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Dalton, TJS, Paton, DA and Needham, DT (2017) The influence of mechanical stratigraphy on multi-layer gravity collapse structures: insights from the Orange Basin, South Africa. Petroleum Geoscience, 438. pp. 211-228. ISSN: 1354-0793

https://doi.org/10.1144/SP438.4

The influence of mechanical stratigraphy on multi-layer gravity collapse structures: insights from the Orange Basin, South Africa.

Abstract

Gravity collapse structures are common features on passive margins and typically have a tripartite configuration including an up-dip extensional domain, transitional domain and a down-dip compressional domain with a common detachment underlying the system. A number of recent studies have classified these systems, yet few document the wide variation in geometry.

This study documents the gravity collapse structures of the Namibian and South African Orange Basin as these represent some of the best imaged examples globally of this important process. Firstly, we demonstrate the geometry and kinematic evolution of these systems, focussing on examples of the tripartite configuration from a typical collapse. We then highlight the significant variability in the systems structures describing features such as; portions of the system with multiple detachments; systems with stacked synchronous detachments; the temporal evolution of faults within the system. Integrating our observations from a number of sections we present a model explaining the spatial and temporal evolution of the system. This enables us to discuss likely causes of collapse structures but also, by placing it into a well constrained stratigraphic context, how the presence of both maximum flooding surfaces and early margin deltaic sequences have a fundamental control on the resulting collapse geometry.

1. Introduction

Deep water fold thrust belts (DWFTB's) and their associated extensional systems occur in many passive margin systems throughout the world and provide an excellent opportunity to study the formation and development of both extensional and compressional faults. A considerable variation in structural style of collapse systems is seen across different margins, this is generally accepted to result from differences in both the driving mechanisms for collapse and the geometry and nature of detachment surface (Rowan et al., 2004, Krueger et al., 2009, Morley et al., 2011).
Morley et al. (2011) classifies DWFTB's into two broad categories those controlled by near field stress systems created by sediment loading and differential uplift/subsidence at passive margins (Type I) and those controlled by far field stress regimes associated with active margins (Type II). Type I DWFTB's are further divided into; Type Ia; shale detachment such as those in the Orange basin and Type Ib salt detachment such as in Angola. Krueger et al. (2009) proposed they be divided similarly based upon those of active margins, caused by subduction, and passive margins similar to far and near field stress system. Krueger et al. (2009) subdivides passive margins into three categories based on the nature of their décollement; regional salt, regional shale and local non-discrete, where the local detachments are discontinuous leading to a regional décollement crossing stratigraphic levels, Figure 1.

In this study we focus on shale detachment systems that, regardless of the driving mechanism commonly comprise three domains (Figure 1); an up-dip extensional domain, dominated by normal faulting; down-dip compressional domain, composed of imbricate thrusts and folds; and a transitional domain. The transitional (sometimes referred to as the translational) domain is not referred to by all authors, but is defined as an area between the extensional and compressional domains which is either a package of largely undeformed sediments (Corredor et al., 2005, Krueger et al., 2009) or an area in which both compressional and extensional features overprint (De Vera et al., 2010, Butler and Paton, 2010). This overprint arises from shifting the location of the point of contact between the compressional and extensional domains. It is commonly difficult to resolve the internal geometry of the transitional domain because of limited seismic imaging. The basic premise of area balancing during deformation would be expected to apply to these coupled systems, however, De Vera et al.'s (2010) and Butler & Paton's (2010) studies in the Orange Basin established a miss-balance between the extension and compression domains of up to 25% in favour of extension leaving a considerable missing contractional strain component yet to be explained.

In this study we look in detail at the three domains along a typical section from the Orange Basin system, before comparing them to other portions of the same collapse structure to observe variations along strike. From these observations we construct a model to explain the temporal evolution of this important margin process. Finally we consider how the margin stratigraphy plays a critical role in the nature of the deformation and propose that this has a significant, and as to now unrecognised control on this process that occurs on many passive margins.

2. Regional Setting

The Orange Basin is the southernmost basin on the West African passive margin; forming during the break-up of Gondwana and subsequent spreading of the South Atlantic Ocean (Muntingh & Brown, 1993, Brown et al.,1995, Paton et al., 2008, Koopman et al., 2014).

It underwent significant rifting during the Upper Jurassic to the Lower Cretaceous forming graben and half grabens infilled with synrift siliclastic and lacustrine sediments (Junslagger et al., 1999; Mohammed, et al., this volume). This is followed by continental breakup in the Barremian and the establishment of a passive continental margin, onto which a thick sedimentary sequence of post-rift stratigraphy was deposited (Gerrard & Smith, 1982). The post-rift sediment thickness ranges from 3 km in the south and north and up to 5.6 km in the centre of the basin. This sediment was largely sourced from the Orange River (Paton et al., 2008) and is broadly separated into two phases; black shales and claystones, deposited during an early drift phase and a latter drift phase depositing a thick succession of interbedded heterolithics composed of shale and claystones (Figure 2). It is within the latter phase that we observe the greatest number of gravity collapse structures. Although much of the margin stratigraphy is claystone, we define the system as shale detachment dominated because the décollement surfaces are shale intervals, these correspond to maximum flooding surfaces with some identified as proven source rocks (van der Spuy et al., 2003). This Cretaceous succession underwent considerable tilting, up to 750 m in inner margin (Paton et al., 2008), at the end of the Maastrichtian producing a considerable proximal unconformity with the overlying Cenozoic sequence. Most of the Cenozoic sequence is deposited outboard into the basin and varies in thickness from 250-450 m on the margin and 500-1400 m out onto the continental slope and beyond. The hydrocarbon system sequence stratigraphy and deeper structures of the Orange Basin are well established (Light et al., 1993, Muntingh & Brown, 1993, Paton et al., 2007, 2008, Hirsch et al., 2010).

3. Data and methods

This study combines 38,480 km of vintage 2D pre-stack time migrated (PSTM) seismic released by the Petroleum Agency of South Africa with 45,386 km of 2D seismic data from Spectrum (see Figure 3) of which 24042 km is pre-stack depth migrated (PSDM) using Fast Track Kirchoff Migration. Maximum recording lengths are in the range of 7-10 seconds two way travel time (TWTT) with the vintage data being acquired between 1976 and 2012 and the Spectrum data acquired in 2012. The line spacing of this deep seismic coverage is between 8-15 km and provides unprecedented data coverage of the basin allowing us to review the entire margin (Paton et al., this volume). In this paper we present our interpretations on the Spectrum PSDM seismic lines because of the better imaging. Well data for the basin is extensive, though limited to the shelf margin, and has been used by previous studies (e.g. Muntingh & Brown, 1993, Paton et al., 2008) to define the margin stratigraphy and to define stratigraphic ages for our seismic intervals (Figure 2). Stratigraphic megasequences and associated regional and local unconformities have been identified using reflection termination, cut-offs and onlap relationships with the sequence stratigraphic system based upon Muntingh & Brown (1993).

As we focus on the detailed architecture of the syn-kinematic packages we subdivide the megasequences into sequences based on variations in seismic character that indicate changes over time in depositional or structural environment and rates. We use a unified stratigraphic system to define these sequences across the margin as presented in Figure 2. We define the packages regionally across the section as megasequences A-E and further subdivide C & D numerically in a temporal succession. However, because they are commonly isolated packages any one section may not show a complete sequence.

4. Regional Sections

We present a 160 km long East-West orientated, regional seismic profile (Figure 4) that illustrates both the main structural elements of the margin and one of the simpler collapse structures in the centre of the Basin (Figure 3). This is shown down to a depth of 6.4 km. Based on the previous regional interpretations of Paton et al. (2008) we divide it into 4 megasequences: Synrift, Late Jurassic to Hauterivian (A), Early Drift, Barremian to Aptian (B), Late Drift, Aptian to Maastrichtian (C & D) and Cenozoic (E). To aid the interpretation across the margin, including structural features and local unconformities, we divide the megasequences into seismic sequences based upon internal seismic reflection character. Although the syn-rift packages are imaged in the dataset (Paton et al., this volume), we focus on the sequences that are stratigraphically above the top syn-rift reflection. The top of the syn-rift package is delineated by a package of high amplitude parallel reflections, present across the basin, onto which Barremian aged stratigraphy is deposited. In this section the nature of the contact is represented by an aggradational sequence of reflections conformable with the top of the synrift package. It is important to note that elsewhere in the basin this boundary has a progradational relationship marked by downlapping reflections onto the synrift sediments, prior to an aggradational phase. The Late Drift megasequence is deposited conformably on the Early Drift package and is definable by its higher reflectivity.

Evident within this Late Drift megasequence are numerous unconformities represented by truncation and onlapping of reflections. These unconformities only occur off the paleo slope margin and are often restricted to fault blocks and are therefore not regional in extent. Reflections within the centre of this package are both folded and faulted. In the proximal portion of the basin westward dipping normal faults are identified by dislocated packages shifting down-dip of one another (eastern end of Figure 4) this is the extensional portion of the gravity collapse structure. Continuing westwards and down-dip the seismic character becomes increasingly chaotic and complex, which we define as the transition domain. The most distal part of the system, the compressional domain, is characterised by a series of east dipping thrust faults that are identifiable by high amplitude steeply dipping reflections that appear to be stacking packages on top of one another.

These structural features will be discussed in more detail subsequently though it is key to note at this point that while the main décollement for this collapse (red reflection in Figure 4) is broadly co-incident with the top of the Early Drift megasequence it is not a constant slope and shows significant changes in dip direction and angle. The décollements are picked based on where faults terminate identified through cut-offs and changes in the dip of reflections. The Late Drift megasequence is capped by a regional truncation at the top of the paleo shelf picked out by the green line (Figure 4), whereas these sequences are conformable at the base of the paleo-slope. The unconformity is, however, still interpretable by a change from a low to a high amplitude reflection along this boundary. The Cenozoic package is defined by a change in the location of deposition from proximal to a more distal position on the continental slope. The package is considerably thinner than the Cretaceous sequence on the central margin and thickens significantly to the west. Now we consider each of the structural elements in more detail.

4.1 Extensional Domain

In order to define structures in more detail we subdivide the megasequence into a number of sequences (A-E, Figure 2). Figure 5 shows a typical interpreted section from the upper portions of the extensional system, where the latest faults form prior to detaching onto a regional décollement. This interpretation focusses on the upper part of the extensional portion of the structure and shows a more detailed breakdown of the Late Drift megasequence (sequences C & D). These packages are defined upon internal seismic facies and reflection termination as described below and reveal relative changes in sediment supply and fault controlled accommodation space (Brown et al., 1995).

Internally sequence C3 has an absence of seismic impedance contrast resulting in limited internal geometry being imaged, but it is conformable with the high amplitude reflections at the base of sequence D1 and it is truncated by sequence D3. This suggests D1 was being deposited as the upper portions of C3 were being eroded, the discontinuity being a direct result of faulting. D3 has a thick package of high amplitude reflections allowing the tracking of several horizons internally. When restored D3 forms a westward thickening wedge. Changes in spacing between traceable horizons in D3 show slight changes in thickness along its length indicating different faults were active at different points. Of particular note are the small changes in thickness in the packages above and below the orange horizon which indicate movement of small faults in the package between two faults with far larger throws. As the largest thickness changes occur on the faults on which D4 and D7 truncate, this implies that deformation tends to concentrate onto a few larger, more widely spaced faults e.g. big faults get big stay big and stop smaller faults from growing. Sequence D4, defined by a package of low amplitude reflections, reinforces this point its presence only in the west of the section abutting a large fault plane implies it grew more rapidly at this point than faults to the east. This created a larger accommodation space which was rapidly infilled as indicated by folding of the reflections into the fault.

D3 is clearly truncated by the base of sequence D5 with a rugose contact that appears to represent the collapse of the top of the fault block. D5 has chaotic and poorly imaged reflectance that infill's the eroded section truncated at the top of D4. As shown in Figure 4, D5 extends for 16km west of the Figure 5 and continues to erode earlier fault blocks. Its chaotic seismic character and erosive base suggest it is a Mass Transport Complex (MTC). Several similar MTC's can be seen throughout the extensional portion of the collapse features (e.g. McGilvery et. al, 2004, Posamentier & Kolla, 2003).

Sequence D7 is defined by a series of reflections that onlap onto the top of D4 & D5 and are clearly imaged on the tops of the fault blocks. This implies that fault movement is outstripping sediment supply at this point. It also appears that several of the faults have switched off by this time. As the supply increases and faults switch off sediments begin to bury the fault blocks and again deformation concentrates into the larger faults where we observe some limited sediment growth into the fault plane.

Truncation of D7 by megasequence E e.g. the boundary between the Late Drift and Cenozoic megasequences, can be seen throughout the paleo continental margin and the upper paleo slope, it is unclear how much sediment has been eroded although Paton et al (2008) suggest it may have been as much as 750 m. Only the two largest faults were active after this unconformity formed although they offset it with only small throws.

4.2 Transitional Domain

The section in Figure 6 shows the point at which the extensional domain changes into the compressional domain and indicates that this occurs predominantly within megasequences C and D.

Megasequence B contains the high amplitude parallel reflections that denote the top of the Early Drift megasequence although immediately beneath the most deformed section of the transition zone these reflections becomes less distinct, probably due to signal attenuation. These reflections are sedimentalogically composed of coarsening upward silt - medium sandstone packages (Paton et al., 2007) that represent the progradation of the Aptian deltaic margin and are capped by a maximum flooding surface which forms the detachment. Sequence C1 is defined as a set of lower amplitude parallel reflections that have a variable relationship across the section with the surrounding sequences. In the east of the section they are conformable with megasequence B, reflections become truncated towards the west by the base of sequence C2. They reappear in the west as a set of reflections conformable with C2 and downlap onto megasequence B. We interpret this as a shift in the depth of the main décollement that is immediately above B in the east and cuts down to an inter mega sequence B layer with the consequence of translating the C1 package downslope by ~2300 m towards the west.

C2 is defined by low amplitude largely discontinuous reflections and is conformable with C3 which is comprised of higher amplitude more continuous reflections. The lowest reflections in C2 to the east and centre downlap onto B and C1, being directly above the detachment at this point they are conformable with C1 in the west as the detachment cuts down sequence as previously described. The division between the C2 and C3 intervals is identifiable by an easily correlatable, high amplitude reflection package. This allows us to define fault cut-offs with confidence. In the east these faults dip steeply landward with normal offsets and detach onto the main basal décollement. Progressing west they become more closely spaced and detach onto a shallow basinward dipping thrust fault located above the regional décollement. Atop this thrust fault a shift from extensional to compressional tectonics occurs. The low amplitudes at the base of C2 is likely a result of the coalescing of multiple faults at this level, causing increased stress at this depth. The thickness of C2/C3 is largely maintained throughout the margin, including the area of intense faulting, which suggests that it is largely a pre-kinematic sequence deposited prior to collapse, though some reflections in the top of C3 show limited thickening into fault planes suggesting some degree of syn-kinesis.

Sequence D1 here is defined by a set of low amplitude reflections that are largely conformable with C3; as in the previous sequence cut-offs are used to define the location of faults. Many packages have wedge like geometries that thicken into fault planes, implying fault growth during deposition making this a syn-kinematic succession. From the position of cut-offs in the region in which normal faulting gives way to thrusting it can be seen that several faults cease moving. The upper boundary is truncated by sequence D2. D2 is defined up by a series of low amplitude continuous reflections which onlap onto the erosional truncation that defined the top of D1. They form a tapering wedge to the east where the formation onlaps onto significant faults with throws of ~120 and ~250 m. Within the formation there appear to be numerous minor truncations of horizons against one another possibly due to limited deposition in what are effectively a mini basins. Some faults do persist into the base of D2 but most are truncated by it implying an erosional episode followed by progressive infill during which limited reactivation occurred causing minor folding as opposed to faulting in the overlying sequence. At this point of the section sequence D3 is defined by low amplitude continuous reflections that downlap onto D2, infilling its uneven upper surface, before latterly adopting a more aggradational geometry. Minor folding of some reflections at the western end of the section imply limited localised reactivations on some thrusts. Sequence D8 is composed of high amplitude reflections conformable with D3 the contact between these two horizons can be traced into the compressional domain down dip. The top of D8 appears conformable with megasequence E (Cenozoic).

The transition domain in the centre of Figure 6 generally picks out a large fold structure detaching onto a thrust above the regional décollement. The back-limb of the fold is cut by normal faults, which progressively become thrust faults towards the crest, some of which are likely to be inverted normal faults. The precise contact between the compressional and extensional domains (e.g. the transitional zone) is narrow similar to modes proposed by Corredor et al. (2005) and Krueger et al. (2009), however the possible inversion may imply a more complex structural style, as suggested by de Vera et al. (2010) and Butler and Paton (2010).

4.3 Compressional Domain

Figure 7 is a typical section from the distal end of the compressional domain and we have divided it into nine packages. Megasequence B, which is correlated from the transitional domain is defined by several near horizontal reflections that have a consistent increase in amplitude from east to west which probably reflects a progressive change in its petrophysical properties. Sequences C1-C3 despite being of varying amplitudes have the same geometry of stacked steeply east dipping reflections (~35º using PSDM data) that shallow and flatten with depth to become parallel with the top reflection of megasequence B. Definable packages of reflections stack with discrete cut-offs that pick out a set of imbricate thrusts. Sequence D1 here is defined by a set of discontinuous low amplitude reflections that onlap the thrust planes and downlap onto C3. Reflections are folded and have been truncated by both sequence D3 and by one another. The variation in thickness and the associated onlap onto anticlines of D1 suggest that thrusting was active during deposition of D3 and was frequently emergent leading to folding and erosion of depositing sediments in a syn-kinematic fashion. This onlap implies that deformation rates were greater than sedimentation rates during this interval. D3 truncates D1 with a set of low amplitude but continuous reflections, these reflections are folded above the underlying thrust planes but are only cut by two of the thrusts with far smaller throws. Whilst it is clear the faults remain active during this period the rate of deformation relative to sedimentation has slowed significantly.

Sequence D8 here truncates the crests of the folded reflections in D3 and onlaps in the synclines formed by the dipping reflections on the back-limb of the thrust faults. This suggests the end of deformation in this part of the compressional domain with this sediment infilling the remnant topography, though sediment supply is insufficient to entirely fill bathymetric lows. The last of these lows are filled by small onlapping packages at the base of sequence D9, which is otherwise conformable with D8. As with the transitional domain the contact between the Late Drift (D) and the Cenozoic megasequences (E) is conformable. In a broader context when viewing the compressional domain in Figure 4 the imbricates have a relatively equal spacing and become progressively less deformed heading away from the transition zone whilst also deforming ever younger sequences. The dips of the faults shallow from the transitional domain, at 40-50º, to the frontal thrusts at 15-25º. They progressively deform younger sequences implying that once thrusts dip becomes too high it is preferential to deform more distal sediments.

5. Variations in DWFTB geometry

Having summarised the structural elements that comprise a typical section for gravity collapse, we now outline how the styles of deformation deviate from this typical section by looking at variations along the margin illustrated with a number of additional sections.

5.1 Lateral variation

The three sections in Figure 8 are modified from Dalton et al. (in press) and show three slip-parallel, 35 km, sections running north-south (Figure 3, a to c), through a DWFTB in the southern portion of the Orange basin. Growth strata indicate collapse initiated during the deposition of the Cenozoic megasequence that detaches onto a maximum flooding surface at the top of the Campanian in the Late Drift megasequence (Paton et al., 2008). Section a) consists of an extensional domain with no corresponding compressional domain. Section b) has a more classical geometry with both extensional and compressional domains detached on to the Campanian décollement, however an additional set of thrusts detach onto the contact between the Cenozoic and Late Drift megasequences. Section c) indicates this upper detachment is far more developed with a separate set of normal faults detaching onto the base of the Cenozoic megasequence. Geometries of reflections in the extensional domain indicate slip occurring along both detachments synchronously, suggesting gravitational driven strain is distributed between both systems. The Campanian detachment has larger throws indicating it has taken more of the strain, though the folding of the Cenozoic reflections in the far west appears to restrict its westerly development. The imbalance between the two detachments in section c) suggest the Cenozoic detachment is more efficient, however as the system grows northwards the Campanian detachment becomes more important. This may relate to local variations in slip potential of the detachment surfaces e.g. changes in thickness, overpressure or lithology.

5.2 Multiple detachments

The presence of multiple detachment horizons in gravity collapse systems has been recognized previously (e.g. Totterdell and Krassey, 2003, Rowan et al., 2004, Corredor et al., 2005, Briggs et al, 2006), but few studies document how the position and interaction of different slip horizons creates a range of complex geometries that indicate changes in timing and location of deformation.

5.2.1 Sub Aptian Failure

The focus of previous studies of the Orange Basin collapse structures (De Vera et al., 2010, Butler et al., 2010) has been on the system that is contained within the Late Drift megasequence. Figure 9 presents a more detailed interpretation of a portion of the extensional domain of the collapse (see Figure 4 for location). We see here the main detachment in the Late Drift megasequence is underlain by a set of thrusts detaching onto the top of the Synrift. This lower detachment is formed along a maximum flooding surface between the top of the Synrift megasequence, (Hauterivian) and base of the Early Drift megasequence, (Barremian) identified by Brown et al. (1995). Folding of the upper detachment and overlying horizons by the developing underlying thrusts imply they formed later. Additionally thickening in sequence D6 into the fault immediately above the fold suggest its inception led to reactivation of this fault. The lack of significant thickening of sequence D7 suggests thrusting had largely ceased by its deposition. This suggests that whilst the upper system initiates first both systems existed coevally. The vergence of these lower thrusts is consistent with the same basinward translation as the upper system. The most proximal normal faults present in east of the section clearly penetrate into the Early Drift megasequence and are likely to link directly to these thrusts although seismic resolution prevents clear confirmation. D6 is not present above these faults (Figure 5) and may have either been eroded out or not been deposited, however the infilling of subsequent D7 into fault planes suggest these faults were most active prior to its deposition.

5.2.2 Stacked detachments

Sections through the far north of the largest collapse structure provide further insights into the multi-layer detachment systems (Figures 3 & 10). In Figure 10 a 83 km long section shows several different detachment surfaces at a number of stratigraphic intervals, picked by the identification of mutual fault terminations. In the east of the section a 30km long package of normal faults extending up to 2.5 km from a detachment layer within the Early Drift megasequence to the Cenozoic horizon. A second smaller extensional domain, which is 12 km long with faults extending vertically 600 m up from the detachment, it is down-dip and is contained entirely within the Early Drift megasequence, representative of an early phase of collapse. Two compressional domains also exist; a lower detachment in the Early Drift megasequence, along the Aptian-Barremian maximum flooding surface (Muntingh & Brown, 1993), containing thrusts penetrating up to 1.1 km into the overlying Late Drift megasequence and an upper 45 km long detachment, along the Cenomanian-Turonian maximum flooding surface (Paton et al., 2007), consisting of widely spaced thrusts extending 700 m up from the detachment, entirely within the Late drift megasequence. Reflections in the lower compressional domain demonstrate two periods of activity one synchronous with the lower extensional system, with which it shares a detachment and a later phase of reactivation leading to thrusting and folding of the Late Drift megasequence. The upper detachment may have been active synchronously with the lower compressional domain but remains active for longer as indicated by thrusting and folding of the dark green sequence. It is interesting to note that the upper system terminates at the location at which the first thrust of the lower system emerges. By altering the upper detachments slope angle at this point it may have rendered further slip inviable along it. The upper extensional system remains active throughout the Late Drift megasequence and clearly transfers considerable strain down-dip. However it is the upper compressional domain that remains active during this period which does not appear to be genetically linked at this point so the process of transmission of strain between the upper and lower compressive domains is not clear. In reviewing parallel sections no genetic link emerges, in fact the lower system disappears relatively rapidly. The transition from extensional to compressional domains along the upper detachment in this section is of a very different character to that seen in Figure 6 appearing as a zone largely deformed sediments as per Corredor et al (2005) & Krueger et al (2009).

6. Discussion

The presence of gravity collapse structures have been documented on many margins and some of the inherent variability has been well discussed (Morley et al., 1996, Rowan et al., 2004, Krueger et al. 2009, Morley et al. 2011). Studies that have focussed on thin shale detachment driven systems generally propose that they are relatively coherent bodies presenting little variation within a single system. We now discuss how the observed lateral variability in geometries we observe influences our understanding of thin shale detachment systems. We also consider the greater complexity observed in these features to synthesis a new temporal model of collapse development in the Orange Basin.

6.1 Model for the temporal evolution of a collapse structure

Variations in style and character of deformation appear consistently across the width of the Orange Basin including the spacing between thrusts, the depth and location of slip detachment surfaces and the nature of the transition zone. Although there is also considerable variation in the thickness of the Upper Cretaceous sediments across the basin the same regional detachment is present throughout. This means the changes in the styles of deformation observed are present within a single DWFTB, meaning that any single end member model is not applicable. Dalton et al. (in press) demonstrate that the extensional domain initiates prior to the formation of a later compressional phase. In this study through growth packages, we show that the earliest phase of collapse is located in the centre around the transition zone. For example D8 in Figure 6 is a post-kinematic horizon but in Figure 7 is clearly a syn-kinematic package and is entirely eroded out to the west where the overlying D9 package, here post-kinematic, becomes syn-kinematic showing later phases of movement occur progressively more distal than the last. Few sequences can be tracked throughout the entire structure as they are either truncated by later sequences or are only locally present. However analyses on the megasequence scale and of larger traceable sequences do reinforce this finding. New faults form and grow at the outer extents of the collapse, although older faults are still active with a reduction in offset. Successive younger faults form out from the transition zone, to the west down-dip in the case of the compressional domain and east up-dip in the case of the extensional domain. The high fidelity of the seismic imaging of our data shows that the transition domain represents a short-wavelength change from extensional to compressional tectonics as opposed to being a zone of overprinted regimes and more importantly appears to remain fixed. In general the position of maximum strain migrates away from the transition domain, although we do observe (Figure 9) fault reactivation occurring.

It is similarly clear that we can relate later more proximal movements to ever more distal thrusts this would reinforce the concept that these regimes preserve the original contact between them as a block of material that ceases to deform allowing translation of strain downslope. Observations of underlying thrust systems and the timing of structures above and beneath in Figure 9, indicate a synchronous relationship between the systems e.g. the overlying detachment was folded by the underlying system which remained active throughout. This infers that they are both part of a single system as opposed to being two stacked systems of different ages.

Whilst many studies make reference to multiple detachment horizons (Rowan et al., 2004, Krueger et al., 2006, Morley et al., 2011, Peel et al., 2014) their presence are generally not included in models of gravity collapse systems. Growth strata indicate that these alternative detachments are often not merely spatially and temporally separate collapses events but are linked integral portions of the same system. They thus have an important role in terms of strain distribution. They preferentially appear on more mature systems, and link to the youngest most proximal normal faults. This implies they form after a point at which continued deformation along the extant distal compressional regime is no longer as efficient as linking a lower detachment. Sequence scale observations show that these structures take a long time to form and go through multiple reactivations which control deposition and erosion along the margin. With is this in mind we have produced a model for the formation and growth of these systems in thin shale detachment systems (Figure 11).

Our model assumes that continued lateral compaction and deformation of the sediments above and ahead of the original detachment reaches a point at which it is no longer the most efficient way of accommodating the gravitationally induced stress. Assuming the underlying sediments are comparatively under-compacted and in the presence of an appropriate alternative slip horizon, strain is now accommodated along a lower décollement. However it is not clear how we transfer strain from normal faults connected to a lower system, with the strain recorded in the upper compressional domain as seen in Figure 10, where both the upper compressional regime and most proximal normal faults deform age equivalent sediments and thus must link. The extensional domain in Figure 8 c) shows two slip surfaces that have been exploited by the same faults at different times it is possible that the same relationships exist in the more mature system in Figure 10 but the continued deformation has made this relationships difficult to ascertain

Brown et al. (1995) indicates our detachment horizons are maximum flooding surfaces presumably composed of low basal friction shales which as long as they are sufficiently thick and continuous will continue to allow slip (Rowan et al., 2004). If the shale thins or is absent from a section the system will lock up. The locking up of the system while overburden builds up sufficient to lead to re-initiation of failure by overcoming frictional cohesion leads to the development of isolated sediment imbalances at the head of fault scarps (De Vera et al., 2010). This in turn leads to the formation of MTC's which rework the sediments of the upper portion of the extensional domain. This explains why we tend to see large scale MTC development only on mature systems prone to more lock-ups. They become more prevalent stepping back towards the coast where fewer shale intervals were deposited to provide potential slip surfaces on what were paleo continental margins.

Initial geometries are controlled by the original local accumulations of sediments that for amenable slip horizons, e.g. the shale with the lowest frictional cohesion will be used primarily above other slip horizons, this cohesion may however vary across the basin as per the original depositional conditions and thus better slip horizons will be used elsewhere (Dalton et al., in press). In the case of the collapse systems in this study they are commonly associated with maximum flooding surfaces or base of slope systems.

6. 2 Stratigraphic controls on margin collapse

Although the majority of the passive margin stratigraphy on the Orange Basin is claystone our observations imply that there is a strong control on the location and evolution of the collapse structures from variations in stratigraphy. The principle slip surfaces have been well documented as being relatively thin (~100 m) organic rich shale horizons (e.g. Muntingh and Brown, 1993; Paton et al., 2008) that acts as a low friction surface. This depositionally controlled variation in the basin can be related to the Krueger et al. (2009) two end member models for gravity collapse structures on shale detachments. One end member suggests slip along a single detachment horizon while in the second the detachment switches between local over pressured shale horizons as variations in depositional occurrence and slip potential allow. In the Orange Basin, as we have shown, examples of both end members are observed with the upper compressional domain in Figure 10 clearly slipping along a single regional plane whilst the easterly extensional domain has a highly undulous character suggestive of smaller localised slip horizons.

The model presented by Morley et al (2011) characterises the collapse systems within the Niger Delta and Orange Basin as being of equivalent types (Type 1a) both being detached on shale and, although there is much discussion as to the existence of shale diapirism there do appear to be distinct differences in the style of deformation between the two basins. The implications of a thick shale interval versus a thin horizon as commented on by Rowan et al. (2004), alters the nature of the failure. Critical wedge concepts (Bilotti & Shaw, 2005, Briggs et al., 2006) would assume propagation of the system oceanwards. As long as there is a low basal friction the system will continue to propagate. If there is a thick detachment layer then it will localise all of the deformation on to the basal system. For example in the Niger delta, where the Akata shale is thick it internally deforms and the whole overburden can behave as a mechanically strong unit (Corredor et al., 2005). This could cause long wavelength folding with some localised faulting (Costa and Vendeville, 2001) and would not require significant intra-stratigraphic deformation. In contrast in the Orange Basin, and other basins dominated by interbedded heterolithics with thin detachments, the mechanically strong unit above the detachment will need to undergo considerable intra-stratigraphic deformation such as folding and intra-layer thrusting to allow it to transfer strain down-dip (Dalton et al., In Press).

Our observations also show that the collapse is controlled not just by detachment thickness but also variations in margin stratigraphy. Existing stratigraphic studies of the Orange Basin (Brown et al., 1995; Paton et al., 2007, 2008) show that there are two key stratigraphic variations in the basin evolution. During the Aptian (megasequence B in this study) the stratigraphy facies comprises a landward stepping clastic front. This results in the landward migration of the delta-foreset to marine shale transition. Overlying the delta system is the main shelf margin sequence with interbedded organic rich shale horizons. This results in a complex distribution of décollement horizons and a corresponding multiphase development, which is described in Figure 12;

a) The stratigraphic distribution of the stable passive margin.
b) extensional faulting initiates on the continental slope, detaching onto an advantageous shale horizon and subsequently leading to thrusting down-dip on the abyssal plain.
c) continued gravitational imbalance on the margin leads to additional faults to form proximal and distal to the original collapse which continues itself to deform.
d) the ability of the upper detachment to redistribute strain down-dip becomes less efficient and so new extensional faults penetrate down to a lower shale horizon to compact lower relatively under compacted sediments.
e) this process continues to exploit lower shale horizons to redistribute strain, the original systems may also continue to deform though lower systems may alter the structural development of the overlying systems. The propagation of the faults to the lower packages is in part controlled by the stratigraphy of the margin, and the location of the delta-front.

Conclusions

Using very well imaged examples of gravity collapse structures from the Namibian and South African Atlantic passive margin we illustrate, the significant variation in structures that are present in these tripartitie systems. This variation includes the typical up-dip extensional faults and down-dip thrust faults but also multi-detachment faulting and folding, stacked detachments, cross-cutting and complex progressive evolution of the system.

As this system is dominated by a series of relatively thin detachments we suggest that the role of stratigraphy, especially the distribution of maximum flooding organic rich units, plays a fundamental role in both the style and spatial distribution of deformation. We propose that such a model helps to explain the differences that occur in thick shale systems, salt systems and thinly bedded heterolithic systems,

Figure Captions

Figure 1 Model of Gravitational collapse (Krueger et al., 2009); upper model describes the typical features and geometry of gravity system controlled by a regional detachment, Lower image describes the geometry where no regional décollement is present.

Figure 2 Chronostratigraphy of Orange Basin adapted from Paton et al. (2008), DWFTB's depths into which systems penetrate across the entire basin.

Figure 3 Map of Orange basin indicating the location of lines used in this study, the location lines used in previous studies (Butler et al., 2010, De Vera et al., 2010, Paton et al. 2007, 2008) and an outline representing the total data coverage used in this study.

Figure 4, PSDM un-interpreted and interpreted section of Line 1 (see Figure 3 for location) shown with a vertical exaggeration of 3:1. The colours correspond to each megasequence (Figure 2) Synrift megasequence is purple, Early Drift in blue, Late Drift in Green and Cenozoic in grey, different shades correspond to discrete packages within each. Detachments are picked out in red.

Figure 5 Detailed interpreted and un-interpreted section of extensional domain in Figure 4. Section is vertically exaggerated 3:1.

Figure 6 Detailed interpreted and un-interpreted section of transitional domain in Figure 4. Section is vertically exaggerated 3:1.

Figure 7 Detailed interpreted and un-interpreted section of the compressional domain from Figure 4, image is vertically exaggerated 3:1.

Figure 8 Three interpreted sections (a-c) from the south of the Orange Basin (Figure 3), adapted from Dalton et al. (in press). All sections are 35 km long and are presented in PSTM with vertical exaggerations of 3:1

Figure 9 Detailed interpreted and un-interpreted section from Figure 4, showing folding of the upper detachment by and lower detachment system. Section is vertically exaggerated 3:1.

Figure 10 Interpreted section taken from northern portion of the same collapse structure as featured in Figure 4. Section is vertically exaggerated 3:1

Figure 11 Multistage model of gravity collapse culminating in the formation of a lower detachment. Orange ellipses represent the distribution of strain within the DWFTB and reflect the findings of Dalton et al. (in press) that considerable compaction of the margin is required prior to the formation of compressional domain.

Figure 12 Model explaining role of deposition on the location and development of detachment horizons; a) section through a typical margin showing 3 stacked sequences two with shale horizons at the base of slope and the upper shale horizon representing a maximum flooding surface. b) development of a simple single detachment system slipping along the maximum flooding surface, c) system matures with the development of additional faults and eventually locks-up d) in response to system locking up alternative slip horizon along lower base of slope shale used instead, e) even lower detachment horizon sought as the shale in d) is depositional restricted .

References

Bilotti, F., and Shaw, J.H., 2005, Deep-water Niger Delta fold and thrust belt modelled as a critical-taper wedge: The influence of elevated basal fluid pressure on structural styles: American Association of Petroleum Geologists Bulletin, v. 89, p. 1475–1491.

Briggs, S.E., Davies, R.J., Cartwright, J.A., Morgan, R., 2006. Multiple detachment levels and their control on fold styles in the compressional domain of the deepwater west Niger Delta. Basin Res. 18, 435–450.

Brown Jr., L.F., Benson, J.M., Brink, G.J., Doherty, Jollands, E.H., Jungslagger, A., Keenen, A., Muntingh, A., van Wyk, N.J.S., 1995. Sequence Stratigraphy in Offshore South African Divergent Basins. An Atlas on Exploration for Cretaceous Lowstand Traps. In: AAPG Studies in Geology, vol. 41. Soekor (Pty) Ltd.

Butler, R.W.H., Paton, D.A., 2010. Evaluating lateral compaction in deepwater fold and thrust belts: how much are we missing from “nature's sandbox”? GSA Today 20 (3), 4–10.

Corredor, F., Shaw, J.H., Bilotti, F., 2005. Structural styles in the deep-water fold and thrust belts of the Niger Delta. AAPG Bull. 89, 735–780.

Costa, E., B.C. Vendeville, 2005, Experimental insights on the geometry and kinematics of fold-and-thrust belts above weak, viscous evaporitic décollement Journal of Structural Geology, 24, pp. 1729–1739

De Vera, J., Granado, P., McClay, K., 2010. Structural evolution of the Orange basin gravity-driven system, offshore Namibia. Mar. Petrol. Geol. 27, 233–237.

Dalton, T.J.S., Paton, D.A., Needham, D.T. and Hodgson, N. Temporal and spatial evolution of deepwater fold thrust belts: Implicaions for quantifying strain imbalance (In Press); AAPG, Interpretation

Gerrard, T., and G. C. Smith, 1982, Post-Palaeozoic succession and structure of the south-western African continental margin: Studies in continental margin geology: AAPG Memoir 34, p. 49–74.

Hirsch, K.K., Scheck-Wenderoth, M., van Wees, J.-D., Kuhlmann, G., Paton, D.A. Tectonic subsidence history and thermal evolution of the Orange Basin (2010) *Marine and Petroleum Geology*, 27 (3), pp. 565-584.

Jungslagger, E.H.A.,1999. Petroleum habitats of the Atlantic margin of South Africa. In: Cameron, N.R., Bate, R.H., Clure, V.S. (Eds.), The Oil and Gas Habitats of the South Atlantic. Special Publication, vol. 153. Geological Society, London, pp. 153–168.

Koopmann, H., Schreckenberger, B., Franke, D., Becker, K., and Schnabel, M.: The late rifting phase and continental break-up of southern South Atlantic: The mode and timing of volcanic rifting and formation of earliest oceanic crust, Geological Society, London, Special Publications, accepted, 2014

Krueger, A.C.V.A., and E. Gilbert, 2006, Mechanics and kinematics of back thrusting in deep water fold-thrust belts – Observations from the Niger Delta (abstract): 2006 AAPG Annual Meeting, Houston, Abstracts Volume, v. 15, p. 58.

Krueger, A., Gilbert, E., 2009. Deepwater fold-thrust belts: not all the beasts are equal. AAPG Search and Discovery, Article #30085. http://www.searchanddiscovery.com/documents/2009/30085krueger/index.html.

Light, M.P.R., Maslanyj, M.P., Greenwood, R.J., Banks, N.L., 1993. Seismic sequence stratigraphy and tectonics offshore Namibia. Tectonics and Seismic Sequence Stratigraphy, Geological Society Special Publications, London, vol. 71, pp. 163–191.

McGilvery, T. A., and Cook, D. L., 2004, Depositional elements of the slope/basin Depositional system Offshore Brunei, Proceedings of the Indonesia Petroleum Association Meeting, Jakarta Indonesia p. 407-420.

Mohammed, M., D. A. Paton, R. E. L. I. Collier, N. Hodgson, and U. Utjavari, Interaction of crustal heterogeneity and lithospheric process in determining passive margin architecture on the Southern Namibian margin: Geological Society of London, Special Publications, in press.

Morley, C.K., Guerin, G., 1996. Comparison of gravity-deformation styles and behaviour associated with mobiles shales and salt. Tectonics 15, 1154–1170.

Morley, C.K., King. R., Hillis, R., Tingay, M., Backe, G., 2011. Deepwater fold and thrust belt classification, tectonics, structure and hydrocarbon prospectivity: a review. Earth Sci. Rev. 104, 41-91.

Muntingh, A., Brown, L.F.J., 1993. Sequence stratigraphy of petroleum plays, post-rift Cretaceous rocks (lower Aptian to upper Maastrichtian), Orange Basin, South Africa, Siliciclastic Sequence Stratigraphy. Recent Developments and Applications. American Association of Petroleum Geologists, Tulsa, Oklahoma, pp. 71–98.

Paton, D.A., di Primio, R., Kuhlmann, G., van der Spuy, D., Horsfield, B., 2007. Insights into the petroleum system evolution of the southern Orange Basin, South Africa. South African Journal of Geology 110, 261–274.

Paton, D.A., van der Spuy, D., di Primio, R., and Horsfield, B., 2008, Tectonically induced adjustment of passive-margin accommodation space; influence on hydrocarbon potential of the Orange Basin, South Africa: AAPG Bulletin, v. 92, p. 589–609.

Peel, F.J., 2014, The engines of gravity-driven movement on passive margins: Quantifying the relative contributions of spreading vs. gravity sliding mechanisms: Tectonophysics, v.633, p.126–142, doi:10.1016/j.tecto.2014.06.023

Posamentier, H. W., and V. Kolla, 2003, Seismic geomorphology and stratigraphy of depositional elements in deep-water settings: Journal of Sedimentary Research, v. 73, p. 367–388.

Rowan, M.G., Peel, F.J., Vendeville, B.C., 2004. Gravity-driven fold belts on passive margins. In: McClay, K.R. (Ed.), Thrust tectonics and hydrocarbon systems: AAPG Mem., 82, pp. 157–182.

Totterdell, J. M. and A. A. Krassay, 2003, The role of shale deformation and growth faulting in the Late Cretaceous evolution of the Bight Basin, offshore southern Australia: Subsurface Sediment Mobilization, Geological Society (London) Special Publication, 216 (2003), pp. 429–442

Van der Spuy, D., T. Ziegler, and M. Bowyer, 2003, Deepwater 2D data reveal Orange Basin objectives off western South Africa: Oil & Gas Journal, v. 101, (April 4, 2003), p. 44– 49.

Welbon, A.I.F., Brockbank, P.J., Brunsden, D. & Olsen, T.S. 2007. Characterising and producing from reservoirs in landslides: challenges and opportunities. Structurally Complex Reservoirs. Geological Society, London, Special Publications, 292, 49–74.