Transform margins: development, controls and petroleum systems – an introduction

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Abstract: This paper provides an overview of the existing knowledge of transform margins including their dynamic development, kinematic development, structural architecture and thermal regime, together with the factors controlling these. This systematic knowledge is used for describing predictive models of various petroleum system concept elements such as source rock, seal rock and reservoir rock distribution, expulsion timing, trapping style and timing, and migration patterns. The paper then introduces individual contributions to this volume and their focus.

Supplementary material: Location table and map of specific transform examples, structural elements of the Romanche transform margin and glossary of terms used in this article is available at https://doi.org/10.6084/m9.figshare.c.3276407

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This volume shows dynamic linkage between new transform margin research and increasing transform margin exploration. It offers a critical set of predictive tools via an understanding of the mechanisms involved in the development of play concept elements at transform margins and their controlling factors. Our introduction to this volume reviews the current knowledge of the transform margins, focusing on their dynamic development, kinematic development, structural architecture, thermal regimes and petroleum systems. The last part of this paper introduces the contributions to this volume.

Dynamic development

Rifted continental margins are typically segmented, consisting of extensional margins (e.g. Falvey 1974; Veevers 1981; Buck 1991; Menzies et al. 2002; Manatschal 2004; Huismans & Beaumont 2005, 2008), transform margins (see the models of Wilson 1965; Freund 1974; Mascale 1976; Mascale & Blarez 1987) and oblique margins (e.g. Cochran 1973; Sims et al. 1999; Bird 2001; Clifton & Schlichte 2001; d’Acremont et al. 2006; Lizarraide et al. 2007). Transform margins, which are also called sheared margins (Rabinowitz & Labrecque 1979; Scrutton 1979) or strike-slip margins (Nagel et al. 1986), and margins developed by dominating transform tectonics are the focus of this special publication.

Worldwide case examples of transform margins allow one to group them into two categories with respect to their initial role in ocean opening. The first group contains transform margins which started as transfer faults between neighboring rift units (see Bally 1982; Gibbs 1984; Bosworth 1985; Lister et al. 1986; Versfelt & Rosendahl 1989; Moustafa 2002). The Central Atlantic Rift System (Schettino & Turco 2009; Fig. 1) is a typical example for this category, having its geometry characterized by numerous offsets of neighboring rift zones. Its faults that were active as transfer faults and subsequent transforms for the longest time are the ones represented by the largest transform margins, separating regions with oceanic crusts of differing age ranges (Müller et al. 1997; Fig. 2). They include the transform margins associated with the Gettysburg–Tarfaya transform fault system (Nemčok et al. 2005 and references therein) and Cobequid–Chedabucto–Gibraltar transform fault,
Fig. 1. Plate reconstruction at 200 Ma (Triassic–Jurassic boundary) (Schettino & Turco 2009). Initial oceanic crust associated with the East Coast and Blake Spur magnetic anomalies is shown in yellow. Orange denotes individual rift branches. Red lines are spreading centres, blue represents oceanic crust.
Fig. 2. Age of the oceanic crust age (Müller et al. 1997). Colour-coding indicates Jurassic, Cretaceous, Palaeogene and Neogene–Present ages.
Fig. 3. Plate reconstruction at 147.7 Ma (Tithonian) (Schettino & Turco 2009). Rifting in northern Central Atlantic evolved to drifting. The Atlas rift branch was aborted and the plate boundary migrated to the north. GiF, Gibraltar Fault; NPF, North Pyrenean Fault. Orange, active rifting; red, sea-floor spreading centres. Blue represents oceanic crusts of the Ligurian and Alpine Tethys oceans.
system (Le Roy & Piqué 2001; Schettino & Turco 2009; Fig. 3). Other examples of transforms dividing neighbouring rift zones are (Table 1): the Cape Range and Wallaby transforms in Western Australia, the Coromandal transform in East India and the Davie transform in East Africa.

The second group contains transform margins that started as transforms connecting oceans opened at different times. Examples come from:

(1) the Sierra Leone and Northern Demerara faults that have divided the Central Atlantic from the Equatorial Atlantic since Albian time (e.g. Loncke et al. 2015; Nemčok et al. 2015 and references therein);
(2) the Pernambuco and Fernando Poo faults that divided the Equatorial Atlantic from the South Atlantic since Albian time (Rosendahl et al. 2005; Nemčok et al. 2012b, 2015a and references therein); and
(3) the Falklands and Agulhas transform faults (Edwards 2013, pers. comm.) representing the southern terminus of the South Atlantic.

What makes transform margins different from extensional margins is the amount of post-breakup uplift. Extensional margins are normally associated with distinct uplift driven by, representing transient uplift mechanisms related to the thermal effects of rifting:

(1) depth-dependent extension (Royden & Keen 1980; Hellingier & Scater 1983; Watts & Thorne 1984; Morgan et al. 1985);
(2) lateral heat flow (Steckler 1981; Cochran 1983; Alvarez et al. 1984; Buck et al. 1988); and
(3) secondary convection under rift shoulders (Keen 1985; Steckler 1985; Buck 1986).

In addition, representing permanent uplift mechanisms, they are driven by:

(1) magmatic underplating (Cox 1980; Ewart et al. 1980; McKenzie 1984; White & McKenzie 1988);
(2) flexural-isostatic uplift (Watts et al. 1982; Gilchrist & Summerfield 1990; Weissel & Karner 1994); and
(3) lithospheric unloading and/or ductile necking (Zuber & Parmentier 1986; Parmentier 1987; Braun & Beaumont 1989; Issler et al. 1989; Chery et al. 1992; Weissel & Karner 1994).

Further contribution comes from:

(1) the erosional unloading of margin coupled with depositional loading of basin (e.g. van Balen et al. 1995; van der Beek et al. 1995; Burov & Cloetingh 1997); and

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**Table 1. Timing data on selected transform margins**

| Transform name | Length of continent-continent clearance (Ma) | Length of spreading ridge-continent clearance (Ma) | Source |
|----------------|--------------------------------------------|-------------------------------------------------|--------|
| **Between two main oceanic segments** | | | |
| Azores-Gibraltar Transform | 45.8 | Reading off the map by Müller et al. (2008) |
| Charlie Gibbs Transform (European side) | 37.5 | Reading off the map by Müller et al. (2008) |
| Charlie Gibbs Transform (N American side) | 18.5 | Reading off the map by Müller et al. (2008) |
| Falklands Transform | 39.7 | Reading off the map by Müller et al. (2008) |
| South Tasman Transform | about 10 | 33.9 | Gibson, 2015 pers. comm.; reading off the map by Müller et al. (2008) |
| **Sort of between two main oceanic segments** | | | |
| St Paul Transform | 4–8 | 12 | Reading off the maps by Müller et al. (1997, 2008); Nemčok et al. 2012b |
| Romanche Transform | | 14 | Reading off the maps by Müller et al. (1997, 2008); Nemčok et al. 2012b |
| **Between two neighbor rift zones** | | | |
| Cape Range Transform | 8.2 | Reading off the map by Müller et al. (2008) |
| Coromandal Transform | 5.5–11 | Reading off the map by Müller et al. (2008); Sinha et al. (2015) |
| Davie Transform | 50.3 | Reading off the maps by Müller et al. (1993, 1997) |
| Wallaby-Zenith Transform | 15.2 | Reading off the map by Müller et al. (2008) |
margin unloading coupled with basin loading due to gravity gliding (Mohriak et al. 2008).

Transform margins, apart from exceptions, in contrast, are not associated with distinct uplift for the following reasons. The strike-slip displacement of the blocks along the transform fault zone does not include any significant removal of the hanging wall from the footwall that would react by the isostatic uplift. An interpreted comparison of uplift related to extensional and transform margins after the breakup comes from the Central Atlantic (Nemčok et al. 2005; Fig. 4). Figure 4a shows the distribution of depositional environments after breakup. It documents that the extensional margin segments in the north and south were undergoing uplift, resulting in mountain ranges preventing siliciclastic material from the hinterland from entering the oceanic basin. Protected from such input, wide carbonate platforms were developing on the shelf. The transform segment in the centre did not undergo any distinct uplift. This allowed sediments to funnel into the oceanic basin through the system of failed pull-apart basins. Figure 4b shows the distribution of depositional environments later after the breakup. It documents that, after the erosional removal of the mountain ranges at extensional margins, sediments had an equal opportunity to enter oceanic basins through transform and extensional margins.

Because pure strike-slip end-members of the transform margins are rather rare, transform margins can have a small component of isostatic uplift.
(see Clift & Lorenzo 1999), which depends on the transform fault dip (see Ring et al. 1998), the size of the normal faulting component in the overall strike-slip fault regime and the flexural rigidity of the continental plate. Another unloading scenario can be provided by erosion (e.g. Basile & Allemand 2002), which depends on factors such as climate, lithological composition and morphology (e.g. Saunders & Young 1983; Kukal 1990; Jordan 1995).

In contrast to extensional systems, the uplift studies in transform systems (e.g. Nemčok et al. 2012b, 2015b) indicate no distinct flexural uplift because, apart from the plate being inherently broken rather than continuous, the thermal regime of the adjacent spreading centre sandwiched by the two progressively widening corridors of newly accreted oceanic crust considerably warms up the transform margin. This results in a significant reduction of its flexural strength.

Heating is a mechanism for some limited transform margin uplift. Early studies of the thermal expansion-driven uplift (e.g. Todd & Keen 1989; Lorenzo et al. 1991; Lorenzo & Vera 1992; Basile et al. 1993; Gadd & Scrutton 1997; Vågnes 1997), drawing from thermomechanical finite-element and finite-difference modelling compared with crustal thinning and erosion estimates constrained by the refraction and reflection seismic, gravity and well data, interpreted a distinct uplift. Subsequent models (e.g. Nemčok et al. 2012b)
found this uplift usually combined with other uplift mechanisms and typically playing a relatively subordinate role.

Another mechanism responsible for uplift at transform margins is the progressive change of the initial strike-slip regime to transpression, which was documented along the Romanche transform margin (e.g. Mascle & Blarez 1987; Antobreh et al. 2009; Nemčok et al. 2012b).

Yet another mechanism potentially resulting in uplift at transform margins is a flow of lower crustal material following pressure gradients when the warming of these margins results in viscosity reduction of specific volumes of lower crustal rocks, which vary in space and time (see Reid 1989; Todd & Keen 1989; Lorenzo et al. 1991; Lorenzo & Vera 1992; Sage 1994; Vagnes 1997; Miles et al. 1998; Radhakrishna et al. 2002; Gunnell et al. 2003; Henk & Nemčok 2015).

Kinematic development

There has been a long-lasting debate with regards to the influence of the pre-breakup continental fault zones on the post-breakup oceanic transforms. The first group of studies interprets the transform faults capitalizing on the locations of the pre-existing continental zones of weakness (e.g. Wilson 1965; Mascle 1976; Wright 1976; Sibuet & Mascle 1978; Rosendahl 1987; Bellahsen et al. 2013; Gibson et al. 2013). The second group interprets transform propagation as unaffected by pre-existing anisotropy (e.g. Taylor et al. 2009; Basile 2015). This group has documented that the initial spreading offsets are often non-transform and that transforms develop either after or during the spreading nucleation. Good natural examples come from the Woodlark Basin (Taylor et al. 2009).

Additional support comes from three-dimensional numerical modelling, which suggests that (Gerya 2013):

(1) spreading segments nucleate in an en echelon pattern in overlapping rift units starting the initial disorganized spreading stage;
(2) subsequently, still during the disorganized spreading stage, proto-transforms sometimes nucleate as oblique to a spreading vector rather than parallel to spreading vector; and
(3) subsequently, still during the disorganized spreading stage, the initially oblique proto-transforms rotate into spreading–parallel position; and
(4) subsequently, during the organized spreading stage, the spreading vector–parallel transform direction becomes a rule governed by space accommodation during the mature ridge-transform spreading that is reached after about the first 5 myr of spreading.

We can look at this discussion about the influence of the pre-breakup continental fault zones from the following point of view. The breakup trajectory in the brittle upper crust usually follows the Andersonian geometry of the normal and strike-slip faults with respect to the stress field that resulted in the breakup, unless their geometry is affected by pre-existing anisotropy. Examples of zig-zag trajectories with normal, oblique-slip or strike-slip faulting-dominated control come from the offshore Gabon (Nemčok et al. 2012a), offshore East India (Nemčok et al. 2012c) and both sides of the Equatorial Atlantic (Nemčok et al. 2012b). Subsequently, the initial spreading centres have to clear their contact with the continental–oceanic transform changing into transform margin segments. The data indicate that there were numerous cases when a spontaneous creation of oceanic transforms reacting to a specific situation with their controlling spreading–centre arrangement had to wait in the vicinity of these continental–oceanic transforms changing to transform margins until the lateral clearance of spreading centres along transform margins with active accommodating strike-slip faults was accomplished. Such delays in spontaneous creation are indicated by the following observations:

(1) in the case of the Wallaby–Zenith continental–oceanic transform, the laterally clearing spreading centre tried to propagate through the adjacent continental ribbon, i.e. Wallaby Plateau, but failed (which can be observed in reflection seismic sections 135-05, 135-06, 135-08 and 310-64);
(2) oceanic crust near the Romanche margin was formed as much thinner than the surrounding normal oceanic crust (Edwards et al. 1997), similar to the thin crust at oceanic transforms (see Canales et al. 2000);
(3) oceanic crust adjacent to an active continental–oceanic transform in the Southern New Zealand underwent distinctive deformation (Paul Mann 2013, pers. comm.); and
(4) thin or even missing oceanic crust near the Falklands Escarpment, explained by the cold-edge effect of the adjacent continental crust, causing a fairly poorly developed, or absent, oceanic crust in the south of the spreading centre migrating along the Falklands transform margin (Rosemary Edwards 2013, pers. comm.).

The kinematic development of the transform margins that underwent spreading centre clearance can be divided into three stages (e.g. Mascle & Blarez 1987):

(1) active continental stage;
(2) active continental–oceanic stage; and
(3) passive margin stage.
The time spans and the termination of individual stages vary along the strike of the transform margin. The termination of the continental stage in specific area of the future transform margin depends on the time required for the adjacent continent to clear this area and for the juxtaposition of this area to laterally clearing oceanic crust (Fig. 5). The passive margin stage initiates in a specific area once the spreading centre clears this area.

Kinematic development of the transform fault zone requires the presence of displacement dissipation structures at both ends. The most typical structure patterns of the already-mentioned receding sides are horse-tail structures, known under various names.

**Fig. 5.** Sketch of the strike-slip fault zone with pull-apart terrains evolving into the ocean. Check point is indicated by small circle. Stages: I, rifting of continental lithosphere and the check point situated at the dextral transform fault; II, the check point still adjacent to continental lithosphere but a landward portion of the future transform margin adjacent progressively to normal continental crust, then thinned continental crust and then oceanic crust; III, the check point adjacent to the spreading ridge; IV, the check point adjacent to oceanic crust and an inactive transform fault. 

d is the distance travelled by the spreading centre and \( l \) is the total length of the transform. \( d = \mu / 2 + t \), where \( \mu \) is the half spreading rate and \( t \) is the time length of ridge travel.

\[ d = \frac{\mu}{2} + t \]
Fig. 6. Surface in 3D seismic cube from the unspecified passive margin showing a fault pattern associated with several horse-tail structures at receding sides of sinistral strike-slip fault zones forming together a transform margin. Note a large number of individual pull-apart basins developed in the horse-tail structure.
names (e.g. Wilcox et al. 1973; Crowell 1974; Christie-Blick & Biddle 1985; Woodcock & Fischer 1986; Sylvester 1988; Fig. 6). They link the transfer zone preceding the transform zone to neighbouring orthogonal or oblique rift units. Field studies of the failed Darag, October, Zeit and Duwi rift units from the Gulf of Suez–Red Sea rift system show that their development was coeval with rift unit-controlling normal faults (Younes & McClay 2002). Interpretation of seismic imagery tied to wells in the successful Krishna–Godavari and Cauvery rift zones linked via the Coromandal transform indicates that they can develop coevally to post-dating the development of the rift zone-controlling faults (Fig. 5b in Nemčok et al. 2012a). The former example documents a development of accommodation zones in the study area as starting with the reactivation of the pre-existing zones of weakness prior or coeval with the propagation of the rift-bound faults and finishing with the arrest of the border faults and transfer of displacement (Younes & McClay 2002). The latter example assumes the development of the transform zone capitalizing on pre-existing anisotropy, based on its parallelism with the boundary between the Archaean cratons and the Proterozoic orogenic belt in the onshore and mezo-scale structural grain of the orogen (Nemčok et al. 2012c; Fig. 7). In this example, the distance between the two rift zones was probably too large for them to be fully linked by the transfer

Fig. 7. Fault map of the East Indian continental margin combined with onshore geological map showing the location of the boundary of the Proterozoic orogen with Archaean cratons (blue or red lines showing alternative interpretations) and its internal structural grain (modified from Nemčok et al. 2012a). Ca, fault pattern of the Cauvery rift zone; Co, fault pattern of the dextral Coromandal accommodation zone; K-G, fault pattern of the Krishna–Godavari rift zone. Note that green faults of the Coromandal zone deform the orange and red faults of the Cauvery rift zone.
zone. This can be interpreted from the observation that, while the faults of the northern horse-tail structure of the transform are fully linked with syn-rift faults of the Krishna–Godavari rift zone, the faults of the southern horse-tail structure first cross-cut the syn-rift faults of the Cauvery rift zone and only subsequently both structures became kinematically linked (Nemčok et al. 2012c; Sinha et al. 2015). The southern transform linkage could have been developed by the east–west Indian Ocean propagation through the evolving rift system. Starting in the east, the Mahanadi rift zone underwent an ocean opening at 145 ± 2 Ma (Berriasian), followed by the offshore remnant of the Cauvery rift zone in the west undergoing an opening at 136 ± 3 Ma (Valanginian) (Sinha et al. 2015). However, the Krishna–Godavari rift zone between them, but significantly offset towards the north, eventually experienced an ocean opening at 125 ± 1 and 123 ± 1 Ma in its eastern and western parts, respectively, causing a ridge jump by capturing the oceanic crust accretion that was taking place further south. Coevally with the breakup in its western portion, the faults of the Coromandal transform fault propagated southward, eventually reaching the Cauvery zone. The active continental and continental–oceanic stages of the transform were interpreted to take place between 123 and 112 Ma (Sinha et al. 2015). The time of 123 Ma ago represents the end of the Krishna–Godavari rift zone propagation. The time of 112 Ma ago roughly marks the end of the Coromandal transform activity.

**Structural architecture**

The structural architecture of transform margins is rather complex. For example, individual structural elements of the Romanche margin include marginal ridge, platform areas, pull-apart regions, dominating transform segments and subordinate rift margin and oblique rift margin segments (Mascle & Blarez 1987; Blarez & Mascle 1988; Mascle et al. 1988; Basile et al. 1993, 1998; Peirce et al. 1996; Edwards et al. 1997; Benkhelil et al. 1998; Sage et al. 2000; Antobreh et al. 2009). The width of the entire fault pattern associated with the Romanche transform fault zone varies between 40 and 70 km depending on area (Basile et al. 1998; Benkhelil et al. 1998; Nemčok et al. 2012b). The structures of this corridor are represented by strike-slip, oblique-slip, normal and reverse faults, folds, releasing bends and offsets, restraining bends and offsets, transpressional and transtensional duplexes, pull-apart basin systems located in both releasing offsets and horse-tail structures. Individual strike-slip faults contain R, P and R′ shears and other building structures developed in succession. They can be present or absent, depending on the development maturity of each individual fault zone. This architecture documents a complex kinematic development of transform faults and the frequent effect of the pre-existing structural grain. One of the best indications of this effect is the overall en echelon arrangement of all main transform faults of the Equatorial Atlantic margins, which are in the P shear position instead of the R shear position, which would be typical for the transform faults developed in the homogeneous crust.

Duplexes and horse-tail structures are characterized by a vertical movement component in the overall strike-slip regime. They are separated from the host rocks either by linked fault zones or by high-density fracture zones (Woodcock & Fischer 1986). Their development indicates that the overall strike-slip regime is maintained with the help of lateral deformation of the surrounding host rocks. Plane strain deformation is disturbed owing to a vertical response of the surrounding areas (Hempton & Neher 1986; Naylor et al. 1986; Woodcock & Fischer 1986). Structures with distinct vertical movements can be located in several strike-slip settings including bends and offsets, on both sides of the strike-slip fault terminations, or even at straight segments of the strike-slip fault zones if controlled by an appropriate combination of shear types (Wilcox et al. 1973; Crowell 1974; Christie-Blick & Biddle 1985; Harding et al. 1985; Hempton & Neher 1986; Naylor et al. 1986; Woodcock & Fischer 1986; Gamond 1987; Homberg et al. 1997; Ohlmacher & Aydin 1997; Sims et al. 1999; Nemčok et al. 2002; Fig. 8a, b).

Perhaps the most dramatic results of the vertical movements are pull-apart basins (Fig. 9), which occur at releasing bends (Burchfiel & Stewart 1966), extensional bridges (Petit 1987), extensional oversteps (Christie-Blick & Biddle 1985; Deng et al. 1986), horse-tail structures (Fig. 6) or releasing junctions (Crowell 1974) of strike-slip fault systems. Their geometry is represented by rhomb-shaped graben, rhomb-shaped half-graben, a series of coalescent basins (e.g. Schubert 1980; Zak & Freund 1981; Aydin & Nur 1985; Sylvester 1988; May et al. 1993) or a more complex combination of fault blocks (Moore 1969; Junger 1976; Howell et al. 1980; Kocák et al. 1981; Royden 1985; Fodor 1995; Sims et al. 1999; Nemčok et al. 2005).

Developed between two controlling strike-slip faults, pull-apart basins record their interaction. Theoretical studies indicate (Segall & Pollard 1980) that the interaction takes place between faults that are separated by a distance that is less than their total depth penetration. Further factors controlling the pull-apart basin shape are the size of a strike-slip overstep (Mandl 1988; Sylvester 1988; Richard et al. 1995), displacement along the boundary strike-slip
Fig. 8. (a) Structures in the strike-slip fault (Woodcock & Fischer 1986). (b) Stress perturbation in extensional strike-slip bridge at St Donats, Bristol Channel (Nemčok et al. 2002). The bridge is associated with dextral strike-slip fault. Figure shows the advanced development stage of the extensional bridge.
faults (Faugère et al. 1986) and relative displacement rates along them (Rahe et al. 1998). Based on their depth of detachment, pull-apart basins can be separated into thick- and thin-skin pull-apart basins (Royden 1985).

The number and geometries of the pull-apart depressions and dividing highs depend on the rheology of the pull-apart detachment horizon. While brittle detachments control a pull-apart basin with clearly defined bounding strike-slip and normal faults (Fig. 10), and a single intra-basin high (Fig. 11a), ductile detachments control a pull-apart basin bounded by oblique-slip faults (Fig. 10), and one or more inter-basin highs (Fig. 11b), which

Fig. 9. Sketch of the pull-apart basin.

Fig. 10. Models of the pull-apart basin detached along a brittle, thin ductile and thick ductile detachments (Sims et al. 1999).
are much more pronounced than the intra-basin high in the brittle case (Sims et al. 1999).

The transition of the transform margin segment into the neighbouring orthogonal rift or pull-apart margin segments is not abrupt. It is represented by strike-slip-dissipating structures such as the horse-tail structure (Fig. 6), which control the change from the narrow shelf and slope of the transform margin segment to the wide shelf and slope of the orthogonal rift/pull-apart margin segment. Therefore, reflection seismic sections through the transform-pull-apart margin segment transitions frequently show a marginal ridge in the cases when they cut the last pull-apart basin landward of the transform margin-controlling strike-slip fault zone (Fig. 12). A good three-dimensional distribution of highs and lows along the entire Coromandal transform with both horse-tail structures is provided by the detailed mapping of basement in a large series of 3D seismic volumes (Nemčok et al. 2008; Fig. 13a). Figure 13b shows that the Coromandal transform fault zone links with the Krishna–Godavari rift zone in the north via an associated horse-tail structure and with the Cauvery rift zone in the south via a joining horse-tail structure. Unlike the transform margins in the Equatorial Atlantic where most of the transform margin segments are neighbours of the coeval pull-apart margin segments, both neighbours of the Coromandal transform zone are orthogonal rift zones, which developed into continental breakup. As discussed earlier, they did not reach the breakup at the same time. It was the

Fig. 11. (a) Scenario of the pull-apart basin detached along a brittle detachment (Sims et al. 1999). Sediments characterized by white–red–blue and orange–red strata are pre-tectonic and syn-tectonic. Note an intra-basin high in the centre of the basin.
Fig. 11. (b) Scenario of the pull-apart basin detached along a thick ductile detachment (Sims et al. 1999). Note a prominent inter-basin high. See (a) for further explanations.
Fig. 12. Offshore Ghana, western, joining, horse-tail structure of the Romanche transform fault zone including a pull-apart basin and marginal ridge (modified from Nemčok et al. 2004).
Cauvery rift zone that reached it first, and only then the Krishna–Godavari developed kinematic linkage with a failed portion of the Cauvery rift zone via the Coromandal transform. As a result, the Krishna–Godavari zone is linked with the Coromandal transform via associated horse-tail structure, while the southern Coromandal horse-tail structure cuts through the pre-existing fault pattern of the Cauvery rift zone, constituting a joining horse-tail structure.

A top-to-basement map of the entire circum-Coromandal region allows one to see the distribution of highs and lows associated with both its horse-tail structures and adjacent portions of rift zones (Fig. 13a). Figure 13a shows the locations of structural highs highlighted by red polygons. These highs are surrounded by a pattern of local pull-apart depressions, which are analogues to the ones shown in Figure 6.

Figure 13a also shows that there is a difference between the geometry of highs in the adjacent rift zones and highs of the horse-tail structures. While the rift zone highs do not tend to plunge along their axis, horse-tail-related highs distinctively plunge in the direction away from the transform fault. Furthermore, the size of the highs of the horse-tail structures is larger or at least equal to that of the rift-related highs. Their finger-shaped ridges plunge into the system of the pull-aparts of the horse-tail structure.

Furthermore, the highs of the associated and joining horse-tail structures (see Fig. 13b for location) are different from each other (Fig. 13a). The
associated horse-tail structure has much larger highs than the joining horse-tail structure.

Analogous to the situation along the Romanche transform margin in the Equatorial Atlantic (see Fig. 12), one of the horse-tail highs forms a marginal ridge, which divides the oceanic crust from the failed basins of the pull-apart terrain. Serial cross-sections through the centre of the Coromandal transform margin segment and its southern horse-tail structure (Fig. 14a–d) show that, while the central portion of the transform fault does not have such a marginal ridge (see Fig. 14a), the southern horse-tail structure does (see Fig. 14b–d). Cross-sections in Figure 14b–d show how its elevation decreases away from the pure transform fault zone and how it is progressively divided from the mainland by a wider and wider pull-apart basin. A similar plunging character of the marginal ridge can be seen in serial cross-sections through the Ghana Ridge in the Equatorial Atlantic.

Another important control of the structural architecture is the fact that faults that are needed
Fig. 14. (a) 2D cut through the portion of the Coromondal transform zone characterized by the maximum total strike-slip displacement (modified from Nemčok et al. 2012b). Note a general lack of pull-apart basins, apart from a small one above the pinching-out lower crust. ucc, upper continental crust; lcc, lower continental crust; poc, proto-oceanic crust.
Fig. 14. (b) 2D cut through the Coromondal horse-tail structure close to the strike-slip fault zone (modified from Nemčok et al. 2008). Note a distinct marginal ridge divided from the continental hinterland by a deep pull-apart basin. BS, Top basement; TS, top syn-rift strata; LK, top lower Cretaceous strata; K, top Cretaceous strata; P, top Palaeogene strata; LM, top Lower Miocene strata; IN, intra-Neogene marker horizon.
Fig. 14. (c) 2D cut through the Coromondal horse-tail structure a bit further from the strike-slip fault zone, in an oblique-slip portion (modified from Nemčok et al. 2008). See (a) and (b) for captions. Note a less distinct marginal ridge divided from the continental hinterland by less pronounced pull-apart basin.
Fig. 14. (d) 2D cut through the Coromondal horse-tail structure rather far from the strike-slip fault zone, somewhere in the transition from the oblique-slip portion to normal fault portion (modified from Nemčok et al. 2008). See (a) and (b) for captions. Note a subtle marginal ridge divided from the continental hinterland by a small pull-apart basin.
Fig. 15. Snap shots of the St Paul and Romanche transform fault activity map done for (a) Albian to Cenomanian/Turonian boundary and (b) Coniacian to Campanian.
for accommodation of the lateral clearance of
the spreading centres along the transform margin
cease their activities at different times, depending
on their location between the landward and ocean-
ward ends of the transform (Fig. 15). While they
are active they further propagate their ends. There-
fore, the oceanward faults of the transform fault
zone cross-cut all pre-existing faults created during
the entire history of the transform fault zone devel-
opment. Furthermore, strike-slip faults that were
needed for the accommodation of the lateral clear-
ance of the spreading centres post-date the breakup
unconformity more and more in an oceanward
direction (Fig. 16). This is contrary to normal fault-
ing at an extensional margin that dies out at breakup
and is topped by the breakup unconformity.

Development of the transform fault zone is
frequently controlled by pre-existing anisotropy
because continental crust always carries a deforma-
tional record of older events. It may provide planar
weaknesses of orientation that can be reactivated
by transform fault zone-controlling stress field.
The smaller-scale examples come from the studies
of Caby et al. (2008), Antobreh et al. (2009), Davieson et al. (2015) and Nemčok et al. (2016) docu-
menting how the propagating Romanche transform
fault system cross-cut or reactivated individual
elements of the pre-existing structural grain devel-
oped by older tectonic events, such as the Pan-
African orogeny.

The larger-scale example comes from the study
of Nemčok et al. (2015a). It documents that the
stress field controlling the initial development of
the transform fault system in initiating the Equato-
rial Atlantic developed a right-stepping system of
ENE–WSW dextral strike-slip fault zones. How-
ever, their P shear instead of R shear orientation,
which is the one that starts the strike-slip fault
development in homogeneous crust represented
by analogue material models (see Christie-Blick &
Biddle 1985; Naylor et al. 1986), indicates that
their positions were affected by prominent crustal
weaknesses of NE–SW strikes associated with
Proterozoic and Palaeozoic orogenies (e.g. Mascle
et al. 1988; Genik 1992; Guiraud & Maurin 1992).
The Late Cretaceous reactivation of the NE–SW
striking Sobral-Pedro II Fault in the Borborema
province and NNE–SSW striking Pan-African shear
zones (Ball 1980; Miranda et al. 1986; Caby 1989)
provides a specific reactivated fault example.

Thermal regime

While the thermal regime of the extensional margin
can be characterized by the rifting-related warm-
ing event reaching its maximum during continental
breakup and subsequent cooling, the thermal regime
of the transform margin includes one more thermal
perturbation, which takes place after the breakup. It
is controlled by the lateral passage of the spreading
centre together with two oceanic crust corridors
along the continental–oceanic transform (Mascle
et al. 1988, 1996; Todd & Keen 1989; Basile et al.
1998; Bouillini et al. 1998; Clift et al. 1998; Clift
& Lorenzo 1999; Bigot-Cormier et al. 2005;

\[ \text{Fig. 15. Continued.} \]
In theory, frictional heating along the transform should also contribute to the thermal perturbation (Carslaw & Jaeger 1959; Lachenbruch & Sass 1980; Kanamori 1994; d’Alesio et al. 2004) although its effect is rather minimal (Todd & Keen 1989; Saffer et al. 2003; Williams et al. 2004, 2005; Zoback et al. 2011). Apart from thermal conduction-controlled perturbation, there is evidence for the effect of hydrothermal fluid flow mechanisms (James & Silver 1988; Kharaka et al. 1988a, b; Torgersen & Clarke 1992; Moore & Reynolds 1997; Lespinasse et al. 1998; Hein et al. 1999; Robinson & Santana de Zamora 1999; Hannington et al. 2001; Pletsch et al. 2001; Kuhn et al. 2003; Zoback & Hickman 2007; Wiersberg & Erzinger 2011), the extent of which, however, remains unknown.

Ignoring the hydrothermal circulation, the main factors controlling the post-breakup thermal history of the transform margin are (Nemčok et al. 2012b, 2015b):

1. the effect of the pre-rift thermal regime on the cooling of all subsequent thermal events;
2. regional cooling of the elevated thermal regime developed by rifting in either pull-apart or rift terrains neighbouring the transform that culminated at continental breakup;
3. heating by the passing-by spreading centre controlled by its passage velocity; and

Fig. 16. The early syn-drift deformation of the previously extended continental crust in offshore Benin (modified from Nemčok et al. 2012b). The dextral-transpression-driven fold has chevron geometry. The Upper Cretaceous–Lower Palaeogene sediments thinning over its crest indicate its Senonian–early Palaeogene growth. The fold shows hardly any erosion as early Palaeogene sediments seem to be removed by gravity gliding. The scale and location are not given for confidentiality reasons. Note that some of the strike-slip faults did not die out during the continental breakup but were active much longer. Their activity was needed for the accommodation of the lateral clearance of the spreading centre along the continental–oceanic transform progressively changing to transform margin.
Petroleum systems

Underappreciated factors controlling petroleum systems elements at transform margins are:

1. post-breakup uplift distribution in space and time;
2. structural architecture and associated topography; and
3. diachronous timing of the transform fault activity.

The distribution of post-breakup uplifts at transform neighbour margins and the transform margin itself controls the occurrence of early syn-drift source, seal and reservoir rocks. In the case of the transform margin that involves minimal post-breakup uplift, the only uplifting mountain ranges preventing terrestrial sediments from reaching the oceanic basin will be located along its pull-apart or rift neighbours. In the case of the transform margin with early syn-drift transpression (see Mascle & Blarez 1987; Antobreh et al. 2009; Nemčok et al. 2012b), the only gaps among marginal uplifts will be located at transform contacts with neighbour pull-apart or rift margins. The gaps could be characterized by enhanced amount of sediment funneled into the oceanic basin. The example comes from the early post-breakup Jubilee depositional system in the Deep Ivorian Basin, just a bit north of the Romanche transform margin.

While transform margins unaffected by such uplift can have sediment entry points located fairly early after the breakup, the sediment transport pathways of those with uplift start with geometries passing along uplifted mountain ranges and change towards being transversal only after their erosional removal. Examples of the former come from offshore Guayana. Examples of the latter come from the Santos basin and Niger Delta regions. The transitional pattern between the initial and final stages can change in time. This is because the erosional removal is characterized by the lateral retreat of the mountain (e.g. van Balen et al. 1995; Burov & Cloetingh 1997; Mohriak et al. 2008) synchronous to its topography reduction owing to lateral erosion being much faster than vertical erosion (see Saunders & Young 1983). Evolving sediment transport patterns react to this change by progressive adjustments. While relatively short transversal pathways may not have sufficient length for cleaning the sediment transported into the oceanic basin, relatively long pathways running around mentioned obstacles may have sufficient length for efficient sorting of the transported sediment.

With regards to potential source rock deposition, marginal uplifts can protect the oceanic deposition from terrestrial input, turning it into more source rock prone. With regards to the distribution of seal rock, marginal uplifts taking their time to undergo erosional removal can considerably slow down the post-breakup transgression of younger sediments on the top-rift unconformity. While extensional margin segments can wait a considerable time for onlapping seals, normal transform margins can have them available much earlier. Further landward of potential marginal uplifts, sag basins on top of the top-rift unconformity at extensional margin segments may provide seals, while such sag basins are missing at transform margin segments.

Ignoring the role of sediment provenances in the character of sediment transported into the oceanic basin for the moment, we can discuss the role of structural architecture. For comparison, extensional margins can be characterized by wide shelf, low-dip and wide slope, and no shortage of various sediment catchment features (Addis et al. 2013a, b; Dorie McGuiness 2013, pers. comm.; Phil Towle 2013, pers. comm.). The sediment transport length from the shoreline to the oceanic basin floor is relatively long. On the contrary, the transform margins have narrow shelves and slopes. Sediment catchment features in both areas can be missing (see Addis et al. 2013a, b; Dorie McGuiness 2013, pers. comm.; Phil Towle 2013, pers. comm.). The sediment transport length from the shoreline to the oceanic basin floor is relatively short. This difference results in rather different depositional models of these two margin types (Fig. 17) with an impact on different reservoir rock distributions.

With respect to migration patterns, pull-apart segments among transform margin segments, owing to their geometry, can undergo divergent to parallel migrations. Transform margin segments can undergo parallel migration. The most typical place...
Fig. 17. Comparison of depositional systems characterizing extensional and transform margins (Addis et al. 2013a, b). (a) The extensional segment is characterized by the palaeobathymetrically lower slope environment and longer sediment transport from shelf to basin floor. Intra-slope basins are typical. Typical is the prograding shelf connection with slope channel–levee complexes. Fairly long compensationally stacked, slope channel–levee complexes are typical. They are wide and rather long. Also typical are narrow and highly sinuous channels with extensive levees. See text for further explanation. (b) The sheared segment is characterized by palaeobathymetrically higher slope environment. It is characterized by a straight shot from shelf to deep basin floor and shorter transport from shelf to basin floor. Intra-slope basins are not typical. Vertically stacked, slope channel–levee complexes are rather narrow and not very long. This case contains robust distributary complexes on the basin floor. See text for further explanation.
of the convergent migration is the associated horse-tail structure (Fig. 13b). According to their structural architectures, the former two provinces can rely on one or two source rock kitchens while the horse-tail structures at both transform margin ends can tap into a larger amount of source rock kitchens.

Further complexity for the transform margin comes from the comparison of a static depositional model in Figure 17b with the transform fault activity distribution model in Figure 16. This brings the third factor, diachronous timing of transform fault activity, into the picture. It allows us to take the distal portions of the depositional system and juxtapose them to the rest of the depositional system along the transform fault zone. Figure 16 would then indicate when these distal portions become disconnected from their feeder system and laterally transported away from it, and when this mechanism stops, which both take place diachronously along the transform. The stoppage is oldest on its

Fig. 18. Late Albian snapshot of the Equatorial Atlantic development. (a) plate reconstruction. (b) Lower Cretaceous sediment thickness map.
landward side and youngest on its oceanward side. For reservoirs still moving away from the feeder system, the top seal may not be a problem, as they can be sealed by deep basin shale deposition in areas away from the feeder system. However, it may become a problem where the continuous coarser sediment transport from the undisconnected feeder system keeps preventing the top seal from deposition.

The diachronous last activity timing of all faults of the transform fault zone results in different depositional conditions at different segments of the transform margin. The ongoing deposition reacts to structural highs, controlled by anticlinal growth associated with wrenching, rise of other types of structural highs driven by various wrenching-related uplift mechanisms and lows controlled by numerous wrenching-related subsidence mechanisms.

This diachronous timing also controls an oceanward-younging age of structural traps along the transform margin. Any early syn-drift transpressional events further complicate this pattern of structural traps in comparison to transform margins without such events.

A lateral passage of spreading centres along the continental–oceanic transform fault zones could have a prominent control on the deposition of potential source rock. This takes place when the early syn-drift centres are either subaerial or shallow marine with distinct positive topography. Together with continents (Fig. 18), laterally clearing their contact along the transform, spreading centres can relatively restrict developing oceanic basins landward of them to the point of them reaching dysoxic conditions. One of the typical natural examples is the distribution of Cenomanian and Turonian source rocks in the Equatorial Atlantic.

The timing of the second thermal transient related to the lateral passage of spreading centres together with sandwiching newly accreted oceanic crust adds quite a complexity to the expulsion timing pattern of source rock kitchens occurring along the transform margin. If completely dependent on this transient, ignoring other interplaying factors for the moment, their timing is oldest on the landward side and youngest on the oceanward side.

**Volume description**

This volume attempts to enhance the knowledge reviewed in the previous text by studying the African and South American conjugates of the Romanche and St Paul transform fault zone (Davison et al. 2015; Dickson et al. 2016; Nemčok et al. 2015b), the Guyana and Suriname transform faults zones (Dickson et al. 2016; Loncke et al. 2015; Nemčok et al. 2015a, c), the French Guyana hyper-oblique margin ( Sapin et al. 2016), transform margins of the Arctic region (Doré et al. 2015), the Andaman Sea transform margin (Morley 2015), the Coromandial transform margin (Sinha et al. 2015), transtensional margin between the Caribbean and North American plates (Sanchez et al. 2015), the Davie transform margin and its neighbour transform margins ( Reeves et al. 2016), studying transform behaviour by a review of multiple worldwide cases (Nemčok et al. 2016) and parametric numerical modelling (Henk & Nemčok 2015).

Discussing dynamic and kinematic developments, Sapin et al. (2016) take on a hyper-oblique margin. They study the subsidence history and its controlling factors, although some of them, such as the bulging in the Andean foreland and Amazon River-related sediment input into the oceanic basin, are area-specific. They document rather localized thinning of this margin type controlling the early syn-rift subsidence. Morley (2015) discusses a progressive development of the region affected by earlier extension and subsequent wrenching and transtension. They document how the transtension occurred in the regions where the initial extension resulted in volumes of ductile middle crust and how the wrenching coincided with either missing or considerably reduced ductile middle crustal layer, representing the regions of increased upper crust/lower crust coupling. Studying a multi-stage ocean opening, Doré et al. (2015) focus on the role of multiple consecutive tectonic events controlled by different stress regimes in microcontinent release. Utilizing a plate reconstruction technique and reflection seismic interpretation, Reeves et al. (2016) also study the role of multiple tectonic events on the breakup geometry. Nemčok et al. (2015a) document a role of wrench faulting in control of the ocean segment terminus. Focusing on the spatial and temporal development of the two competing rift zones trying to control the continental breakup, Sinha et al. (2015) discuss the role of the transform fault in the release of the microcontinent from the overlap region between the two zones. They document the kinematic evolution of all main transform zone faults in space and time, including their linkage to fault patterns of the two former rift zones. The theme of the role of wrenching in the release of microcontinents is further evolved in the review-based study of Nemčok et al. (2016), who focus on the microcontinent release in the settings, including the competing rift zones, competing horsetail structure elements, competing wrench faults and occurrence of multiple consecutive tectonic events.

Structural architecture with its controlling factors is documented by Loncke et al. (2015) who focused on distribution and the timing of gravity gliding in the circum-transform regions and associated depositional processes and fluid seepage in space and time. Sanchez et al. (2015) discuss the
evolution of the transtensional margin and its effect on the resultant structural architecture. Similarly to Davison et al. (2015), they observe the effect of a pre-existing weakness reactivation on the final structural architecture.

The role of various factors controlling the thermal regime of transform margins is discussed by Nemčok et al. (2015b), who focus on the role of pre-rift thermal regimes, cooling patterns of the pull-apart terrains and connecting transforms, heating by the passing-by spreading centre and heating by the passing-by newly accreted oceanic crust. The theme of thermal regimes at transform margins is expanded into their role in the spatially and temporally varying distribution of ductile lower crust by Henk & Nemčok (2015).

Discussing petroleum systems of transform margins, Davison et al. (2015) go through documented multiple reactivation events of pre-existing pre-rift, syn-rift and post-rift faults of the transform margin as it continues evolving in time, affecting the distribution of reservoir rock and traps by spatially and temporally varying sediment pathways and located erosional events. It is also interesting to read Dickson et al. (2016) documenting the distribution of potential source rock depositional environments indicating significant asymmetry with respect to their presence at conjugate sides. Important asymmetry further seems to affect the distribution of syn-rift architecture, i.e. syn-rift related play concept elements. Petroleum migration at transform margins is the focus of Nemčok et al. (2015c), who observe lateral migration-dominated regions with post-rift strata at transform margin segments undisturbed by gravity gliding, which are typical for these segments and vertical migration-dominated regions overlapping with an occurrence of strata deformed by gravity gliding that either occurs rather late in transform margin history or never.

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