



**HAL**  
open science

## Transform marginal plateaus

L. Loncke, W.R. Roest, F. Klingelhoefer, C. Basile, D. Graindorge, A. Heuret,  
Boris Marcaillou, T. Museur, A. S. Fanget, M. Mercier de Lépinay

► **To cite this version:**

L. Loncke, W.R. Roest, F. Klingelhoefer, C. Basile, D. Graindorge, et al.. Transform marginal plateaus. *Earth-Science Reviews*, 2020, 203, pp.102940. 10.1016/j.earscirev.2019.102940. hal-02532555

**HAL Id: hal-02532555**

**<https://hal.science/hal-02532555>**

Submitted on 13 Apr 2021

**HAL** is a multi-disciplinary open access archive for the deposit and dissemination of scientific research documents, whether they are published or not. The documents may come from teaching and research institutions in France or abroad, or from public or private research centers.

L'archive ouverte pluridisciplinaire **HAL**, est destinée au dépôt et à la diffusion de documents scientifiques de niveau recherche, publiés ou non, émanant des établissements d'enseignement et de recherche français ou étrangers, des laboratoires publics ou privés.

---

## Transform marginal plateaus

Loncke L. <sup>1,\*</sup>, Roest Walter <sup>2</sup>, Klingelhofer Frauke <sup>2</sup>, Basile C. <sup>3</sup>, Graindorge David <sup>4</sup>, Heuret A. <sup>5</sup>,  
Marcaillou B. <sup>6</sup>, Museur Thomas <sup>2,4</sup>, Fanget Anne-Sophie <sup>1</sup>, Mercier De Lépinay M. <sup>1</sup>

<sup>1</sup> Univ. Perpignan Via Domitia, Centre de Formation et de Recherche sur les Environnements Méditerranéens (CEFREM), UMR 5110, 52 Avenue Paul Alduy, 66860 Perpignan, France

<sup>2</sup> IFREMER, Unité de Recherche Géosciences Marines, 29280 Plouzané, France

<sup>3</sup> Univ. Grenoble Alpes, Univ. Savoie Mont Blanc, CNRS, IRD, IFSTTAR, ISTERRE, 38000, France

<sup>4</sup> Laboratoire Géosciences Océan (UMR CNRS 6538), Université de Bretagne Occidentale and Institut Universitaire Européen de la Mer, Place Nicolas Copernic, 29280 Plouzané, France

<sup>5</sup> Université de Guyane, Géosciences Montpellier, 97300 Cayenne, France

<sup>6</sup> Université Côte d'Azur, CNRS, Observatoire de la Côte d'Azur, IRD, Géoazur, 250 Rue Albert Einstein, 06560 Valbonne, France

\* Corresponding author : L. Loncke, email address : [lies.loncke@univ-perp.fr](mailto:lies.loncke@univ-perp.fr)

---

### Abstract :

Numerous submarine plateaus form highstanding bathymetric highs at continent to ocean transitions. Due to their proximity to continents, they have been frequently labelled “marginal plateaus”, although this term has not been clearly defined or associated with a specific geology or geodynamic process. Until now, these elevations have been interpreted as submerged thinned continental fragments detached from continents, basaltic buildups formed by hotspots, volcanic margins or oceanic plateaus. Many of these plateaus formed at transform margins connecting oceanic basins of contrasted ages.

We propose for the first time to define and review a class of marginal plateaus related to a specific tectonic setting: “Transform Marginal Plateaus” (TMPs). Based on a compilation of 20 TMPs around the world, we show that most of them have a polyphased history and have undergone at least one major volcanic phase. Our review highlights in particular a hitherto unrecognized close link between hotspots, volcanic activity and transform margins.

We also propose that, due to their polyphased history, TMPs may contain several successive basins and overlooked long-lived sedimentary archives. We finally highlight that, because these TMPs were transform plate boundaries perpendicular or oblique to surrounding rifts, many of them were close to last-contact points during final continental breakup and may have formed land bridges or bathymetric highs between continents. Therefore, we discuss broader scientific issues, such as the interest of TMPs in recording and studying the onset and variations of oceanic currents or past biodiversity growth, bio-connectivity and lineage evolution.

**Keywords** : transform marginal plateau, transform margin, volcanic margin, Large Igneous Province, continental breakup, paleoceanography, land bridge, gateway

## Introduction:

Many anomalous elevations occur in oceanic domains: submerged continental fragments detached from adjacent continents (e.g., the Jan Mayen ridge), ancient island arcs (e.g., the Aves Ridge), basaltic buildups formed by hotspots and spreading centers (e.g., the Cocos Ridge, Iceland) (review in Nur & Ben Avraham, 1982). Most intraplate oceanic plateaus relate to hotspot activity (e.g., Kerguelen, Broken Ridge, Ontong-Java) and their formation is relatively well-understood (refs à completer, Self et al., 2015?).

In turn, many submarine plateaus are emplaced near continent to ocean transition domains, forming spurs at the scale of 50 to up to 1000 km in length and their nature and origin is less well understood and may involve important issues related to continental breakup conditions, for example.

Although not strictly defined, the term “Marginal Plateau” has frequently been used in literature to describe those submarine elevated features located close to continents: the term “marginal plateau” first appeared in a publication by Ewing et al. (1971) to describe the Falkland submarine plateau near the Falkland-Malvinas margin. Later, Veevers and Cotterill (1978) use the same term to describe the Northern Scott, Exmouth, Naturaliste and Wallaby plateaus, which look like continental prolongations offshore Australia. In 1979, Dingle and Scrutton defined the Goban Spur SW of Ireland as a marginal plateau. They described it as a “rectangular-shaped upstanding block which juts oceanwards from the southwestern edge of the Celtic continental margin”. Eldholm et al. (2002) use the same term to characterize the Vøring Plateau, which is bounded by the Jan Mayen Fault Zone and was the location of a major volcanic event during breakup. These authors write that “Off central Norway the slope is broken by the large Vøring marginal plateau at 1000-1500 m water depth”. The term “marginal plateau” can also be found in hydrographic international charts. They are defined in the joint international IHO report as “flat or nearly flat elevations of considerable extent, dropping off abruptly on one or more sides” (IHO, 2008). The crustal nature of these “marginal plateaus” and their relation with breakup is debated: some authors relate them to thinned continental crust domains (Demerara plateau, Greenroyd et al., 2007 and 2008). Others propose that they could correspond to thickened oceanic crust domains (Falkland-Malvinas plateau, Schimschal and Jokat, 2018) or underplated and intruded continental crust (Vøring plateau, Berndt et al., 2001).

More recently, Mercier de Lépinay et al. (2016) observed, in a worldwide transform continental margins review, the presence of anomalous submarine plateaus for at least a quarter of the transform margins they identified. This resulted in defining a specific class of plateaus with, for the first time, both physiographic and tectonic criteria (Mercier de Lépinay et al., 2016). These “marginal plateaus” have rectangular or triangular shapes and are systematically bounded by a transform margin on one of their sides (Figure 1 - and KML suppl. file). Mercier de Lépinay et al. (2016) also showed that many of these marginal plateaus are located at the intersection of large ocean basins with different breakup ages and/or opening

directions. Therefore, the authors proposed that marginal plateaus may consist of fragments of continental crust, thinned and individualized by several successive rifting and transform deformation episodes.

We propose in this review to follow the definition of marginal plateaus suggested by Mercier de Lépinay et al. (2016), but to rename them “Transform Marginal Plateaus” (TMPs) to be more accurate and separate them from other potential plateaus not associated with transform margins. We also propose to re-evaluate the TMP genesis hypothesis proposed by Mercier de Lépinay et al., 2016, and to integrate the fact that several authors have emphasized that many of these marginal plateaus were not only transform but also volcanic margins (examples of the Demarara, Vøring, Exmouth, Rockall marginal plateaus, Reuber et al., 2016; Klingelhoefer et al., 2017; Eldhom et al., 2002; Berndt et al., 2001; Sydmonds et al., 1998; Elliott and Parson, 2008, Elliott et al., 2017).

In this context, the first aim of this paper is to draw a synthesis of what we know about TMPs: their nature, their geodynamic significance, their relation to large igneous provinces and the sedimentary archives they may host. Because these TMPs were transform plate boundaries, likely forming last-contact points between separating continents, we also propose to evaluate their role in paleoceanography and the conditions in which some may have formed land bridges for animals.

In the following sections we (1) update and improve the inventory of transform marginal plateaus of Mercier de Lépinay et al., 2016, (2) review what is known about the nature and tectono-magmatic history of each of the identified TMPs, (3) compare them to each other in order to discuss their structure and the role of crustal thinning and volcanism in their evolution, and (4) discuss the role that these plateaus, which have a high elevation compared to the surrounding ocean floor, might have on paleogeography and oceanic circulation.

## **1. Updated inventory of Transform Marginal Plateaus**

### **1.1. *Reminder of definition criteria***

Our inventory of TMPs follows the definition proposed by Mercier de Lépinay et al. (2016), defining TMPs as “deep, planar and sub-horizontal plateaus located between the platform and the lower continental slope, (...) and bounded on one of their sides by a transform margin”. This notion is illustrated in Figure 2 by the example of the Demerara transform marginal plateau where the landward edge of the marginal plateau is located beyond the shelf break and the seaward edge along the continental slope upper break. This typical TMP bathymetric profile is compared to the bathymetric profile off the Ceara transform margin (Brazil) devoid of a marginal plateau. In order to constrain the landward edge of TMPs in our compilation, the 300 and 500 m isobaths were used as guides. Bathymetric sections perpendicular to the coast were used to define and localize the slope-break between the upper slope and marginal plateau, based on the GEBCO 2014 digital bathymetry data set (The GEBCO\_2014 Grid, version 20141103, <http://www.gebco.net>).

The transform edges of marginal plateaus mostly follow the transform margins identified by Mercier de Lépinay et al. (2016) (KML files available in supplementary material 1 and 2). Finally, the geometry of each transform marginal plateau was completed in general conformity with the first two limits (landward and transform edges). Where appropriate, the global continent-ocean boundary compilation by Seton et al., 2012, and updated by Müller et al., 2016, was used to assist in locating the seaward edges. This was particularly the case when the junction between the lower slope and the deep ocean floor was not clearly expressed morphologically, such as along the eastern edge of the Sao Paulo Plateau or the southern edge of the NE Greenland Plateau. Note that, since the transform margin traces identified by Mercier de Lépinay et al. (2016) are placed at the junction of the lower continental slope and the deep ocean floor located beyond, the lower slope is included within the general shape of the TMPs in our inventory.

Once the general shape of each TMP was established (KML file available in supplementary file 2), some basic parameters were determined to provide morphological characteristics for each plateau. Table 1 shows the TMPs identified in this compilation with the relevant characteristics. The bathymetric profiles that are roughly perpendicular and parallel to the margin for each TMP are shown in supplementary file 3. Along each of these profiles, we identify the different elements where expressed: the shelf break, marginal plateau, lower continental slope and the ocean floor beyond (rise or abyssal plain).

Given their sometimes irregular plan shape, for each of the TMPs we provide an estimate of the length (measured perpendicular to the margin, in the general direction of opening) and width (measured parallel to the margin). For a more quantitative analysis, we also calculated, using the outlines digitized as shape files in QGIS, the total surface area, the average depth and its standard deviation as simple parameters that allow us to compare different marginal plateaus. The average depth and its standard deviation were determined for each TMP using the GEBCO 2014 grid. We consider that standard deviation is a better measure for the variability in depth of each marginal plateau than minimum and maximum depths, as these extreme values may represent outlier values in the grid, which could be artefacts.

In addition to the average depth of each TMP, we also calculated the average depth of the seafloor surrounding it. A buffer was created for each plateau, expanding its geographic shape outward. This buffer was calculated using the QGIS “create buffer” plug-in. The average depth of the surrounding ocean floor was then calculated within a polygon that represents only the seaward part of the buffer area. For each TMP, the extent of this buffer area is shown in the inset map in supplementary file 3 as a transparent light blue band seaward of each marginal plateau. Table 1 lists the average depth of the ocean floor surrounding each TMP thus calculated.

Although the exact outline of each of the TMPs can be questioned, slight variations in the outlines would only lead to small uncertainties in the average depth and surface extent of each marginal plateau. We therefore feel confident that the parameters provided in Table 1 are robust within an error margin of a few percent.

The construction of a detailed outline of each transform marginal plateau also led us to slightly update the transform margin inventory of Mercier de Lépinay et al. (2016) (KML file in supplementary file 1). Using the same methodology, with higher resolution bathymetric and gravity grids, or updated information on some continent to ocean transitions, the location and orientation of some segments were slightly adjusted. The new TMP inventory and the updated transform margin inventory are provided in KML format in the supplementary data section.

## 1.2. *Modifications and limits*

Compared to the compilation of Mercier de Lépinay et al. (2016), we have removed the following four plateaus, initially identified as transform marginal plateaus, for the reasons outlined below:

- The Newfoundland Plateau: this plateau is located at the junction of two ocean basins of different ages and is bounded by a transform margin. However, the Newfoundland Plateau is in fact a very wide and shallow continental shelf. Unlike for TMPs, where the shelf break is located landward of the plateau, in the case of the Newfoundland Plateau, the shelf break occurs at the seaward edge of the plateau identified by Mercier de Lépinay et al. (2016). (See supplementary file 4 for bathymetric profiles crossing the Newfoundland Plateau, illustrating its shallow depth all along its sides).
- The Sierra Leone Plateau was also dismissed, being, like the Newfoundland Plateau, a wide and shallow continental shelf rather than a marginal plateau. (See supplementary file 4 for bathymetric profiles crossing the Sierra Leone Plateau, illustrating its shallow depth all along its sides).
- Blake Plateau was removed, due to significant uncertainties in the transform margin location, drawn parallel to the Blake Spur Fracture Zone by Mercier de Lépinay et al. (2016) but not clearly expressed in the bathymetry or gravimetry.
- Given the lack of knowledge related to the Bruce Rise offshore Antarctica, notably concerning its tectonic evolution and the presence of a transform segment, we decided not to include it in our current compilation of TMPs.

On the other hand, we have added the Liberia, Sao Paulo and Morondova TMPs in our compilation. They were not included in the Mercier de Lépinay et al., 2016, inventory, but they meet all the defined bathymetric and tectonic criteria, as will be discussed later.

We also considered the cases of other marginal plateaus that we identified in bathymetry data and that may look like TMPs or non-transform marginal plateaus: respectively the Cormorin, Galicia, Landes, Mozambique, Madagascar, Campbell, Clatham and Challenger plateaus:

- The status of Galicia Bank and Landes Plateau could not be resolved because one of their edges, potentially transform, has been obliterated by plate boundary reactivation.
- Uncertain kinematic reconstructions and lack of geological knowledge did not allow us to decide if the Cormorin Plateau, South of India, meets the criteria.
- Although the Mozambique Plateau shares all the physiographic characteristics of TMPs, this plateau is bounded by an oceanic FZ and not by a transform margin.
- The Madagascar Plateau represents a similar case, and is, moreover, not clearly and sharply bounded by a fracture zone.
- Several continental wide marginal plateaus surround New-Zealand: to the west, the Challenger Plateau, to the east, the Campbell Plateau and Chatham Rise (Sutherland, 1999; Mayes et al., 1990). The western Challenger Plateau is clearly not a transform margin (see Cretaceous reconstructions in Sutherland, 1999). Its eastern edge is complex and involved in subduction. The Campbell Plateau has never been described as prolonging a transform margin: (1) its western edge is interpreted as a slightly oblique Eocene passive margin emplaced along the Southeast Tasman Emerald Basin propagator (review in Lebrun et al., 2003), (2) its southeastern edge, a conjugate of Marie Byrd Land, is clearly not a transform margin, (3) its very northeastern tip could have been a transform margin in the prolongation of the Ballons Island and Antipode Fracture Zone (see Cretaceous reconstructions in Sutherland, 1999). However, the Bounty Trough and the onset of subduction along Hikurangi interrupt the continuity of this potential margin and it is difficult to postulate about the transform nature of this marginal plateau. Finally, the northern edge of the Chatham Rise has been reactivated by the Hikurangi Plateau (LIP) subduction and is even more difficult to classify.

Overall, we dismissed several potential TMPs from our inventory due to poor geodynamic knowledge or complex tectonic reactivations. We decided not to include marginal plateaus in our compilation in case of doubt as to their transform margin character.

## **2. Review of the geological nature and history of transform marginal plateaus**

The geological knowledge of these transform marginal plateaus is highly variable. The plateaus in the Arctic and Antarctic are the least well known due to the

paucity of data (limited accessibility due to climate conditions). Other plateaus, such as the Falklands-Malvinas, reach such important sizes that their knowledge is understandably only partial. In some rare instances, crustal-scale images are available along both the transform and the divergent/oblique borders. A short description of the identified TMPs is presented below. The main characteristics of each plateau are summarized in Table 1, which also includes a list of related key references. Three figures support this description: Figure 3, which positions each of the identified TMPs on a global breakup age map (ages from Müller et al., 2008, Müller et al., 2016, completed by Marks and Tikku (2001), Gibbons et al. (2013), Labails et al. (2010) and Davison (2010)) ; Figure 4, which shows detailed bathymetric views centered on each of the TMPs (labelled seafloor ages from Müller et al., 2008, Müller et al., 2016, completed by Marks and Tikku (2001), Gibbons et al. (2013), Labails et al. (2010) and Davison (2010)); and Figure 5, which presents, where available, cross-sections with crustal information (all sections are presented at the same scale and with a common legend). Supplementary file 3 shows bathymetric profiles crossing each TMP.

## 2.1. Arctic and North Atlantic Oceans

### Morris Jesup Rise and Yermack

The Yermack and Morris Jesup TMPs are conjugate plateaus bounded by two oceans of different ages: their northern borders and the southeastern border of Morris-Jesup are bounded by a Paleocene to Oligocene oceanic crust belonging to the Arctic Ocean, while their southern borders are bounded by the early-middle Miocene Fram Strait oceanic crust. The Fram Strait opened in a very oblique mode which resulted in the formation of conjugated transform margins and the oblique separation of the Yermack and Morris-Jesup plateaus (Figure 4 (1-2)).

There is a lack of data for both plateaus and our knowledge about their structure is sparse. They underwent a polyphased history with a first deformation in the late Paleozoic (Barents Sea) followed by Cretaceous stretching (Engen et al., 2008). To the south of them, a Paleocene-Eocene breakup occurred in combination with a major volcanic event (North Atlantic Igneous Province) and transform motion south of the plateaus. Their separation took place in the early-middle Miocene during the final opening of the Fram Strait, following a prolonged period of NW-SE oblique rifting (Engen et al., 2008, Faleide et al., 2008; Dossing et al., 2010). Whether the Morris Jesup and Yermack plateaus represent protrusions of stretched continental crust or Eocene-late Oligocene thickened oceanic crust is still debated (Jackson et al., 1984; Engen et al., 2008; Faleide et al., 2008). Based on gravity and magnetic data, Engen et al. (2008) proposed that these plateaus are made of stretched continental crust and that their high amplitude magnetic anomalies relate to volcanics (Arctic large igneous province?) (Figure 5-A). ODP leg 151 drills only reached the Pliocene sedimentary cover of the Yermack Plateau (Table 1). Dredges along Yermack Plateau basement highs and borders recovered magmatic and metamorphic rocks and sandstone (Jokat, 1999 and Riefstahl et al., 2013). Those samples have been attributed to a stretched continental crust strongly affected by alkaline magmatism, about 51 Myrs old (Riefstahl et al., 2013). The Morris Jesup Rise has never been sampled by drilling.



## Vøring and NE Greenland

The Vøring TMP emplaced along the North Atlantic Cenozoic ocean (Figure 4-4) and is bounded to the south by a transform margin prolonged towards the ocean by the Jan Mayen Fault Zone (JMFZ, Figure 4-4). It is the conjugate of Greenland volcanic margin and Jan Mayen Ridge (Table 1).

Based on wide-angle and deep-penetrating seismic profiles, the Vøring Plateau is proposed to be formed of stretched and highly intruded continental crust covered by early Eocene basalts and underplated by mafic intrusions (Faleide et al., 2008; Abdelmalak et al., 2017). The conjugated continental margins off Norway and Greenland experienced a long history of post-Caledonian extension (since the Devonian) until breakup in early Cenozoic times (Brekke, 2000; Skogseid et al., 2000; Hamann et al., 2005; Tsikalas et al., 2005a and b; Faleide et al., 2008). Proterozoic inherited transfer zones govern the segmentation of those margins and the shape of the Vøring and NE Greenland marginal plateaus. The complete lithospheric breakup occurred during the Paleocene-Eocene transition and was associated with a major magmatic event. Volcanics cover large areas of the continent-ocean transition on both sides (Eldholm et al., 2002; Faleide et al., 2008; Upton et al., 1995). These magmatic structures are part of the North Atlantic Large Igneous Province.

Berndt et al. (2000 and 2001) investigated the crustal structure of the outer Vøring Plateau using wide-angle (WAS) and multi-channel seismic (MCS) data. They propose that this outer domain consists of a 20 km thick underplated continental crust (Figure 5-B). A 10 km thick and 50 km long high-velocity body was identified near the Jean Mayen Fault Zone (JMFZ), providing evidence for a local decrease of volcanism in the vicinity of the JMFZ bounding the southern Vøring Plateau in comparison to the nearby divergent highly volcanic segments (Møre and Vøring margins respectively, Figure 4-3). The authors propose that the incipient JMFZ developed during breakup in a cold lithosphere setting between two thermal upwelling zones under the Møre margin rift zone and a pull-apart basin further west, where volcanism is abundant. DSDP, ODP and IODP drills (Table 1 for details) allowed us to sample pieces of the Vøring TMP, in particular the “landward flows” and inner SDRs corresponding to subaerially emplaced flood basalts (Hole 642E) (Eldholm et al., 1989, Mutter et al., 1982, seismic volcanostratigraphy terminology defined in Planke et al., 2000) and the initial deeply intruded sedimentary basin emplaced just underneath those SDRs (Lower Series Flows, Abdelmalak et al., 2015) (Hole 642E).

No deep-penetrating seismic data (WAS and/or MCS) are available along the NE Greenland TMP (located in Figure 4-3). However, close to the TMP in the South, Schlindwein and Jokat, 1999, found evidence for a 30 km thick continental crust displaying magmatic underplating.

## Faroe-Rockall

The Faroe-Rockall TMP is a composite domain that comprises the Faroe, Rockall and Hatton geological structures (Figure 4-5). This plateau is bounded to the north and the west by the Cenozoic North Atlantic domain, and to the south by the older Cretaceous North Atlantic oceanic domain. The southern border of the Faroe-

Rockall Plateau corresponds to a transform margin formed at the junction of those two oceanic domains of contrasted ages (Figure 4-5). This transform margin corresponds and is extended by the Charlie Gibbs Fault Zone (CGFZ, Figure 4). The conjugates of this plateau are the South Greenland and Grand Banks/Labrador margins (Table 1). This TMP is bounded to the south by the CGFZ that separates an early stretched and magma-poor hyperextended margin to the south from a highly magmatic margin including the Hatton and Rockall banks to the north (Welford et al., 2012). The eastern CGFZ seems to abort towards the Porcupine Bank but trends parallel to the onshore Variscan front and Iapetus suture (Welford et al., 2012).

The Hatton-Rockall domain comprises a series of Mesozoic rift basins hosted within a metamorphosed Proterozoic-Palaeozoic Caledonian and Variscan basement (Welford et al., 2012). The details of the deep structure of this part of the plateau remain largely unknown due to the presence of a thick succession of Cenozoic flood basalts that cap it (Elliot, 2017). Elliot and Parsons (2008) highlight, however, an important segmentation of the plateau along the Hatton and Rockall banks (Figure 4-5). The authors differentiate three tectono-magmatic segments flanked by oceanic crust: (1) a southernmost domain displaying inner and outer Seaward Dipping Reflector (SDR) packages, an elongated outer high following the western edge of Hatton Bank (terminology following Planke et al., 2000 seismic volcanostratigraphy), and volcanic cones still visible in bathymetry, (2) a central domain interpreted as continental with no inner SDRs - the Hatton Bank block, and (3) a northernmost domain, showing a wider region of SDRs and volcanic cones. This segment is closest to the Iceland hotspot. These different segments seem to have formed diachronically according to the northward propagation of the Atlantic opening (Elliott and Parsons, 2008).

Deep-penetrating Multi Channel Seismic (MCS) and Wide-Angle Seismic (WAS) experiments show the lateral variability of the Hatton-Rockall domain, consisting of different crustal blocks with frequent underplating and SDRs along the western border of this plateau (Figure 5C-D-E-F). The Hatton Bank is characterized by a 20 to 25 km thick crust of continental nature underlain by a high-velocity body ( $V_p > 7.2$  km/s) on its western side (White et al., 2008, White and Smith, 2009, Figure 5F). To the north, towards Lousy Bank (Figure 5D), there is an ambiguity concerning the location of the Continent-Ocean Boundary (COB) due to important intrusions between the oceanic crust and the heavily stretched continental crust. West of the Lousy Bank, Klingelhoefer et al. (2005) and Funck et al., 2008 show a reduction in crustal thickness from ~25 km to 16 km with velocities suggestive of oceanic crust affinity and underplating (Figure 5D). The Lousy Bank consists of a 24 km thick continental crust, while the Rockall Trough is made of a 13 km thick stretched continental crust (Figure 5E). Volcanic underplating is also observed next to the Hebrides continental shelf (Klingelhoefer et al., 2005).

The Faroe block is located further north, including the Faroe Islands and surrounding shelf, and the Faroe-Iceland Ridge (FIR). The FIR is composed of a 30 km thick oceanic crust, with a gradual transition to ancient continental crust toward the Faroe Islands (Richardson et al., 1998). The upper 5 km of the Faroe Island crust consist of Tertiary basalts generated during continental breakup, overlying the continental crust. The lower crust, which is poorly constrained by seismic data, exhibits high seismic velocities (7.1-7.6 km/s) attributed to underplating or intrusions by mafic melts during continental breakup in the early

Tertiary (Richardson et al., 1998). More recent experiments image a 25 to 28 km thick continental crust along the Fugloy Ridge (emplaced between the Faroe Islands and the Hebrides continental shelf) with a high-velocity body ( $V_p > 7.2$  km/s) in the lower crust (White et al., 2008) (Figure 5C). Funck et al. (2008) show that the Faroe bank located southwest of the Faroe Islands consists in a 27 km thick continental crust (location in Figure 4-5).

DSDP drills reached the top basement along the Edora and Rockall banks (Table 1). Basalts were recovered in both sites. Along the Edora Bank, hole 554A drilled the “outer high” recovering subaerially emplaced volcanoclastic conglomerates and sandstones interbedded with lava flows. Dredges recovered early Proterozoic gneiss and granulites along the Rockall Bank and Paleocene/Eocene volcanic series along the Hatton basin and bank (Hitchen, 2004, Table 1).

## 2.2. Equatorial Atlantic

This oceanic domain opened around 110 Myr ago, between two older ocean basins: the Jurassic Central Atlantic Ocean to the north and the Early Cretaceous South Atlantic Ocean to the south (Basile et al., 2005). It is, therefore, called the “Equatorial Atlantic Gateway” in some publications.

### Demerara

The Demerara TMP is bounded to the north and east by the Cretaceous Equatorial Atlantic domain, and to the west by the older Jurassic Central Atlantic domain (Figure 4-6). This plateau derives from a double opening history: (1) its western boundary was formed at Jurassic times during the Central Atlantic opening, (2) its eastern and northern boundaries were formed during the transform-dominated Cretaceous Equatorial Atlantic opening (Gouyet, 1988; Campan, 1995; Greenroyd et al., 2007; Basile et al., 2013; Mercier de Lépinay, 2016). The Demerara Plateau is the transform margin conjugate of the Guinea Plateau: before the formation of the transform margin, the Guinea Plateau was located to the north of the Demerara Plateau (Table 1).

The structure and history of the Demerara Plateau is quite well known due to the availability of industry and ODP drill cores (Gouyet, 1988; ODP leg 159, Mosher et al., 2007), long offset deep-penetrating reflection and refraction seismic data (Greenroyd et al., 2007, 2008; Mercier de Lépinay, 2016, Reuber et al., 2016; Klingelhoefer et al., 2017), as well as academic exploration and sampling cruises (Loncke et al., 2009, 2016, Pattier et al., 2013; 2015; Tallobre et al., 2016). The basins hosted by this plateau have been particularly well described in comparison with those of other TMPs.

Two regional profiles, based on the interpretation of industry and academic seismic data, cross the entire Demerara Plateau (Mercier de Lépinay, PhD, 2016) (Figure 6). These lines illustrate the deep structure of Demerara, showing two breakup unconformities: one related to the first Jurassic opening event (in purple on Figure 6), the second related to the Cretaceous transform-dominated opening event (in red on Figure 6). The first breakup unconformity, proposed to be of Sinemurian age in this interpretation, seals a complex of seaward-dipping reflectors that have been interpreted by Mercier (2016) and Reuber et al. (2016) as being related to major

volcanic activity during the Jurassic breakup and the formation of the Central North Atlantic Ocean. Reuber et al. (2016) relate this volcanic event to the Bahamas hotspot. Recent work by Basile et al. (in prep), based on the dating and geochemistry of dredged volcanic rocks and kinematic reconstructions, proposes that these volcanic rocks formed in relation with the Central Atlantic Magmatic Province (CAMP) hotspot. The oldest sediments recovered by drilling are Jurassic in age and correspond to post-rift carbonates (unit D in blue in Figure 6) then overlain by more detritic series (unit E in green Figure 6) (Gouyet, 1988; Mercier de Lépinay, 2016). These detritic series are extremely thick (up to 2s TWT) towards the Jurassic margin (Figure 6A and B) and are involved in compressional wrench tectonics approaching the Demerara Cretaceous transform border (Figure 6C). These deformed detritic sediments are eroded, producing the prominent upper Albian unconformity (in red on Figure 6) (Mercier de Lépinay, 2016). Once the Demerara marginal plateau passed into a passive stage (post-Albian sediments), sedimentation rapidly evolved towards shallow to deep marine conditions (Gouyet, 1988; Pattier et al., 2013). Sediment thickness increases towards the continent, i.e. along the continental shelf and upper marginal plateau (Figure 6) and decreases towards the ocean. The outer marginal plateau, at a distance of nearly 200 km from the shoreline, is considered to be nearly sediment starved at present (Mosher et al., 2007), except along a contourite drift that parallels the edges of the plateau (Tallobre et al., 2016; Fanget et al., submitted). The contourite drift is shown in Figure 6C (labelled CDS for Contourite Depositional System).

All in all, four successive basins record periods of subsidence in the history of Demerara: basin 1 in the SDR unit records the Jurassic volcanic breakup, basin 2 consists of post-breakup passive margin deposits, basin 3, near the Cretaceous transform boundary, comprises syn-cretaceous transform-dominated breakup deposits, and basin 4 contains post-breakup passive margin deposits (Figure 6 A, B, C).

The crust along the Demerara Plateau was first interpreted as a 15 to 30 km thick thinned continental crust based on a WAS line (Greenroyd et al., 2007, 2008). More recently, additional long offset MCS data show that the Demerara crust more probably corresponds to a highly volcanic intruded continental margin (Reuber et al., 2016; Klingelhofer et al., 2017) (Figure 5-G). Extremely thick SDR packages reaching 20 km in thickness rest on a thin 10 to 15 km thick lower crustal body (Reuber et al., 2016; Klingelhofer et al., 2017) (Figure 5-G). Klingelhofer et al., 2017 demonstrate, on the basis of a well-constrained wide-angle seismic dataset, the presence of a ~7 km thick high-velocity layer in or underneath the lower crust. ODP and industry drills have documented sedimentation down to the Jurassic, however neither the SDR packages nor the underlying crust have been reached. 173 Myr old magmatic rocks have been dredged along the northern transform border of the Demerara Plateau, interpreted as relating to the CAMP hotspot (Basile et al., in prep) (Table 1).

## Guinea

The Guinea TMP is bounded to the west by the Jurassic Central Atlantic oceanic domain and to the south by the Cretaceous Equatorial Atlantic. Its southern border

is a transform margin, the conjugate of the Demerara Plateau transform border (Figure 4-7, Table 1).

The Guinea Plateau has been less well investigated than its conjugate. It has been investigated by single-channel seismic acquisition (e.g. Benkhelil et al., 1995), analysis of industry data (drill-holes and seismic lines): Mercier de Lépinay, 2016; Ye et al., 2017; Olyphant et al., 2017) and dives of the manned submersible Nautilie (Masclé et al., 1992). Mercier de Lépinay (2016) shows that both the Demerara and Guinea plateaus share a common history until the opening of the Cretaceous Equatorial Atlantic: SDRs dipping towards the Jurassic Central Atlantic form the basement of both plateaus, which were first volcanic margins. Subsequently, a deep-sea fan and carbonate platform emplaced on the margin. The margin was then covered by Aptian detrital rocks (sandstones and mudstones). Aptian basalts have been drilled on the shelf (Olyphant et al., 2017). Sedimentary series then record the Cretaceous opening of the Equatorial Atlantic through complex vertical movements. A prominent upper Albian unconformity seals most deformations related to the transform-dominated opening of the equatorial Atlantic (Mercier de Lépinay, 2016). Albian syn-rift volcano-clastics have been drilled on the shelf (Olyphant et al., 2017).

Finally, post-rift volcanism affects the Guinea Plateau (Olyphant et al., 2017; Ye et al., 2017). This volcanism is characterized by a series of circular volcanic edifices probably related to the Sierra Leone hotspot. Volcanism particularly affects the southern border of the plateau, frequently masking its transform border on seismic data (Benkhelil et al., 1995; Mercier de Lépinay, 2016). The exact age of these volcanic rocks is debated: Olyphant et al. (2017) and Ye et al. (2017) propose a Cretaceous age based on industry drilling of volcanic rocks on the shelf, while Jones et al. (1991) date the nearby eastern Sierra Leone Rise at 53-54 Ma, suggesting that these volcanic edifices could be Cenozoic in age. In any case, magma feeding these volcanic edifices may have been guided to the surface by transform-related lithospheric discontinuities as suggested further west along the Sierra Leone Rise (Jones et al., 1991). The basement of this TMP has never been sampled by drills or dredges (Table 1).

## **Liberia**

This TMP is part of the Equatorial Atlantic domain. It is bounded to the south by a transform margin prolonged in the ocean by the St Paul Fault Zone (SPFZ on Figure 4-8). Behrend et al. (1974) underline that the location of this transform margin is probably strongly controlled by older Eburnean (~2,000 Ma) structures that acted as weakness zones during breakup. This plateau has been little investigated by academia since the seventies and the only available information about the structure and history originates from industry studies. Bennett and Rusk (2002) observe on industry seismic data that the basins developed in two phases, a syn-rift phase and a passive margin phase, both significantly overprinted by wrenching associated with Atlantic transform fault systems. No deep-penetrating seismic and wide-angle data exist along this margin and no drills or dredges are available along this TMP (Table 1).

## Côte d'Ivoire Ghana

The Côte d'Ivoire Ghana TMP also belongs to the Cretaceous Equatorial domain. It is bounded to the south by a transform margin extended by the Romanche Fault Zone (RFZ) (Figure 4-9). It is the conjugate of the Ceara transform margin off Brazil (Table 1).

The Côte d'Ivoire Ghana Plateau has been the locus of numerous scientific investigations, including MCS and WAS acquisitions, gravity, ODP drills and dives (Emery et al., 1975; Mascle, 1976, Mascle and Blarez, 1987, Basile et al., 1993; Sage et al., 2000; Mascle et al., 1998; Attoh et al., 2004, Antobreh et al., 2009). Its southern edge is bounded by a marginal ridge (MR on Figure 4-8 - bathymetric and structural high of debated origin - see the review by Basile, 2015) that is prolonged by the Romanche FZ that has been drilled (Mascle et al., 1998). It consists of exhumed syn-transform Cretaceous deformed sediments (Table 1). This plateau clearly consists of stretched continental crust formed at the intersection of a divergent and a transform margin. At the location of the marginal ridge, the continental crust is ~20 km thick (Sage et al., 2000) (Figure 5-H). The origin of the marginal ridge bounding the plateau seems to be flexural (Basile and Allemand, 2002). This ridge acts as a structural high trapping sediments along the plateau (Figure 4-9).

## Potiguar

This TMP, also called “Rio grande do Norte Plateau” emplaced between the Early Cretaceous South Atlantic domain and the younger Early Cretaceous Equatorial Atlantic domain (Figure 4-10). It is bounded to the north by a transform margin, prolonged by the Fernando de Noronha FZ. It is the conjugate of the Nigerian transform margin.

The Rio Grande do Norte Plateau is very poorly known. Fainstein and Miliman 1979 and Gouyet 1988 describe it as a piece of stretched continental crust on the basis of vintage seismic data. Dredges carried out on the eastern edge of the plateau recovered granites and amphibolites dated between 459 and 492 Ma (Guazelli et al., 1977). The nearby Potiguar rifted basin, which prolongs the transform margin onshore, has been studied in more detail (De Castro et al., 2012 and 2015). It developed along Precambrian shear zones reactivated during the Early Cretaceous. Two magmatic events are associated with the basin evolution: (1) the Ceara-Mirim dike swarm that was emplaced during the Late Jurassic south of the Potiguar basin (Araujo et al., 2001 and Sial et al., 1981) and which preceded the rift phase, and (2) the Macau formation that developed from 45 Ma to 7 Ma (Almeida et al., 1988; Araujo et al., 2001, Mizusaki et al., 1998).

### 2.3. South Atlantic Ocean

#### Sao Paulo

The Sao Paulo TMP (also called the “Sao Paulo Santos Plateau” in Moulin et al., 2013 and Evain et al., 2015) is located in the South Atlantic domain between the

“Central segment” to the north and the “Southern segment” to the south (Moulin et al., 2010; 2013) (Figure 3 and Figure 4-11). The Sao Paulo Plateau is bounded to the south by the Florianopolis FZ (FFZ, Figure 4-11). The Continent-Ocean Boundary of this domain is complex, partly hidden by a thick sedimentary cover and salt body masking deep structures (Meisling et al., 2001; Torsvik et al., 2009). Müller et al., 2008 first proposed a COB parallel to the coast. Later on, Torsvik et al., 2009 (followed by Müller et al., 2016), based on the observation of dextrally shifted SDRs and the lateral extension of Aptian salt proposed a COB parallel to the FFZ. This COB has only recently been constrained by WAS experiments (Evain et al., 2015). Those experiments suggest that most of the plateau is made of thinned continental crust, with a sharp transition to oceanic crust along the FFZ (Klingelhoefer et al., 2014; Evain et al., 2015) (Figure 5-1). The Sao Paulo Plateau is the “transform” conjugate of the Walvis TMP and the “divergent” conjugate of the Angolan margin.

The Sao Paulo Plateau separates two very different domains of the South Atlantic domain: south of the FFZ is the magma-rich southern segment of the South Atlantic that opened between 139.5 and 130 Ma, whereas north of the FFZ, is the Central segment that opened 18 Myrs later (ages from Rabinowitz and Labrecque, 1979; Austin and Uchupi, 1982; Curie, 1984; Torsvik et al., 2009 ; Moulin et al., 2010; Heine et al., 2013), at the end of evaporite deposition, in a less magmatic context. Therefore, the Sao Paulo Plateau has been interpreted as a kinematic “buffer” domain by Moulin et al., 2013.

The nature of the crust along the Sao Paulo Plateau has long been a matter of debate: Leyden et al., 1971 first concluded on the oceanic nature of this plateau, poorly surveyed until the early 2010s. Zalan et al., 2011, Moulin et al., 2013 and Evain et al., 2015 on the basis of recent WAS, gravimetry and magnetic data propose that the vast majority of the Sao Paulo Plateau is in fact continental in nature (Figure 5-1). Evain et al., 2015, based on a robust set of 5 WAS and MSC profiles, show that this plateau is mainly made of thinned continental crust and that normal oceanic crust is reached south of the FFZ. They underline that there is a certain heterogeneity in the Central part of the plateau with a 2-4 km thick upper crust layer characterised by velocities between 6.0 and 6.5 km/s overlying a high velocity layer (7 to 7.8 km/s) and the absence of Moho reflections at its base. This body could be interpreted as a portion of atypical oceanic crust (extinct spreading center?), as exhumed continental crust overlying altered mantle or as thinned and intruded continental crust. In any case, the limits of the crustal blocks defined by Evain et al., 2015 trend nearly parallel to the Florianopolis FZ, and the authors therefore associate the Sao Paulo Plateau emplacement with oblique rifting.

The Sao Paulo Plateau is located in close vicinity to the Parana Large Igneous Province. In the kinematic reconstructions of Torsvik et al., 2009, South America and Africa were first affected, at 130 Ma, by an important magmatic pulse that

gave rise to the Parana-Etendeka LIP. This LIP localized on a major lithospheric shear zone called the Parana Etendeka FZ (FPEZ). The FFZ would later prolong the PEFZ. At the same time, the southern part of the Southern domain of the South Atlantic begins to open. Breakup occurs at 126 Ma South of the FFZ. North of the FFZ, a proto Central segment appears between the early and late Aptian. Salt filled in this Central segment during the late Aptian at 112 Ma. The definitive opening of the Central segment north of the FFZ is proposed to be late Aptian (Moulin, et al., 2013). SDRs and volcanics are present in the Santos Basin, sealed by a sag basin, under the Albian salt deposits (Meisling et al., 2001; Rowan, 2014).

One DSDP drill sampled the Sao Paulo Plateau, only reaching lower Miocene sediments (Table 1). The existence of dredges of “mafic igneous rocks” is mentioned by Kumar (1977) (Table 1).

### Walvis

The Walvis TMP is located in the South Atlantic Cretaceous domain. The Walvis Plateau is a particular case: it corresponds to the northeastern end of the Walvis Ridge, which is bounded by a transform fault zone to the north: the Florianopolis Fracture Zone (Figure 4-12). This domain is interpreted as a Cretaceous volcanic margin associated with the Tristan Da Cunha hotspot trail (Elliott et al., 2009; Fromm et al., 2017; Planert et al., 2017). It is believed to be the conjugate of the Sao Paulo Platform, and was probably a volcanic segment of the African margin before transform motion occurred (Gladchenko et al., 1998; Fromm et al., 2017) (Table 1). The Walvis Plateau corresponds to the northern terminus of the SW African volcanic rifted margins. On this plateau, SDRs dip towards the north and are cut by the transform fault. Elliott et al. (2009) propose that this plateau formed in two stages: (1) first, it records the thermal uplift and melt associated with the Tristan Da Cunha Hotspot and the formation of an E-W oriented spreading center, (2) then it might have been reactivated and individualized in its present shape by a left-lateral transform movement followed by an eastward ridge jump. If this is true, this margin shows a very interesting case of transform initiation and localization within a weak lithosphere.

Moreover, Fromm et al., 2017 and Planert et al., 2017 conducted MCS and WAS experiments highlighting the crustal structure of the Walvis TMP. This plateau mainly consists of a 26 to 35 km thick intensively intruded transitional continental crust. This crust displays a more than 300 km long high-velocity lower crust (HVLC) with  $V_p$  seismic velocities comprised between 7.0 and 7.5 km/s (Figure 5-J). This HVLC also exists further south along the Namibia volcanic margin but has shorter lengths (150 to 200 km in length, Fromm et al., 2017).

The very western part of the Walvis TMP consists in a 26 km thick oceanic crust “of Icelandic type” (Fromm et al., 2017) (Figure 5-J). Crustal thickness changes abruptly north of the FFZ passing to a 12 km thick oceanic crust (Planert et al., 2017). Fromm et al., 2017 and Planert et al., 2017 conclude that the observed HVLC represent mafic intrusions formed by hotspot-related magmatism and the onset of rifting. Given that the observed HVLC is the largest beneath the Walvis Plateau, they propose that it was located above the hotspot during its formation but they do not favor the hypothesis of a large plume head - observations seem



more in accordance with the activity of a smaller pre-existing plume trail. Planert et al., 2017 also conclude that the influence of the hotspot on igneous continental growth is limited to the margin south of the FFZ, which probably either indicates a secondary lateral shift of formerly created volcanic margins or the fact that the FFZ acted as a thermal buffer during hotspot activity.

DSDP and ODP drills did not reach the basement of this TMP (Table 1) but basaltic rocks were dredged (Hekinian and Thompson, 1976 - Table 1).

### **Falklands-Malvinas**

The Falkland-Malvinas TMP is bounded to the north by the Early Cretaceous South Atlantic oceanic domain, to the east by the Late Jurassic Georgia Basin and to the south by the Oligocene Scotia Sea (Figure 4-13).

Three major tectonic episodes affected the plateau: (1) during the Middle Jurassic, a volcanic margin formed between Antarctica, South America and Africa. This domain comprises the current Argentine continental shelf to the west and the Maurice Ewing Bank (MEB on Figure 4-10) to the east; (2) during the Early Cretaceous, the final Gondwana breakup between Africa and South America created a 1200 km long E-W trending continental ocean transform margin to the north and a NE-SW trending passive rifted margin along the eastern side of the Maurice Ewing Bank; and, (3) during the Oligocene, the southern Falkland-Malvinas Plateau became an active transform plate boundary associated with the Sandwich subduction and the opening of the Scotia Sea (Figure 4-13).

The Falkland-Malvinas Plateau is a complex structure in most kinematic reconstructions, at the juncture of southern Africa, South America and eastern Antarctica before the Gondwana breakup (Table 1). It underwent an intricate breakup history and is now involved in oblique convergence with the Scotia Sea. It is the largest transform marginal plateau in the world, 1180 km in length and 510 km in width. Seismic refraction experiments (Ewing and Ewing 1959; Ewing et al. 1971; Barker, 1999; Schimschal and Jokat 2018), Deep Sea Drilling Project (DSDP) results (Barker et al. 1977; Ludwig 1983), reflection seismic analysis (Lorenzo and Mutter 1988; Platt and Philip, 1995; Barker, 1999), and gravity and magnetic analysis (Barker, 1999) reveal that the Falkland Plateau consists of a complex juxtaposition of crustal blocks of debated origin: (1) the Falkland/Malvinas islands (located on Figure 4-10) correspond to Devonian, Carboniferous and Permian clastic sediments intruded by Early Jurassic dolerite dykes. A small outcrop of Precambrian basement is also mentioned (Barker, 1924; Greenway, 1972; Musset and Taylor, 1994; Thistlewood et al., 1997), which consists of a 34 km thick continental crust (Schimschal and Jokat, 2018) (Figure 5-K). (2) The Maurice Ewing bank is interpreted as a continental basement of late Pre-cambrian metasediments overlain by Callovian-Oxfordian beach sands and younger pelagic and hemipelagic marine sequences (Barker, 1999, DSDP drill results and references in Table 1). (3) The basin province between them is designated as “a thinned continental section or a somewhat thicker than normal oceanic section” by Lorenzo and Mutter (1988). More recently, Schimschal and Jokat (2018), based on WAS data, propose that the Falklands Basin is floored by a thick oceanic crust (Moho depth between 19 and 29 km) (Figure 5-K).

Platt and Philip (1995) describe a series of Mesozoic and Tertiary basins with sedimentary sections ranging from 2 to 10 km in thickness west and east of the Falkland/Malvinas Islands. The first one is the “Malvinas Basin” southwest of the Falklands-Malvinas Islands, which develops above the Late Jurassic volcanics that cap the Paleozoic basement. Immediately south of the Falkland-Malvinas Islands, they observe the same structure with sedimentary series strongly dipping southward and involved in wrench tectonics. In the “Falkland Basin” province, east of the Falklands-Malvinas Islands, igneous intrusives may correspond to Jurassic volcanics or a younger rifting event. Barker (1999), based on magnetic, gravity, sonobuoy and MCS data, considers that the western Falkland Plateau is a Jurassic volcanic rifted margin related to the Karoo hotspot province.

### **Agulhas Bank**

The Agulhas Bank TMP directly extends the South African continental shelf. The plateau is bordered to the west by the South Atlantic Cretaceous oceanic domain and to the south by the southwestern Indian oceanic domain that formed between the Jurassic and Cretaceous. Its southern border corresponds to a transform margin extended by the Agulhas-Falkland Fracture Zone (AFFZ) (Figure 4-14). The conjugates of this plateau are the Falkland-Malvinas TMP and San Jorge Plateau off Argentina (Table 1). The Agulhas Bank, which is a TMP, should not be confused with the Agulhas Plateau emplaced on the Agulhas abyssal plain further south (location on Figure 4-14).

The Agulhas Bank results from two crustal stretching episodes: (1) the first, during Jurassic times (169-155 Ma) associated with the beginning of the breakup between Africa and Antarctica, is interpreted as having influenced the entire Outeniqua Basin (Figure 4-14); (2) the second episode, Cretaceous (136 Ma), is associated with a transtensional component of shear motion along the AFFZ and the opening of the South Atlantic and Indian oceans. The Outeniqua basin faults are orthogonal to oblique to the AFFZ. Before these crustal stretching episodes, South Africa hosted basaltic lava forming the Karoo Large Igneous Province at  $183 \pm 1$  Ma (Cole, 1992; Duncan et al., 1997). This volcanism has been attributed to the Bouvet plume (Hawkesworth et al., 1999) that accompanied the first movements between the eastern and western parts of Gondwana and initial stretching (Eagles and König, 2008).

Ben Avraham et al. (1997) and Parsiegli et al. (2009) propose that this plateau consists of thinned continental crust bounded by the Agulhas-Falkland Fracture Zone (AFFZ) (Figure 5-L). Crustal velocities directly beneath the Outeniqua Basin are consistent with the interpretation of the Paleozoic Cape Supergroup continental rocks underlying most parts of the basin (Parsiegli et al., 2009). No high-velocity body has been imaged using WAS data.

This TMP has never been drilled. 58 Myrs old alkaline rocks (tuffs, trachybasalts, trachytes) have been dredged along the Agulhas Bank upslope of the TMP. They are proposed to relate to the “Alphard Tertiary Igneous Province” that prolongs onshore (Dingle, 1973; Dingle and Gentle, 1972) (Table 1). These intrusions are

associated with small post-Cretaceous faults but the origin of this volcanic event does not seem to have been elucidated (Dingle, 1973; Dingle and Gentle, 1972).

## 2.4. Southern Ocean

### Gunnerus Ridge

The Gunnerus Ridge TMP forms a promontory that extends north from Antarctica (Figure 4-15). It is bounded on its eastern side by a transform fault that is in the continuity of the Davie Fracture Zone along Madagascar. The Gunnerus Ridge is believed to be of continental or volcanic origin and kinematic reconstructions by Reeves and DeWit (2000) place it between Mozambique, South Madagascar, India and Sri Lanka before the Gondwana breakup. It separated from Africa at ~140 Ma and from South Madagascar/Sri Lanka India between 112 and 99 Ma (Reeves and DeWit, 2000, Jokat et al., 2010, König and Jokat, 2010; Ali and Krause 2011). Hotspot volcanic products are reported immediately north of the Gunnerus Ridge on the CCGM geological world map (Bouysse, 2018). Leitchenkov et al., 2008, based on the analysis of MCS data and sonobuoy records, propose a continental origin for the Gunnerus Ridge. They propose a 28 km thick continental crust model that fits with the observed free air gravity anomaly signal. They underline that the seismic velocities of the lower crust are very poorly constrained. Inferred autochthonous gneissic and pegmatitic debris have been dredged on site (Saki et al., 1987) (Table 1).

## 2.5. Indian Ocean

### Morondova

This transform marginal plateau is located west of Madagascar between the Somalia and Mozambique oceanic basins which are Jurassic in age. It is bounded to the west by the Davie Ridge that follows the Davie Fracture Zone (DFZ-DR on Figure 4-16). The conjugate of this plateau is proposed to be, depending on the authors, the Kenya, Tanzania or Somalia margin.

This plateau probably experienced a first Jurassic continental magmatic phase related to the Bouvet plume forming the Karoo Large Igneous Province. The Somalia and Mozambique basins started to open later in the Jurassic. The Davie FZ became inactive in Aptian times (120 Ma) together with the cessation of the oceanic accretion of the Somalia Ridge, while the Mozambique Basin pursued its opening further South.

The Davie FZ was later intruded by Cretaceous volcanics that are contemporaneous with later onshore volcanic traps related to the Marion plume activity (Storey, 1995; Bardintzeff et al., 2001, 2010). The associated Cretaceous Morondava flood basalts are interpreted as resulting from the reactivation of a lithospheric scale shear zone that led to the Madagascar/Greater India continental breakup. Thus, the Morondova Plateau area first underwent a long period of crustal extension and Karoo volcanism, then Madagascar sheared away from Kenya/Somalia in the Middle

to Upper Jurassic by a dextral transcurrent movement along the Davie Ridge. Finally, Upper Cretaceous volcanism associated with the dislocation of Madagascar and Greater India intruded and capped the domain (Bardintzeff et al., 2010). Sparse Miocene volcanism also occurs along the Davie Ridge and Morondova Plateau (Cucciniello et al., 2017).

The crustal structure of this domain is poorly known, never having been investigated by MCS or WAS experiments. The nearby onshore Morondova domain consists of Precambrian terranes affected by Karoo volcanic units and Cretaceous volcanic units (Bardintzeff et al., 2010). These basement series, capped by Cretaceous volcanic formations, may form the submerged offshore plateau. The basement of this plateau has never been sampled by drills or dredges (Table 1).

### **Tasman**

The Tasman TMP is located south of Tasmania and is usually called “the South Tasman Rise” (STR). It is bounded to the west by the Early Cretaceous Indian oceanic domain, to the south by the Paleocene Indian oceanic domain and to the west by the Late Cretaceous Tasman Sea (Figure 4-17). The conjugates of this plateau are eastern Antarctica (Oates Land and George V Land) and Zealandia (Lord Howe Rise) (Table 1).

These structures derived from the non-magmatic East Gondwana breakup and the dispersal of Australia, Antarctica and Zealandia (Storey, 1995).

The crust of the South Tasman Rise is of thinned continental nature linked to Tasmania, which underwent Jurassic magmatism at 180 Ma (Ferrar and Tasman dolerites - Hergt et al., 1989; Storey, 1995). It was affected by northwest-southeast strike-slip motion in the Late Cretaceous and north-south extension in the Tertiary. Numerous basins of high petroleum interest developed in this domain (Hill et al., 2001; Exon et al., 2007). Hill et al. (2001) mention recent volcanics related to the Balleny plume activity. According to Royer and Rollet (1997), the South Tasman Rise is composed of two distinct terranes. The western domain rifted away from Antarctica in the late Paleocene/early Eocene and underwent severe wrench deformation. The eastern domain rifted from Tasmania and the East Tasman Plateau/Lord Howe Rise and underwent Cretaceous tectonics related to the separation from the Lord Howe Rise and later tectonic phases related to the separation from Antarctica. DSDP drills recovered quartz micashists along the southern part of the Tasman TMP at 160 mbsf (Table 1). Altered alkaline intraplate basalts were dredged in several locations within the STR (Crawford et al., 1997). Dolerites with strong affinities with the Jurassic dolerites of Tasmania were dredged in the central section of the STR (Crawford et al., 1997, Table 1).

### **Naturaliste**

The Naturaliste TMP is located in the southwest of Australia. It is bounded to the north by an Early Cretaceous oceanic domain (Perth abyssal domain part of the Indian Ocean, Figure 4-18) and to the south by the Late Cretaceous Indian oceanic domain. The plateau is bounded by a transform boundary to the north and by two

transforms to the south, respectively prolonged by the Naturaliste Fault Zone (NFZ, Figure 4-18) and Leeuwin Fault Zone (LFZ, Figure 4-18).

Breakup in the vicinity of the Naturaliste Plateau occurred in two stages: first, in a very oblique mode, between India and Australia-Antarctica, culminating in Early Cretaceous times (140-132 Ma; Direen et al., 2007); and second, between Australia and Antarctica, resulting in Late Cretaceous breakup (83-73 Ma; Sayers et al., 2001) (Halpin et al., 2008).

The Naturaliste Plateau has been alternatively interpreted as a continental fragment or as an oceanic plateau related to the Kerguelen mantle plume. This TMP is probably one of the best sampled by drills and dredges (see complete review in Table 1). Dredge hauls have yielded basaltic rocks, intrusive rocks that range from mafic to felsic compositions (Coleman et al., 1982; Direen et al., 2017) and high-grade felsic gneisses (Beslier et al., 2004). Some basaltic rocks are 120-130 Ma in age (Coleman et al., 1982; Storey et al., 1992), which makes them compatible with syn-breakup volcanism. Those ages are similar to the western Australia onshore Bunbury basalts and suggest a common origin (Bunbury plume, Olierook et al., 2016). For some authors, the Bunbury plume corresponds to a proto-Kerguelen plume (see review in Olierook et al., 2016). Direen et al., 2017, on the basis of dating and the geochemical analysis of dredged magmatic rocks, propose that an extensive igneous event related to the Kerguelen hotspot initiation occurred between 130 and 128 Myrs ago. They propose that the Naturaliste Plateau is part of a Large Igneous Province including the Naturaliste, Mentelle Basin, Wallaby, Enderby, Princess Elizabeth, Bunbury and Cornei-Cona areas before breakup. Storey et al., 1992 describe SDR packages between the Naturaliste Plateau and Bengale conjugate.

Beslier et al. (2004), based on dredges and the interpretation of geophysical data, suggest that much of the Naturaliste Plateau may be underlain by thinned (12.5-16 km) continental crust capped by Cretaceous volcanic rocks (Coleman et al., 1982; Direen et al., 2007). This assumption is confirmed by Olierook et al., 2016, who propose a present-day 20 to 25 km thick total crustal thickness based on gravity inversions. In parallel, IODP drills reached the top of those Cretaceous volcanic rocks. The oldest extrusive rocks reached by drilling correspond to a mix of subaerial and marine flows deposited close to sea level (Table 1).

### **Wallaby-Cuvier**

The Wallaby-Cuvier TMP is located between the Cuvier and Perth abyssal plains, all Cretaceous in age and part of the Indian Ocean (Figure 4-19). The plateau is bounded to the south by a transform margin prolonged by the Wallaby-Zenith Fracture Zone (WZFFZ, Figure 4-19) that formed between the shifted Cuvier and Perth abyssal plains.

Symonds et al. (1998) first described SDRs and patches of flood basalts along the Wallaby-Cuvier Plateau (Figure 5-M). However, until the early 2010s, available data were sparse and the nature of the plateau was debated. Alternative interpretations included continental core (Symonds and Cameron, 1977), volcanic plateau (Veervers and Cotterill, 1978), volcanic edifice derived from a mantle plume (Mihut

and Müller, 1998) or a mixed model of extended continental crust covered by syn-breakup extrusive flood basalts (Planke et al., 1996, Symonds et al., 1998).

Several recent geophysical data acquisitions and dredge analyses indicate with more confidence that this plateau corresponds to a continental block highly deformed by different extensional tectonic phases and affected by volcanism. SDRs and a high velocity lower continental crust underlie the Wallaby saddle (Colwell et al., 1994; Goncharov and Nelson, 2012). The timing between breakup and volcanism is more or less constrained depending on the plateau domains: (1) In the eastern part of the plateau, closest to Australia, SDRs are imaged on deep-penetrating seismic lines (Figure 5-M). They are interpreted by Symonds et al. (1998) as volcanic bodies associated with breakup, but they have not been drilled and directly dated; (2) further west, dredges were collected along the WZFZ and Sonne Seamount (see location of dredges, pink stars, on Figure 4-19-20). Volcanic rocks were recovered, dated and analyzed for their geochemistry. Olierook et al. (2015) show that these lavas show some affinity with continental flood basalts. These basalts formed at 124 Ma, i.e. 6 Myrs after oceanic spreading began in the adjacent abyssal plains. They are thus interpreted as resulting from a plume head that only formed after breakup and delivered lavas to the surface following the major Wallaby-Zenith lithospheric discontinuity. As mentioned before, for Direen et al., 2017, the Wallaby-Cuvier Plateau is part of a proto-Kerguelen Large Igneous Province including the Naturaliste TMP, Mentelle basin, Enderby, Princess Elizabeth, Bunbury and Cornei-Cona areas.

Incidentally, Stilwell et al. (2012) characterized a series of Berriasian to Barremian-Aptian sedimentary rocks along the Wallaby escarpment that bounds the plateau to the south. They associate these series with a possible pre-breakup sedimentary section. Later, Owens et al. (2018) show that, further north, this plateau hosts a thick (up to 19 km thick) depocenter of Permian to Cenozoic sediments. Two major unconformities record two major phases of rifting: the first, Permian, and the second, Early Jurassic ending with the transform-dominated Early Cretaceous breakup.

## Exmouth

The Exmouth TMP emplaced at the intersection of the Jurassic Argo abyssal plain and the Cretaceous Gascoyne abyssal plain (Figure 4-19-20). It corresponds along its southern edge to a transform margin boundary.

The Exmouth Plateau is an unusually broad region of continental crust, deformed during the first stages of rifting from the mid-Carboniferous to early Permian and Late Triassic (Frey et al., 1998). This deformation is followed by a Jurassic rifting and breakup (~161 Ma) period that formed the Argo abyssal plain and preceded Early Cretaceous (~130 Ma) oblique seafloor spreading in the adjacent Indian Ocean and more particularly the Gascoyne abyssal plain (Powell, 1976; Veevers and Cotterill, 1978, Larson et al., 1979; Lorenzo et al., 1991). Results from ODP drilling show that the northern Exmouth Plateau is partly composed of syn-rift, shallow-water, Tithonian to Valanginian terrigenous sequences resting unconformably on a Triassic prerift basement (Von Rad et al., 1992). Igneous intrusions and extrusions

are associated with this first Triassic rifting (Table 1). In the southern Exmouth Plateau, a Lower Cretaceous erosional hiatus spanning the Barremian-Hauterivian (~10 Ma) separates the sequence from about 1 km of postrift hemipelagic to eupelagic sediment deposits (Table 1). Lorenzo et al. (1991) imaged the Exmouth Plateau orthogonally to its transform border using WAS and MCS data and showed an unusual 10 km thick high-velocity layer (7.2/7.4 km/s) interpreted as magmatic underplating (Figure 5-N). They propose a model that implies a first phase of rifting and transform segmentation, and a second phase of drifting during which the accretionary ridge sweeps along the Exmouth transform boundary and causes local underplating. Symonds et al. (1998) describe SDRs and flood basalts on the western edge of the Exmouth Plateau (Figure 5-O). They relate them to voluminous transient breakup volcanism regionally expressed along the Western Australian margins. These SDRs and flood basalts have not been drilled but a basalt sill was reached along site 766 of ODP 123 in the southwestern corner of the Exmouth Plateau (Table 1).

### 3. Synthesis and discussion

#### *Nature of the crust of transform marginal plateaus (Table 1 and Figure 5)*

Mercier de Lépinay et al. (2016), based on the observation that most of the transform marginal plateaus share a multi-stage evolution and especially a first stretching and thinning phase before transform motion and separation, proposed that these domains correspond to thinned continental crust orthogonally reworked by transform motion. They suggested that the observed bathymetry of marginal plateaus derives from the isostatic compensation of this thinned continental crust.

Our inventory, updated and expanded from the compilation of Mercier de Lépinay, 2016 (Table 1), underlines that only two transform marginal plateaus are indisputably composed of non-volcanic thinned continental crust (Côte d'Ivoire Ghana and Agulhas). The Côte d'Ivoire Ghana Plateau has been explored by deep-penetrating multichannel reflection as well as refraction seismics which show that it is composed of a 20 to 25 km thick continental crust emplaced at the junction between divergent and transform margin segments (Sage et al., 2000, Figure 5-H). The thinning of this plateau is therefore attributed to syn-rift extensional tectonics. The Agulhas Plateau shares the same characteristics with significant extensional faulting and thinning along the Outenequia Basin emplaced at the divergent/transform junction of the margin (Parsiegli et al., 2007, Figure 5L).

For six TMPs, the lack of data does not allow us to state definitively about the nature of their crust: stretched continental crust, intruded or not, or thickened oceanic crust (Morris Jesup, Yermack, NE Greenland, Liberia, Potiguar, Gunnerus).

For eleven plateaus (Vøring, Faroe-Rockall, Demerara, Guinea, Sao-Paulo, Walvis, Morondova, Tasman, Naturaliste, Wallaby-Cuvier, Exmouth), the crust is believed to be continental but highly intruded and/or capped by volcanics. Seven of these plateaus have been investigated by modern deep-penetrating seismic and refraction methods (Table 1 and Figure 5, Vøring, Faroe-Rockall, Demerara, Sao-Paulo, Walvis, Wallaby, Exmouth). Their crustal thicknesses vary from 20 to 40 km. High-velocity lower crustal bodies ( $V_p > 7$  km/s) and/or underplating have been

suggested for all of them (Table 1) (Berndt et al., 2001; Klingelhoefer et al., 2005; White et al., 2008; Klingelhoefer et al., 2017; Evain et al., 2015; Fromm et al., 2017; Planert et al., 2017; Lorenzo et al., 1991). Interestingly, Bauer et al. (2000) define immediately south of the Walvis Plateau a transitional domain between the normal oceanic crust and the extended intruded continental crust that displays high P-wave velocities  $>7$  km/s on the lower half of the crust and SDRs in the upper crust. They interpret this transitional domain as a margin parallel magmatic belt consisting exclusively of igneous material. They remind us of the results of Sobolev et al. (1998) that correlate such P-wave velocities with olivine-rich high Mg magmas typically associated with a 20 to 25% partial melting of peridotites - which is made possible in plume influenced domains.

For the Falkland Plateau, recent wide-angle experiments image an 8 km thick basin overlying a 20 km thick oceanic crust east of the Falkland-Malvinas Islands. This exceptionally thick igneous crust with high lower crustal velocities (up to 7.4 km/s) is attributed to a regional mantle plume active during its formation (Schimschal and Jokat, 2018). With the same data set, authors interpret the eastern Falkland-Malvinas domain as a 34 km thick continental crust structured as a volcanic margin. A 26 km thick “Icelandic type” thickened oceanic crust is mentioned for the outer Walvis TMP (Fromm et al., 2017). This overthickened oceanic crust is interpreted as resulting from the Tristan Da Cunha hotspot trail. Fromm et al., 2017 also describe an overthickened “Icelandic-type” oceanic crust in the western Walvis marginal plateau.

For the Faroe-Rockall, Sao-Paulo, Walvis Falkland-Malvinas, Tasman, Wallaby and Exmouth plateaus, important lateral variability in the nature and structure of the crust is underlined, the plateaus being formed of several crustal slivers of contrasted nature and structure (White et al., 2008; Klingelhoefer et al., 2005; Evain et al., 2015; Fromm et al., 2017; Ewing et al., 1971; Exona et al., 1997; Rollet et al., 1996; Olierook et al., 2015; Schimschal and Jokat, 2018).

### ***Depth and shape of marginal plateaus***

Marginal plateaus are found at depths ranging from several hundreds of meters to 4000 m water depth, with an average depth of 2025 m (table 1). They comprise wide planar domains whose mean slope value is generally less than  $0.5^\circ$ . The size of transform marginal plateaus is highly variable, but they represent a significant surface area, of around 2 700 000 km<sup>2</sup> in total (Table 1). Despite their different origin and evolution, their average depth seems characteristic, systematically shallower than the nearby oceanic domains (Table 1 and Supplementary data 3) and may originate from and express a particular crustal type. Understanding the particular depth of these marginal plateaus will need more combined WAS and MCS data acquisitions, drills and isostatic models, but three main circumstances may explain the depth of marginal plateaus: (1) the presence and isolation of thinned continental crust slivers as proposed by Mercier de Lépinay et al. (2016) (Côte d’Ivoire, Agulhas TMPs - 20 to 30 km thick, no underplating or volcanics); (2) underplating under volcanic margins. As recalled by Lorenzo et al. (1991), “a consequence of underplating is a permanent crustal isostatic rise, as the mantle is



replaced by less dense igneous rocks". Many TMPs are in fact volcanic margins and therefore host underplated material, erupted flood basalts or both. Underplated material typically reaches P-wave velocities of  $> 7$  km/s and important densities (3 to  $3.3 \text{ gr/cm}^3$ ), volcanoclastics, basalt flows and hyaloclastites typically forming SDRs are usually characterized by P-wave velocities ranging from 3 to 5.5 km/s, densities from 2 to 2.8 (Eldholm et al., 1995; Planke et al., 2000; Bauer et al., 2000). The respective thicknesses of these two extrusive and intrusive end-members probably define the depth of the considered marginal plateaus. (3) The thickening of oceanic crust. This typically occurs along oceanic plateaus related to hotspot activity. This is the interpretation that Schimschal and Jokat (2018) propose for a large part of the Falklands Plateau (between the Falkland-Malvinas Islands and the Maurice Ewing Bank, Figure 5I). Fromm et al. (2017) propose the same interpretation for the outer Walvis Plateau. In this respect, important partial melting of peridotites in plume environments may be responsible for transitional dense and purely igneous crustal types (Bauer et al., 2000) that resemble oceanic crusts with overthickened layer 3 lower crust domains.

In any case, the intermediate depth of transform marginal plateaus (between the shelf and the abyssal plain, Table 1 and Supplementary file 3) can be explained by intermediate crustal thickness between continents and oceanic crust thicknesses. However, transform marginal plateaus share a striking feature as they all represent flat but deep surfaces. We can understand this flatness by two successive processes:

- A regional flat erosional surface is a common feature in the sedimentary record of transform marginal plateaus. As an example, an upper Albian unconformity erodes the older sediments below the whole Demerara Plateau (Figure 6). This unconformity has been related to sub-aerial erosion (Basile et al., 2013; Mercier de Lépinay et al., 2016), indicating a regional uplift of the plateau. One possible mechanism may be a thermal uplift above a mantle plume, which can be postulated in many cases from the occurrence of magmatism. Thermal uplift and associated sub-aerial erosion could have initiated the flat surface.

- This flat surface is subsequently expected to thermally subside deeper and be overlain by prograding sediment wedges. In divergent continental margins, shelf progradation commonly occurs and should result in building sedimentary prisms or continental slope over the older structures. Because marginal plateaus represent spurs out of divergent margins, their outer parts are relatively starved of detrital input from adjacent continents. In the example of Demerara, the erosion of South America (and especially of the northern Andes) feeds two very voluminous detrital sedimentary systems, the Orinoco and the Amazonia deep sea fan, on both sides of the Demerara Plateau (Fig. 4.5). On the plateau itself, the shelf progradation is fed only by the littoral drift of fines, and only covers the southern half of the flat erosional surface (Fig. 6C), maintaining the flat shape on its northern edge. Ancient flat surfaces can be preserved as marginal plateaus because they were relatively isolated from detrital input from the neighboring continents.

### ***Compared tectono-magmatic evolution of TMPs***

An important and new result of the description and comparison of these TMPs is that at least 65% (13 out of 20) of transform marginal plateaus have experienced

volcanism during at least one of their stages of evolution (Table 1, figure 7). Six plateaus have been insufficiently investigated to decide conclusively about their nature (NE Greenland, Liberia, Potiguar, Gunnerus, Morris Jesup Rise and Yermack). However, Morris Jesup Rise, Yermack, NE Greenland, Potiguar and Gunnerus are strongly believed to be affected by volcanics (Engen et al., 2008, Bouysse, 2018). If so, the percentage of TMPs that experienced volcanism is even larger, reaching 90%. Volcanic activity is presently unknown only for two plateaus out of twenty (10%): the Côte d'Ivoire Ghana and the Agulhas transform marginal plateaus (Table 1).

Concerning the TMPs that host volcanics, different cases exist (Table 1, Figure 7) depending on the timing of volcanism in reference to the two stages of rifting experienced by each TMP:

- (1) Most plateaus underwent volcanism during their first stretching and thinning phase before breakup. It is the case of the Sao-Paulo, Walvis, Falklands, Morondava, Tasman and possibly the Morris-Jesup, Yermack, NE Greenland, Demerara, Guinea, Gunnerus plateaus, all emplaced in the vicinity of the Parana-Etendeka, Central Atlantic Magmatic Province, Jurassic Karoo or Ferrar-Tasman continental Large Igneous Province respectively (Storey, 1995).
- (2) Other plateaus first formed as volcanic margins before a second phase of transform-dominated opening in a propagating ocean context. It is the case of the Demerara and Guinea plateaus that underwent strong Jurassic volcanism associated with the Central Atlantic Magmatic Province (CAMP) or the Bahamas hotspot and developed as volcanic margins (Mercier de Lépinay, 2016; Reuber et al., 2016; Basile et al., submitted). These plateaus were definitively separated at Cretaceous times during the equatorial Atlantic opening that connected the South and Central Atlantic oceans. Limited volcanism is mentioned during this second phase (Gouyet, 1988; Olyphant et al., 2017). The western Falklands Plateau equally falls into this category as Barker (1999) and Schimschal and Jokat (2018) defines it as a volcanic margin structured in relation with the Karoo hotspot and later dextrally shifted from Africa during the final Gondwanan Early Cretaceous breakup. In those three cases, transform faults allowed the connection of the shifted Jurassic oceanic domains (Central Atlantic to the north, Africa-South America oceanic domains to the south - Figure 3) during later Cretaceous opening. The Walvis and Sao-Paulo Plateaus also probably formed as Neocomian volcanic margins before the opening of the Central segment of the South Atlantic. The time duration between the initial volcanic breakup phase and later transform-dominated opening is about 65 Myrs for Demerara-Guinea, 50 Myrs for the western Falklands domain and 18 Myrs for the Sao-Paulo and Walvis plateaus (Table 1). The Naturaliste Plateau also formed in two successive phases, the first one clearly associated with volcanism (Bunbury plume, probably related to the Kerguelen proto LIP - Dieren et al., 2017). The second breakup phase, in that case, was oblique and may have produced additional volcanics associated with the Kerguelen plume.
- (3) Few plateaus experienced volcanism and transform segmentation during the same tectono-magmatic phase. This is the case of the Vøring Plateau belonging to the North Atlantic Igneous Province associated with the Iceland

hotspot onset (62 Ma) and possibly also of the Wallaby-Cuvier Plateau (Symonds et al., 1998). The Vøring Plateau is the best investigated of those plateaus. Berndt et al. (2001) propose that the transform fault that bounds the southern Vøring Plateau developed during breakup in a cold lithosphere setting between two thermal upwelling zones that were offset from one another. The Walvis Plateau can, at first glance, be included in the same category. It corresponds to a volcanic margin that formed in an oceanic domain of apparent homogeneous age (Figure 3). Elliott et al., 2009 show, however, in a more focused study some diachronism between volcanism and transform motion: the Walvis Plateau corresponds to a volcanic margin developed in relation with the Tristan Da Cunha hotspot, an event closely followed by a westward ridge jump and associated transform emplacement. This transform fault seems to be emplaced along an abandoned spreading center and sharply delimits the volcanic Walvis Plateau from the northern non-volcanic African margin.

- (4) Other plateaus underwent volcanism after a first breakup and seafloor spreading phase. This is the case of the Faroe-Rockall Plateau, and possibly the Gunnerus Plateau. For Faroe-Rockall, the situation is relatively simple: the transform that bounds the plateau to the south corresponds to the northern terminus of the Early Cretaceous Southern North Atlantic (Figure 4). The plateau itself formed in a second phase associated with the Paleocene highly magmatic opening of the northern North Atlantic (North Atlantic Igneous Province). The transform that bounds the southern plateau thus sharply separates a volcanic margin segment to the south from a non-volcanic margin segment further north. For the Exmouth Plateau, the situation is less clear. Lorenzo et al. (1991) described a 10 km thick high-velocity body interpreted as underplating under the transform border of the plateau. They relate this underplating as resulting from the passage of the accretionary ridge along the transform margin during early seafloor spreading. The age of the emplacement of this magmatic underplated body is, however, poorly constrained and Symonds et al. (1998) relate the onset of volcanism to the Cretaceous breakup with SDRs dipping from the Exmouth Plateau toward the Southern Cuvier abyssal plain.
- (5) Finally, some plateaus clearly experienced volcanism after the final stage of breakup and the onset of seafloor spreading. It is the case of the Morondava, Tasman, Wallaby-Cuvier, Guinea, and possibly Potiguar and Naturaliste, plateaus. The Morondava Plateau and nearby Davie Ridge experienced volcanism in the Late Cretaceous (Cenomanian-Turonian, 99-88 Ma) a long time after the Somalia Basin began opening (Late Jurassic ~160 Ma) and the Davie FZ activity ceased (120 Ma). Bardintzeff et al. (2010) attribute this late volcanic activity to the final dislocation of Madagascar and Greater India accompanied by intense volcanic activity all over the island of Madagascar. Morondava basin volcanism would result in this context from the reactivation of a lithospheric scale shear zone that guides volcanics to the surface (Bardintzeff et al., 2009). In the case of the Wallaby-Cuvier Plateau, a small mantle plume seems to be responsible for volcanism after breakup (Olierook et al., 2015). Lavas preferentially followed the Wallaby-Zenith major lithospheric discontinuity. In the case of the Guinea Plateau, Cretaceous to late Cenozoic volcanism has been recorded after breakup and seafloor

spreading following transform fracture zones (Benkhelil et al., 1995; Jones et al., 1991; Olyphant et al., 2017).

To summarize, this study reveals a large variety of contexts that link hotspot volcanism and transform marginal plateaus.

A first conclusion is that most TMPs, forming protruding marine elevations extending the continental shelves, are partly volcanic and derived from hotspot activity before, during or after breakup.

A second conclusion is that a close and hitherto unrecognized link exists between hotspots, volcanic margins and transform fault localization. However, we cannot propose any simple or common links between all TMPs. Different hypotheses and questions arise from our review:

- (1) First, hotspots are known to localize breakup (Storey, 1995) and form volcanic margins (Geoffroy et al., 2015). The Gondwana breakup typically initiated with the onset of the Bouvet plume that was at the origin of the Karoo LIP. From numerical models of transform fault imitation, Basile and Braun (2015) suggested that in the case of oblique rifting (i.e. when the plate boundary is oblique relative to plate displacement), extensional boundaries (i.e. rifts) propagate from lithospheric weak zones (as hotspots can be) before being connected from one to the other by perpendicular transform faults. This hypothesis has been suggested for the formation of the Vøring TMP where volcanic series drastically thin toward the Jan Mayen transform FZ (Berndt et al., 2001).
- (2) However, in some cases transform motion largely postdates volcanic margin formation (Demerara-Guinea, Falklands) and the thermal state of the lithosphere may not allow us to consider these margins as weak domains. When transform faulting occurs during a second rift phase, it connects a laterally shifted propagating rift with an older oceanic basin, where an oceanic accretion axis is already active. As recalled by Basile (2015), continental transform plate boundaries seem to occur for obliquities ranging from 60 to 90° with the plate displacement direction. They probably preferentially locate along pre-existing inherited lithospheric heterogeneities (case of the Charlie Gibbs FZ off Hatton-Rockall, Welford et al., 2012). A cross section of the Cretaceous transform edge of the Demerara Plateau shows that the transform is exactly located where the Jurassic SDRs laterally vanish from more than 20 km to less than 1 km (Mercier de Lépinay, 2016). This suggests that while magmatic centers may be weak points during the formation of magmatic margins, they may represent strong points tens of million years later, and the relative softer parts of the lithosphere where transform may localize should be on the side of the magmatic centers. In this case the localization of transform may be inherited, not from the old continental structures, but from the magmatic segmentation of the previous magmatic margin.
- (3) Another interesting case is that of the Walvis and Sao Paulo conjugate TMPs. There, transform motion probably only slightly postdates volcanic margin formation (18 Myr time lapse between the magma-rich Southern segment and the Central segment opening). Planert et al., 2017 emphasized the sharpness of the volcanic crust transition along the Walvis TMP toward the

Florianopolis transform. Elliott et al., 2009 also show that SDRs dip toward the Florianopolis FZ. They therefore propose that the Florianopolis transform was first the location of an E-W spreading center and volcanic margin, later abandoned and reactivated as a transform plate boundary. In their hypothesis, transform faulting would have initiated in a highly thinned and probably hot crust. This could be an easy way to produce transforms. As demonstrated by Gerya, 2010, Liao and Gerya 2015, and Ammann et al., 2018 on the basis of 3D thermo-mechanical modelling, transform faults more easily form in thinned, detached continental or oceanic crust domains or in pre-existing weakness zones parallel to extension.

- (4) In some rare cases, volcanism clearly occurred during the second transform-dominated breakup. It is the case of the Faroe-Rockall TMP associated with the onset of the North Atlantic Large Igneous Province and bounded to the South by the Charlie Gibbs Fault zone (CGFZ). Welford et al. (2012) show, on the basis of gravity inversions and kinematic reconstructions at the initiation of Cretaceous seafloor spreading, that the CGFZ already separated a hyper-extended, stretched and poorly magmatic continental crust to the south from a less-stretched continental crust to the north. Later, the magmatic centers that shaped the Faroe-Rockall TMP seem to have been limited and bounded to the south by the CGFZ. Preexisting transform faults juxtaposing crustal slivers of different thickness may impact mantle melting amounts above an ascending plume. Phipps Morgan et al. (1995) proposed, for example, that the amount of mantle melting above the Hawaiian hotspot was associated with discrete changes in the age of the overriding lithosphere along transform faults. Where the lithosphere is the shallowest, volcanism seems enhanced.
- (5) Finally, after breakup, when passing on plume pathways, transform faults and associated fracture zones seem to be preferentially reactivated and guide volcanic lavas to the surface (examples of Guinea, Wallaby, Morondava, Potiguar lava flows). Transform lithospheric faults may be considered as efficient structures for guiding lava to the surface.

These observations and questions underline how poor our understanding is of the thermo-mechanical conditions that are necessary for transform initiation and reactivation. As mentioned by Gerya (2012) “both numerical and analog models of transform faults are still relatively scarce and a number of first order questions concerning the origin and evolution of these faults remain unanswered”.

### ***Conjugate TMPs***

Many TMPs (at least 8 out of 20, Table 1) can be paired as conjugate margins, i.e. adjacent blocks before rifting (e.g., Morris-Jesup/Yermack, Demerara/Guinea, Sao Paulo/Walvis, Falklands-Malvinas/Aghulas TMPs). Other TMPs are clearly devoid of a conjugate TMP (at least 7 out of 20, Table 1). We cannot state definitively for 5 TMPs. For example, the Western Australia TMP conjugates belong to the collided Greater India or even subducted. We also suspect that the Jan Mayen Ridge could be considered as an isolated re-rifted conjugate TMP of the Voring TMP.

It is difficult to strictly compare these conjugates as the available crustal information is always partial (Figure 5). However, some pairs of TMPs are clearly

asymmetric. For example, the Falkand Plateau basin has been interpreted as being built on a thick (20 km) oceanic crust (Schimschal and Jokat, 2018), while the conjugated Agulhas Plateau appears to be devoid of magmatism. The fact that transform margins may form preferentially on the edge of older magmatic centers could explain this asymmetry.

### ***TMP sedimentary archives***

In many cases, TMPs have been largely overlooked from a sedimentary point of view, compared with other parts of continental margins, either because they were believed to be oceanic plateaus or because they were not defined and understood properly in all their complexity. In some cases, they constituted such vast domains that their comprehensive and continuous exploration remains difficult.

More than half of the defined TMPs developed at the junction of oceans of contrasted ages (Morris Jesup, Yermack, NE Greenland, Faroe-Rockall, Demerara, Guinea, Potiguar, Sao-Paulo, Walvis, Falklands-Malvinas, Gunnerus, Tasman, Naturaliste, Exmouth, review in Figure 7).

Therefore, TMPs record a polyphased history, and may host long-lived sedimentary archives that are important pieces in the puzzle of understanding breakup processes and their timing, ocean circulation onset, and the biological connectivity between continents.

The Demerara Plateau, situated at the junction of the Jurassic Central Atlantic and the Cretaceous Equatorial Atlantic illustrates well the diversity and polyphased character of basins that develop on TMPs. The internal structure of the Demerara Plateau displays two major breakup unconformities individualizing four successive basins of contrasted extents, geometries and infills (Figure 6). A first basin corresponds to the syn-rift bodies associated with extensive volcanics. A second basin develops after this first breakup stage as a Jurassic passive margin. A part of this margin is then reactivated orthogonally by a transform-dominated breakup responsible for complex vertical movement and the development of rapidly subsiding wrench basins (Mercier de Lépinay, 2016). Finally, the plateau once again becomes a passive margin. During this last stage, large prograding clinoforms develop between the shelf break and the upper marginal plateau (Fig 6-C, basin 4). More distally, the plateau becomes nearly sediment starved due to its distance to the continent except along its very outer edge where contourites (CDS in fig 6-C) deposit in relation with the North Atlantic Deep Water mass. As recalled by Rebesco et al., 2014, “currents tend to flow parallel to large-scale bathymetry - i.e. along the continental margins”. This is especially the case along outstanding TMPs where they are accelerated and build important contourite depositional systems (Tallobre et al., 2016).

### **General Implications: role of TMPs in oceanography and biology**

#### ***TMPs as “land bridges”***

The understanding of the formation and timing of land bridges during continental breakup has many implications in terms of past biodiversity growth, bio-connectivity and lineage evolution, in particular for terrestrial vertebrates and

non-aquatic mammals (Darwin, 1859; Ali, 2018). Most transform marginal plateaus form prominent and outstanding elevations that were the last-contact points between emerged areas during final continental breakup and may be considered as land bridges for terrestrial vertebrates and mammals. Also, once continents definitively separate, these domains become narrow gateways between initially disconnected oceanic domains, therefore playing key roles in oceanic connection, ventilation and circulation (Figure 8 - example of the Equatorial Atlantic gateway, Walvis/Sao-Paulo margins and Falklands). In this section, we propose to review the conditions under which transform marginal plateaus may have played the role of land bridges, and their role in oceanic oxygenation and circulation.

Many transform marginal plateaus were the last, or close to last, contacts during final continental breakup, some of them probably emerged, especially those that were volcanic margins that typically evolve in their first stages as aerial to very shallow basins (Planke et al., 2000; Geoffroy et al., 2015). It is probably the case of the Vøring Plateau lately connected to the Jan Mayen microcontinent/Greenland margin and the Faroe-Rockall/Newfoundland plateaus also connected until the late opening of the North Atlantic domain. The Walvis Plateau, Walvis Ridge and Sao-Paulo Plateau may also be good candidates for having formed another land bridge during the South Atlantic Ocean opening (Figure 8). The Wallaby Plateau could also have been a land bridge between Western Australia and Greater India in the Early Cretaceous (see reconstructions by Olierook et al., 2016). The Tasman Plateau is also described as a shallow land bridge between Australia and Antarctica by Exon et al., 2002.

In this context, it is important to remember why transform margins form last contact points between continents: at the time when oceanic accretion starts in divergent basins, the two continents are still in contact at a lithospheric scale on each side of the transform fault. This contact is lost at the end of the intra-continental transform stage, for a time equal to the length of the transform divided by the spreading rate (Basile, 2015). For a spreading rate of 2 cm/year, the last continental contact is lost 5 Myrs after oceanic spreading started for a 100 km long transform, and 25 Myrs for a 500 km long transform, which represents a significant time lapse for bio-connectivity. In contrast, without obliquity (rift direction perpendicular to plate displacement) no transform fault will develop, and the last contact between continents should be simultaneous at the geological scale all along the rift, whatever the segmentation of the rift may be. Consequently, non-transform marginal plateaus, if they exist, are not expected to have any influence on this topic of continental connectivity.

However, lithospheric connectivity should be distinguished from land connectivity. Being on continental lithosphere does not necessarily imply positive elevation, especially in rift settings. As shown by the Dead Sea basin, being below sea level does not exclude land continuity, which also depends on the elevation of the connection with the open sea. The relief of transform faults is characterized by along-strike depressions (transform valley), ten to a few tens of kilometers wide, bounded on one or two sides by uplifted shoulders. For continental transform faults (intra-continental transform stage), the valley is expected to be filled by detrital sediments coming from the shoulder and transported along-strike. Because of this detrital input, the transform valley is not expected to be deep, even when located

at sea level (e.g. Côte d'Ivoire-Ghana transform margin, Basile et al., 1998). Intra-continental transform valleys should not be a barrier to the circulation of plants and wildlife. Along transform margins (i.e. when the transform fault is between continental and oceanic lithospheres), the active transform fault is expected to be in a transform valley on the edge of the oceanic crust, at the base of the continental slope. The edge of the margin is uplifted to form a marginal ridge which is generally eroded, but which can stand above sea level for some million years after the end of the intracontinental stage (e.g. circa 10 Myr for the Côte d'Ivoire-Ghana transform margin, from the late Albian to Turonian: Basile et al., 1998).

Interestingly, once the transform margin becomes passive, the oceanic transform valley still can emerge, for example in response to local and far-field stress changes (e.g. St Paul and St Peter islets along the St Paul transform, Maia et al., 2016 or eroded ridges along the Romanche FZ, Honnorez et al., 1991). However, these emerged lands are not expected to be continuous from one continent to the other, but likely represent a series of aligned islands where the distance between two emerged areas is reduced. This may facilitate some migration of wildlife and plants across straights after the continents pulled apart, but may not represent true land bridges.

Finally, because most TMPs seem to form in relation to mantle plumes, they may present additional topographic elevations associated with extrusive volcanic cones and flood basalts and a topographic swell produced by the rising of hot mantle material (initial thermal uplift) and underplating (more permanent uplift due to compositional buoyancy) (Phipps Morgan et al., 1995). Inner SDRs and outer high extrusives frequently observed along TMPs are proven to form in subaerial conditions (Planke, 2000).

As modelled by Phipps Morgan et al. (1995) the thermal heating associated with shallow crustal magma emplacement probably decays rapidly (1 to 4 Myrs for the Hawaiian hotspot) provoking a rapid first stage of subsidence of uplifted domains. A second and slower stage of subsidence then occurs as deeper mantle swell root heat perturbations decay (in the order of 60 to 80 Myrs for the Hawaiian swell). Those time values were calculated for an intraplate oceanic hotspot and depend on the thickness of the lithosphere overriding the hotspot plume and also on the moving rates of this overriding plate (Phipps Morgan et al. 1995). They probably need to be specifically evaluated for each hotspot-related TMP.

Hotspots may affect TMPs in a different way, depending on the relative timing of magmatism and transform faulting:

- (1) When transform faulting occurs a long time after the onset of a large igneous province, the thermal effect of the mantle plume is expected to have vanished, and transform faulting and underplating should be the main process that controlled the elevation.

The Demerara and Guinea conjugate plateaus slightly subsided after Jurassic magmatic breakup, which is attested by the deposition of platform carbonates and later detritic sediments on most of the surface of both plateaus (Mercier de Lépinay, 2016). All post-Jurassic series are, however, affected by a regional upper Albian unconformity that seals most syn-transform deformations and attest to subaerial erosion of the plateaus (see seismic interpretations and drill results in



Gouyet, 1988; Basile et al., 2013; Mercier de Lépinay, 2016). Those plateaus thus probably formed an emerged domain between Africa and America until the Sinemurian and re-emerged later on, during the upper Albian. Considering the case of the Falklands-Malvinas Plateau (Figure 8), its northern border is bounded by a transform whose duration of activity is ~24 Ma (Mercier de Lépinay et al., 2016) and which was responsible for the formation of an important marginal ridge called the “Falklands-Malvinas Escarpment” (Lorenzo and Mutter, 1988). As along the Demerara and Guinea, a regional post deformation unconformity has been described along this marginal ridge (Lorenzo and Mutter, 1988; Lorenzo and Wessel, 1997). This unconformity may record aerial erosion and the past existence of dry land on the plateau - however, this ridge has not been drilled. South of this ridge, marine sediments have infilled the Falkland Basin since Jurassic times. Most paleogeographic reconstructions consider that the Falkland-Malvinas Plateau corresponds to a shallow marine environment in the Early Cretaceous (Cao et al., 2017). However, depending on the mechanisms that formed the Falklands-Malvinas marginal ridge (thermal bulge and underplating associated with the lateral passage of the Atlantic ridge as proposed by Lorenzo and Mutter, 1988? Flexural mechanism associated with the margin erosion as suggested by Basile and Allemand 2005?) a land bridge or at least a series of successive islands might have existed between southern Africa and South America along the northern Falkland-Malvinas Plateau domain during the Early Cretaceous. At least, as mentioned by several authors, the Falkland-Malvinas TMP formed a bathymetric high influencing the onset of oceanographic circulation.

- (2) When magmatism is coeval with transform fault formation or reactivation (e.g. the Vøring Plateau, Hatton-Rockall), large parts of the plateaus are expected to be emerged and rise above adjacent areas (Vøring: Abdelmalak et al., 2016; Planke et al., 2017).
- (3) When magmatism occurs after transform faulting (likely along the transform margin or associated fracture zone), it is expected to form individual volcanoes, eventually lined up as along the edge of the plateau, but not true land bridges. It was probably also the case for Morondova, Guinea, Wallaby-Cuvier or Tasman, where the most probable hypothesis is that those volcanic edifices only formed marginal isolated islands in open oceanic domains. The Morondava Plateau and the nearby Davie Ridge may have formed a land bridge during Cretaceous and later Miocene volcanic pulses due to the extreme proximity of Africa and Madagascar. McCall (1997) underlines the paradoxical presence of several mammalian groups on Madagascar, groups that postdate the Cretaceous breakup of Gondwana. Based on geodynamic and biological arguments, he proposes that the Davie Fracture Zone was a land bridge from the mid-Eocene to early Miocene.

To conclude, transform marginal plateaus may have been land bridges allowing late biological connectivity between continents after Gondwana or later breakup, on the condition that they were either totally or partially emerged. This condition was probably fulfilled when those plateaus were volcanic in nature, and when

magmatism was coeval with transform faulting (Figure 7 for review). Inner flows, inner SDRs, outer highs and the regional erosional surfaces commonly observed along those plateaus may be interpreted as past emersion witnesses and their mapping and dating may give interesting clues for past biological dispersal and isolation.

### ***TMPs: their role on oceanic circulation***

The notion of land bridges associated with transform-dominated breakup also implies that narrow straits existed between those marginal plateaus just after separation - the best actual examples being the Fram Strait between the Yermack/Morris Jesup and NE Greenland/Vøring plateaus and the Davie Strait between Somalia and Morondava. Jakobsson et al. (2007) emphasized the early evolution of the Fram Strait and the late connection between the Arctic and North Atlantic oceans (Figure 9). They evidenced the fact that the Arctic Ocean underwent (1) an oxygen-poor “lake stage” that they compared to the Black Sea environment during the last glacial lowstands, (2) a transitional estuarine sea stage during the first phases of opening of the Fram Strait, and (3) a fully ventilated phase. These different stages were completed within only 0.7 Myrs. Engen et al. (2008) underline the fact that the onset of this last phase is a major paleoceanographic event since it indicates the onset of deep-water exchange due to the opening and deepening of the Fram Strait, which is the only gateway for deep waters between the Arctic and the world oceans.

They also present early contemporaneous contourite buildups immediately south of the strait, indicating the persistent action of bottom currents that are driven by thermohaline circulation. Along the strait itself, acting as a physiographic bottleneck for currents, strong erosion is believed to have occurred and removed all terrigenous sediments that were possibly trapped in the immature rifted continental crust initially separating the two plateaus. Eiken and Hinz (1993) show that contourites later deposited in the Fram Strait itself. The age of the oldest Fram Strait contourites is estimated to be Miocene to Pliocene, i.e. deposited 4 to 7 Myrs after the initial connection between the Arctic and Atlantic oceans. Those contourites rest on a long hiatus. Contourite deposition has continued through the late Neogene and Quaternary, but vertical changes in the pattern suggest that current intensities have decreased through time.

As informative as this particular case study may be, the oceanographic and sedimentary evolution observed in the vicinity of the Fram Strait should not be strictly extrapolated to the other transform marginal plateau domains. Indeed, most of them evolved under different stages of breakup and oceanic opening (Early Jurassic opening of the Indian ocean for Morondava, Gunnerus, Exmouth - Figure 3), climatic conditions (some plateaus formed under icehouse conditions, others under greenhouse conditions, such as Demerara, Guinea, Côte d'Ivoire-Ghana, Falklands, Cuvier-Wallaby, Naturaliste) and latitudes. In many cases, the timing of the onset of regional oceanographic circulation and associated contourites appears to be different.

For example, in the Demerara Plateau, which formed one of the last gateways before complete connection of the Central and Equatorial Atlantic oceans (Campan, 1995; Moulin et al., 2010), Fanget et al. (2017) defined three

evolutionary stages: (1) a first pre-contourite stage probably lasting from the late Albian to early Miocene, during which no clear influence of bottom currents could be identified on seismic data. During this stage, sedimentation passed from Cenomanian and Turonian highly organic black shales, deposited in a still highly confined Equatorial Atlantic basin, to Campanian oxic deposits as the Equatorial Atlantic Gateway definitively opened and allowed the South Atlantic Intermediate Water to invade the proto North Atlantic (Friedrich and Erbacher 2006; Mosher et al., 2007; Donnadieu et al., 2016; Friedrich et al., 2012). During the Eocene-Oligocene, despite the fact that this period is marked by the beginning of Cenozoic icehouse conditions (Séranne, 1999) and the onset of global circulation in the Atlantic (Katz et al., 2011; Uenzelmann-Neben et al., 2016), no clear contourite deposits could be mapped. Fanget et al. (2017) postulate that this record may have been removed by later mass transport deposits described by Pattier et al. (2013) or that circulation was still weak, as suggested by Uenzelmann-Neben et al. (2016), (2) During a second transitional middle Miocene to early Pliocene stage, regional erosion and slope destabilization marked major changes in oceanic circulation that led to the permanent establishment of modern thermohaline circulation (Niemi et al., 2000; Pfuhl and McCave, 2005; Hernandez-Molina et al., 2009; Herold et al., 2012; Uenzelmann-Neben et al., 2016). At this time, a first contourite terrace shapes the plateau, (3) Finally, during a Plio-Quaternary contourite stage the plateau is dominantly shaped by bottom-currents, respectively the north-flowing AAIW (Antarctic Intermediate Water) in the upper plateau, the south-flowing North Atlantic Deep Water (NADW) on the outer plateau and the deeper north-flowing Antarctic Bottom Water at the foot of the plateau on the Demerara abyssal plain (Tallobre et al., 2016; Fanget et al., 2017).

Further south, along the southern Argentine continental slope and the Falkland-Malvinas escarpment, Hernandez et al. (2009) also describe a wide contourite depositional system (CDS) based on seismic stratigraphy. This CDS evolved following three stages: (1) a first development at the Eocene/Oligocene boundary, potentially coeval with the opening of the Drake passage and the onset of the Antarctic Bottom Current, (2) an early to middle Miocene aggradation phase, and (3) a late Miocene to present stage during which more vigorous intermediate and deep circulation are recorded and interpreted as related to the extension of the NADW in the southern hemisphere.

Finally, Exon et al., 2002, based on ODP drills, show that the Tasman TMP played the role of a shallow land bridge during Australia and Antarctica's last phase of separation. They evidence three evolutive phases: (1) between 43 to 37 Myrs, the Tasman Plateau played the role of a shallow shelf environment landbridge between the western proto Southern Ocean and the Tasman sea, (2) between 37 and 33.5 Myrs, a shallow gateway opened along the south Tasman TMP allowing the emplacement of strong and shallow currents and the change from organic-rich to organic-poor sedimentation, (3) after the definitive separation of Tasman TMP and Antarctica (33.5 Myrs onward), they underline important paleoenvironmental changes with the onset of the Antarctic Circumpolar Current with both shallow and deep circulation leading to global cooling and some formation of ice-sheets.

At present, since the final establishment of modern deep water circulation, contourite deposits have been described at the foot or along most TMPs (see

complete reviews and references in Rebesco et al., 2014 and Thran et al., 2018). This observation is not conclusive, since Rebesco et al. (2014) underline that those bottom-current controlled features are ubiquitous in marine basins. Loncke et al. (2016) and Talloire et al. (2016) propose that the presence of transform marginal plateaus that extend the continental shelf seaward and reach important depths could affect deep current dynamics in constraining and accelerating deep current flows. TMPs may therefore be considered as key domains able to record slight variations in current intensity especially in transitional periods of climatic-related current intensification or reduction.

## Conclusions

This review allows us to propose new key observations and starting hypotheses for the understanding of transform marginal plateaus:

- (1) 65 to 90% of TMPs have undergone volcanic activity during their history. Our review highlights in particular a close and hitherto unrecognized link between hotspots, volcanic activity and transform margins.
- (2) Some plateaus (10%) are underlain by thinned continental crust as stated by Mercier de Lépinay et al. (2016), but the great majority of them seems to consist of deformed, intruded and underplated continental crust. Others may be formed of thickened oceanic or transitional igneous crusts.
- (3) 75% of TMPs formed at the junction of ocean basins of contrasted ages. For these marginal plateaus, transforms always occur in a second breakup phase and allowed the connection of spatially shifted oceanic domains.
- (4) These TMPs, especially when located at the junction of ocean basins of contrasted ages, have undergone a complex polyphased history and host different generations of sedimentary basins and associated long-term sedimentary archives.
- (5) TMPs that underwent syn-breakup volcanism and stayed in a high-standing emerged position probably played the role of land bridges during the Gondwana and later breakup phases.
- (6) TMPs and their neighboring margins seem to be key domains for studying the onset and variations of oceanic currents.

Based on these new observations, several scientific challenges may have to be addressed in the future:

- (1) What are the geodynamic and thermo-mechanical conditions for transform margin initiation? What is the link between hotspot activity and transform margins?
- (2) For a long time, it was considered that oceanic plateaus, due to their thickness, buoyancy and rheology, could not enter into subduction (Cloos, 1993). However, since then the subduction of oceanic plateaus and its consequences on continental growth have been proven, like for the Laramide orogeny (Liu et al., 2010). Oceanic plateaus can cause flat subduction, with associated modified volcanism (Gutscher et al., 1999). Similarly, TMPs may also have geodynamic consequences in convergence, subduction, collision and continental growth that need to be explored.
- (3) What can we learn by studying the sedimentary and paleoenvironmental archives contained in and along TMPs, in particular concerning the high-resolution record of oceanographic variations through time?

- (4) Can we exemplify how those structures influence biodiversity growth, bio-connectivity and lineage evolution, in particular for terrestrial vertebrates and non-aquatic mammals during successive continental breakup phases?

### Figure Captions:

Table 1: List of marginal plateaus with names - numbers (Figure 1) and main characteristics: width, length, surface, water depth, surrounding water depth, conjugated margins, nature of the crust, available information on the crust, chronology of tectonic events and volcanism, duration between two openings (if relevant), associated LIPs (if relevant), drills, dredges and key references.

Figure 1: Location of Transform Marginal Plateaus (TMPs) on a world bathymetric and topographic map (ETOPO-1 model, Amante and Eakins 2009). Transform margins are underlined by a black thick line (from Mercier et al. 2016) (see also the TMP KML file in supplementary data). In blue, oceanic fracture zones from Matthews et al., 2011.

1. Morris Jesup Rise; 2: Yermarck; 3: NE Greenland; 4: Vøring Plateau; 5: Faroe-Rockall; 6: Demerara; 7: Guinea; 8: Liberia; 9: Côte d'Ivoire-Ghana; 10: Potiguar; 11: Sao Paulo; 12: Walvis; 13: Falklands-Malvinas; 14: Agulhas; 15: Gunnerus Ridge; 16: Morondava, 17: Tasman; 18: Naturaliste; 19: Wallaby-Cuvier; 20: Exmouth.

Figure 2: Bathymetric profiles of transform margins with and without marginal plateau domains (modified from Mercier de Lépinay et al., 2016). A/ Demerara Transform Plateau, offshore Suriname, with marginal plateau, B/ Ceara transform margin, offshore Brazil, with no marginal plateau. In the case of the Demerara transform marginal plateau, two slope domains (the shallow slope and the continental slope) separate the marginal plateau from the shelf and abyssal plain respectively.

Figure 3: Transform margins (thick black lines), transform marginal plateaus (red), and breakup ages (modified from Mercier de Lépinay et al., 2016). Most breakup ages are deduced from Müller et al., 2008 and Müller et al., 2016. Indian-Madagascar-Antarctic separation domains are from Marks and Tikku (2001) and Indian-Australia-Antarctic separation ages are from Gibbons et al. (2013). The Central Atlantic breakup age comes from Labails et al. (2010) and Davison (2010). In blue, oceanic fracture zones from Matthews et al., 2011.

1. Morris Jesup Rise; 2: Yermarck Plateau; 3: NZ Greenland Plateau; 4: Vøring Plateau; 5: Faroe-Rockall Plateau; 6: Demerara Plateau; 7: Guinea Plateau; 8: Liberia; 9: Côte d'Ivoire-Ghana; 10: Potiguar Plateau; 11: Sao Paulo Plateau; 12: Walvis Plateau; 13: Falklands-Malvinas Plateau; 14: Agulhas Plateau; 15: Gunnerus Ridge; 16: Morondava Plateau, 17: Tasman Plateau; 18: Naturaliste Plateau; 19: Wallaby-Cuvier Plateau; 20: Exmouth Plateau.

Domain abbreviations: B.Bay=Baffin Bay; L.Sea=Labrador Sea; N.Atl (N)=Northern North Atlantic; N.Atl(S)=Southern North Atlantic; Equ Atlantic = Equatorial Atlantic; A.-S-A=Africa-South America; A.-M.-A.=Africa-Madagascar-Antarctic; I.A.=India-

Antarctic; A.-A.=Australia-Antarctic; G.I.-A.=Greater India-Australia; Argo=Argo abyssal plain; T.Sea=Tasman Sea.

Figure 4: Bathymetric details along the 20 transform marginal plateaus (modified from Mercier de Lépinay et al., 2016). The yellow lines indicate the continent-ocean boundary along transform margins. The brown dashed line indicates the overall outline of each marginal plateau (completed by the KML supplementary file). The dashed black line corresponds to the main fracture zones and boundaries between two oceanic domains, generally of distinct ages. Bathymetric basemaps correspond to the Global Multi-Resolution Topography synthesis (Ryan et al., 2009) generated by GeoMapApp (<http://www.geomapapp.org>). Deep crustal lines shown in Figure 5 are shown in pink. Most breakup ages are deduced from Müller et al., 2008 and Müller et al., 2016. Arctic, North Atlantic and Fram strait breakup ages are from Engen et al., 2008. The Central Atlantic breakup age comes from Labails et al. (2010) and Davison (2010). Indian-Australia-Antarctic separation ages are from Gibbons et al. (2012). Breakup ages for Gunnerus are from Marks and Tikku 2001. JMFZ: Jan Mayen Fault Zone, CGFZ: Charlie Gibbs Fault Zone, SPFZ: Saint Paul Fault Zone, RFZ: Romanche Fault Zone, MR: Marginal Ridge, FFZ: Florianopolis Fault Zone, AFFZ: Agulhas Falklands Fault Zone, DFZ: Davie Fault Zone, DR: Davie Ridge, STR: South Tasman Rise, ETP: East Tasman Plateau, NFZ: Naturaliste Fault Zone, LFZ: Leewin Fault Zone, WZFZ: Wallaby Zenith Fault Zone, ZP: Zenith Plateau, CP: Cuvier Plateau.

Figure 5: Crustal-scale cross sections of TMPs investigated by deep-penetrating reflection seismic data and/or wide-angle seismic data. All cross-sections are displayed at the same scale and with the same colour code. Vertical exaggeration is 3.5.

Figure 6: Synthetic diagram summarizing the relative chronology of margin volcanism and segmentation along transform marginal plateaus (details and references in table 1). Volcanic events are underlined in red.

Figure 7: Regional cross-sections along the Demerara Plateau offshore French Guiana and Suriname. Interpretation and seismic stratigraphy are from Mercier de Lépinay (2016). A and B are oriented orthogonally to the Jurassic volcanic margin: Unit A corresponds to the basement, B corresponds to SDR sealed by the Jurassic breakup unconformity. This passive margin then hosts a thick sedimentary basin (units C, D, E, forming basin 2). All these series are later deformed and sealed by a top Albian breakup unconformity. The plateau then definitively becomes a passive margin. C is oriented orthogonally to the Cretaceous transform boundary of the plateau. Syn-Cretaceous wrench deformations are particularly well-expressed and sealed by Rd8. Syn and post-transform basins are indicated (respectively basins 3 and 4). Vertical exaggeration is at least 10 considering a mean seismic velocity of 2km/s.

Figure 8: Plate kinematic reconstruction of Africa/America respective positions at 110 Ma with South America fixed present-day bathymetry along TMPs and margins. Rotation poles and COB are from Matthews et al., 2016 and Torsvik et al., 2009.

Figure 9: Plate reconstruction of Engen et al., 2008 centered on the Fram Strait that formed between the Morris-Jesup and Yermack plateaus in a transform-dominated motion. Illustration of the (1) Arctic Ocean oxygen-poor “lake stage” as defined by Jakobsson et al., 2007 and the (2) opening of Fram Strait associated with complete ventilation of the Arctic Ocean (after Jakobsson et al., 2007).

## Supplementary

Data:

### Supplementary Data 1 and 2:

Two KML files are supplied with this paper. The first one, TransformMarginalPlateaus\_Loncke\_2019\_Final contains the outline of each of the identified transform marginal plateaus. For each plateau the following information is provided: an ID number, the surface area in km<sup>2</sup>; the average depth in meters; the standard deviation of the average depth. The second KML file, TransformMarginsInventory\_Update2019\_Final, provides an update to the transform margin inventory of Mercier de Lépinay et al., 2016. In some cases, modifications were made to adjust the transform margin trace to the outline of the adjacent marginal plateau. Also, transform margin traces were added for the additional transform marginal plateaus identified in the current paper. For each transform margin, the length of the transform segment is given in km, as well as its association to a marginal plateau and a marginal ridge, as applicable.

**Supplementary Data 3:** In order to further illustrate their morphological character, in this document we present images of bathymetric profiles derived from the GEBCO 2014 grid. For each of the transform marginal plateaus identified in this paper, a first profile extends from the shelf to the deep ocean floor, along the length of the plateau. A second profile crosses the plateau along its width, in a direction generally parallel to the coast. These profiles help appreciate the intermediate depths of TMPs, located between the shelf break and the lower slope.

**Supplementary Data 4:** In this document we present bathymetric profiles derived from the GEBCO 2014 grid for the Newfoundland and the Sierra Leone plateaus. These two plateaus were removed from the inventory of TMPs as they are in fact part of the continental shelf itself and do not constitute bathymetrical highs embedded within the slope. The profiles along and across these plateaus clearly demonstrate that they do not satisfy all the criteria of a TMP.

## Acknowledgements

Earthbytes open-source Gplates software was used for plate kinematic reconstructions (Müller et al., 2016) and Generic Mapping Tools software (GMT) was used to draft several figures (Wessel and Luis, 2017). We thank TOTAL and the ANRT for funding Marion Mercier de Lépinay's PhD thesis. We thank CNRS for funding of the Workshop "Marginal plateaus: from magmatic divergent margins to transform margins, and from the continental shelf to the pelagic and contouritic

setting" that took place in Grenoble, November 28-29th 2017 and was the starting point of numerous discussions about transform marginal plateaus.

## References

- Abdelmalak, M.M., Andersen, T.B., Planke, S., Faleide, J.I., Corfu, F., Tegner, C., Shephard, G.E., Zastrozhnov, D., Myklebust, R., 2015. The ocean-continent transition in the mid-Norwegian margin: Insight from seismic data and an onshore Caledonian field analogue. *Geology* 43, 1011-1014.
- Abdelmalak, M.M., Faleide, J.I., Planke, S., Gernigon, L., Zastrozhnov, D., Shephard, G.E., Myklebust, R., 2017. The *T* -Reflection and the Deep Crustal Structure of the Vøring Margin, Offshore mid-Norway: Vøring Margin Deep Crustal Structure. *Tectonics* 36, 2497-2523.
- Ali, J.R., Krause, D.W., 2011. Late Cretaceous bioconnections between Indo-Madagascar and Antarctica: refutation of the Gunnerus Ridge causeway hypothesis. *Journal of Biogeography* 38, 1855-1872.
- Almeida, F.F.M. de, Brito Neves, B.B. de, Dal Ré Carneiro, C., 2000. The origin and evolution of the South American Platform. *Earth-Science Reviews* 50, 77-111.
- Amante, C., and B. W. Eakins (2009), ETOP01 1 arc-minute global relief model: Procedures, data sources and analysis, NOAA Tech. Memo. NESDIS NGDC-24, 19 pp.
- Ammann, N., Liao, J., Gerya, T., Ball, P., 2018. Oblique continental rifting and long transform fault formation based on 3D thermomechanical numerical modeling. *Tectonophysics* 746, 106-120.
- Antobreh, A.A., Faleide, J.I., Tsikalas, F., Planke, S., 2009. Rift-shear architecture and tectonic development of the Ghana margin deduced from multichannel seismic reflection and potential field data. *Marine and Petroleum Geology* 26, 345-368.
- Araújo, M.G.S., Brito Neves, B.B., Archanjo, C.J., 2001. Idades 40AR/39AR do magmatismo básico Meso-Cenozóico da Província Borborema oriental, Nordeste do Brasil. In: Simpósio de Geologia do Nordeste, Natal, Brasil, pp. 260-261(Extended abstract).
- Attoh, K., Brown, L., Guo, J., Heanlein, J., 2004. Seismic stratigraphic record of transpression and uplift on the Romanche transform margin, offshore Ghana. *Tectonophysics* 378, 1-16.
- Bardintzeff, J.-M., Bonin, B., Rasamimanana, G., 2001. The Cretaceous Morondava volcanic province (West Madagascar): mineralogical, petrological and geochemical aspects. *Journal of African Earth Sciences* 32, 299-316.
- Bardintzeff, J.-M., Liégeois, J.-P., Bonin, B., Bellon, H., Rasamimanana, G., 2010. Madagascar volcanic provinces linked to the Gondwana break-up: Geochemical and isotopic evidences for contrasting mantle sources. *Gondwana Research* 18, 295-314.
- Barker, P.F., 1999. Evidence for a volcanic rifted margin and oceanic crustal structure for the Falkland Plateau Basin. *Journal of the Geological Society* 156, 889-900.



- Barker, P.F., Dalziel, I.W.D., et al., 1977. Initial Reports of the Deep Sea Drilling Project, 36, Initial Reports of the Deep Sea Drilling Project. U.S. Government Printing Office.
- Basile, C., 2015. Transform continental margins – part 1: Concepts and models. *Tectonophysics* 661, 1-10.
- Basile, C., Mascle, J., Benkhelil, J., Bouillin, J.-P., 1998. 11. Geodynamic evolution of the Côte d'Ivoire-Ghana transform margin: an overview of leg 159 results, in: *Proceedings of the Ocean Drilling Program, Scientific Results*.
- Basile, C., Mascle, J., Guiraud, R., 2005. Phanerozoic geological evolution of the Equatorial Atlantic domain. *Journal of African Earth Sciences* 43, 275-282.
- Basile, C., Braun, J., 2016. The initiation of pull-apart basins and transform continental margins: results from numerical experiments of kinematic partitioning in divergent settings. *Terra Nova* 28, 120-127.
- Basile, C., Maillard, A., Patriat, M., Gaullier, V., Loncke, L., Roest, W., Mercier de Lépinay, M., Pattier, F., 2013. Structure and evolution of the Demerara Plateau, offshore French Guiana: Rifting, tectonic inversion and post-rift tilting at transform-divergent margins intersection. *Tectonophysics* 591, 16-29.
- Basile, C., Mascle, J., Popoff, M., Bouillin, J.P., Mascle, G., 1993. The Ivory Coast-Ghana transform margin: a marginal ridge structure deduced from seismic data. *Tectonophysics* 222, 1-19.
- Bauer, K., Neben, S., Schreckenberger, B., Emmermann, R., Hinz, K., Fechner, N., Gohl, K., Schulze, A., Trumbull, R.B., Weber, K., 2000. Deep structure of the Namibia continental margin as derived from integrated geophysical studies. *Journal of Geophysical Research: Solid Earth* 105, 25829-25853.
- Ben-Avraham, Z., Hartnady, C.J.H., Kitchin, K.A., 1997. Structure and tectonics of the Agulhas-Falkland fracture zone. *Tectonophysics* 282, 83-98.
- Benkhelil, J., Mascle, J., Tricart, P., 1995. The Guinea continental margin: an example of a structurally complex transform margin. *Tectonophysics* 248, 117-137.
- Berndt, C., Mjelde, R., Planke, S., Shimamura, H., Faleide, J.I., 2001. Controls on the tectono-magmatic evolution of a volcanic transform margin: the Vøring Transform Margin, NE Atlantic. *Marine Geophysical Researches* 22, 133-152.
- Berndt, C., Skogly, O.P., Planke, S., Eldholm, O., Mjelde, R., 2000. High-velocity breakup-related sills in the Vøring Basin, off Norway. *Journal of Geophysical Research: Solid Earth* 105, 28443-28454.
- Beslier, M.-O., Royer, J.-Y., Girardeau, J., Hill, P.J., Boeuf, E., Buchanan, C., Chatin, F., Jacovetti, G., Moreau, A., Munsch, M., 2004. A wide ocean-continent transition along the south-west Australian margin: first results of the MARGAU/MD110 cruise. *Bulletin de la Société Géologique de France* 175, 629-641.
- Bouysse, P., 2018. ©CCGM/CGMW July 2014 ISBN 978-2-917310-24-3- Geological Map of the World at 1:35 000 000, 3rd revised edition.

- Brekke, H., 2000. The tectonic evolution of the Norwegian Sea continental margin, with emphasis on the Voring and More basins. *Special Publication-Geological Society of London* 167, 327-378.
- Bryan, S.E., Ferrari, L., 2013. Large igneous provinces and silicic large igneous provinces: Progress in our understanding over the last 25 years. *Geological Society of America Bulletin* 125, 1053-1078.
- Campan, A., 1995. Analyse cinématique de l'Atlantique équatorial: implications sur l'évolution de l'Atlantique Sud et sur la frontière de plaques Amérique du Nord/Amérique du Sud (PhD Thesis). Paris 6.
- Civile, D., Lodolo, E., Vuan, A., Loreto, M.F., 2012. Tectonics of the Scotia-Antarctica plate boundary constrained from seismic and seismological data. *Tectonophysics* 550-553, 17-34.
- Cloos, M., 1993. Lithospheric buoyancy and collisional orogenesis: Subduction of oceanic plateaus, continental margins, island arcs, spreading ridges, and seamounts. *Geological Society of America Bulletin* 105, 715-737.
- Cole, D., De Wit, M., Ransome, I., 1992. Evolution and development of the Karoo Basin. Inversion tectonics of the Cape Fold Belt, Karoo and cretaceous basins of Southern Africa 87-99.
- Coleman, P.J., Michael, P.J., Mutter, J.C., 1982. The origin of the Naturaliste Plateau, SE Indian Ocean: implications from dredged basalts. *Journal of the Geological Society of Australia* 29, 457-468.
- Colwell, J.B., Symonds, P.A., Crawford, A.J., 1994. The nature of the Wallaby (Cuvier) Plateau and other igneous provinces of the west Australian margin. *Journal of Australian Geology and Geophysics* 15, 137-156.
- Crawford, A.J., Lanyon, R., Elmes, M., Eggins, S., 1997. Geochemistry and significance of basaltic rocks dredged from the South Tasman Rise and adjacent seamounts. *Australian Journal of Earth Sciences* 44, 621-632.
- Cucciniello, C., le Roex, A.P., Jourdan, F., Morra, V., Grifa, C., Franciosi, L., Melluso, L., 2018. The mafic alkaline volcanism of SW Madagascar (Ankililoaka, Tulear region):  $^{40}\text{Ar}/^{39}\text{Ar}$  ages, geochemistry and tectonic setting. *Journal of the Geological Society* 175, 627-641.
- Curie, D., 1984. Ouverture de l'Atlantique Sud et discontinuités intra-plaques: une nouvelle analyse (PhD Thesis).
- Dailly, P., Henderson, T., Hudgens, E., Kanschhat, K., Lowry, P., 2013. Exploration for Cretaceous stratigraphic traps in the Gulf of Guinea, West Africa and the discovery of the Jubilee Field: a play opening discovery in the Tano Basin, Offshore Ghana. *Geological Society, London, Special Publications* 369, 235-248.
- Darwin, C., 1859. *On the Origin of Species by Means of Natural Selection* (J. Murray, London).

- Davison, I., Dailly, P., 2010. Salt tectonics in the Cap Boujdour Area, Aaiun Basin, NW Africa. *Marine and Petroleum Geology* 27, 435-441.
- De Castro, D.L., Bezerra, F.H.R., Sousa, M.O.L., Fuck, R.A., 2012. Influence of Neoproterozoic tectonic fabric on the origin of the Potiguar Basin, northeastern Brazil and its links with West Africa based on gravity and magnetic data. *Journal of Geodynamics* 54, 29-42.
- De Castro, D.L. and F.H.R. Bezerra, 2015. Fault evolution in the Potiguar rift termination equatorial margin of Brazil, *Solid Earth*, 6, 185-196.
- Dingle, R.V., 1973. Post-Palaeozoic stratigraphy of the eastern Agulhas Bank, South African continental margin. *Marine Geology* 15, 1-23.
- Dingle, R.V., Gentle, R.I., 1972. Early Tertiary volcanic rocks on the Agulhas Bank, South African Continental Shelf. *Geol. Mag.* 109, 127-136.
- Direen, N.G., Borissova, I., Stagg, H.M.J., Colwell, J.B., Symonds, P.A., 2007. Nature of the continent-ocean transition zone along the southern Australian continental margin: a comparison of the Naturaliste Plateau, SW Australia, and the central Great Australian Bight sectors. Geological Society, London, Special Publications 282, 239-263.
- Direen, N.G., Cohen, B.E., Maas, R., Frey, F.A., Whittaker, J.M., Coffin, M.F., Meffre, S., Halpin, J.A., Crawford, A.J., 2017. Naturaliste Plateau: constraints on the timing and evolution of the Kerguelen Large Igneous Province and its role in Gondwana breakup. *Australian Journal of Earth Sciences* 64, 851-869.
- Døssing, A., Stemmerik, L., Dahl-Jensen, T., Schindwein, V., 2010. Segmentation of the eastern North Greenland oblique-shear margin—regional plate tectonic implications. *Earth and Planetary Science Letters* 292, 239-253.
- Duncan, R.A., Hooper, P.R., Rehacek, J., Marsh, J., Duncan, A.R., 1997. The timing and duration of the Karoo igneous event, southern Gondwana. *Journal of Geophysical Research: Solid Earth* 102, 18127-18138.
- Eagles, G., König, M., 2008. A model of plate kinematics in Gondwana breakup. *Geophysical Journal International* 173, 703-717.
- Eiken, O., Hinz, K., 1993. Contourites in the Fram Strait. *Sedimentary Geology* 82, 15-32.
- Eldholm, O., Thiede, J., Taylor E., 1989. Evolution of the V\*ring Volcanic Margin. Proc. ODP, Sci. Results 104, 1033-1065.
- Eldholm, O., Tsikalas, F., Faleide, J.I., 2002. Continental margin off Norway 62-75 N: Palaeogene tectono-magmatic segmentation and sedimentation. Geological Society, London, Special Publications 197, 39-68.
- Elliott, G., 2017. An overview of the Hatton-Rockall plateau. ISTERre - Workshop "Marginal plateaus: from magmatic divergent margins to transform margins, and

from the continental shelf to the pelagic and contouritic setting”, November 28-29th, 2017.

- Elliott, G.M., Berndt, C., Parson, L.M., 2009. The SW African volcanic rifted margin and the initiation of the Walvis Ridge, South Atlantic. *Marine Geophysical Researches* 30, 207-214.
- Elliott, G.M., Parson, L.M., 2008. Influence of margin segmentation upon the break-up of the Hatton Bank rifted margin, NE Atlantic. *Tectonophysics* 457, 161-176.
- Emery, K.O., Uchupi, E., Phillips, J., Bowin, C., Mascle, J., 1975. Continental Margin Off Western Africa: Angola to Sierra Leone. *AAPG Bulletin* 59, 2209-2265.
- Engen, Ø., Faleide, J.I., Dyreng, T.K., 2008. Opening of the Fram Strait gateway: A review of plate tectonic constraints. *Tectonophysics* 450, 51-69.
- Evain, M., Afilhado, A., Rigoti, C., Loureiro, A., Alves, D., Klingelhofer, F., Schnurle, P., Feld, A., Fuck, R., Soares, J., de Lima, M.V., Corela, C., Matias, L., Benabdellouahed, M., Baltzer, A., Rabineau, M., Viana, A., Moulin, M., Aslanian, D., 2015. Deep structure of the Santos Basin-São Paulo Plateau System, SE Brazil: Santos Basin-São Paulo Plateau Structure. *Journal of Geophysical Research: Solid Earth* 120, 5401-5431.
- Ewing, J., Ewing, M., 1959. Seismic-refraction measurements in the Atlantic Ocean basins, in the Mediterranean Sea, on the Mid-Atlantic Ridge, and in the Norwegian Sea. *Geological Society of America Bulletin* 70, 291-318.
- Ewing, J.I., Ludwig, W.J., Ewing, M., Eittreim, S.L., 1971. Structure of the Scotia sea and Falkland plateau. *Journal of Geophysical Research* 76, 7118-7137.
- Exon, N.F., Kennett, J.P., Malone, M.J. (Eds.), 2004. *Proceedings of the Ocean Drilling Program, 189 Scientific Results, Proceedings of the Ocean Drilling Program. Ocean Drilling Program.*
- Exon, N.F., Kennett, J.P., Malone, M.J., et al. (Eds.), 2001. *Proceedings of the Ocean Drilling Program, 189 Initial Reports, Proceedings of the Ocean Drilling Program. Ocean Drilling Program.*
- Exon, N.F., Lafoy, Y., Hill, P.J., Dickens, G.R., Pecher, I., 2007. Geology and petroleum potential of the Fairway Basin in the Tasman Sea. *Australian Journal of Earth Sciences* 54, 629-645.
- Exon, N.F., Moore, A.M.G., Hill, P.J., 1997. Geological framework of the South Tasman Rise, south of Tasmania, and its sedimentary basins. *Australian Journal of Earth Sciences* 44, 561-577.
- Exon, N., Kennett, J., Malone, M., Brinkhuis, H., Chaproniere, G., Ennyu, A., Fothergill, P., Fuller, M., Grauert, M., Hill, P., Janecek, T., Kelly, C., Latimer, J., McGonigal, K., Nees, S., Ninnemann, U., Nuernberg, D., Pekar, S., Pellaton, C., Pfuhl, H., Robert, C., Röhl, U., Schellenberg, S., Shevenell, A., Stickley, C., Suzuki, N., Touchard, Y., Wei, W., White, T., 2002. Drilling reveals climatic consequences of Tasmanian Gateway Opening. *Eos Trans. AGU* 83, 253.

- Fainstein and Milliman 1979. Structure and Origin of Three Continental Margin Plateaus, Northeastern Brazil, AAPG Bulletin, Vol 63, Issue 2, 218-238.
- Faleide, J.I., Tsikalas, F., Breivik, A.J., Mjelde, R., Ritzmann, O., Engen, O., Wilson, J., Eldholm, O., 2008. Structure and evolution of the continental margin off Norway and the Barents Sea. *Episodes* 31, 82-91.
- Frey, Ø., Planke, S., Symonds, P.A., Heeremans, M., 1998. Deep crustal structure and rheology of the Gascoyne volcanic margin, western Australia. *Marine Geophysical Researches* 20, 293-311.
- Friedrich, O., Erbacher, J., 2006. Benthic foraminiferal assemblages from Demerara Rise (ODP Leg 207, western tropical Atlantic): possible evidence for a progressive opening of the Equatorial Atlantic Gateway. *Cretaceous Research* 27, 377-397.
- Friedrich, O., Norris, R.D., Erbacher, J., 2012. Evolution of middle to Late Cretaceous oceans—a 55 my record of Earth's temperature and carbon cycle. *Geology* 40, 107-110.
- Fromm, T., Jokat, W., Behrmann, J.H., 2017. Interaction between a hotspot and a fracture zone: The crustal structure of Walvis Ridge at 6° E. *Tectonophysics* 716, 108-120.
- Funck, T., Andersen, M.S., Keser Neish, J., Dahl-Jensen, T., 2008. A refraction seismic transect from the Faroe Islands to the Hatton-Rockall Basin. *Journal of Geophysical Research* 113.
- Geoffroy, L., Burov, E.B., Werner, P., 2015. Volcanic passive margins: another way to break up continents. *Scientific reports* 5, 14828.
- Gerya, T., 2010. Dynamical Instability Produces Transform Faults at Mid-Ocean Ridges. *Science* 329, 1047-1050.
- Gerya, T., 2012. Origin and models of oceanic transform faults. *Tectonophysics* 522, 34-54.
- Gibbons, A.D., Barckhausen, U., van den Bogaard, P., Hoernle, K., Werner, R., Whittaker, J.M., Müller, R.D., 2012. Constraining the Jurassic extent of Greater India: Tectonic evolution of the West Australian margin. *Geochemistry Geophysics Geosystems* 13.
- Gibbons, A.D., Whittaker, J.M., Müller, R.D., 2013. The breakup of East Gondwana: Assimilating constraints from Cretaceous ocean basins around India into a best-fit tectonic model. *Journal of geophysical research: solid earth* 118, 808-822.
- Gladczenko, T.P., Skogseid, J., Eldhom, O., 1998. Namibia volcanic margin. *Marine Geophysical Researches* 20, 313-341.
- Goncharov, A., Nelson, G., 2012. From two way time to depth and pressure for interpretation of seismic velocities offshore: Methodology and examples from the Wallaby Plateau on the West Australian margin. *Tectonophysics* 572, 26-37.

- Gouyet, S., 1988. Evolution tectono-sédimentaire des marges guyanaise et nord-brésilienne au cours de l'ouverture de l'Atlantique Sud (PhD Thesis).
- Gradstein, F.M., Ludden, J.N., et al. (Eds.), 1992. Proceedings of the Ocean Drilling Program, 123 Scientific Results, Proceedings of the Ocean Drilling Program. Ocean Drilling Program.
- Greenroyd, C.J., Peirce, C., Rodger, M., Watts, A.B., Hobbs, R.W., 2008. Demerara Plateau - the structure and evolution of a transform passive margin. *Geophysical Journal International* 172, 549-564.
- Greenroyd, C.J., Peirce, C., Rodger, M., Watts, A.B., Hobbs, R.W., 2007. Crustal structure of the French Guiana margin, West Equatorial Atlantic. *Geophysical Journal International* 169, 964-987.
- Greenway, M.E., 1972. The geology of the Falkland Islands. Cambridge, British Antarctic Survey, 42pp.
- Guazelli, W., Kosta M.P.A and Kowsmann R.O., 1977. Cruzeiro, platôs marginais do Nordeste Brasileiro : resultados geologicos preliminares. In : simposio de Geologia do Nordeste Brasileiro, 8, Camina Grande. Petrobras.
- Gutscher, M.-A., Olivet, J.-L., Aslanian, D., Eissen, J.-P., Maury, R., 1999. The “lost inca plateau”: cause of flat subduction beneath peru? *Earth and Planetary Science Letters* 171, 335-341.
- Halpin, J.A., Crawford, A.J., Direen, N.G., Coffin, M.F., Forbes, C.J., Borissova, I., 2008. Naturaliste Plateau, offshore Western Australia: A submarine window into Gondwana assembly and breakup. *Geology* 36, 807-810.
- Hamann, N.E., Whittaker, R.C., Stemmerik, L., 2005. Geological development of the Northeast Greenland Shelf. Geological Society, London, Petroleum Geology Conference series 6, 887-902.
- Hawkesworth, C., Kelley, S., Turner, S., Le Roex, A., Storey, B., 1999. Mantle processes during Gondwana break-up and dispersal. *Journal of African Earth Sciences* 28, 239-261.
- Heine, C., Zoethout, J., Müller, R.D., 2013. Kinematics of the South Atlantic rift. *Solid Earth* 4, 215-253.
- Hekinian, R., Thompson, G., 1976. Comparative geochemistry of volcanics from Rift Valleys, transform faults and aseismic ridges 18. *Mineralogy and Petrology*, Volume 57, Issue 2, pp.145-162.
- Hergt, J.M., Chappell, B.W., McCulloch, M.T., McDougall, I., Chivas, A.R., 1989. Geochemical and isotopic constraints on the origin of the Jurassic dolerites of Tasmania. *Journal of Petrology* 30, 841-883.
- Hernández-Molina, F.J., Paterlini, M., Violante, R., Marshall, P., de Isasi, M., Somoza, L., Rebesco, M., 2009. Contourite depositional system on the Argentine Slope: An exceptional record of the influence of Antarctic water masses. *Geology* 37, 507-510.

- Herold, N., Huber, M., Müller, R.D., Seton, M., 2012. Modeling the Miocene climatic optimum: Ocean circulation. *Paleoceanography and Paleoclimatology* 27.
- Hill, P.J., Exon, N.F., Keene, J.B., Smith, S.M., 1998. The continental margin off east Tasmania and Gippsland: structure and development using new multibeam sonar data. *Exploration Geophysics* 29, 410-419.
- Hill, P.J., Moore, A.M.G., Exon, N.F., 2001. Sedimentary basins and structural framework of the South Tasman Rise and East Tasman Plateau.
- Hitchen, K., 2004. The geology of the UK Hatton-Rockall margin. *Marine and Petroleum Geology* 20. Volume 21, Issue 8, 993-1012.
- Honnorez, J., Mascle, J., Basile, C., Tricart, P., Villeneuve, M., Bertrand, H., 1991. Mapping of a segment of the Romanche Fracture Zone: A morphostructural analysis of a major transform fault of the equatorial Atlantic Océan. *Geology*, 19 (8): 795-798.
- IHO, S., 2008. 44. IHO Standards for Hydrographic Surveys.
- Jackson, H.R., Johnson, G.L., Sundvor, E., Myhre, A.M., 1984. The Yermak Plateau: Formed at a triple junction. *Journal of Geophysical Research: Solid Earth* 89, 3223-3232.
- Jakobsson, M., Backman, J., Rudels, B., Nycander, J., Frank, M., Mayer, L., Jokat, W., Sangiorgi, F., O'Regan, M., Brinkhuis, H., 2007. The early Miocene onset of a ventilated circulation regime in the Arctic Ocean. *Nature* 447, 986.
- James A. Austin, J., 1982. Continental-Oceanic Crustal Transition Off Southwest Africa. *AAPG Bulletin* 66, 1328-1347.
- Jokat, W., 2000. The expedition ARKTIS-XV/2 of "Polarstern" in 1999. *Berichte zur Polarforschung (Reports on Polar Research)* 368.
- Jokat, W., Boebel, T., König, M., Meyer, U., 2003. Timing and geometry of early Gondwana breakup. *Journal of Geophysical Research: Solid Earth* 108.
- Jokat, W., Nogi, Y., Leinweber, V., 2010. New aeromagnetic data from the western Enderby Basin and consequences for Antarctic-India break-up. *Geophysical Research Letters* 37.
- Jones, E.J.W., Goddard, D.A., Mitchell, J.G., Banner, F.T., 1991. Lamprophyric volcanism of Cenozoic age on the Sierra Leone Rise: implications for regional tectonics and the stratigraphic time scale. *Marine Geology* 99, 19-28.
- Katz, M.E., Cramer, B.S., Toggweiler, J.R., Esmay, G., Liu, C., Miller, K.G., Rosenthal, Y., Wade, B.S., Wright, J.D., 2011. Impact of Antarctic Circumpolar Current development on late Paleogene ocean structure. *Science* 332, 1076-1079.
- Klingelhoefer, F., Evain, M., Afilhado, A., Rigoti, C., Loureiro, A., Alves, D., Leprêtre, A., Moulin, M., Schnurle, P., Benabdellouahed, M., Baltzer, A., Rabineau, M., Feld, A., Viana, A., Aslanian, D., 2014. Imaging proto-oceanic crust off the Brazilian Continental Margin. *Geophysical Journal International* 200, 471-488.

- Klingelhoefer, F., Edwards, R.A., Hobbs, R.W., England, R.W., 2005. Crustal structure of the NE Rockall Trough from wide-angle seismic data modeling. *Journal of Geophysical Research: Solid Earth* 110.
- Klingelhoefer, F., Musser, T., Roest, W.R., Graindorge, D., Chauvet, F., Loncke, L., Basile, C., Poetisi, E., Deverchere, J., Lebrun, J.F., Perrot, J., Heuret, A., 2017. Tectonic History and Deep Structure of the Demerara Plateau from Combined Wide-Angle and Reflection Seismic Data and Plate Kinematic Reconstructions. *AGU Fall Meeting Abstracts* 51.
- König, M., Jokat, W., 2010. Advanced insights into magmatism and volcanism of the Mozambique Ridge and Mozambique Basin in the view of new potential field data. *Geophysical Journal International* 180, 158-180.
- Kumar, N., Gamboa, L.A.P., Schreiber, B.C., Mascle, J., 1977. Geologic history and origin of Sao Paulo Plateau (Southeastern Brazilian Margin), comparison with the Angolan margin and the early evolution of the Northern South Atlantic. *Initial Reports of the Deep Sea Drilling Program* 39, 927-945.
- Labails, C., Olivet, J.-L., Aslanian, D., Roest, W.R., 2010. An alternative early opening scenario for the Central Atlantic Ocean. *Earth and Planetary Science Letters* 297, 355-368.
- Larson, R.L., Mutter, J.C., Diebold, J.B., Carpenter, G.B., Symonds, P., 1979. Cuvier Basin: a product of ocean crust formation by Early Cretaceous rifting off Western Australia. *Earth and Planetary Science Letters* 45, 105-114.
- Lebrun, J.-F., Lamarche, G., Collot, J.-Y., 2003. Subduction initiation at a strike-slip plate boundary: The Cenozoic Pacific-Australian plate boundary, south of New Zealand: subduction initiation at strike-slip boundary. *J. Geophys. Res.* Vol 108. B9, 2453.
- Leitchenkov, G., Guseva, J., Gandyukhin, V., Grikurov, G., Kristoffersen, Y., Sand, M., Golynsky, A., Aleshkova, N., 2008. Crustal structure and tectonic provinces of the Riiser-Larsen Sea area (East Antarctica): results of geophysical studies. *Marine Geophysical Researches* 29, 135-158.
- Leyden, R., Ludwig, W.J., Ewing, M., 1971. Structure of Continental Margin off Punta del Este, Uruguay, and Rio de Janeiro, Brazil. *AAPG Bulletin* 55, 2161-2173.
- Liao, J., Gerya, T., 2015. From continental rifting to seafloor spreading: Insight from 3D thermo-mechanical modeling. *Gondwana Research* 28, 1329-1343.
- Liu L., Gurnis M., Seton M., Saleeby, J., Müller R.D., and J.M. Jackson, 2010, The role of oceanic plateau subduction in the Laramide orogeny, *Nature Geoscience*, vol 3, 353-357.
- Livermore, R., 1994. Scotia Sea tectonics from high-resolution satellite gravity. *Earth and Planetary Science Letters* 14.
- Loncke, L., Droz, L., Gaullier, V., Basile, C., Patriat, M., Roest, W., 2009. Slope instabilities from echo-character mapping along the French Guiana transform margin and Demerara abyssal plain. *Marine and Petroleum Geology* 26, 711-723.



- Loncke, L., Maillard, A., Basile, C., Roest, W.R., Bayon, G., Gaullier, V., Pattier, F., de L epinay, M.M., Grall, C., Droz, L., 2016. Structure of the Demerara passive-transform margin and associated sedimentary processes. Initial results from the IGUANES cruise. Geological Society, London, Special Publications 431, 179-197.
- Lorenzo, J.M., Mutter, J.C., 1988. Seismic stratigraphy and tectonic evolution of the Falkland/Malvinas Plateau. *Revista Brasileira de Geociencias* 18, 191-200.
- Lorenzo, J.M., Mutter, J.C., Larson, R.L., 1991. Development of the continent-ocean transform boundary of the southern Exmouth Plateau. *Geology* 19, 843-846.
- Lorenzo, J.M., Wessel, P., 1997. Flexure across a continent-ocean fracture zone: the northern Falkland/Malvinas Plateau, South Atlantic. *Geo-Marine Letters* 17, 110-118.
- Ludden, J.N., Gradstein, F.M., et al. (Eds.), 1990. Proceedings of the Ocean Drilling Program, 123 Initial Reports, Proceedings of the Ocean Drilling Program. Ocean Drilling Program.
- Ludwig, W.J., Krasheninnikov, V.A., et al., 1983. Initial Reports of the Deep Sea Drilling Project, 71, Initial Reports of the Deep Sea Drilling Project. U.S. Government Printing Office.
- Maia, M., Sichel, S., Briais, A., Brunelli, D., Ligi, M., Ferreira, N., Campos, T., Mougel, B., Brehme, I., H emond, C., Motoki, A., Moura, D., Scalabrin, C., Pessanha, I., Alves, E., Ayres, A., Oliveira, P., 2016. Extreme mantle uplift and exhumation along a transpressive transform fault. *Nature Geosci* 9, 619-623.
- Marks, K.M., Tikku, A.A., 2001. Cretaceous reconstructions of East Antarctica, Africa and Madagascar. *Earth and Planetary Science Letters* 186, 479-495.
- Mascl e, J., 1976. Atlantic-type continental margins - Distinction of two basic structural types. *Anais da Academia Brasileira de Ciencias* 48, 191-197.
- Mascl e, J., Blarez, E., 1987. Evidence for transform margin evolution from the Ivory Coast-Ghana continental margin. *Nature* 326, 378.
- Mascl e, J., Guiraud, M., Basile, C., Benkhelil, J., Bouillin, J.-P., Cousin, M., Mascl e, G., 1993. La marge transformante de C ote-d'Ivoire-Ghana: premiers r esultats de la campagne Equanaute (Juin 1992). *Comptes rendus de l'Acad emie des sciences. S erie 2, M ecanique, Physique, Chimie, Sciences de l'univers, Sciences de la Terre* 316, 1255-1261.
- Mascl e, J., Lohmann, G.P., Moullade, M. (Eds.), 1998. Proceedings of the Ocean Drilling Program, 159 Scientific Results, Proceedings of the Ocean Drilling Program. Ocean Drilling Program.
- McCall, R.A., 1997. Implications of recent geological investigations of the Mozambique Channel for the mammalian colonization of Madagascar. *Proceedings of the Royal Society of London B: Biological Sciences* 264, 663-665.
- Mayes, C.L., Lawver, L.A., Sandwell, D.T., 1990. Tectonic history and new isochron chart of the south Pacific. *J. Geophys. Res.* 95, 8543.

- Matthews, K. J., Maloney, K. T., Zahirovic, S., Williams, S. E., Seton, M., & Mueller, R. D. (2016). Global plate boundary evolution and kinematics since the late Paleozoic. *Global and Planetary Change*, 146, 226-250. Matthews et al. 2016
- Meisling K.E., Cobbold, P.R., Van Mount S. 2001. Segmentation of an Obliquely Rifted Margin, Campos and Santos Basins, Southeastern Brazil, AAPG Bulletin, 85(11) : 1903-1924.
- Mercier de Lépinay, M., 2016. Inventaire mondial des marges transformantes et évolution tectonosédimentaire des plateaux de Demerara et de Guinée, Oceanography (PhD Thesis). Thèse de Doctorat, Université de Perpignan Via Domitia, Perpignan.
- Mercier de Lépinay, M., Loncke, L., Basile, C., Roest, W.R., Patriat, M., Maillard, A., De Clarens, P., 2016. Transform continental margins-Part 2: A worldwide review. *Tectonophysics* 693, 96-115.
- Mihut, D., Müller, R.D., 1998. Volcanic margin formation and Mesozoic rift propagators in the Cuvier Abyssal Plain off Western Australia. *Journal of Geophysical Research: Solid Earth* 103, 27135-27149.
- Mosher, D.C., Erbacher, J., Malone, M.J. (Eds.), 2007. Proceedings of the Ocean Drilling Program, 207 Scientific Results, Proceedings of the Ocean Drilling Program. Ocean Drilling Program.
- Moulin, M., Aslanian, D., Rabineau, M., Patriat, M., Matias, L., 2013. Kinematic keys of the Santos-Namibe basins. *Geological Society, London, Special Publications* 369, 91-107.
- Moulin, M., Aslanian, D., Unternehr, P., 2010. A new starting point for the South and Equatorial Atlantic Ocean. *Earth-Science Reviews* 98, 1-37.
- Müller, R. D., M. Sdrolias, C. Gaina, and W. R. Roest (2008), Age, spreading rates, and spreading asymmetry of the world's ocean crust, *Geochem. Geophys. Geosyst.*, 9, Q04006, doi:10.1029/2007GC001743.
- Müller, R.D., Seton, M., Zahirovic, S., Williams, S.E., Matthews, K.J., Wright, N.M., Shephard, G.E., Maloney, K.T., Barnett-Moore, N., Hosseinpour, M., Bower, D.J., Cannon, J., 2016. Ocean Basin Evolution and Global-Scale Plate Reorganization Events Since Pangea Breakup. *Annu. Rev. Earth Planet. Sci.* 44, 107-138.
- Müller, R.D., Gaina, C., Roest, W.R., Hansen, D.L., 2001. A recipe for microcontinent formation. *Geology* 29, 203-206.
- Müller, R.D., Seton, M., Zahirovic, S., Williams, S.E., Matthews, K.J., Wright, N.M., Shephard, G.E., Maloney, K.T., Barnett-Moore, N., Hosseinpour, M., Bower, D.J., Cannon, J., 2016. Ocean Basin Evolution and Global-Scale Plate Reorganization Events Since Pangea Breakup. *Annual Review of Earth and Planetary Sciences* 44, 107-138.

- Müller, R. D., Cannon, J., Qin, X., Watson, R. J., Gurnis, M., Williams, S., et al. 2018. [GPlates: Building a virtual Earth through deep time](#), in: *Geochemistry, Geophysics, Geosystems*, 19. doi:10.1029/2018GC007584.
- Mutter, J.C., Talwani, M., Stoffa, P.L., 1982. Origin of seaward-dipping reflectors in oceanic crust off the Norwegian margin by “subaerial sea-floor spreading.” *Geology* 10, 353-357.
- Mizusaki, A.M.P., Thomaz-Filho, A., Césero, P., 1998. Ages of the magmatism and opening of the South Atlantic Ocean. *Pesquisas* 25 (2), 47-57.
- Mussett, A.E., Taylor, G.K., 1994.  $^{40}\text{Ar}$ - $^{39}\text{Ar}$  ages for dykes from the Falkland Islands with implications for the break-up of southern Gondwanaland. *Journal of the Geological Society* 151, 79-81.
- Niemi, T.M., Ben-Avraham, Z., Hartnady, C.J., Reznikov, M., 2000. Post-Eocene seismic stratigraphy of the deep ocean basin adjacent to the southeast African continental margin: a record of geostrophic bottom current systems. *Marine Geology* 162, 237-258.
- Nur, A., Ben-Avraham, Z., 1982. Oceanic plateaus, the fragmentation of continents, and mountain building. *Journal of Geophysical Research: Solid Earth* 87, 3644-3661.
- Olierook, H.K., Merle, R.E., Jourdan, F., Sircombe, K., Fraser, G., Timms, N.E., Nelson, G., Dadd, K.A., Kellerson, L., Borissova, I., 2015. Age and geochemistry of magmatism on the oceanic Wallaby Plateau and implications for the opening of the Indian Ocean. *Geology* 43, 971-974.
- Olyphant, J.R., Johnson, R.A., Hughes, A.N., 2017. Evolution of the Southern Guinea Plateau: Implications on Guinea-Demerara Plateau formation using insights from seismic, subsidence, and gravity data. *Tectonophysics* 717, 358-371.
- Owens, R., Borissova, I., Southby, C., Hall, L., Bernardel, G., Grosjean, E., Mitchell, C., 2018. Tectono-stratigraphic development of the northern Houtman Sub-basin, Perth Basin. *ASEG Extended Abstracts 2018*, 1-8.
- Parsiegla, N., Stankiewicz, J., Gohl, K., Ryberg, T., Uenzelmann-Neben, G., 2009. Southern African continental margin: Dynamic processes of a transform margin. *Geochemistry, Geophysics, Geosystems* 10.
- Patchett, P.J., Chase, C.G., 2002. Role of transform continental margins in major crustal growth episodes. *Geology* 30, 39-42.
- Pattier, F., Loncke, L., Gaullier, V., Basile, C., Maillard, A., Imbert, P., Roest, W.R., Vendeville, B.C., Patriat, M., Loubrieu, B., 2013. Mass-transport deposits and fluid venting in a transform margin setting, the eastern Demerara Plateau (French Guiana). *Marine and Petroleum Geology* 46, 287-303.
- Pattier, F., Loncke, L., Imbert, P., Gaullier, V., Basile, C., Maillard, A., Roest, W.R., Patriat, M., Vendeville, B.C., Marsset, T., 2015. Origin of an enigmatic regional Mio-Pliocene unconformity on the Demerara plateau. *Marine Geology* 365, 21-35.

- Pfuhl, H.A., McCave, I.N., 2005. Evidence for late Oligocene establishment of the Antarctic Circumpolar Current. *Earth and Planetary Science Letters* 235, 715-728.
- Phipps Morgan, J., Morgan, W.J., Price, E., 1995. Hotspot melting generates both hotspot volcanism and a hotspot swell? *J. Geophys. Res.* 100, 8045-8062.
- Planert, L., Behrmann, J., Jokat, W., Fromm, T., Ryberg, T., Weber, M., Haberland, C., 2017. The wide-angle seismic image of a complex rifted margin, offshore North Namibia: Implications for the tectonics of continental breakup. *Tectonophysics* 716, 130-148.
- Planke, S., Symonds, P.A., Alvestad, E., Skogseid, J., 2000. Seismic volcanostratigraphy of large-volume basaltic extrusive complexes on rifted margins. *Journal of Geophysical Research: Solid Earth* 105, 19335-19351.
- Planke, S., Millett, J.M., Maharjan, D., Jerram, D.A., Abdelmalak, M.M., Groth, A., Hoffmann, J., Berndt, C., Myklebust, R., 2017. Igneous seismic geomorphology of buried lava fields and coastal escarpments on the Vøring volcanic rifted margin. *Interpretation* 5, SK161-SK177.
- Platt, N.H., Philip, P.R., 1995. Structure of the southern Falkland Islands continental shelf: initial results from new seismic data. *Marine and Petroleum Geology* 12, 759-771.
- Powell, D.E., 1976. The geological evolution of the continental margin off northwest Australia. *The APPEA Journal* 16, 13-23.
- Rabinowitz, P.D., LaBrecque, J., 1979. The Mesozoic South Atlantic Ocean and evolution of its continental margins. *Journal of Geophysical Research: Solid Earth* 84, 5973-6002.
- Reeves, C., De Wit, M., 2000. Making ends meet in Gondwana: retracing the transforms of the Indian Ocean and reconnecting continental shear zones. *Terra Nova* 12, 272-280.
- Reuber, K.R., Pindell, J., Horn, B.W., 2016. Demerara Rise, offshore Suriname: Magma-rich segment of the Central Atlantic Ocean, and conjugate to the Bahamas hot spot. *GeoScienceWorld Interpretation* 4, T141-T155.
- Richardson, K.R., Smallwood, J.R., White, R.S., Snyder, D.B., Maguire, P.K.H., 1998. Crustal structure beneath the Faroe Islands and the Faroe-Iceland Ridge. *Tectonophysics* 300, 159-180.
- Riefstahl, F., Estrada, S., Geissler, W.H., Jokat, W., Stein, R., Kämpf, H., Dulski, P., Naumann, R., Spiegel, C., 2013. Provenance and characteristics of rocks from the Yermak Plateau, Arctic Ocean: Petrographic, geochemical and geochronological constraints. *Marine Geology* 343, 125-145.
- Roberts, D.G., 1975. Marine geology of the Rockall Plateau and Trough. *Phil. Trans. R. Soc. Lond. A* 278, 447-509.

- Rollet, N., Royer, J.-Y., Exon, N.F., Hill, P.J., 1996. Le plateau Sud-Tasman (Australie): collage de deux fragments du Gondwana oriental? *Comptes rendus de l'Académie des sciences. Série 2. Sciences de la terre et des planètes* 323, 865-872.
- Rowan, M.G., 2014. Passive-margin salt basins: hyperextension, evaporite deposition, and salt tectonics. *Basin Research* 26, 154-182.
- Royer, J.-Y., Rollet, N., 1997. Plate-tectonic setting of the Tasmanian region. *Australian Journal of Earth Sciences* 44, 543-560.
- Ryan, W.B.F., Carbotte, S.M., Coplan, J.O., O'Hara, S., Melkonian, A., Arko, R., Weissel, R.A., Ferrini, V., Goodwillie, A., Nitsche, F., Bonczkowski, J., Zemsky, R., 2009. Global Multi-Resolution Topography synthesis: GLOBAL MULTI-RESOLUTION TOPOGRAPHY SYNTHESIS. *Geochemistry, Geophysics, Geosystems* 10, n/a-n/a.
- Sage, F., Basile, C., Mascle, J., Pontoise, B., Whitmarsh, R.B., 2000. Crustal structure of the continent-ocean transition off the Côte d'Ivoire-Ghana transform margin: implications for thermal exchanges across the palaeotransform boundary. *Geophysical Journal International*, vol 143, issue 3, 662-678.
- Saki, T., Tamura, Y., Tokuhashi, S., 1987. Surveys off Queen Maud Land, East Antarctica, in: *Proc. NIPR Symp. Antarct. Geosci.* pp. 23-40.
- Schimschal, C.M., Jokat, W., 2018. The crustal structure of the continental margin east of the Falkland Islands. *Tectonophysics* 724, 234-253.
- Schubert, G., Sandwell, D., 1989. Crustal volumes of the continents and of oceanic and continental submarine plateaus. *Earth and Planetary Science Letters* 92, 234-246.
- Self, S., Coffin, M.F., Rampino, M.R., Wolff, J.A., 2015. Chapter 24 - Large Igneous Provinces and Flood Basalt Volcanism, in: Sigurdsson, H. (Ed.), *The Encyclopedia of Volcanoes (Second Edition)*. Academic Press, Amsterdam, Elsevier, pp. 441-455.
- Séranne, M., 1999. Early Oligocene stratigraphic turnover on the west Africa continental margin: a signature of the Tertiary greenhouse-to-icehouse transition? *Terra Nova-Oxford* 11, 135-140.
- Seton, M., Müller, R.D., Zahirovic, S., Gaina, C., Torsvik, T., Shephard, G., Talsma, A., Gurnis, M., Turner, M., Maus, S., Chandler, M., 2012. Global continental and ocean basin reconstructions since 200Ma. *Earth-Science Reviews* 113, 212-270.
- Simpson, E.S.W., Schlich, R., et al., 1974. Initial Reports of the Deep Sea Drilling Project, 25, *Initial Reports of the Deep Sea Drilling Project*. U.S. Government Printing Office.
- Skogseid, J., Planke, S., Faleide, J.I., Pedersen, T., Eldholm, O., Neverdal, F., 2000. NE Atlantic continental rifting and volcanic margin formation. *Geological Society, London, Special Publications* 167, 295-326.
- Sobolev, S. V., N. Laube, K. Bauer, A. Schulze,, R. Trumbull, 1998. Signature of a mantle plume in the seismic velocity structure of the Namibian continental margin. Presented at the IAVCEI International Volcanological Congress, Cape Town, South Africa,.

- Stilwell, J.D., Quilty, P.G., Mantle, D.J., 2012. Paleontology of Early Cretaceous deep-water samples dredged from the Wallaby Plateau: new perspectives of Gondwana break-up along the Western Australian margin. *Australian Journal of Earth Sciences* 59, 29-49.
- Stone, P., Richards, P.C., Kimbell, G.S., Esser, R.P., Reeves, D., 2008. Cretaceous dykes discovered in the Falkland Islands: implications for regional tectonics in the South Atlantic. *Journal of the Geological Society* 165, 1-4.
- Sutherland, R., 1999. Basement geology and tectonic development of the greater New Zealand region: an interpretation from regional magnetic data. *Tectonophysics* 308, 341-362.
- Symonds, P.A., Cameron, P.J., 1977. The structure and stratigraphy of the Carnarvon Terrace and Wallaby Plateau. *The APEA Journal* 17, 30-41.
- Symonds, P.A., Planke, S., Frey, O., Skogseid, J., 1998. Volcanic evolution of the Western Australian continental margin and its implications for basin development. *The Sedimentary Basins of Western Australia* 2, 33-54. Copyright © 2016 by Petroleum Exploration Society of Australia (PESA).
- Tallobre, C., Loncke, L., Bassetti, M.-A., Giresse, P., Bayon, G., Buscail, R., de Madron, X.D., Bourrin, F., Vanhaesebroucke, M., Sotin, C., 2016. Description of a contourite depositional system on the Demerara Plateau: Results from geophysical data and sediment cores. *Marine Geology* 378, 56-73.
- Thistlewood, L., Leat, P.T., Millar, I.L., Storey, B.C., Vaughan, A.P.M., 1997. Basement geology and Palaeozoic-Mesozoic mafic dykes from the Cape Meredith Complex, Falkland Islands: a record of repeated intracontinental extension. *Geological Magazine* 134, 355-367.
- Thran, A.C., Dutkiewicz, A., Spence, P., Müller, R.D., 2018. Controls on the global distribution of contourite drifts: Insights from an eddy-resolving ocean model. *Earth and Planetary Science Letters* 489, 228-240. <https://doi.org/10.1016/j.epsl.2018.02.044>
- Torsvik, T.H., Rousse, S., Labails, C., Smethurst, M.A., 2009. A new scheme for the opening of the South Atlantic Ocean and the dissection of an Aptian salt basin. *Geophysical Journal International* 177, 1315-1333.
- Tsikalas, Filippou, Eldholm, O., Faleide, J.I., 2005. Crustal structure of the Lofoten-Vesterålen continental margin, off Norway. *Tectonophysics* 404, 151-174.
- Tsikalas, F., Faleide, J.I., Eldholm, O., Wilson, J., 2005. Late Mesozoic-Cenozoic structural and stratigraphic correlations between the conjugate mid-Norway and NE Greenland continental margins, in: Geological Society, London, Petroleum Geology Conference Series. Geological Society of London, pp. 785-801.
- Uenzelmann-Neben, G., Weber, T., Grützner, J., Thomas, M., 2017. Transition from the Cretaceous ocean to Cenozoic circulation in the western South Atlantic—A twofold reconstruction. *Tectonophysics* 716, 225-240.
- Upton, B.G.J., Emeleus, C.H., Rex, D.C., Thirlwall, M.F., 1995. Early tertiary magmatism in NE Greenland. *Journal of the Geological Society* 152, 959-964.

- Veevers, J.J., Cotterill, D., 1978. Western margin of Australia: evolution of a rifted arch system. *Geological Society of America Bulletin* 89, 337-355.
- Von Rad, U., Haq, B.U., et al. (Eds.), 1992. Proceedings of the Ocean Drilling Program, 122 Scientific Results, Proceedings of the Ocean Drilling Program. Ocean Drilling Program.
- Welford, J.K., Shannon, P.M., O'Reilly, B.M., Hall, J., 2012. Comparison of lithosphere structure across the Orphan Basin-Flemish Cap and Irish Atlantic conjugate continental margins from constrained 3D gravity inversions. *Journal of the Geological Society* 169, 405-420.
- Wessel, P., Luis, J.F., 2017. The GMT/MATLAB Toolbox: THE GMT/MATLAB TOOLBOX. *Geochemistry, Geophysics, Geosystems* 18, 811-823.
- White, R.S., Smith, L.K., Roberts, A.W., Christie, P.A.F., Kusznir, N.J., 2008. Lower-crustal intrusion on the North Atlantic continental margin. *Nature* 452, 460.
- Zalan, P.V., n.d. An Entirely New 3D-View of the Crustal and Mantle Structure of a South Atlantic Passive Margin - Santos, Campos and Espírito Santo Basins, Brazil, #30177 (2011) 12.

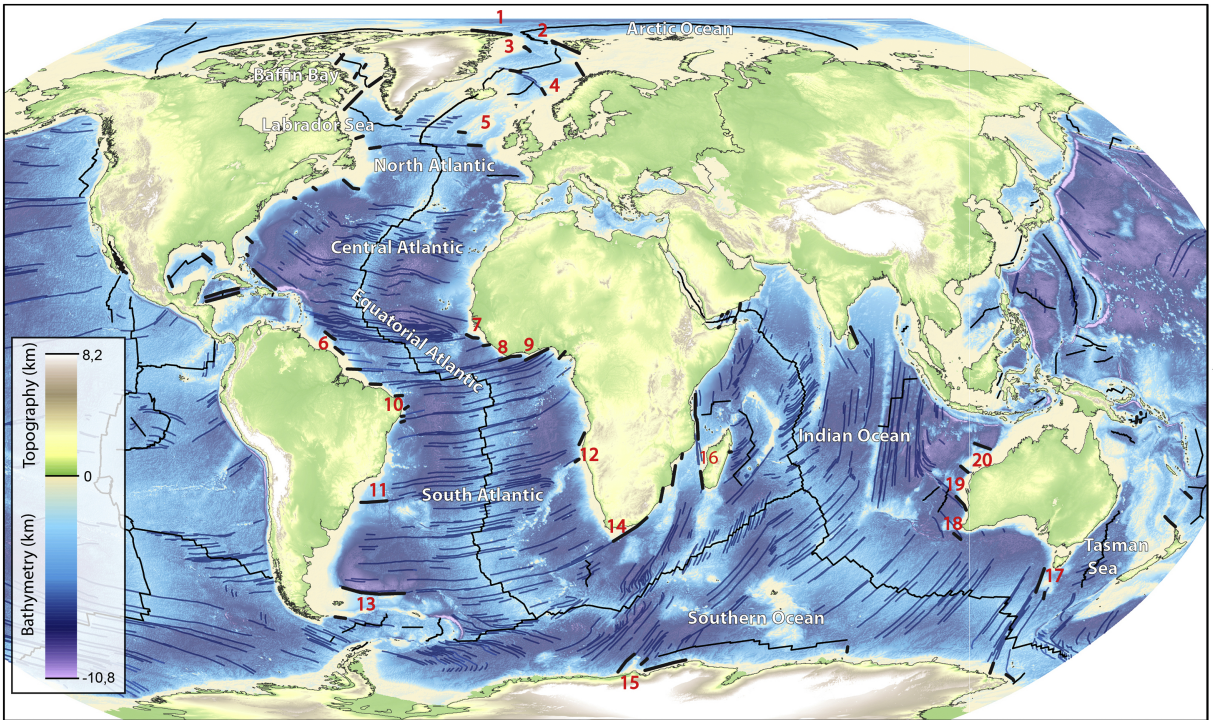


Figure 1



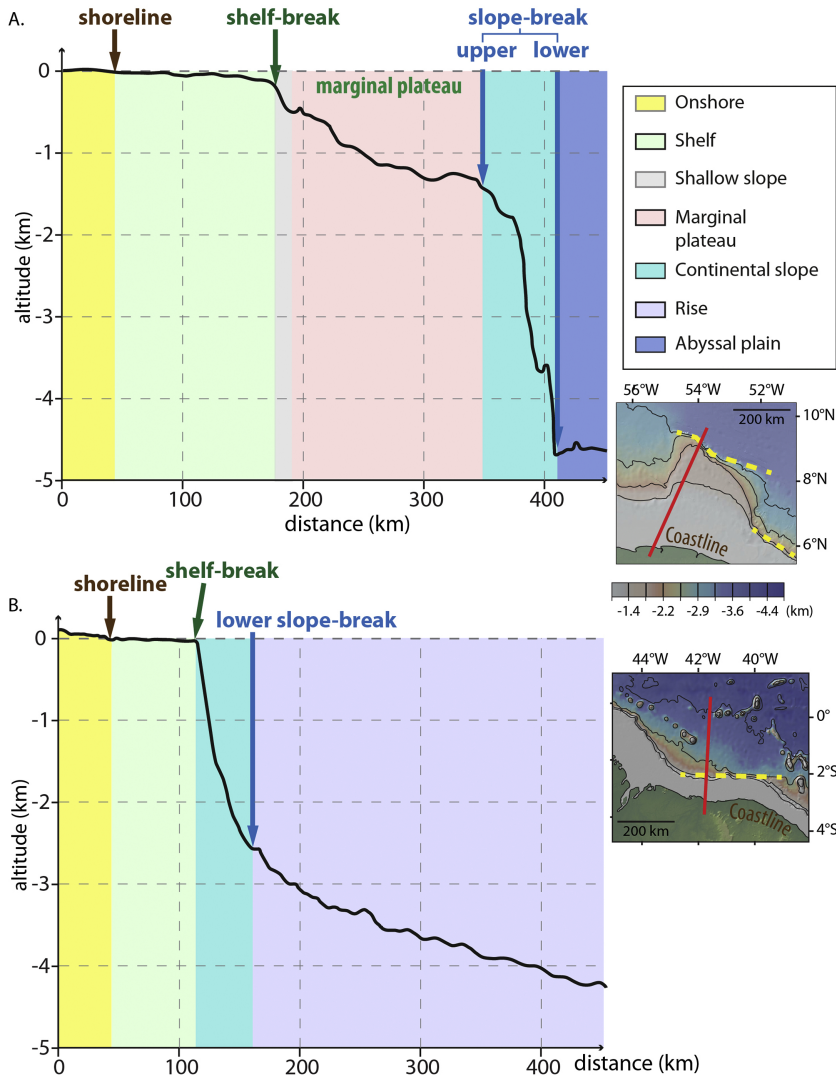


Figure 2

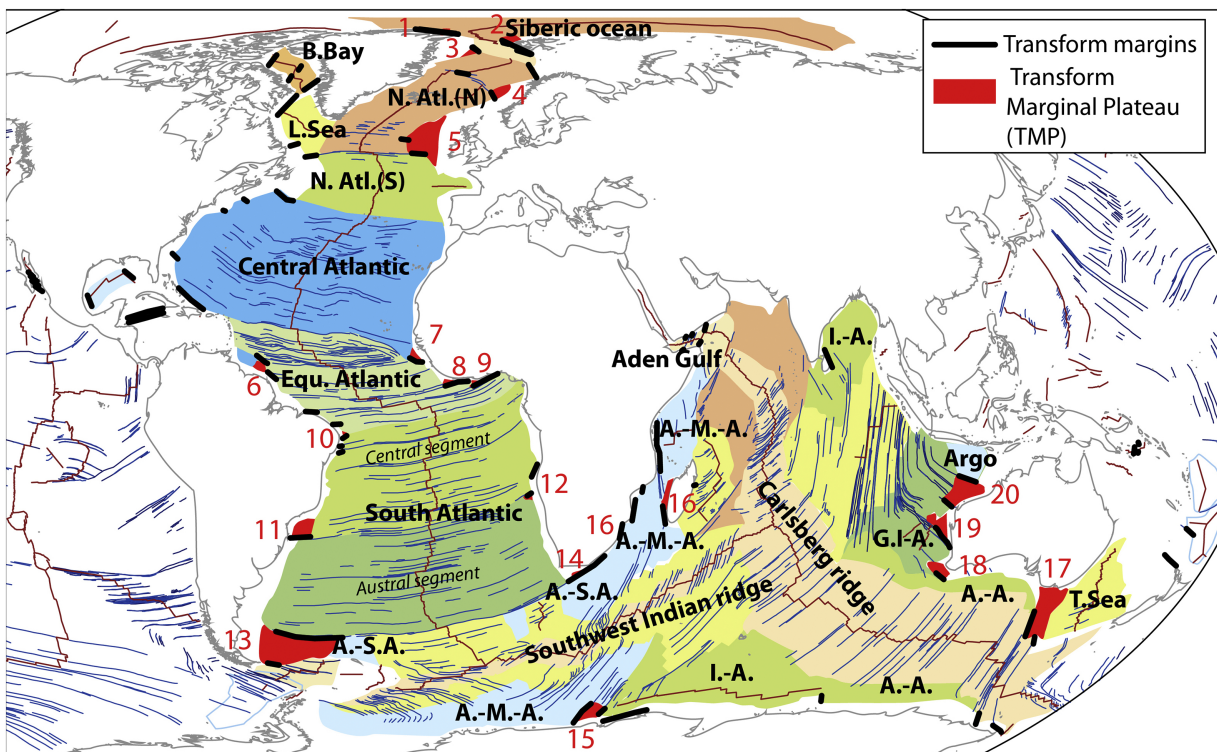


Figure 3

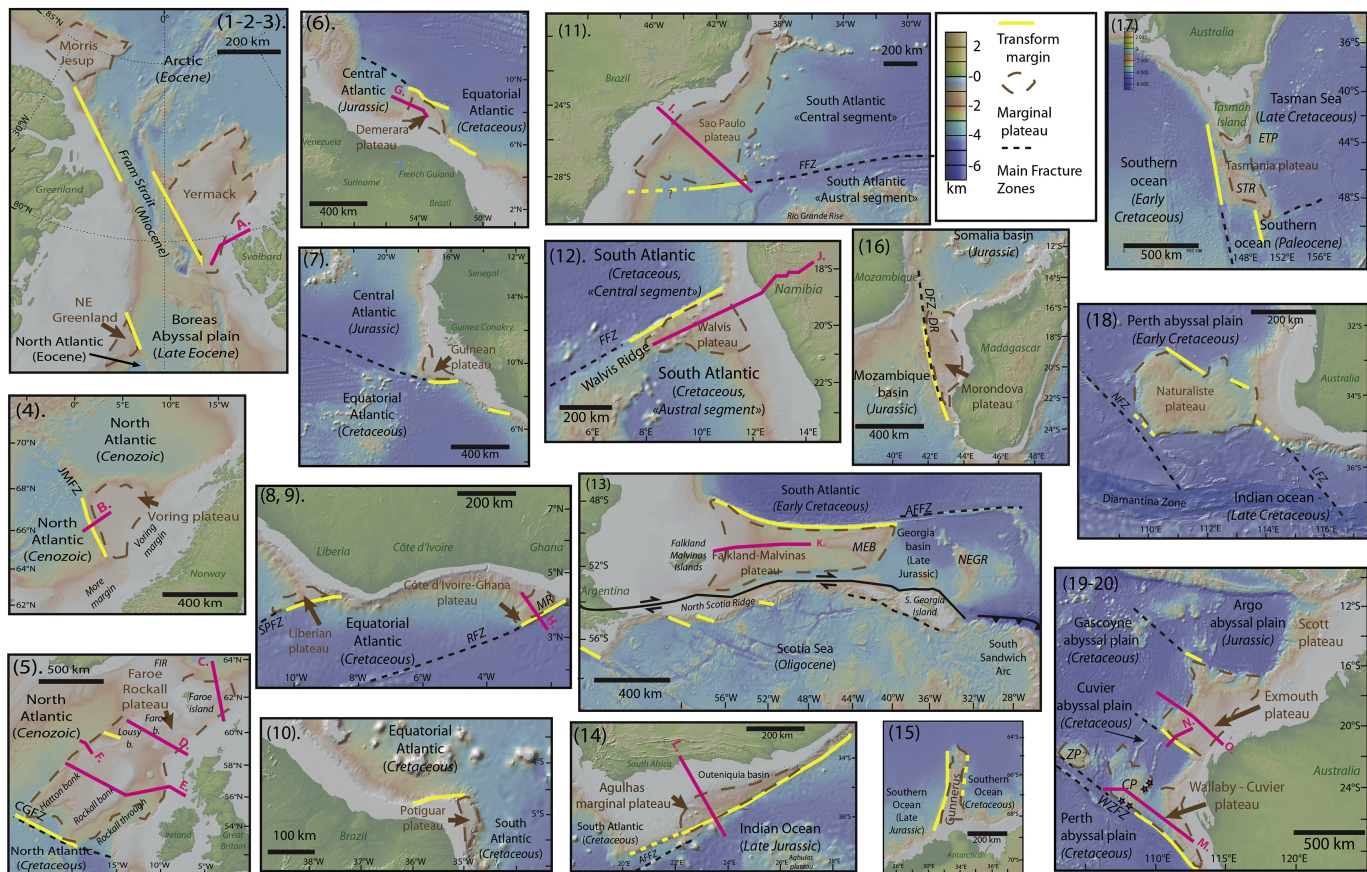


Figure 4

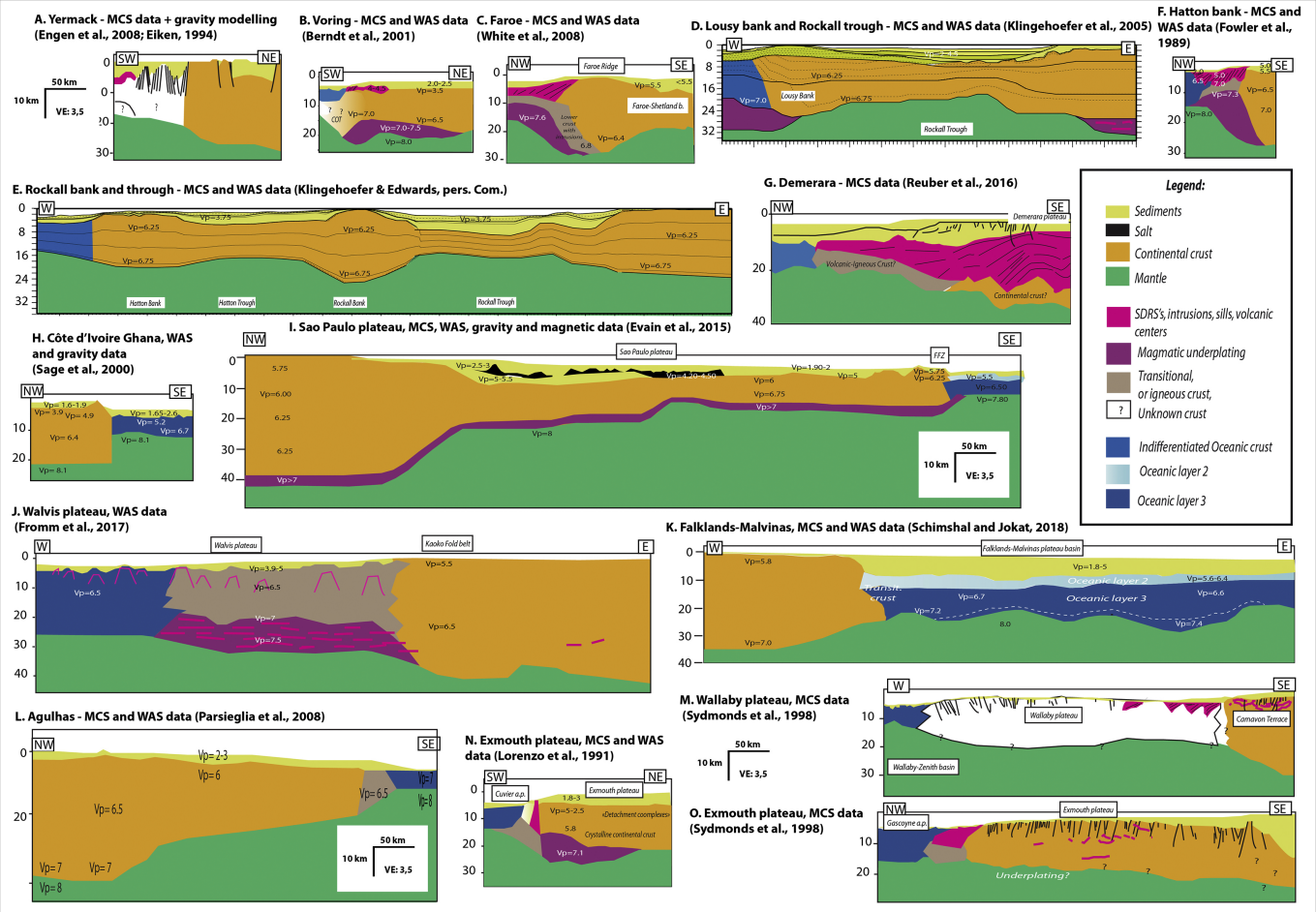


Figure 5

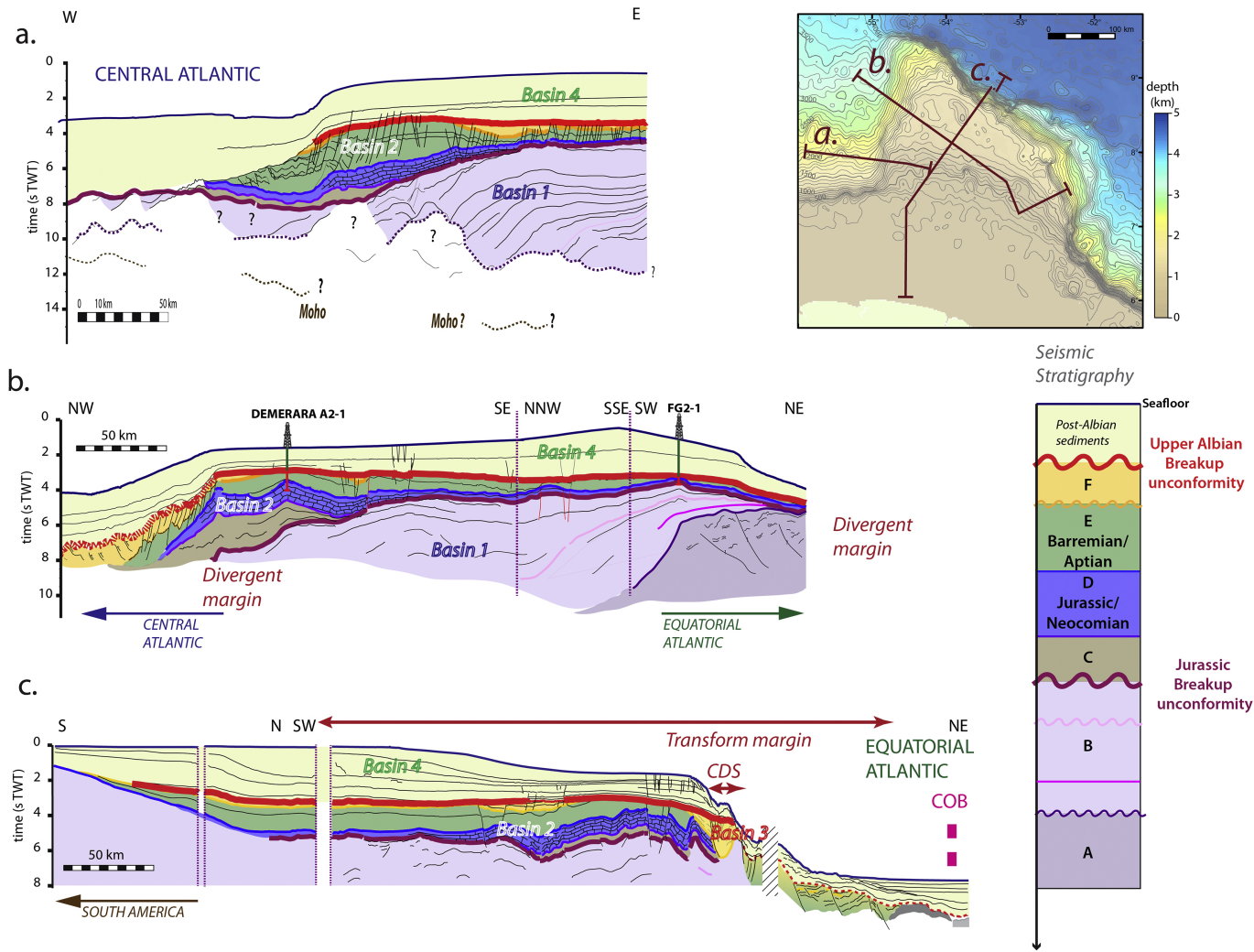
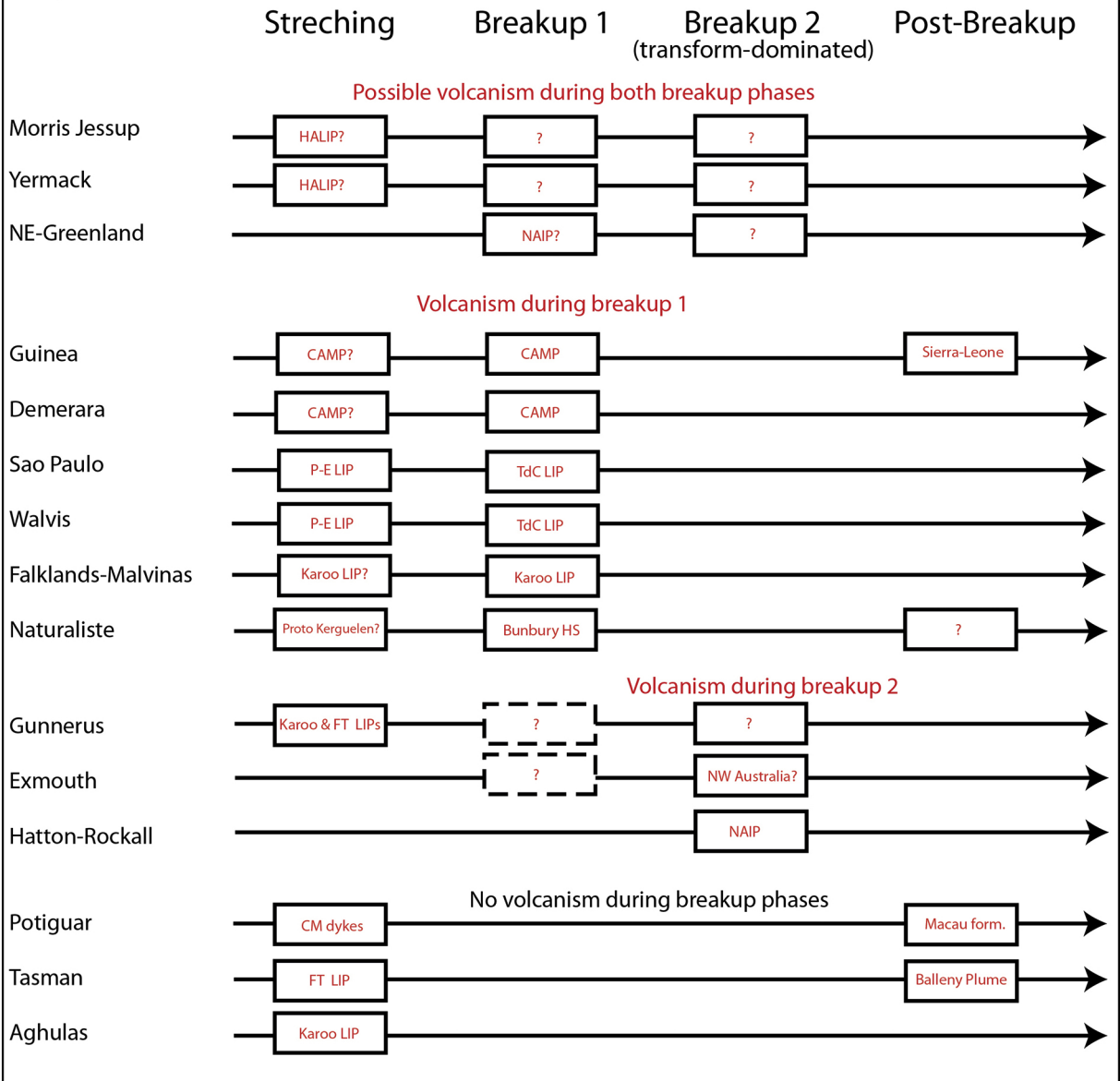


Figure 6

## Polyphased breakup



## Single breakup and segmentation phase

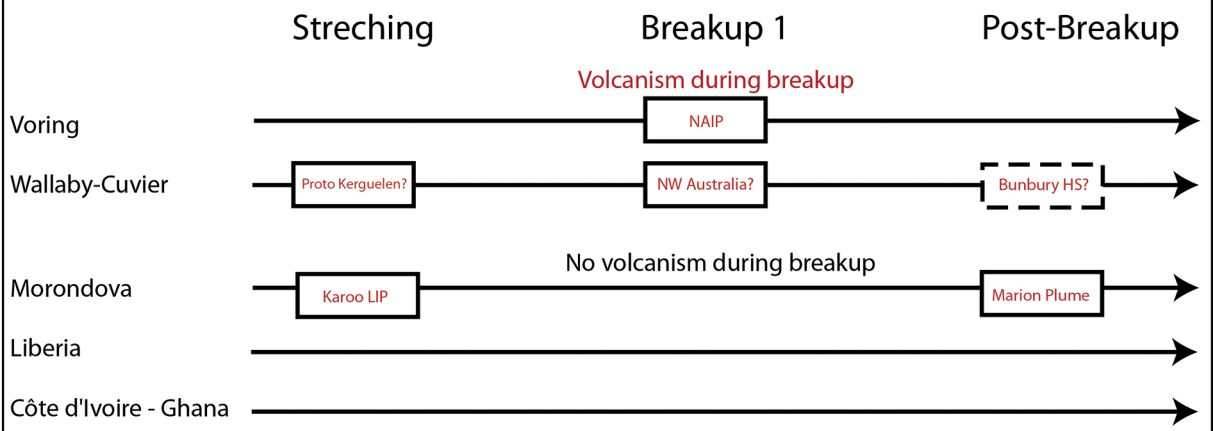


Figure 7

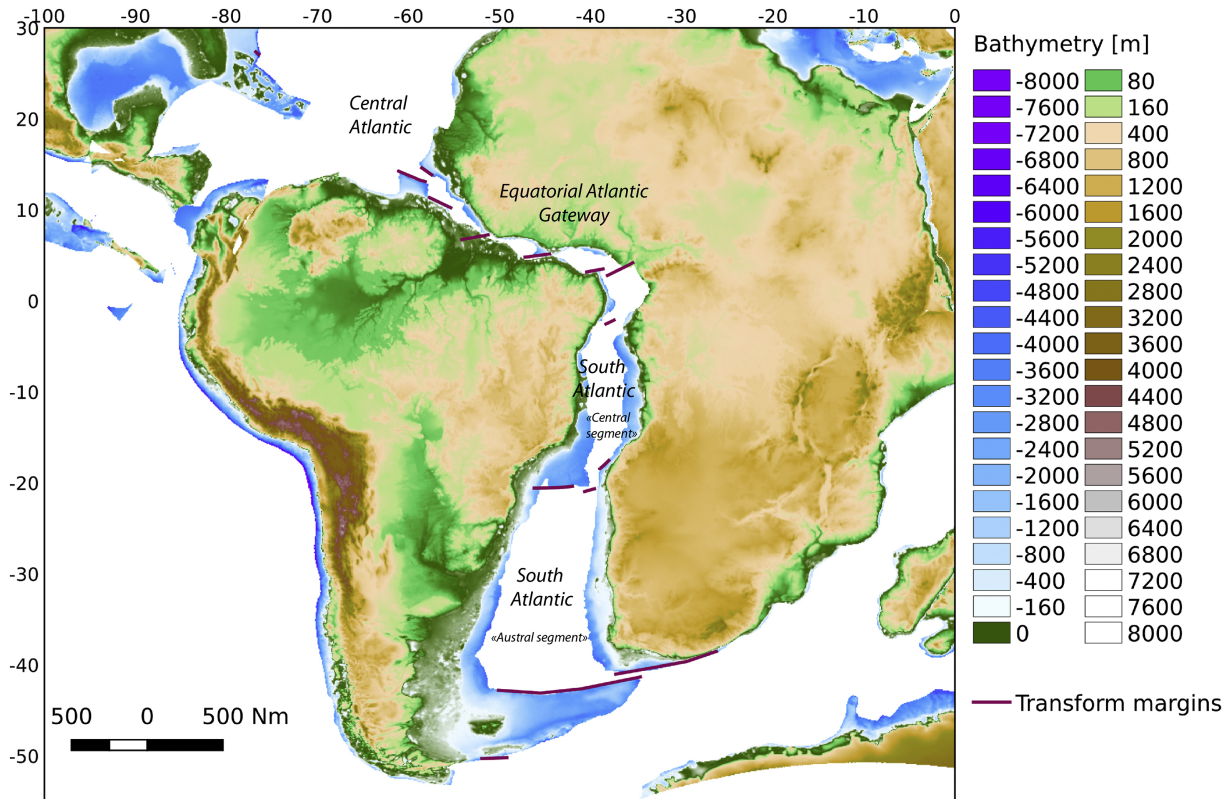
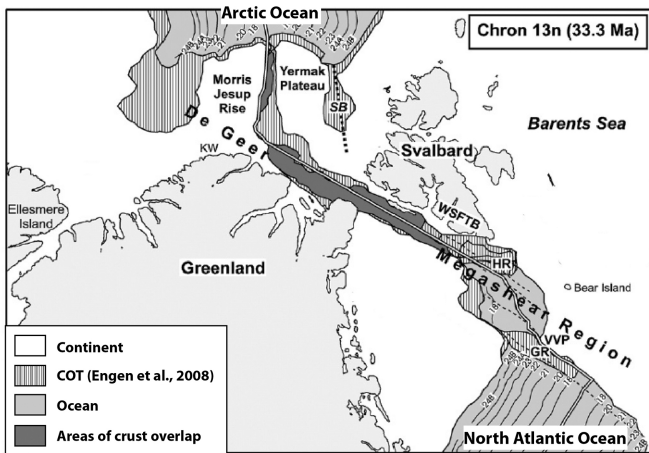


Figure 8

## (1) Arctic Oxygen-poor «lake stage»



## (2) Fully-ventilated stage - Open Fram Strait

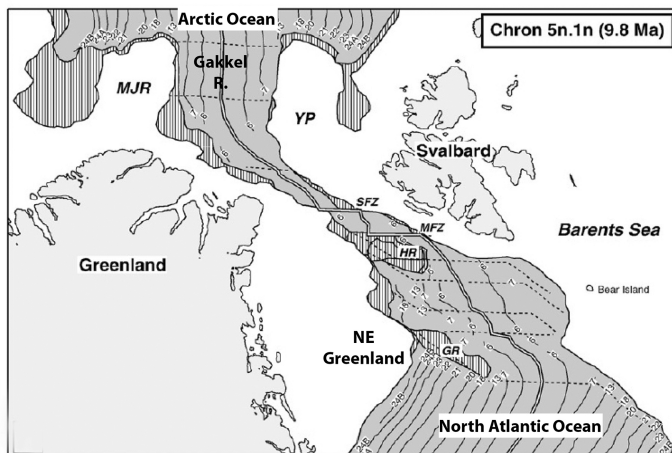


Figure 9