H., Tenthorey, E., Consoli, C., & Nicol, A. (2015). Polygonal faulting and seal integrity in the Bonaparte Basin, Australia. Marine and Petroleum Geology, 60, 120–135. https://doi.org/10.1016/j.marpgeo.2014.10.012

SEG. (2019). Dictionary: Polarity standard [Online]. SEG. https://wiki.seg.org/wiki/Dictionary:Polarity_standard

Sibson, R. H. (1996). Structural permeability of fluid-driven fault-fracture meshes. Journal of Structural Geology, 18, 1031–1042. https://doi.org/10.1016/0191-8141(96)00032-6

Sibson, R. H. (2000). Fluid in involvement in normal faulting. Journal of Geodynamics, 29, 469–499. https://doi.org/10.1016/S0264-3707(99)00042-3

Soliva, R., & Benedicto, A. (2005). Geometry, scaling relations and spacing of vertically restricted normal faults. Journal of Structural Geology, 27, 317–325. https://doi.org/10.1016/j.jsg.2004.08.010

Soliva, R., Benedicto, A., & Maerten, L. (2006). Spacing and linkage of confined normal faults: Importance of mechanical thickness. Journal of Geophysical Research, 111. https://doi.org/10.1029/2004J B003507

Soliva, R., Schultz, R. A., & Benedicto, A. (2005). Three-dimensional displacement-length scaling and maximum dimension of normal faults in layered rocks. Geophysical Research Letters, 32. https://doi.org/10.1029/2005GL023007

Stewart, S., & Argent, J. (2000). Relationship between polarity of extensional fault arrays and presence of detachments. Journal of Structural Geology, 22, 693–711. https://doi.org/10.1016/S0191-8141(00)00004-3

Stuevold, L. M., Faerseth, R. B., Arnesen, L., Cartwright, J., & Möller, N. (2003). Polygonal faults in the Ormen Lange field, More basin, offshore mid Norway. Geological Society, London, Special Publications, 216, 263–281. https://doi.org/10.1144/SP.2003.216.01.17

Takizawa, S., & Ogawa, Y. (1999). Dilatant clayey microstructure in the Barbados décollement zone. Journal of Structural Geology, 21(1), 117–122.

Tindale, K., Newell, N., Keall, J., & Smith, N. (1998). Structural evolution and charge history of the Exmouth Sub-basin, northern Curnarvon Basin, Western Australia. Petroleum Exploration Society of Australia (PESA).

Vendeville, B. C., & Jackson, M. P. (1992). The rise of diapirs during thin-skinned extension. Marine and Petroleum Geology, 9, 331–354. https://doi.org/10.1016/0264-8172(92)90047-I
Vétel, W., Le Gall, B., & Walsh, J. J. (2005). Geometry and growth of an inner rift fault pattern: The Kino Sogo Fault Belt, Turkan Rift (North Kenya). *Journal of Structural Geology*, 27, 2204–2222. https://doi.org/10.1016/j.jsg.2005.07.003

Victor, P., & Moretti, I. (2006). Polygonal fault systems and channel boudinage: 3D analysis of multidirectional extension in analogue sandbox experiments. *Marine and Petroleum Geology*, 23, 777–789. https://doi.org/10.1016/j.marpetgeo.2006.06.004

Waghorn, K. A., Pecher, I., Strachan, L. J., Crutchley, G., Bialas, J., Coffin, R., Davy, B., Koch, S., Kroeger, K. F., Papenberg, C., & Sarkar, S. (2018). Paleo-fluid expulsion and contouritic drift formation on the Chatham Rise, New Zealand. *Basin Research*, 30, 5–19. https://doi.org/10.1111/bre.12237

Watterson, J., Walsh, J., Nicol, A., Nell, P., & Bretan, P. G. (2000). Geometry and origin of a polygonal fault system. *Journal of the Geological Society*, 157(1), 151–162.

Woodside. (2011). Moyet-1 final well completion report.

Wrona, T., Magee, C., Jackson, C. A., Huuse, M., & Taylor, K. G. (2017). Kinematics of polygonal fault systems: Observations from the northern North Sea. *Frontiers in Earth Science*, 5, 101. https://doi.org/10.3389/feart.2017.00101

**How to cite this article:** Ireland, M. T., Morley, C. K., & Davies, R. J. (2021). Systematic spacing and topological variations in layer-bound fault systems. *Basin Research*, 33, 2745–2762. https://doi.org/10.1111/bre.12582

---

**APPENDIX A**

**FIGURE A1** Crossline XL4600 (a) and Inline IL3600 (b) seismic sections, showing the sections used for scan line analysis. See Figure 1d for location.