Shelf stability and mantle convection on Africa's passive margins (Part 1)

Neil Hodgson[1*] and Karyna Rodriguez[2] demonstrate that dynamic topography offers a mechanism to contextualise basin stability and provides a framework for the generation of gravity structures.

Gravity slides and source rocks

Many clastic wedges prograding from the coast in Africa's passive margin basins display extraordinary gravity-driven collapse structures described variously as gravity-driven linked systems, fold and thrust belts or megaslides (Butler and Turner, 2010, Scarselli et al 2016). These features form relatively slowly and are distinct from instantaneous collapse submarine landslides or Mass-Transport Complexes (MTC) that reflect sudden catastrophic shelf collapse in response to seismicity, gas hydrate destabilization or high sedimentation rates.

Megaslides occur on giant scales from hundreds to thousands of square kilometres in extent, and are characterized by up-dip listric growth fault rollover systems in extensional zones, and a corresponding down-dip shortened section comprising multiple imbricate toe thrust faults and duplexes often referred to as fold-and thrust belts (FTB's). Separating the structured material from largely undeformed coherently bedded strata below, is a zone of planar, sub-horizontal detachment, or décollement.

One model that seeks to explain the induced instability that initiates these features infers that the décollement surface comprises a layer that is organic-rich. Sediments prograding out over this layer eventually bury this unit to a temperature and pressure that induces early hydrocarbon generation. This increases the inter-grain pore pressure and reduces the strength of the unit, so that previously stable sediments, in a stable angle of repose begin to slide under gravity basin-ward above this décollement surface.

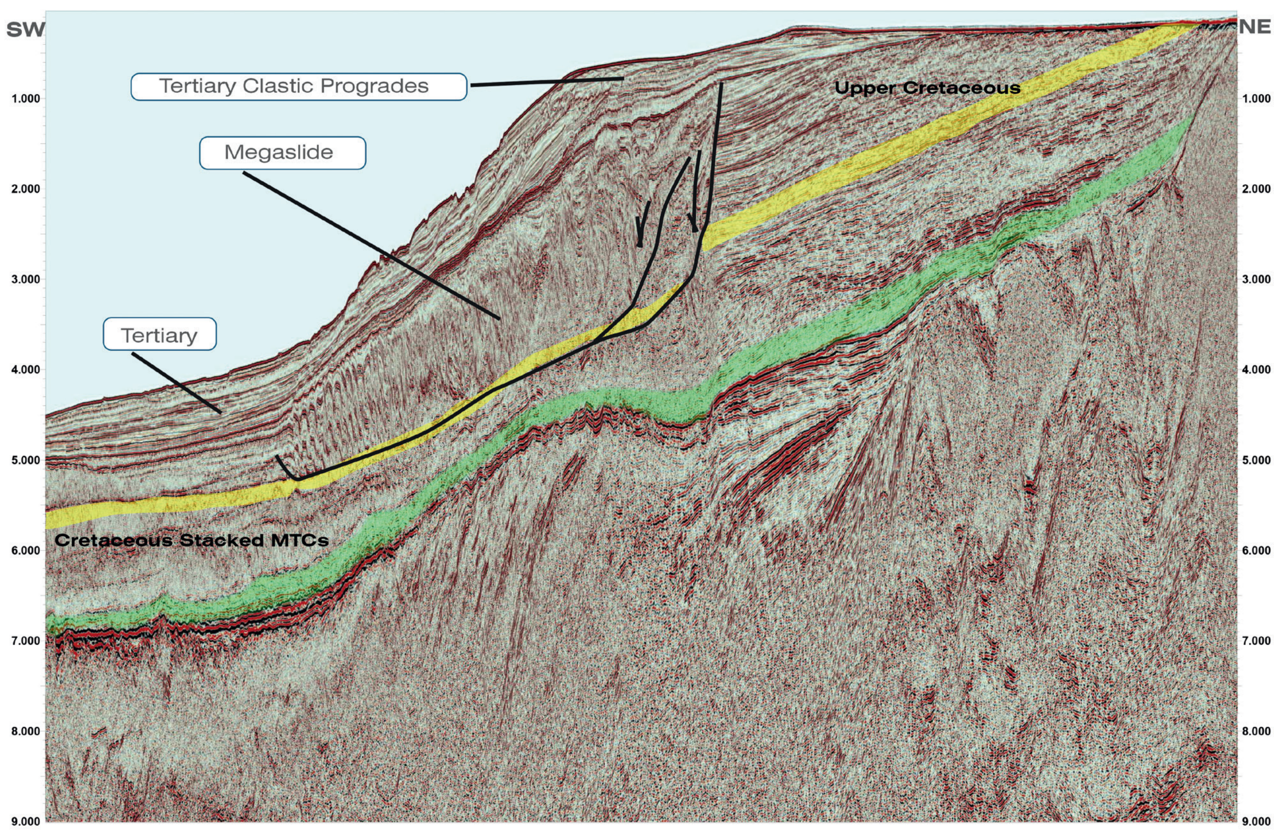

Figure 1 A slowly formed megaslide, Orange River Basin, Namibia and South Africa. PSDM section in TWT 405 km long. Green unit Aptian, Yellow unit, Cenomanian – Turonian.

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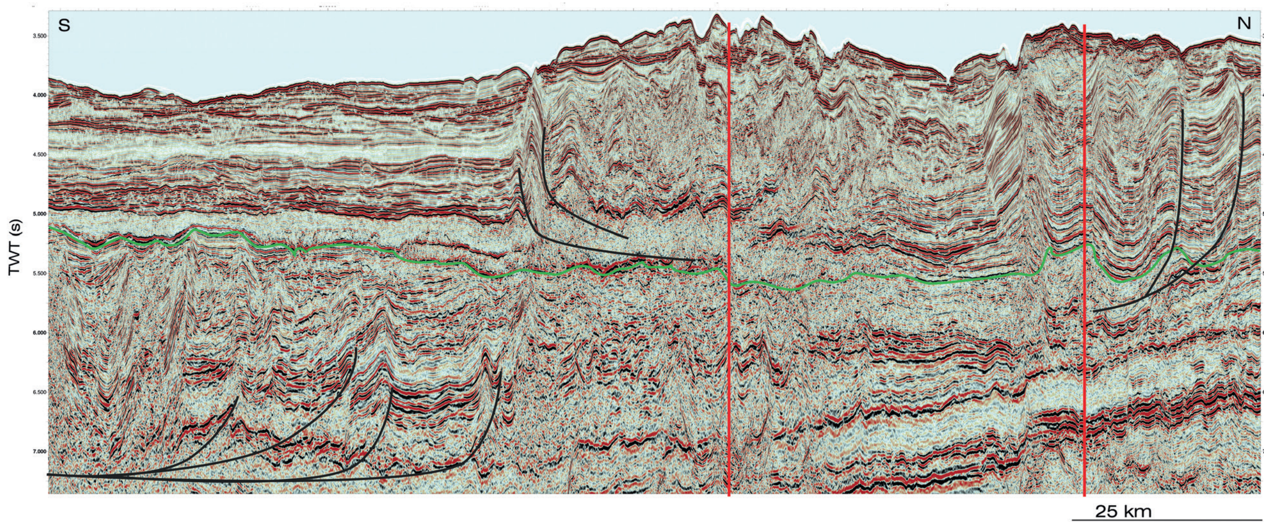

Figure 2 Two orthogonal lines from Juba-Lamu Basin, Somalia, showing two slowly formed megaslides with décollement surfaces in Cretaceous and Palaeogene sediments. PSTM sections in TWT, upper W-E line 157 km long, lower S-N line 169 km long.

This is an appealingly simple model, and indeed the décollement in Figure 1 from Namibia corresponds broadly to the Cenomanian-Turonian 'source rock' penetrated in many wells drilled on the shelf. The two megaslides in Figure 2 from Somalia show décollements in the Late Cretaceous and Eocene respectively which may also be source rock horizons.

This model of megaslide formation may be used as a proxy for the identification of organic-rich horizons (i.e. potential source rocks) in undrilled basins. While this could be a powerful tool in de-risking prospectivity, there appears to be two significant drawbacks of such an assumption. Firstly, it is not at all clear that every megaslide has a potential source rock as a décollement surface; for example at Palaeocene-Eocene level below the megaslides of the Rovuma delta Mozambique the décollement surface is thought to be a shale overpressured by disequilibrium compaction (Mahanjane and Franke, 2014). Indeed alternative décollement surfaces could be provided by mobile salt layers, ductile shales, or even chaotically organised shelf collapse sediments (MTC's).

The second disadvantage of such an assumption is that, were source rocks ubiquitous with megaslides, would it then follow that having 'no megaslide' is an indication of 'no potential source rock'? The lack of megaslides in some proven hydrocarbon basins on the Atlantic passive margins (e.g. Sergipe, Brazil), would suggest that this is not the case, and further, in the Orange Basin itself the proven Aptian source rock (ref HRT wells of 2014) appears not to be associated with a slide décollement at that level (Figure 1). Similarly, the (albeit unproven) Jurassic syn-rift source in Somalia (Figure 2) appears not to be associated with décollement.

The Orange Basin megaslide in Figure 1 is comprised of Cretaceous sediments, and sliding occurred approximately at the end of the Cretaceous. Post sliding, Tertiary clastics prograde again but they do not restart a second phase of sliding. This is curious behaviour if loading and source rock maturation is the only mechanism controlling destabilization and we suggest there is another control that is capable of switching off the instability, or indeed may prevent the instability from occurring in the first place.

Stabilization in this context may reflect the rate of sediment accumulation perhaps or an adjustment to the gradient of the slope between the shelf and basin floor, compensating for décollement ductility.

Shelf stability history and processes

Comparison of the sedimentary architecture of the Namibian Orange River Delta in Figure 1 reveals more information regarding the change in shelf stability of this basin with time. In the inboard section, a fairly continuous well-bedded and coherent sequence of Early Cretaceous carbonates through Late Cretaceous to Tertiary clastics is preserved. Outboard however, above Aptian source rocks deposited on the newly drifting oceanic crust, the succeeding Late Cretaceous sequence comprises mostly MTCs – the accumulated chaotic debris of collapsed clastic shelves. On top of these Cretaceous MTCs sits a well-bedded coherent prograding Tertiary clastic wedge, and between these two lies the megaslide which formed at the end of the Cretaceous (Figure 1).

The presence of extensive MTCs indicates that while the Namibian shelf was stable in the Aptian early drift phase, it became unstable during the Late Cretaceous, repeatedly failing catastrophically, yet the system returns to stability again in the Tertiary. The megaslide sits at the transition between these two states.

Stability of the Orange River clastic prism, reflects the changing gradient of the slope, i.e relative basin subsidence for a given sedimentary infill. When the basin subsides faster than the shelf, the gradient of the slope increases and prograding sediments over-steepen and collapse forming MTCs. When the basin subsides slower than the shelf, the prograding sediments never over-steepen; indeed they are self-stabilizing during such times. In this way, we can map in time the 'stability' of the outer shelf, or rather the change in gradient of the slope from outer shelf to basin floor. Change of slope gradient then is an important engine for the gravity structuration in the basin; both the dramatic and instantaneous shelf collapse, slowly developed megasliding, or in contrast, the relative stability that allows sequences to be deposited without over-steepening.

By comparison, in the southern Somalian Juba-Lamu Basin, we see that after a relatively stable Jurassic syn-rift and Cretaceous drift section, the margin destabilises, forming the first of the megaslides at the end of the Cretaceous (with décollement on a Cenomanian-Turonian source rock). This is followed by a second period of relative stability, a Tertiary section progrades as a stable package, then this too slides on a Palaeocene/Eocene décollement (and candidate source rock), and MTCs are subsequently common, suggesting the shelf is relatively unstable. We propose a time-stability series for the Juba-Lamu Basin that shows a transition from stable to unstable shelf at the end of the Cretaceous, followed by a variable but generally unstable Tertiary. As in Namibia, the shelf stability is a reflection of the steepness of the gradient from shelf to basin floor, and changes in this slope gradient owing to basin uplift or subsidence would be the main drivers of structuration, assisted by organic-rich décollement surfaces.

The Somalian stability history is different than that observed in the Orange River Basin. Does every basin have its own stability history reflecting localized unconnected variations in basin subsidence or could there be a connecting mechanism that is controlling divergent events?

Dynamic topography

Since the 1930s it has been recognised that mantle convection processes should affect the topography of the crust, and there are many examples of how isolated mantle plumes have influenced sedimentation in basins, including the appearance of the Forties Formation in the Central Graben of the UKCS (Mudge, 2014) and the change of facies north of the Indian plate and the deposition of the basin floor Pab sandstone formation (Eschard et al., 2004). More recently, however, in an extraordinary paper, Hoggard et al. (2016) published observations on the depth of oceanic crust in the world basins. They measure the deviation in depth of the observed position of the crust (corrected for sediment loading and crust thickness) from a theoretical cooling curve through time. The measurement is related as a 'residual depth' either '+ve' (higher than predicted) or '–ve' (lower than predicted). Many global observations by the authors allow the calculation of a residual distribution, which shows topographic variations of +/-1000 m on a 1000 km lateral scale. These authors interpreted the residuals to reflect mantle convection cells, either upwelling and positively lifting the crust, or down-welling and negatively depressing the crust. A map of the observed global residual distribution calculated by Hoggard et al is shown in Figure 3.

Spectrum's global library of long record length data allows us to confirm the observations of Hoggard et al in terms of oceanic crust residuals, and calculate new residuals on basins previously unevaluated. We observe that the residual for the oceanic crust under the Juba-Lamu Basin of Somalia today is some 1km deeper than it should be for oceanic crust of Jurassic age (1 km –ve residual), while in Namibia, as noted by Hoggard et al, the residual for oceanic crust outboard of the Orange River Delta today is some 1 km higher than it should be for Early Cretaceous crust (1 km +ve residual).

Basin stability and residuals

We assume a relationship between slope gradient and residual, such that a subsiding basin with –ve residual will have an increasingly steeper slope than a basin with a +ve residual where the basin is becoming less deep compared to the shelf and therefore the slope becomes less steep. It seems reasonable to assume that an increasingly steeper slope is less stable then an increasingly less steep slope, and that this is accentuated reflecting a difference between oceanic and continental crust in the response to a given mantle convection stimulus.

Such a linkage allows us to propose a residual history for basins by reference to the observed seismic structuration or stability profile. Within such a model we postulate that Namibian oceanic crust displayed a –ve residual through the Upper Cretaceous, leading to the super-unstable shelf and deposition of a dominantly MTC sequence in deepwater, whilst Somalian oceanic crust through the Upper Cretaceous displayed a +ve residual, leading to stable clastic progradation. The projected residuals in the Upper Cretaceous for these two basins are the reverse of their observed residuals today (Figure 3). A +ve to –ve transition implies an extraordinary 1-2 km of relative subsidence and shelf destabilization of the basin floor. The reverse, a –ve to +ve residual transition of such a magnitude, reflects a reduction of slope having a stabilising effect.

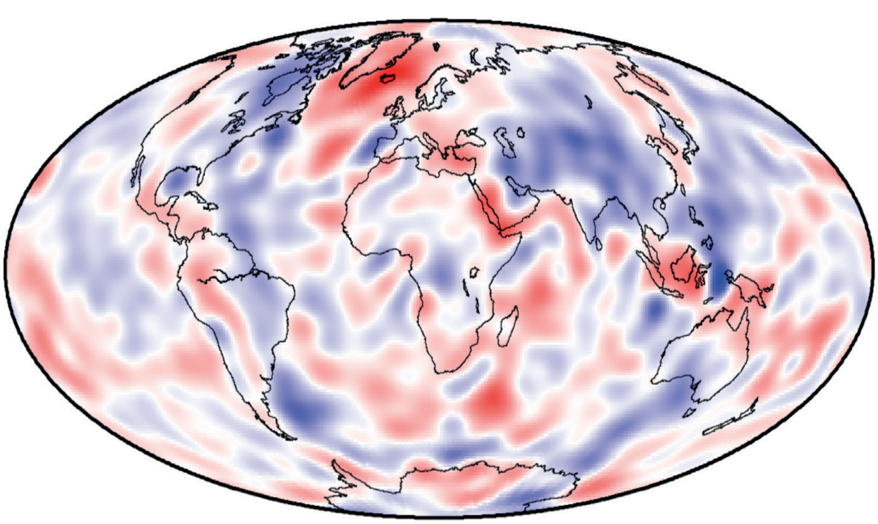

Figure 3 Present-day global observed dynamic topography. Spherical harmonic model of residual up to l=30 of global measurements of depth to top crust (Source: Hoggard et al., 2016). Positive residuals are red, negative are blue.

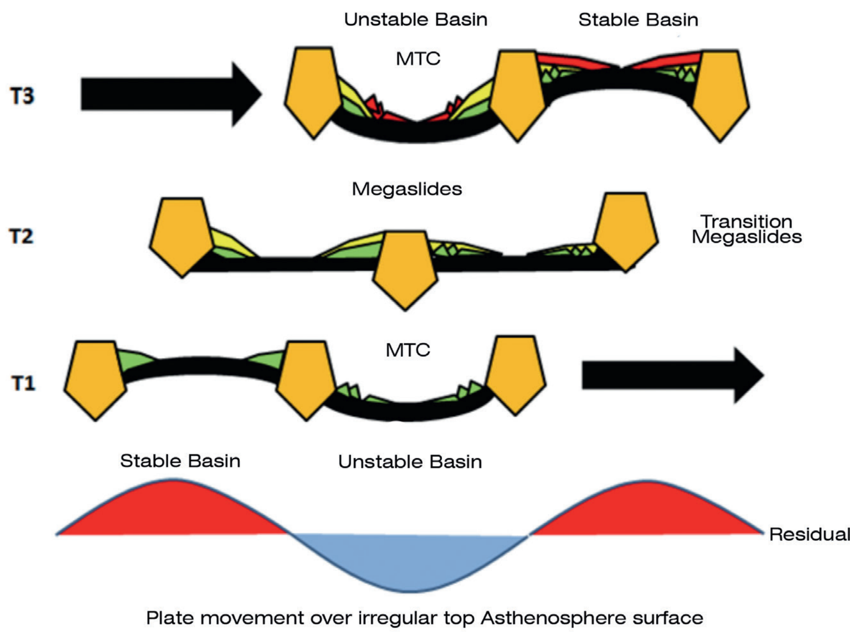

Figure 4 Sketch of Lithosphere moving over asthenospheric residual surface.
T1: Left hand basin Stable, Right Hand Basin Unstable.
T2: Transition Period, Megaslide development T3: Left-hand basin Unstable, Right-hand Basin Stable.

To be clear, in such a model we need not propose that the underlying convection cell 'changes direction', rather, we interpret the change in observed (or interpreted) residual reflects the movement of the African plate over the underlying mantle convection cell framework.

By mapping the movement of plates through time (tied to the mantle plume reference frame), and describing a series of inferred paleo-residuals, we can therefore map out the inferred variation in residual with time for each basin, and by examining a dataset of all the world's basins, can, by inference in effect create a dynamic version of Figure 3 back through time.

Curiously, when the inferred residuals from stability analysis are reviewed, assuming African plate movement over the past 120 Ma traced relative to the mantle plume framework, close similarity is observed to the present-day residual map (Figure 5). Surprisingly, this suggests that asthenospheric convection cells (at least around Africa's margins) may themselves be geographically static or stable relative to the plume framework over relatively long periods of time.

That the mantle convection cells are relatively stable through time is of interest and consistent with some observations and models at least for mantle convection derived from an independent and discrete set of earth-observations (e.g. Torsvik et al., 2010). This needs further corroboration, although it seems relatively intuitive that upper mantle convection cells 700+ km deep and 1000 km in diameter are largely unaffected by the relatively thin and brittle crust comprising a 25 to 60 km thick surface layer. Although the wavelength of the residual pattern seems quite short, it is not yet clear at what depth the mantle convection that causes oceanic plate residuals lies, what is apparent is that the convection cells are much smaller than the overlying plates, making one reassess the mechanism for driving plate movement.

For Somalia, the imposed time-series residual pattern matches well with the inferred time-series residual pattern from the expression of structuration and stability observed on seismic (Figure 5). For Namibia, a similar good correlation is observed between interpreted residual and static convection cell pattern

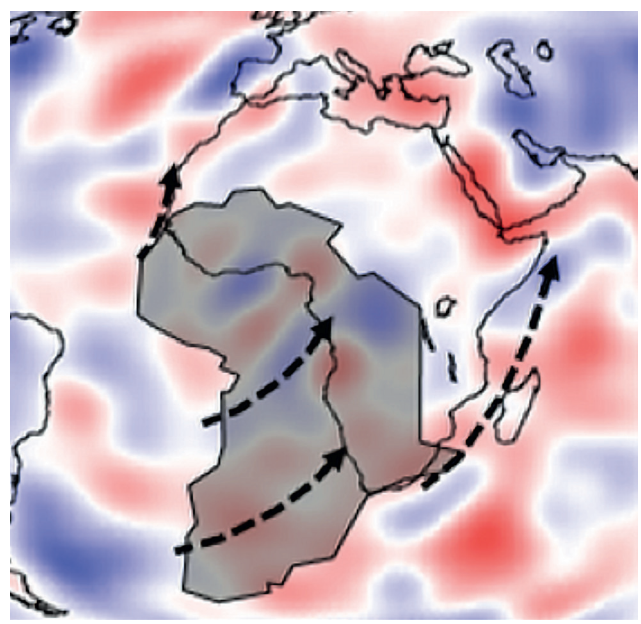

Figure 5 Present-day dynamic topography with Africa migration path indicated for the last 120 Ma. An assumption of mantle convection cell stability correlates well with observed patterns of basin stability, based on seismic observations.

(Figure 5). However, this falls down in the Aptian (the time of Gondwana break-up and rifting). Although the underlying convection cell framework could well have affected syn-rift topology, flexural isostasy post break-up still has to exert its influence. Yet in the subsequent drift-phase, after a short period of relative stability, very rapidly crustal basin topography is controlled again by the underlying mantle convection cell pattern.

During the initiation of drift, the residual fabric will determine if a given margin will be rapidly submarine or subaerial, hence determining the style of magmatism (Jackson et al., 1999). Rifting above a –ve residual experiences rapid marine inundation, preventing flood basalt formation and yielding a SDR poor/free 'amagmatic' margin such as observed in South Gabon. Alternatively if a part of the rift is underlain by a positive residual, then it may remain above sea level for a considerable period of time, such that erupted plateau lavas build an extensive SDR transitional crust, such as observed on the 'magmatic' Namibian margin. Asymmetric positioning of rift relative to residual may yield asymmetric architecture between conjugate margins (Lentini et al 2010). Again, this approach gives a mechanism to infer paleo residuals for rifted passive margins.

While it is early days with this analysis, it is interesting that the Congo/Gabon part of the West African plate margin would track a path along a down-welling mantle –ve residual all the way through the Cretaceous showing no uplift during this period. Remaining over a single phase cell for this period, one would also predict a limited number of unconformities after the flexural isostasy corrections post rifting within the shelf section through the Cretaceous and Tertiary sections. In a relative sense the lack of unconformities post rifting isostasy adjustment in the offshore succession of Gabon is evident compared to the Namibian or Somalian margins. Tying the stability of margin and the occurrence of unconformities to the movement of plates over the mantle convection cell framework could provide a powerful predictive tool.

Conclusion

Margin instability is recorded in the stratigraphic section by both catastrophic collapse MTCs and slowly forming megaslides underlain by low strength decollement surfaces (organic-rich or over-pressured horizons). Yet such gravity induced instability can be counteracted , or advanced by changes in basin margin slope. We propose that the gradient of the margin slope can be significantly affected by lateral basin migration over the irregular asthenosphere as mapped by Hoggard et al.

Dynamic topography offers a mechanism to contextualise basin stability, and provides a framework for the generation of gravity structures by increasing continental slope angle (causing destabilisation), or reducing slope gradients (stabilising the section).

In Part 2 of this work, we will develop the use of paleo-dynamic topography as a proxy to understand SDR distribution and margin asymmetry along the Atlantic margins, and examine the mechanism for plate movement.

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