Mechanical Stratigraphy Controls Normal Fault Growth and Dimensions, Outer Kwanza Basin, Offshore Angola

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Abstract Mechanical stratigraphy controls the growth patterns and dimensions of relatively small normal faults, yet how it influences the development of much larger structures remains unclear. Here, we use 3D seismic reflection data from the Outer Kwanza Basin, offshore Angola to constrain the geometry and kinematics of several normal faults formed in a deep-water clastic succession. The faults are up to 6.3-km long and 1.9-km tall and have up to 44 m of throw. Aspect ratios and lower-tip throw gradients are greater for faults that terminate downward at a c. 100 m thick, mass-transport complex (MTC; up to 5.2 and 0.12) than for those that offset it (up to 2.7 and 0.01). Faults that offset the MTC invariably have >30 m of throw. Based on their geometric properties and throw patterns, we interpret that the faults nucleated above the MTC and propagated down toward it. Upon encountering this unit, which we infer behaved in a more ductile manner than encasing strata, tip propagation was halted until tip stresses were sufficiently high (corresponding to minimum throw of c. 30 m) to breach it. Faults with smaller throw were unable to breach the MTC. We argue that using only geometric criteria to determine fault growth patterns can mask the significant control mechanical stratigraphy has on fault kinematics. Mechanical stratigraphy is therefore a key control on the growth of large, seismic-scale normal faults, in a similar way to that observed for far smaller structures.

1. Introduction

Understanding the geometry and growth of normal faults is crucial for a range of geoscience disciplines. For example, these structures control landscape evolution and sediment transport pathways in areas of continental extension (e.g., Childs et al., 2003; Eliet & Gawthorpe, 1995; Gawthorpe & Hurst, 1993; Gawthorpe & Leeder, 2000; Gibbs, 1984; Leeder & Gawthorpe, 1987; Mcclay, 1990; Peacock & Sanderson, 1994; Trudgill & Cartwright, 1994), the magnitude and recurrence interval of potentially hazardous earthquakes (e.g., Cowie & Scholz, 1992b; Soliva et al., 2008; Walsh et al., 2002; Wilkinson et al., 2015), and the occurrence and viability of hydrocarbon and geothermal resources (e.g., Athmer & Luthi, 2011; Bodvarsson et al., 1982; Corbel et al., 2012; Fairley & Hinds, 2004; Rotevatn et al., 2007; Serié et al, 2017).

Fault growth models were originally derived from global compilations of displacement (D) and length (L) data. More specifically, the recognition that D and L were positively correlated over several-orders-of-magnitude in the form D = cL^n, where c is a constant and n is between 1 and 1.5, led to the development of the so-called “propagating” fault model, in which faults grow via a broadly synchronous increase in their displacement and length (e.g., Cartwright et al., 1995; Nicol et al., 2005, 2020; Rotevatn et al., 2018; Walsh et al., 2003; Walsh & Watterson, 1988). More recent observations, primarily from 3D seismic reflection data imaging sedimentary basins, support an alternative model for normal fault growth, the so-called “constant-length” fault model. In this model, faults rapidly establish their near-final length before accumulating significant displacement (e.g., Rotevatn et al., 2018; Walsh et al., 2002). Rotevatn et al. (2018) also propose a hybrid growth model in which rapid fault lengthening by tip propagation, and ultimately segment linkage (characteristics most consistent with the propagating fault model) principally occurs during the initial 20–30% of the fault lifespan (Stage 1), with this period followed by a second, more prolonged period of displacement accrual, during which time the fault does not significantly lengthen (characteristics more characteristic of the constant-length model; Stage 2).

Fault geometry and growth can be controlled by the so-called “mechanical stratigraphy” of the host rock, although this is not explicitly illustrated in global D–L datasets, which incorporate faults formed in a wide range of lithologies. The term “mechanical stratigraphy” encompasses: (i) the varying material properties of rock strata (e.g., measured properties such as compressive and tensile strengths, Young’s modulus), (ii) the thicknesses of
the mechanical layers, and (iii) the character and frictional properties of the transitions or boundaries between mechanical layers (e.g., sharp vs. gradational boundaries, and smooth vs. rugose contacts; cf., Ferrill & Morris, 2008; Ferrill et al., 2017; Groshong, 2006; Laubach et al., 2009). In this context, “strong,” “stiff,” or “competent” units (e.g., igneous rock, cemented sandstone or carbonate) tend to resist deformation, maintain bed length and thickness during deformation, accommodate little deformation before brittle failure, or have low ductility, whereas “weak,” “soft,” or “incompetent” strata (e.g., mudstone, salt) tend to deform relatively easily, change bed length and thickness during deformation, accommodate significant deformation before brittle failure, or have high ductility (see Ferrill et al., 2017). Within this framework, the vertical and lateral propagation of faults that nucleate in “strong” layers can be impeded vertically and laterally as they approach, deform, and mechanically interact with “weak” layers, a situation common in heterogeneous sedimentary sequences. This behavior can result in anomalous fault aspect ratios (i.e., the ratio between maximum fault height and length) and a deviation from the expected \( D-L \) relationship (Soliva et al., 2005a; Soliva & Benedicto, 2005; Wilkins & Gross, 2002). For example, a fault may preferentially propagate along-strike (i.e., laterally) in competent layers, thereby restricting its height and increasing its aspect ratio, as well as limiting its ability to accumulate displacement; such faults may thus appear “under-displaced.” Short- and long-range stress interactions between salt bodies and faults (e.g., Tvedt et al., 2013), and between neighboring faults (e.g., Dawers & Anders, 1995; Martel, 1999; Peacock, 2002; Peacock & Sanderson, 1991; Soliva et al., 2006), can also influence fault growth and geometry. To-date, however, most studies have focused on the control of mechanical stratigraphy on the geometry and growth of relatively small normal faults (i.e., displacements of <1- and <100-m long; e.g., Soliva et al., 2005a; Soliva & Benedicto, 2005; Wilkins & Gross, 2002), and it is not clear if and how this control might scale with increasing mechanical unit thickness and fault size.

In this study, we use high-quality 3D seismic reflection data acquired by Compagnie Générale de Géophysique (CGG) from the Outer Kwanza Basin, offshore Angola to quantify the geometry of and displacement patterns on several normal faults (Figures 1 and 2). These data allow us to assess how host rock mechanical properties, inferred from seismic facies analysis, influence fault growth and ultimate geometry, thereby allowing us to test fault growth models. These faults occur in a mudstone host rock, within which mechanical layering is imposed by a regionally developed mass-transport complex (MTC). We reveal marked differences in the geometry and kinematics of the faults, and explore the role of host rock heterogeneity and local mechanical interactions on their growth patterns. We show that the aspect ratios and displacement patterns vary on even closely spaced faults forming in the same host rock. We explore why faults can deviate from \( D-L \) scaling laws as they grow, even when they presently have \( D-L \) scaling relationships that broadly adhere to global compilations \( \left( D = cL^2 \right) \); Figure 2). Our study shows that \( D-L \) scaling (i.e., geometric) data alone do not allow us to determine the style of fault growth; detailed kinematic analysis using growth strata (e.g., Childs et al., 2003; Jackson & Rotevatn, 2013; Tvedt et al., 2013), and the analysis of faults of varying scales within a single population, together provide more robust kinematic constraints.

2. Data and Methods
2.1. Data

The study area is imaged by a high-quality, pre-stack, depth-migrated (PDSM) 3D survey with an aerial extent of 2915 km² (Figure 1). The inline and crossline spacing for the survey is 25 m, with a vertical sampling interval of 2 m. The data are normal polarity (a peak indicates a downward increase in acoustic impedance) and zero phase, with a vertical resolution of c. 6 m at the seabed and c. 30 m at a depth of 3.5 km, approximated by a quarter of the dominant wavelength of the data (assuming average P-wave velocity of 2,500 m/s; see discussion by Evans & Jackson, 2019). In the absence of well-data, the lithology and absolute ages of the mapped seismic-stratigraphic units are constrained by using seismic facies analysis, seismic-stratigraphic principles, and by correlating their bounding horizons with three regionally mapped horizons (the Top Albian, Top Eocene and Top Miocene) presented by Hudec and Jackson (2004) (see also Hudec & Jackson, 2003, 2004; Jackson et al., 2005; Jackson & Hudec, 2008; Marton et al., 2000; Serie et al., 2017; Spathopoulos, 1996). The ages of other mapped horizons are inferred relative to these regional surfaces and documented geological events (e.g., periods of continental uplift and related salt tectonics). We use these age estimates to infer at which stratigraphic levels fault throw maxima occur, which may relate to the depth at which the faults nucleated (e.g., Hongxing & Anderson, 2007).
2.2. Seismic Interpretation and Fault Characterization

We study a c. 70 km² area that is located between a c. 10 km elongate ridge of salt (i.e., salt wall) and a sub-circular dome of salt that has pierced the overlying strata (i.e., salt stock), the latter only being partially imaged along the western edge of the seismic survey (Figures 1 and 3). Within this area we focus on 16, exceptionally well-imaged normal faults that form part of a larger array (Figures 4 and 5). These faults are also selected because we can comprehensively map numerous (18) horizons in their flanking host rocks and, therefore, generate 3D throw strike-projections that allow us to constrain throw distributions across their surfaces (e.g., Baudon & Cartwright, 2008; Duffy et al., 2015; Walsh et al., 2003). We accurately mapped the faults on closely spaced (c. 50 m) seismic sections trending normal to fault strike. We therefore constrain their three-dimensional geometry, including their aspect ratio (i.e., length/down-dip height), with reasonable precision (Figure 6; Nicol et al., 1996; Polit et al., 2009). Where fault-related folding of the host rock is present adjacent to the fault surfaces, the cutoffs are extrapolated to remove the effects of this ductile or continuous deformation (e.g., Hongxing & Anderson, 2007; Jackson et al., 2017; Walsh & Watterson, 1987).

Shallow stratigraphy (c. 0–300 m below seabed) in the study area contains deep-water channels that incise up to 200 m into underlying stratigraphy (Figure 7). This makes it difficult to confidently and regionally map...
near-seabed growth sequences that are now eroded (i.e., F6 and 7; Figure 7). Since the upper tips of many of the faults have been eroded by channels, their present aspect ratios may be slightly higher than their original aspect ratios. We follow the method of Walsh and Watterson (1991) to construct throw–distance (t–x) plots that show how throw varies along-strike on a given horizon, as well as strike-projections that show the throw distribution across the entire fault surface (e.g., Figure 8) (Collanega et al., 2018; Duffy et al., 2015; Torabi et al., 2019; Tvedt et al., 2013). We also calculate throw gradients (i.e., change in throw/distance on fault surface over which the change occurs) across the fault surface, noting that relatively high gradients (>0.15) may reflect retardation of fault tip propagation due to mechanical interactions with adjacent faults or host rock layers (e.g., Nicol et al., 1996; Walsh & Watterson, 1988). To determine where and when faults intersected the free surface during their growth, we calculate the expansion indices (EI) of packages of growth strata. EI is calculated by dividing the hanging-wall thickness of the package by its adjacent footwall thickness (e.g., Jackson et al., 2017; Tvedt et al., 2013). Finally, we define down-dip fault height (H) as the (sub)vertical distance from the fault upper to lower tip, and fault length (L) as the maximum (sub)horizontal distance between the fault’s lateral tips (Figure 6). Using these geometric data we determine fault aspect ratio (fault length/down-dip height of fault), which again can be used to infer whether the growth of a fault has been inhibited by mechanical interactions with adjacent faults and/or changes in host rock mechanical properties (Nicol et al., 1996; Soliva et al., 2005b).

3. Geological Setting

3.1. Outer Kwanza Basin

The formation of the Inner and Outer Kwanza basins, offshore Angola initiated in the Early Cretaceous when rift pulses between the African and South American plates, and ultimately the breakup of Gondwana, resulted in the opening of the South Atlantic Ocean (Brice et al., 1982; Hudec & Jackson, 2004; Jackson et al., 2005; Serié et al., 2017; Figure 1).

During the Early Cretaceous, intermittent marine incursions resulted in restricted marine conditions and the deposition of thick, Aptian-Albian evaporites in the Inner and Outer Kwanza basins (e.g., Bate et al., 2001; Karner and Gambôa, 2007). Flow of this evaporite sequence from the Late Cretaceous until present controlled the post-breakup structural evolution of the Kwanza basins (Hudec & Jackson, 2004; Jackson & Hudec, 2008; Marton et al., 2000). For example, basinward tilting of the margin during the Late Cretaceous initiated the first post-salt phase of extension, causing the formation of predominantly seaward-dipping, salt-detached normal faults (Jackson et al., 2005). Post-salt deposition was initially dominated by relatively fine-grained deep-water sediments of the lobe Group (Hudec & Jackson, 2003, 2004; Jackson & Hudec, 2008; Serié et al., 2017).

The second phase of significant post-salt extension occurred during the Neogene, triggered by onshore craton uplift (Hudec & Jackson, 2004; Jackson et al., 2005). Basement-involved uplift of the continental shelf and upper slope in the Miocene increased the rate of seaward translation of salt and its overburden, from c. 0.2 mm/yr to 1 mm/yr (Evans & Jackson, 2019; Hudec & Jackson, 2004; Jackson & Hudec, 2008, Jackson et al., 2005). This resulted in an additional 20 km of extension in the Outer Kwanza Basin (Evans &
Jackson, 2019; Hudec & Jackson, 2003; Jackson & Hudec, 2008). This regional extension may have driven the faulting studied here, although local extensional stresses triggered by coeval diapir rise and minibasin subsidence may also have played a role. Since the Miocene, sedimentation in the Outer Kwanza Basin has been dominated by hemipelagic clays and silts, with coarser-grained sediments being deposited in the basin in turbidite-fed channels and lobes (Howlett et al., 2020; Serié et al., 2017). Seismically chaotic, sharp-based, low-amplitude bodies of unknown composition are also observed in the post-Miocene sedimentary sequence; these are geophysically distinct from the fine- and coarse-grained slope deposits described above, which are typically characterized by moderate-to-high amplitude, continuous seismic reflections (Figure 3) (see Howlett et al., 2020; Serié et al., 2017). Based on their seismic expressions and the deep-water depositional setting, we suggest that these bodies represent mass-transport complexes (MTCs; Moscardelli & Wood, 2008; Wu et al., 2019). We lack borehole data to constrain their lithology, but we tentatively suggest that they may be mudstone-dominated, based on the fact that the post-Miocene slope of Angola comprise very fine-grained rock types (e.g., Fraser et al., 2005).

Figure 4. (a–d) Depth structure maps of the Intra-Pliocene, Top Miocene, Top Eocene, and Top Albian surface, respectively. The temporal and spatial presence of each of the studied faults is also highlighted on each of the structure maps. The Top Albian surface is not offset by any of the faults and shows the location of seismic sections for Figure 5.
4. Results

The normal fault array studied here sits within the axis of a N-trending mini-basin (Figures 3 and 4). The largest faults dip to the north and offset Cenozoic stratigraphy (Figures 3 and 4). Within the fault array we can define three distinct fault groups based on their geometrical characteristics and inferred kinematic behavior.

4.1. Planar Faults (F1–F5)

4.1.1. Observations

Planar faults are developed in the north of the study area, striking E-W and dipping 54–58° (Figures 4 and 5). Striking orthogonal to the trend of adjacent salt wall, all of the faults dip northwards, with the exception of a single antithetic fault that dips southward (F1). The faults are 750–950 m tall and have maximum trace lengths of 1.8–4 km (Figure 9). The maximum throw in this group is ~20 m, with the throw maxima occurring in Pliocene stratigraphy (Figures 8 and 10). The upper tips of the planar faults are in Pliocene stratigraphy and their lower tips are located in Miocene stratigraphy (Figure 5). F1 and F2 are quasi-elliptical, whereas F3 and F4 are elliptical (Figures 8b–8e). The throw maxima for F1 and F2 are located near their centers, with throw ultimately decreasing toward their tips (Figures 8b and 8c). In contrast, throw maxima for F3 and F4 are slightly offset toward their lower halves (Figures 8d and 8e). This difference in the position of their throw maxima is reflected in their lower-tip throw gradients. For example, F1 and F2 show a relatively low throw gradient of 0.01, whereas F3 and F4 display throw gradients that are an order of magnitude greater (i.e., 0.12 and 0.11, respectively). The aspect ratio for planar faults with a throw maximum occurring along the Top Miocene horizon (i.e., F3 and F4) ranges from 3.5 to 5.2, whereas the ratio for those with a throw maximum in shallower, Pliocene stratigraphy is 2.7 (F1 and 2; Figures 8 and 10). The lower tips of faults that exhibit relatively high aspect ratios (>3.5), and which have maximum throws of <20 m, are located at the top of a laterally extensive, seismically chaotic unit that we interpret as an MTC (F3 and F4; Figure 11a; Moscardelli & Wood, 2008).

Only F1, which has the largest throw and shallowest upper tip of all the planar faults, is associated with growth strata. In this case, Pliocene growth strata have an EI of 1.2 (Figure 10). Stratigraphic packages adjacent to all of the other faults in the planar fault group have EI < 1.1 making it difficult to determine if these structures were ever growth faults that breached the free surface, or whether they were always buried (e.g., Baudon & Cartwright, 2008; Jackson et al., 2017).

4.1.2. Interpretations

Their quasi-elliptical geometry, relatively low aspect ratio, and concentric throw distribution suggest F1 and F2 grew as mechanically unrestricted structures (cf. F3 and F4 below; see also Figure 8; Nicol et al., 1996). Figure 11a shows that the lower tips of F3 and F4 coincide with the MTC. Furthermore, these faults have aspect ratios greater than those of F1 and F2. Given that these relatively closely spaced faults formed in similar host rock in response to broadly similar driving stresses, one plausible interpretation for the greater aspect ratios of F3 and F4 is that their basal tips mechanically interacted with the MTC near the base of the faulted sequence. This could have resulted in their tips being pinned, inducing the preferential lengthening of these faults and thus giving rise to the development of relatively high
aspect ratios (Figures 9, 11a, and 12). This interpretation is supported by the observation that the throw maxima for F3 and F4 are slightly offset toward their lower tips, and that lower-tip throw gradients are higher for these faults than F1 and F2 (Figures 8d and 8e). These geometrical characteristics are consistent with enhanced strain accumulation in these areas as a function of vertical tip restriction during fault propagation (Figures 8 and 12; Wilkins & Gross, 2002).

4.2. Bifurcating Faults (F6–F7)

4.2.1. Observations

These faults are the longest in the studied array (>5 km). They are characterized by a single fault segment at depth and two segments within shallower stratigraphy (F6 and F7; Figures 4, 5, and 13). Bifurcating faults (F6 and F7) strike WNW-ESE and dip 54–60° northward (Figures 4 and 5). The bifurcating faults contrast with the planar faults in two key ways; (i) they have larger throws (>30 m; Figure 11) and (ii) their basal tips lie beneath rather than above the MTC (Figures 3 and 13).

F7 bifurcates upwards with a sub-horizontal, c. 1.2 km long branchline coinciding with the top of the MTC (Figures 4, 5, and 13b). Maximum throw on F7 (c. 32 m) is also located near the top of the MTC and the throw gradient below the MTC is relatively low (0.08; Figure 13b). The maximum down-dip height and trace length of F7 are 1.75 and 6.4 km respectively, resulting in an aspect ratio of 3.6 (Figures 9 and 13b).

At the structural level of the Pliocene, F6 consists of two parallel, overlapping fault segments separated by a narrow (~70 m) relay zone (Figures 4 and 5). At depth these segments link and form a single fault surface (Figures 4c, 5, and 13a). Similar to F7, the c. 1 km long, sub-horizontal branchline for the two upper segments of F6 coincides with the top of the MTC (Figure 13a). In contrast to F7, the throw maximum for F6 occurs beneath (rather than above) the MTC, within Oligocene-Miocene stratigraphy (Figure 13a). The distribution of throw on F6 defines a broad, U-shaped (depth) throw maximum that encircles the relay zone and its associated throw minimum, and which extends below the MTC and branchline (blue area; Figure 13a). Relatively high throw gradients of up to c. 0.2 occur in the relay zone where the faults overlap (sharp transition from blue to white; Figure 13a). F6 has a maximum down-dip height and length of 1.9 and 6.3 km, respectively, resulting in an aspect ratio of 3.3.
4.2.2. Interpretations

The position of the maximum throw on F7 at the top of the MTC suggests the fault: (i) nucleated at this structural level and/or (ii) nucleated elsewhere, possibly within the shallower stratigraphy, with the position of maximum throw migrating downwards with time. Irrespective of where the fault nucleated, it was able to propagate through the MTC (Figure 13b). This observation, coupled with the distinct decrease in throw and throw gradient immediately below the MTC, suggest that the MTC may have initially retarded downward or upward propagation of F7, causing strain to accumulate along the top of the MTC before the fault could propagate onward (Figures 13b and 14a). We therefore suggest that in a similar way to F3 and F4, the tips of smaller, precursor segments were initially vertically restricted along the top of the MTC (Figure 14a). However, unlike in the case of F3 and F4, the accumulation of additional displacement meant tip stresses around F7 became high enough to allow the fault to breach the MTC (Figure 14a; Wilkins & Gross, 2002).

In contrast to F7, we suggest F6 nucleated below (rather than above) the MTC based on the fact that maximum throw on the structure occurs at least partly below the MTC (Figure 13a). Having nucleated below the MTC,
F6 was then able to propagate upwards toward and ultimately through the MTC, locally bifurcating as it did (Figure 14b). The throw distribution presently observed on F6 is consistent with numerically modeled throw distributions where a bifurcating planar frontal segment is present (Figure 13a; Soliva et al., 2008). We therefore suggest that the high throw values broadly encircling the relay zone, which is also associated with relatively high throw gradients (0.2), are as a result of tip interaction between the overlapping shallow segments during bifurcation and upward growth (Figures 13a and 14b) (Nicol et al., 1996, 2017; Soliva et al., 2008).

### 4.3. Arcuate Faults (A1–A9)

#### 4.3.1. Observations

The group is characterized by faults defined by smooth, along-strike changes in their orientation, resulting in a bow-shaped, convex trace in map-view (Figure 4). This group of faults are concentrated in the south of the study area, on the northern rim of a locally deep part of the minibasin on the west of the salt wall; this deep area is best expressed at the Pliocene and Miocene structural level (Figures 4a and 4b, respectively). Close to the salt wall these faults strike E-W, perpendicular to the salt-sediment interface, changing to a NE-SW strike ≥2 km away from the salt wall; this change in strike defines their plan-view geometry, with the faults following the overall shape of the edge of the deep part of the minibasin (Figures 4a and 4b). Maximum throws on arcuate faults decrease with increasing distance from the salt wall; for example, A4 has a maximum throw of 44 m, whereas A9 has a maximum throw of ~10 m at 0.8 and 3.5 km distance from the salt wall, respectively (Figure 15a; see also Figures 4a and 4b). We also note that arcuate faults appear overdisplaced, achieving similar maximum throws at only 60%–70% of the strike length of other fault types (Figure 15b). For example, A3 is 2.6 km long and has c. 35 m maximum throw, whereas F6 has comparable throw (c. 38 m) but is significantly longer (6.3 km; Figure 15b). Arcuate faults also have notably lower aspect ratios (i.e., 1.4–2.7) compared to the linked and planar faults (i.e., 2.7–5.2; Figure 9).

![Figure 9](image9.png)

**Figure 9.** Fault height–length data for all fault groups within the studied array. Overall, a trend is observed of fault height increasing proportionally with fault length. However, faults which tip out down dip at the Top MTC surface and have high aspect ratios (F3–4) show increasing fault length without a proportional increase in fault height (red), as predicted for vertically restricted faults (Figure 12).

![Figure 10](image10.png)

**Figure 10.** A seismic section of the position of unrestricted faults (F1–2) within the stratigraphy. Their lower tips do not interact with the MTC layer. A t–z plot for F1 where the EI on the upper tip is c. 1.2 (left). t–z plots (F1–4) showing that restricted faults exhibit throw maxima in Miocene stratigraphy and unrestricted faults exhibit throw maxima in Pliocene stratigraphy (right). Data courtesy of CGG Multi-Client.
We can sub-divide the arcuate fault group based on the depth at which they occur, their detailed geometry, and their overall throw distribution. The first subgroup faults strike broadly E-W (i.e., similar to the overall trend of the fault array) and are at greater depths (i.e., below −1,900 m) than the second subgroup (A1-4; Figures 4a–4c). They exhibit down-dip heights of 1–1.9 km and maximum lengths of 1.5–4.7 km, yielding aspect ratios of 1.4–2.4 (Figure 9). The largest faults in this subgroup, A3 and A4, have maximum throw values of 36 and 44 m, respectively, and offset the MTC, whereas those with throws <30 m do not (Figures 11b and 11c and 15).

The second subgroup occurs at a shallower depth, in post-Miocene stratigraphy (A5-9; Figures 3 and 4a). The upper tips of faults in this subgroup are located just 100 m beneath the seabed and they typically tip-out ∼150 m above the Top Miocene horizon (Figure 16). Four faults in this subgroup dip toward the south, toward the center of the deep part of the minibasin described above, whereas one dips to the north (Figure 4). A5–9 have down-dip heights which range between 400 and 670 m and maximum lengths of 930–1,730 m, yielding aspect ratios of 2.1–3.9 (Figure 9). The maximum throw on these faults are generally less than that observed on A1–4 (i.e., maximum throws of ∼10–15 m; Figure 15a).

Figure 11. (a) A seismic cross-section showing the tips of F3–4 tipping out downdip on top of the MTC surface and the adjacent bifurcating faults (F6–7) offsetting the Top MTC surface. Above, the maximum throw of the respective faults are displayed and the proposed accumulated strain needed to overcome the threshold for onward growth through the MTC (gray box). (b) Throw-length data for the fault array displaying that faults with maximum throws >30 m offset the MTC, while faults with throw values <30 m which interact with the MTC are vertically restricted. $t-x$ regression analysis for the array's data also suggests the constant length model for fault growth [i.e., $T > 0$ where $L = 0$]. (c) $t-x$ plots for faults restricted by the MTC (F2–3) and some which achieve onward propagation (F6–7, A4). Data courtesy of CGG Multi-Client.

Figure 12. A sketch diagram portraying the effects of vertical restriction by a mechanical layer (MTC) during fault growth and its predicted effect on fault geometrical data as a result of growth with preferential lengthening. Adapted from Soliva et al. (2005b).
Figure 13. Strike projections for the birfurcating faults (F6 and F7), displaying the geometry of, and throw distribution on, the fault surfaces with the Intra-Pliocene, Top Miocene, Top MTC and Top Eocene surfaces, respectively with depth. (a) F6: Maximum throw lies beneath the Top MTC surface and relay branchline (blue area). Overall, greater throw values encircle the relay zone which is associated with a throw minimum (white area). High throw gradients (0.2) are present at the transition from the fault surface to the relay zone (sharp transition from blue to white area). (b) F7: Maximum throw lies coincident with the Top MTC surface on the main fault surface. Overall, greater throw values are present along strike at the structural level of the Top MTC surface.

Figure 14. Sketch diagrams of proposed growth histories for bifurcating faults. (a) F7: Individual precursor fault segments become vertically restricted by the regionally extensive MTC during downward propagation. As a result of this restriction precursor faults preferentially lengthen via segment linkage and the throw maxima migrate downwards as strain accumulation occurs at the restricted tip. Eventual strain accumulation overcomes the MTC restricting influence and continues downward propagation. (b) F6: Similarly to F7, an initial vertical restriction results in strain accumulation and throw maxima migration toward the MTC. Following overcoming the MTC’s restricting influence, bifurcation at the MTC boundary occurs.
4.3.2. Interpretations

Since the arcuate faults closely parallel the outer rim of the deep part of the minibasin (Figure 4) we suggest they formed to accommodate outer arc-style bending and stretching of the minibasin-fill during differential subsidence and salt withdrawal (Figure 16); geometrically similar faults are observed in association with mobile mud withdrawal (Stewart, 2006). We propose these faults became overdisplaced for two reasons. First, the larger faults in the first subgroup (A1–4) terminate very near the salt-sediment interface suggesting that these faults became restricted laterally due to not being able to propagate into the weak salt (e.g., Tvedt et al., 2013). As a result, these faults become overdisplaced as they continued to accumulate throw. Second, the shallower second subgroup are more closely spaced than other faults within the extensional array and are sub-parallel (Figure 4a). We propose that these faults experienced mechanical interactions between their fault tips, meaning they could not propagate laterally and as a result of continued displacement accumulation, became overdisplaced. While the cause (i.e., fault interaction) for fault restriction differs to that of the larger arcuate faults (i.e., salt interaction), the result (i.e., the formation of overdisplaced faults) is the same (e.g., Crider & Pollard, 1998; Martel, 1999; Nicol et al., 2017).

5. Discussion and Implications

5.1. Origin and Kinematics of Planar Normal Faults: The Case for Tip Restriction by Host Rock Properties

The occurrence of throw maxima within Miocene and Pliocene strata suggest the planar faults nucleated within Neogene stratigraphy (Figure 10; Hongxing & Anderson, 2007). Whereas all other planar faults never reached the free surface, F1 represented a surface-breaching growth fault during at least the Pliocene (Figure 10). During the Middle Miocene, salt and overburden translation rates increased significantly from c. 0.2 mm/yr to 1 mm/yr within the Outer Kwanza Basin, principally due to margin uplift (Evans & Jackson, 2019; Hudec & Jackson, 2004). Depending on the structural position on the margin, this resulted in extensional (i.e., typically in updip, proximal locations) or contractional (i.e., typically in downdip, distal locations) deformation of salt and its overburden (Evans & Jackson, 2019; Hudec & Jackson, 2004). Despite our study area being in the present-day transitional domain (Figure 1), increases in downdip salt flux could have resulted in local extension of this area due to outer-arc bending and translation across base-salt relief (Evans & Jackson, 2019). The planar faults may therefore have formed in the Middle Miocene in response to this increased rate of basinward translation of salt and its overburden, eventually breaching the seabed in the Pliocene, as evidenced by the presence of the growth strata in the hanging wall of F1 (Evans & Jackson, 2019).

Field-based studies and modeling results show that mechanical stratigraphy can influence the growth and ultimate geometry of normal faults; for example, multi-layered sequences defined by different rock competencies controls, down-dip linkage, fault restriction, and fault aspect ratio (e.g., Schöpfer et al., 2006; Soliva & Benedicto, 2005;
Wilkins & Gross, 2002). For example, Nicol et al. (1996) suggest vertically restricted faults display greater aspect ratios. More recent work proposes various mechanisms by which vertical restriction might occur, showing that faults growing in mechanically more homogenous host rocks assume circular to elliptical shapes throughout their lives, whereas those in multi-layered sequences can have their vertical growth impeded, thus growing from a circular to an elliptical geometry, resulting in an increase in aspect ratio (Figure 12) (Soliva & Benedicto, 2005; Soliva et al., 2005a, 2006). One consequence of the latter style of fault growth is the development of high throw gradients at restricted (vertical) tips (Figures 8d and 8e; Wilkins & Gross, 2002). Within the planar fault group, F3 and F4 have relatively high aspect ratios (3.5 and 5.2, respectively) and high throw gradients (0.12 and 0.11, respectively) at their lower tips; these values are an order-of-magnitude greater than those of F1 and F2 (0.01). The migration of throw to the restricted tip and consequent high throw gradients is shown by the presence of their throw maxima being slightly offset toward their lower tips, which is characteristic of mechanically bound faults (Figures 8d, 8e, and 10; e.g., Soliva & Benedicto, 2005; Soliva et al., 2005b; Wilkins & Gross, 2002). In contrast, relatively closely (<1 km) spaced faults within the bifurcating fault group, which have throw maxima of >30 m and that offset the MTC, do not have these geometric characteristics (Figure 11). This pattern, where only faults with \( T_{\text{max}} > 30 \) m offset the MTC, can be observed in the arcuate fault group in the extensional array (Figures 11b). An important observation is that despite being <1 km from F3 and 4 and although they offset similar stratigraphy (Figure 5), the lower tips of F1 and 2 lie above the MTC; as a result, these faults do not have the same high aspect ratios or throw gradients at their lower tips (Figures 5, 7, and 9).
The variable geometric relationships between the faults and the MTC suggests that it acted as a mechanical layer, restricting the downward and upward propagation of the fault tips and overall enlargement of the surfaces of faults that accumulated <30 m of throw (Figure 11b). This resulted in higher aspect ratios and lower tip throw gradients on those faults not able to propagate, as well as influencing the growth, geometry and throw distribution of bifurcating faults (Figures 12 and 14). Because drilling data are not available to determine the MTCs lithology or mechanical properties, we do not know if this unit retarded tip propagation by: (i) acting as a relatively stiff layer that was hard to strain or (ii) acting in a ductile manner, with strain being internally distributed in more diffuse manner. The latter case has been observed in field-based studies of normal faults, mechanically layered host rocks, with faults seemingly nucleating and propagating only in more competent, stiff layers (typically >0.5-m thick; Soliva et al., 2005b). Underlying and overlying, more compliant layers (<0.5-m thick) behave in a ductile manner and are characterized by diffuse strain and high strain gradients. Faults essentially decouple in these layers, with high aspect ratios and underdisplaced fault populations arising as a function of “forced” lateral propagation within the stiff layers. We propose that this case may also apply at a substantially larger, seismic-scale, such as that documented here. Deep water MTC’s are predominantly mudstone-dominated (Wu et al., 2019) meaning that it could be the case that they act in a ductile manner with strain being more diffuse, as documented in mechanically heterogeneous faulted sequences (e.g., Bürgmann et al., 1994; Nicol et al., 1996; Roche et al., 2012; Soliva & Benedicto, 2005; Soliva et al., 2005a, 2005b; Wilkins & Gross, 2002). In the absence of drilling data, we propose that either of these hypotheses could be the case for the MTC’s mechanical control on fault growth.

5.2. Kinematics of Bifurcating Faults: Overcoming Tip Restriction and Continued Fault Growth

Fault linkage and bifurcation control throw distributions on normal fault surfaces (e.g., Dawers & Anders, 1995; Mansfield & Cartwright, 2001; Peacock, 2002; Peacock & Sanderson, 1991; Soliva & Benedicto, 2004; Soliva et al., 2008). We noted that the laterally restricted branchline of F7 occurs at the top of the Miocene MTC, and that the throw maximum also lies at this structural and stratigraphic level. The throw maximum is restricted to the larger, rearward (i.e., footwall side) fault and does not continue onto the frontal (i.e., hangingwall side) splay (Figure 13b). In Section 5.1 we discussed that the vertical restriction of the lower tips within the planar fault group resulted in high throw gradients (Jackson et al., 2014; Wilkins & Gross, 2002). If a similar process occurred during growth of the fault F7, which bifurcates upwards, it could follow that pinned precursor segments nucleated above the MTC as physically separate structures, with high stressing building-up at their lower tips before they were able to link and propagate onward, resulting in the present throw pattern (Figures 13b and 14a). The same process could occur for a segmented fault system nucleating below and ultimately propagating upwards through a mechanical layer (Wilkins & Gross, 2002).

An alternative interpretation for the throw patterns on F7 is based on our observation that the main, rearward segment accommodates a laterally extensive, sub-horizontal throw maxima, whereas the frontal splay has overall lower throw values and lacks a discrete throw maximum (Figure 13). Thus, we may interpret that the main fault surface nucleated above the MTC and started to grow before the splay, becoming temporarily mechanically pinned along the MTC, resulting in the development of throw maxima before propagating further downwards (Wilkins & Gross, 2002; Figure 13). The splay segment, which also nucleated above the MTC, may not have achieved the critical tip stresses required for onwards propagation, instead linking down-dip with the main fault surface near the top of the MTC (Soliva & Benedicto, 2005; Soliva et al., 2005b; Wilkins & Gross, 2002; Figure 13).

Bifurcating fault F6 shows a distinct concentration of high throw near the MTC (Figure 13a); this is consistent with the hypothesis of stress and strain concentration at the MTC prior to onward fault propagation (Figure 14). Moreover, a second distinct area of high throw occurs on the frontal fault of F6, immediately east of the relay zone (Figure 13a). This is consistent with throw distributions typically seen within relay zones, where tip interaction occurs between adjacent, upward-propagating segments (Freitag et al., 2017; Long & Imber, 2012; Peacock, 2002; Peacock & Sanderson, 1991, 1994; Trudgill & Cartwright, 1994). Fault bifurcation can occur as a result of stress field reorientation or non-uniformity, as well as irregular tip-line propagation due to intra-host rock mechanical heterogeneities, such as an intra-stratal MTC (Soliva et al., 2008). This would explain the bifurcation and position of the branchline on the top of the MTC (Figures 5a and 13). Our data suggest that only faults that accumulated >30 m of throw were able to propagate through the MTC; below this value, the tips were pinned due to the build-up of insufficient tip stresses to allow onward propagation (Figures 11a and 11b; Wilkins & Gross, 2002).
5.3. Normal Fault Growth and D–L Scaling Relationships

D–L scaling relationships derived from global datasets have historically been used to support a normal fault growth model by synchronous lengthening and displacement accumulation (i.e., “propagating model”; Childs et al., 2017; Rotevatn et al., 2018; see also e.g., Watterson, 1986; Walsh & Watterson, 1988; Walsh et al., 2002). More recently, however, studies using 3D seismic reflection data have integrated stratigraphic data in an otherwise purely geometrical analyses. This has given rise to an alternative fault growth model that states fault lengthening (which may include linkage of initially isolated segments) occurs before the accumulation of significant displacement (i.e., “constant-length model”; Childs et al., 2017; Jackson & Rotevatn, 2013; Nicol et al., 2020; Rotevatn et al., 2018).

We show here that host-rock heterogeneities, and near- and far-field stress regimes and their interactions, can all result in the formation of under- or over-displaced faults (i.e., faults that significantly deviate from global D–L scaling relationships and thus produce scatter within these datasets), and/or faults with anomalously low or high aspect ratios (Cowie & Scholz, 1992a; Dawers & Anders, 1995; Peacock, 2002; Soliva & Benedicto, 2005; Soliva et al., 2005a, 2005b). This is in addition to fault linkage, which typically results in underdisplaced faults likely characterized by anomalously low aspect ratios (Cartwright et al., 1995). Our study shows how detailed analysis of faults that do not reach the free surface (and thus lack growth strata) and which are of varying sizes, but which developed within the same overall tectono-stratigraphic setting, can shed light on the kinematics of fault growth; such studies can complement those focused on growth faults flanked by syn-kinematic strata (e.g., Duffy et al., 2015; Jackson et al., 2017; Tvedt et al., 2013). A key observation is that the studied faults would fall within the general scatter present in global D–L scaling datasets (Figure 2), and that this would mask the not insignificant variability in their geometric properties (i.e., D–L relationship, aspect ratio) and inferred growth patterns, and the relationship of these to local stratigraphic factors (i.e., an intra-stratal MTC).

6. Conclusions

3D seismic reflection data from the Outer Kwanza Basin, offshore Angola image a basement-decoupled normal fault population that deform a clastic-dominated, deep-water succession. These faults are up to 6.3-km long, 1.9-km tall, and have up to 44 m of throw. Aspect ratios (i.e., fault height to fault length), maximum throw, and lower-tip throw gradients vary between faults, being greater (aspect ratio up to 5.2; maximum throw <30 m; throw gradients up to 0.12) for faults that terminate downwards at an intra-stratal mass-transport complex (MTC) than those that offset it (aspect ratio up to 3.6; maximum throw >30 m; throw gradients up to 0.08). We interpret that the faults nucleated above and propagated down toward the MTC. Upon encountering this unit, which we infer was weaker than the encasing strata, tip propagation was halted until tip stresses were sufficiently high to breach it. We suggest that such stresses were reached when a fault accumulated a maximum throw of at least 30 m. The displacement–length (D–L) relationship for all faults fall within the general scatter present in global D–L scaling datasets, regardless of their geometry or growth patterns. In the absence of growth strata, D–L scaling relationships must be used with caution when studying fault kinematics, given the role mechanical stratigraphy plays in controlling fault propagation, size, finite throw (or displacement), and geometry. Mechanical stratigraphy therefore has a key control on the growth of large, seismic-scale normal faults in a similar way to that observed for far smaller structures.

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