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Structure and evolution of the Atlantic passive margins: a review of existing rifting models from wide-angle seismic data and kinematic reconstruction

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Abstract :

Deep seismic data and plate kinematic reconstructions help understand the mechanisms of rifting and opening of new oceans, basic principles of plate tectonic cycles. In this study, available deep wide-angle seismic velocity models from the Atlantic margins are reviewed and plate reconstructions used to define conjugate model pairs. The main objective was to study the question of how magma-rich and magma-poor margins develop and the role of inheritance in the break-up. We also studied the question of the mechanism of formation and the origin of transform marginal plateaus, which are typically found at the border of two ocean basins of different ages and are mostly characterized by at least one volcanic phase during their formation. The results of the study include the comparison of crustal thickness, oceanic plate thickness and the influence of volcanism along the Atlantic margins. The conjugate profiles image different degrees of asymmetry of the Atlantic Margin rifts. Marginal plateaus might form when rifting stops at barriers leading to the accumulation of heat in the mantle and increased volcanism directly before or after halting of the rifting.

Highlights

► The comparison of published deep seismic velocity models, plate kinematic reconstructions and geological information allows to assess rifting styles and volcanic influence along the Atlantic Margins. ► Zones of upper mantle exhumed at the seafloor and serpentized by water intruding through cracks and faults have been discovered only at magma-poor margins. ► At magma-rich margins the volume of magma is largest close to transfer zones and decreases in the direction of rifting in each individual margin segment. ► Sheared margins are often associated with transform marginal plateaus, which have undergone at least one volcanic phase.

1. Introduction

1.1 Rifting and the break-up of continents

Rifting, the break-up of continents, and opening of an oceanic basin, are subjects of strong interest for Earth sciences (e.g. Franke et al., 2013; Geoffroy et al., 2015; Brune et al., 2017a). Passive continental margin basins around the world have good oil and gas exploration prospects. Hydrocarbons discovered in offshore passive continental margin basins especially in deep water have been increasing during the last 20 years (e.g. USA Gulf of Mexico, Angola, Brazil, Egypt, Nigeria, Guyana and Suriname). Additionally, divergent continental margins pose substantial geohazards. Submarine landslides possibly resulting in Tsunamis (e.g. Krastel et al., 2012) and massive CO₂ degassing during rifting have been proposed as elements contributing to change climate through time (Brune et al., 2017b). Continental flood basalts associated with rifting have been suggested to be contributors to mass extinctions due to the release of volatiles, aerosols and dust into the atmosphere (Courtillot and Renne, 2003). The focus of our study is the Atlantic Ocean which represents a natural laboratory for investigating and testing rifting models and the formation of divergent, oblique and transform continental margins.

Geodynamic models of rifting and the genesis of rifted margins have been developed alongside the investigation techniques. In the 1970s and 1980s, two types of models of passive continental margins were proposed: conservational model (McKenzie, 1978; Wernicke, 1985) where the masses are kept balanced and non-conservational models where masses are allowed to be exchanged between the crust and mantle (Keen and Beaumont, 1990). Among the first type, the pure shear model (McKenzie, 1978) suggests that extension results in the symmetric thinning of the continental crust leading to the formation of tilted blocks in the upper crust. Thinning is proposed as being directly proportional to horizontal dilatation (Figure 1). Simple shear models (Wernicke, 1985), on the other hand, do not assume symmetrical or uniform deformation and propose the existence of a lithospheric detachment fault crossing the crust and reaching into the upper mantle along which crustal thinning and exhumation take place leading to asymmetric margin formation (Figure 1). Non-conservational models, on the contrary, include an exchange of material between the mantle and the crust and propose additional factors, such as the cause of thinning and in which the mantle has a significant role in the process of passive margin genesis (Keen and Beaumont, 1990; Aslanian et al., 2009; Huisman and Beaumont, 2011 and references therein).

This simple view did change and most modern models of rifting and break-up include several stages comprising both pure shear and simple shear phases as well as interaction between the

crust and the mantle. Based on geophysical and geological observations from the west Iberian margin, and geological mapping in the Alps, Whitmarsh et al. (2001) present a conceptual model of magma-poor rifted margins including a first phase of lithospheric necking and diffuse faulting, followed by a second phase of localization of rifting and the onset of seafloor spreading. Later models tried to explain the apparent discrepancy between displacement along the faults and the amount of extension through exhumation of mid-crustal and mantle material (Lavie and Manatschal, 2006) or through a later second phase of deformation setting in after embrittlement of the crust (Reston, 2007). Today, a consensus exists that rifting is generally polyphased and results from crust-mantle interactions (Peron-Pinvidic et al., 2019). Depending on the rheology, rifting might accelerate abruptly at break-up due to the rapidly decreasing rift strength and slow down afterwards, when seafloor spreading sets in (Brune et al., 2016).

By regarding the amount of volcanism produced during rifting, two major types of divergent rifted margins can be identified: volcanic and magma-poor margins (Figure 1). Although all margins show at least some volcanic and magmatic influence, magma-poor margins show broad zones of continental thinning including faulting and the formation of tilted blocks (Manatschal, 2004; Reston, 2007; Peron-Pinvidic et al., 2019). Hyper-extended margins are termed those in which the crust becomes extremely stretched and embrittled, with fluids penetrating to the mantle through faults and serpentinizing the upper mantle material (Doré and Lundin, 2015; Tugend et al., 2015). For this type, conjugate margins are often asymmetric, with one margin significantly wider than its conjugate (Lister et al., 1991; Hopper et al., 2004). For these margins, the crustal break-up often precedes the lithospheric mantle break-up leading to the exhumation of upper mantle material (Manatschal, 2004). Rifted magma poor margins are frequently associated with slow extension (e.g. Bown and White, 1995; Perez-Gussinyé et al., 2006; Doré and Lundin, 2015).

Volcanic margins are characterized by large amounts of extrusive lavas forming thick seaward dipping reflector sequences and intrusives in the form of magmatically overprinted high-velocity zones in the lower crust (e. g. Thybo et al., 2013). In contrast with magma-poor margins, volcanic margins show crustal thinning in a narrow zone (~20 km) and the deposition of volcanic products and magmatic additions to the lower crust (Hinz et al., 1993; Eldholm and Grue, 1994; Geoffroy et al., 2015). Here, the mantle lithosphere breaks before or during crustal break-up allowing the production of large volumes of magmatic products (Franke, 2013). At the outer part of these margins, exhumed lower crust can be in direct contact with the extrusives (Geoffroy et al., 2015).

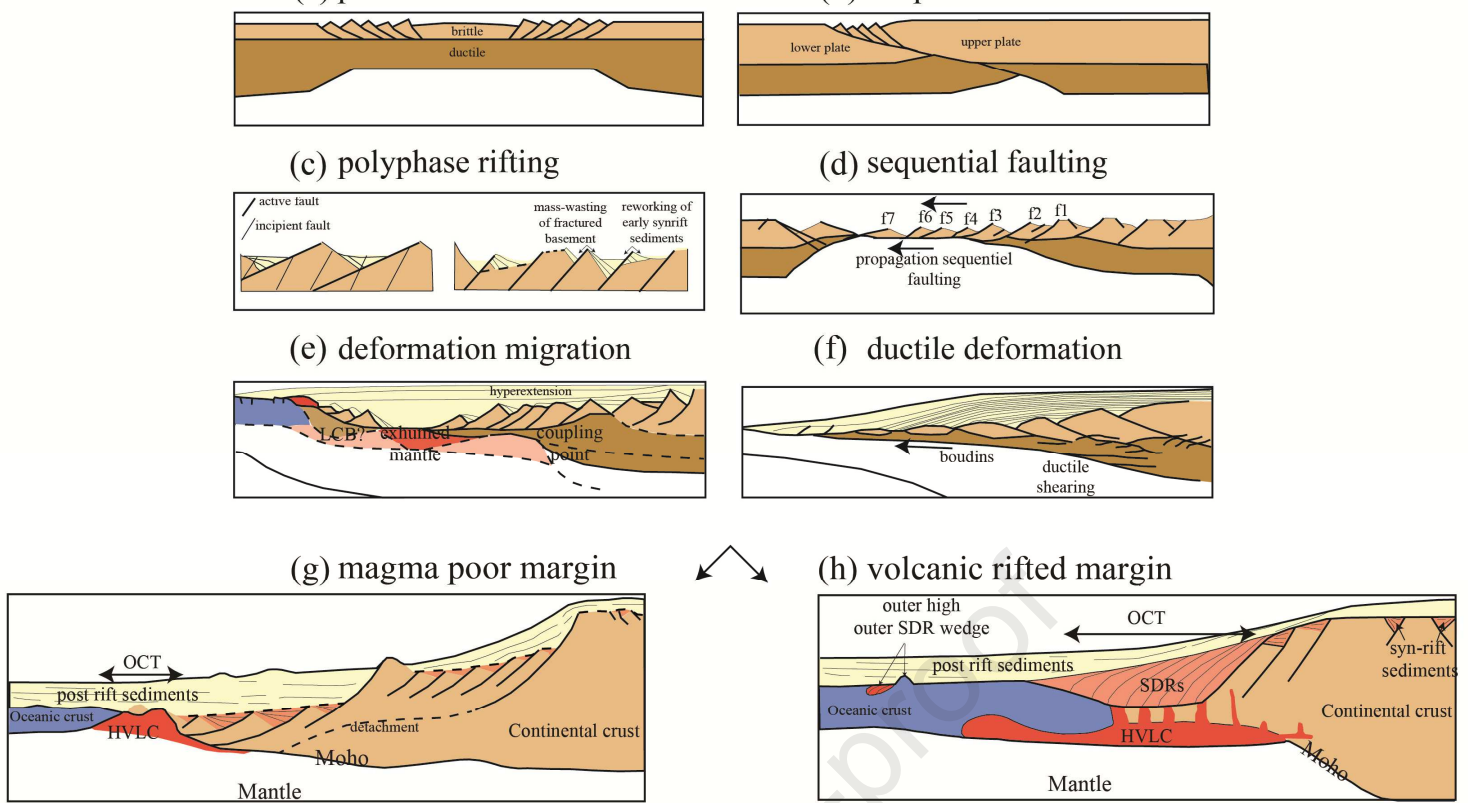


Figure 1: Schematic representation of the progress of knowledge from top to bottom :the early view of (a) Mackenzie, 1978 and (b) Wernick, 1985, new concepts of (c) Reston 2005, (d) Ranero and Perez-Gussinye 2010, (e) Peron-Pinvidic et al., 2013 and (f) Clerc et al., 2018). (g) and (h) shows schematic representation of the architecture of a slowly extended magma-poor margin and a rapidly extended magma-rich margin (after Franke., 2013; Peron-Pinvidic et al., 2013; Sutra et al., 2012; Doré & Lundin, 2015; Peron-Pinvidic and Manatschal, 2019)

A second group of continental margins that bounds parts of the Atlantic are transform margins, along which the movement of the plates is shear or oblique shear instead of extensional. Transform margins might be one of the main mechanisms for the break-up of continents as less force is required to break obliquely through cratonic crust (Lundin and Doré, 2019; Brune et al., 2012). The kinematic evolution of a transform margin is suggested to include three stages (Mascle and Blarez, 1987): 1) the intra-continental transform fault stage such as for example the San Andreas and Dead Sea faults; 2) the active transform margin stage, when continental lithosphere slides against oceanic lithosphere; and 3) the passive transform margin, after the spreading center has passed along the margin and the transform margin becomes inactive. The length of a transform margin depends on three parameters: the obliquity between regional trend and relative plate motion, the width of the continental rift and timing of the deformation partitioning (Basile, 2015). One characteristic feature of many transform margins is the existence of marginal ridges parallel to the transform margin (Basile et al., 1996). Several hypotheses have tried to explain the existence of these marginal ridges (a) crustal thickening from transpression in the intra-continental transform zone (e.g. Huguen et al., 2001); (b) thermal uplift due to heating of the continental lithosphere by the hotter oceanic lithosphere (e.g. Scrutton, 1979; Reid, 1989); and (c) flexural response of the lithosphere due to unloading by erosion (e.g. Basile and

Allemand, 2002). Modern numerical studies suggest that most margins do open in an oblique manner (e.g. Brune, 2014; Brune et al., 2018). Based on plate kinematic calculations, and the position of the continent-ocean transition, the authors propose that more than 70% of all rift segments exceed an obliquity of 20%.

However, some fundamental questions regarding rifting and opening of oceanic basins, remain highly debated. With this systematic review of deep seismic profiles along the Atlantic margins we aim to highlight the main open questions and put them into context. Three main questions that will be addressed in this contribution are.

1) What are the factors leading to the development of either volcanic or magma-poor margins and during which extension phase does this occur?

Volcanic margins have often been associated with Large Igneous Provinces and elevated potential mantle temperature (Smallwood and White, 2002). However, the thermal influence of a hotspot should be largest at its center, which is not the case at the US volcanic margins (Austin et al., 1990) and at the southern South Atlantic continental margins, where the opening was directed toward the proposed hotspot (Franke et al., 2013). It is frequently suggested that magma-poor margins result from slow extension, so it could be assumed that volcanic margins are associated with rapid extension. However, the South and Central Atlantic, which are both characterized by magma-rich zones, have been opening at ultra-slow to slow rates, and also, for example, the volcanic NE-Atlantic margin, which opened at a rate of about 25 mm/yr half spreading rate, subsequently rapidly decreasing to half-spreading rates of 10 mm/yr (Le Breton et al., 2012). There are other explanations for magmatic rifting, rather than purely the mantle temperature during extension. The amount of volcanism may also be controlled by other factors such as excess decompressional melting, tectono-magmatic interactions or pre-rift rheology of the lithosphere (e.g. Koopmann et al., 2014a; Davies et al., 2017; Tugend et al., 2018; Gouiza and Paton, 2019).

Recent work has proposed to generalize the model developed for magma-poor rifting to a general model valid for all types of Atlantic rifted margins (Mjelde et al., 2002; Peron-Pinvidic et al., 2013; Tugend et al., 2018), but questions as to the existence of serpentinized upper mantle material underneath the SDR sequences on volcanic margins remain highly debated. Lundin et al. (2018) have also shown that some if not all volcanic margins started to evolve from a typical non-volcanic margin, changing to a typical volcanic margin sometimes during the rifting period and that the influence of a plume or hotspot is not necessary to be invoked. If rifts are associated with LIP formation, the eruption happened close to a break-up and thus was preceded by rifting, indicating that rifting was initiated by tectonic forces (Buiter and Torsvik, 2014). Alternate models propose a significant difference between the development of volcanic and magma poor margins, where crustal thinning leads to the exhumation of serpentinite at magma-poor margins and lithospheric pure-shear leads to hot active melting at sub-Moho levels (Geoffroy et al., 2015; Zastrozhnov et al., 2020).

2) What is the influence of inheritance in margin break-up?

The role of tectonic inheritance in the development of rifts and break-up is still a highly debated subject (e. g. Piqué and Laville, 1996; Petersen and Schiffer, 2016). The present-day margins of the Central and North Atlantic resulted from at least two supercontinent cycles and subsequent break-ups including assembly of the supercontinent Rodinia, opening and closure of the Iapetus ocean, assembly and fragmentation of Gondwana and opening of the Atlantic (Wilson, 1966; Thomas, 2006). Rifts often develop along old suture zones (Buiter and Torsvik, 2014). Fossil subduction zones with remnant slabs from earlier oceans might survive rifting with parts ending up on both conjugate margins (Schiffer et al., 2015). Material from fossil serpentinized subduction zone mantle wedges might be present in some ocean-continent transition zones or at the base of the thinned continental crust. However, rifts also break through existing cratons, for example the Greenland craton.

Reuber and Mann (2019) found that where orogenic fabric trends parallel to the north–south South Atlantic rift direction narrow (55–90 km) rift zones with modest continental beta factors of 2.5–3.5 develop, while basement fabric trending near-orthogonally to the north–south South Atlantic rift direction results in wider (185–220 km) rift zones with higher beta factors of 4.3–5. These authors explain that greater amounts of stretching are needed to rupture the orthogonal, orogenic, basement fabric, while smaller amounts of rifting are needed to stretch the weaker and parallel, orogenic, basement fabric.

3) What are the origins and mechanisms of formation of Transform marginal plateaus (TMPs)?

Transform marginal plateaus (TMPs) are bathymetric highs at continental margins, situated at the limit between ocean basins of different ages. One side of the conjugate margin is associated with a transform, or, at least, it is a highly oblique margin (Loncke et al., 2020). The formation of TMP and its role in plate tectonics are still poorly understood. Examples in the Atlantic Ocean are the Falkland, Demerara and Guinea PLATEAUS (Schimschal and Jokat, 2018 and 2019; Mercier de Lepinay, 2016) and their conjugates the Guinea and Agulhas PLATEAUS (Parsiegla et al., 2008; Olyphant et al., 2017). 65–90% of TMPs have seen volcanic activity during their tectonic history and only 10% are underlain by continental crust; most of these have complex crusts overprinted by volcanism (Loncke et al., 2020). TMPs that remained in a high position during break-up due to volcanism might have formed the last land bridges between ocean basins and have therefore probably controlled the onset and variation of ocean currents (Friedrich and Erbacher, 2006). Oceanic transforms can act as a long-lived magmatic pathway for melts in the lithosphere (Gernigon et al., 2009) and might deform due to stresses induced by an increase in melt supply at the neighboring mid-ocean ridge (Maia et al., 2016). The systematic study of these edifices might answer questions about the link between transform margin initiation and volcanic activity (Burke and Dewey, 1973; Basile et al., 2020; Loncke et al., 2020).

1.2 Atlantic Ocean: Rifting, break-up and associated volcanism

The Atlantic has been an important region for plate kinematic studies because of its easily identifiable magnetic anomalies. The first kinematic reconstruction of the Atlantic was published in 1912 by Alfred Wegener, who used the shorelines and fauna on the African and South American continents to propose that Africa and South America were once joined together (Wegener, 1912). Later, Bullard et al., 1965 was the first to propose a kinematic reconstruction of the continents around the Atlantic using computational methods based on the hinge line position. During the 1970s and 1980s other elements were incorporated into the models and kinematic reconstructions, as for example seafloor magnetic anomalies, aspects of regional geology and intra-plate deformation (e.g. Le Pichon and Fox 1971; Pitman and Talwani, 1972; Olivet, 1978; Sibuet and Mascle., 1978; Wissmann and Roeser, 1982; Olivet, 1984). Modern reconstructions use continuously closing plate circuits and include features such as crustal thinning and mountain building (e.g. Matthews et al., 2016; Barnett-Moore et al., 2018; Gurnis et al., 2018; Müller et al., 2018) helping to calculate the position and extent of slab remnants in the mantle (e.g. van der Meer et al., 2018).

There is general agreement that rifting in the Atlantic started first in the Central Atlantic Ocean (CAO) (North America/NW-Africa) at about 195 Ma ago (Figure 2a; Table 1) with the break-up of the Gondwana continent in latest Triassic-earliest Jurassic (e. g. Olivet, 1978, Klitgord and Schouten, 1986; Sahabi et al., 2004). Obliquity of opening in the Central Atlantic was low (Brune et al., 2018). The next portion of the Atlantic Ocean that opened was the South Atlantic, starting first in the southern Austral segment (Argentina/Namibia-South Africa) during the Early Cretaceous with the break-up of western Gondwana (Figure 2b) around 140 Ma (e. g., Granot and Dymant, 2015). Extension then propagated to the north of the South Atlantic where break-up began in the late Early Cretaceous (112 Ma ago) between the Eastern Brazilian and Western African continental margins (Figure 2d). The southern North Atlantic opened in two phases (Late Triassic to Early Jurassic and Late Jurassic through Early Cretaceous) with a first phase of undistributed rifting and a second phase when rifting propagated from south to north with final break-up occurring in the Aptian-Albian around 133 Ma (Figure 2c) (e.g. Tucholke and Sibuet 2007). The Equatorial Atlantic started opening 118 Ma ago as a highly oblique margin (e.g. Burke and Dewey, 1973; Sibuet and Mascle, 1978; Unternehr et al., 1988). The full oceanic connection between the Central and North Atlantic occurred during Albian–Turonian times (Figure 2e) about 106 Ma ago (e.g. Pletsch et al., 2001). Finally, the north-eastern Atlantic started opening between East Greenland and NW Europe at 55.5 Ma (Storey et al., 2007) (Figure 2f).

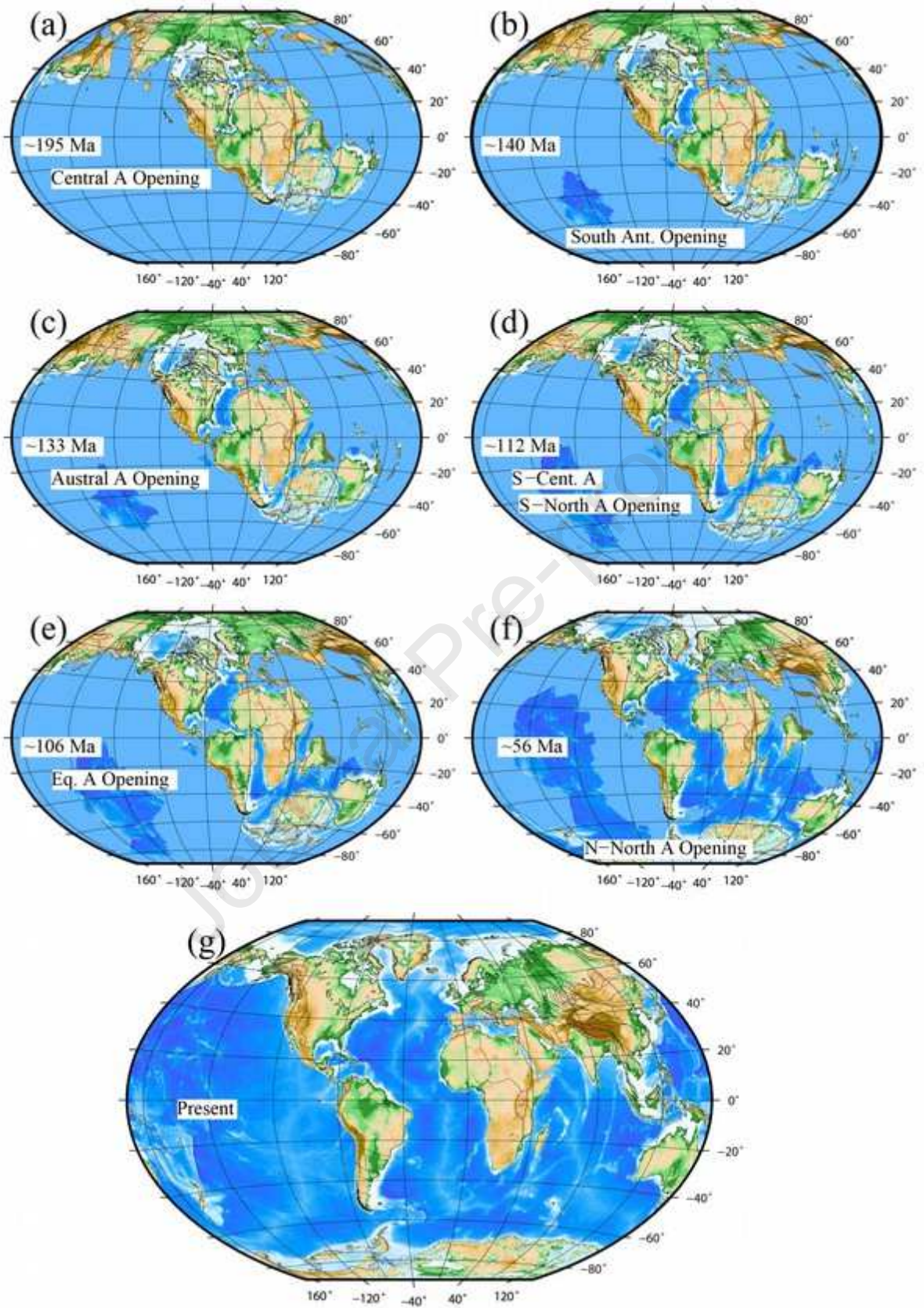
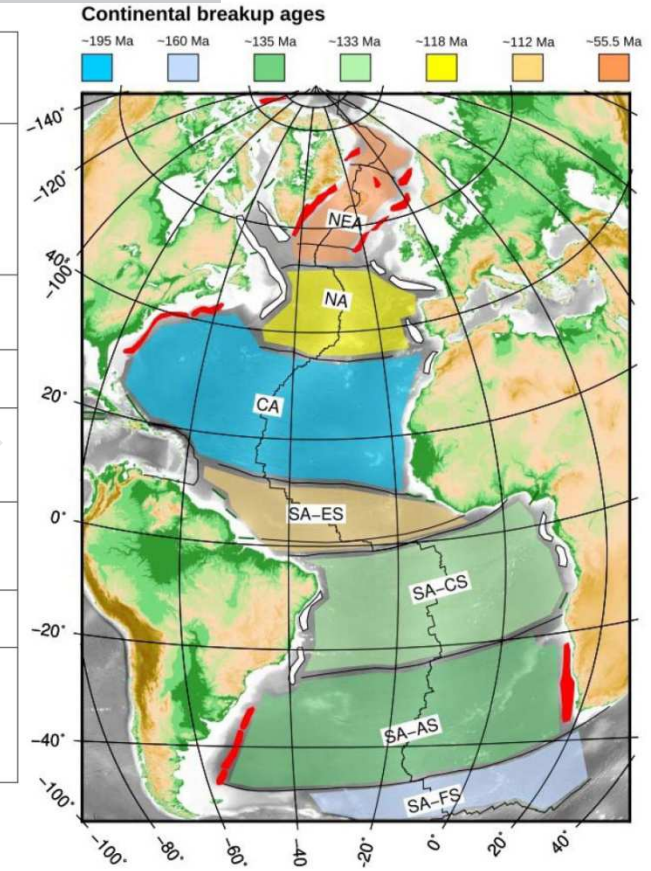


Figure 2: (a) - (f) Reconstruction of the Central Atlantic with rotational poles from Matthews et al., 2016. The fixed plate is Africa. (g) Topography and bathymetry map of the Atlantic region.

Geographic domain	Age of opening	Full spreading rate at opening	Margin type	Salt basin	Rifting obliquity ⁽¹⁴⁾	Transform marginal Plateau ^{(15),(16)}	Associated magmatic province
Central Atlantic	~195 Ma ⁽¹⁾ ~175 Ma ⁽²⁾	0.8 cm/yr ⁽⁹⁾	North: Magma-poor South: Heavily magmatic (SDR, Underplate)	Am. : Sable, Shelburne, and Georges Bank basins Af. : Seine, Meseta, Mauritania, Casamance, Caroline basins	Divergent (0-20°)	-	CAMP
South Atlantic/Falkland segment	~160 Ma ⁽³⁾	1-2 cm/yr ⁽¹⁰⁾	Magmatic	Danmarks Havn Basin (NE Greenland)	Transform or highly oblique (60-90°)	Falkland-Malvinas Agulhas	KLIP
South Atlantic/Austral segment	~135 Ma ⁽⁴⁾	7 cm/yr ⁽¹⁰⁾	High magmatism (SDR, Underplate)	-	Divergent (0-20°)	Sao Paulo, Walvis	Paraná-Etendeka LIP
South Atlantic/Central segment	~118 Ma ⁽⁷⁾	3.8 cm/yr ⁽¹¹⁾ 4.7 cm/yr ⁽¹²⁾	Moderate magmatism.	Am. : Esperito Santo, Campos, Santos basins Af.: Gabon, Kwanza, Benguela basins	Divergent (0-20°)	-	CAMP
South Atlantic/Equatorial segment	~112 Ma ⁽⁵⁾	1 cm/yr ⁽¹⁰⁾	Magmatic	-	Transform or highly oblique (60-90°)	Demerara, Guinea, Liberia, Côte d'Ivoire-Ghana	CAMP
Southern North Atlantic	~133 Ma ⁽⁶⁾	1 cm/yr ⁽¹⁰⁾	Magma-poor	-	Divergent (0-30°)	-	NAIP
Northeast Atlantic	~55.5 Ma ⁽⁸⁾	1.5-2.5 cm/yr ⁽¹³⁾	Strong magmatism (SDR, Underplate)	- Paleozoic salt basin- Danmarkshavn Basin (NE Greenland)	Oblique (40-55°)	Morris Jesup Rise Yermack Plateau, NE Greenland, Vøring, Faroe-Rockall	NAIP



- (1) [Sahabi et al., 2004]
 (2) [Klitgord and Schouten, 1986]
 (3) [Tarling, 1972]
 (4) [Granot and Dymant, 2015]
 (5) [Rabinowitz and Labrecque, 1979]
 (6) [Tucholke and Sibuet, 2007]
 (7) [Burke and Dewey, 1973]
 (8) [Storey et al., 2007]

- (9) [Labails et al., 2009]
 (10) [Müller et al., 2008]
 (11) [Moulin et al., 2010]
 (12) [Torsvik et al., 2009]
 (13) [Gaina et al., 2009]
 (14) [Brune et al., 2018]
 (15) Mercier de Lepinay et al., 2016
 (16) Loncke et al., 2020

Table 1: Summary of opening phases and their characteristics of Atlantic basins and tectonic setting of the continental margins with location map. Red and white polygons are volcanic margin parts and salt basins from Lundin et al., 2018. Green polygons are TMPs as defined in Mercier de Lépinay et al., 2016 and Loncke et al., 2020.

1.3 Observations constraining rifting and opening in the marine environment

The first observations used for the reconstruction of the opening of the Atlantic Ocean were bathymetric data (Figure 3a), allowing to constrain the basic structures of plate tectonics, such as mid-ocean ridges, transform faults and hinge lines along passive margins (e. g. Wegener, 1912; Bullard et al., 1965). Second, magnetic anomalies were used for dating the oceanic blocks, to identify the continent-ocean transition zone, the onset of oceanic crustal accretion, as well as the velocity of seafloor spreading (Sahabi et al., 2004; Collier et al., 2017) (Figure 3b). Over time, marine magnetic anomalies, combined with other geological information (fracture zones, marginal salt basins), provided the framework for the kinematic reconstructions of the break-up, the genesis of the Atlantic rifted margins and the geometry of seafloor spreading (Collette et al., 1984; Olivet et al., 1984; Klitgord and Schouten, 1986; Sahabi et al., 2004; Eagles, 2007; Labails et al., 2010; Moulin et al., 2010, Perez-Diaz and Eagles, 2014).

Volcanism during rifting and break-up leads to the emplacement of seaward-dipping reflector sequences (SDR) comprising up to 20km-thick layers of mainly volcanic material which often extend up to hundreds of kilometers. It typically also leads to the emplacement of high-velocity material underneath the crustal layer, as has been imaged along the US continental margin (Lase study group, 1986) or East Greenland and Norway (Mjelde et al., 1998). The underplate layer typically shows up as a high-velocity lower crustal layer (>7.2 km/s). The continental margin of the Hatton Bank in the Northeast Atlantic is the most well-known example where this layer is thought to be caused by magmatism above a mantle thermal anomaly during continental break-up (White et al., 1987). Several studies in volcanic margins, then, interpret this high-velocity lower crustal layer as a magmatic underplate, produced by excess mantle melting (Barton and White, 1997, Klingelhoefer et al., 2005). For other authors, however, the high-velocity lower crustal layer represents continental crust intruded by sills (White et al., 2008). A study in the Austral South Atlantic suggests that it can be both, heavily intruded lower crust during the early evolution of the volcanic rifted margin, and underplated bodies thereafter, during maturation of volcanic rifting after break-up (Becker et al., 2014).

These volcanic structures are one of the most characteristic features of volcanic passive margins (Planke et al., 2000; Geoffroy, 2005) and are emplaced during break-up and the initial stage of seafloor spreading (e.g Austin and Uchupi, 1982; Mutter et al., 1982; Mutter, 1985; Planke and Eldholm, 1994; Gladczenko et al., 1998, Geoffroy, 2005).

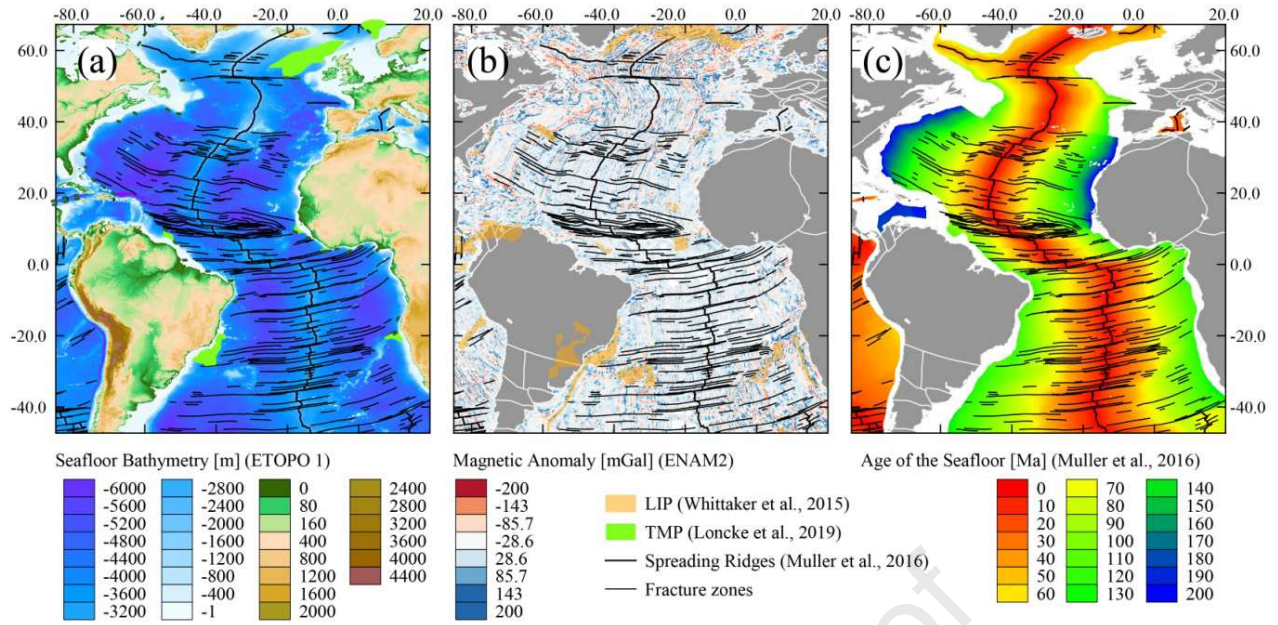


Figure 3: (a) Seafloor bathymetry from the ETOPO1 dataset (b) Magnetic anomalies of the seafloor from the EMAG2 grid ([doi:10.7289/V5H70CVX](https://doi.org/10.7289/V5H70CVX)). (c) Age of the seafloor (Müller et al., 2016).

1.4 Methodology of this study - wide-angle seismic data help constraining rifting history

Deep sounding seismic reflection and wide-angle data provide information about the deep crustal structure of passive margins and thereby help to answer questions about rifting and the onset of seafloor spreading. Using these data, continental crust, transitional zones and oceanic domains can be better separated and characterized. The structures revealed by the seismic data are a major input for plate kinematic reconstructions and help to further describe the evolution of passive margins. In this study, published seismic reflection and refraction data along the Atlantic continental margins are reviewed. We compare profiles along each margin segment with profiles along their conjugate, in order to estimate the degree of symmetry in the original break-up as well as the influence of volcanism during break-up on the formation of a single margin or conjugate margin pair. By doing so, the geometry of the crustal structure and the degree of crustal thinning of all conjugate profiles can be compared and a model of the opening mechanism proposed. However, uncertainties in the identification of magnetic lineations as seafloor spreading features, especially in the oldest parts of oceanic basins where they could belong to continent-ocean transition (COT) zones and formed by serpentinization processes (west Iberia, Sibuet et al, 2007), uncertainties in paleomagnetic data of Iberia during rifting and early stages of oceanic accretion (Vissers et al., 2016), which were later contested (Barnett-Moore et al., 2016), uncertainties in the identification and age of the oldest oceanic lineations (e.g. Eagles et al., 2015; Causer et al., 2020 and references therein) raise new problems concerning the relationship between rifting and oceanic accretion, which can overlap through time and the nature, extent and role of COTs.

To identify conjugate margin pairs of seismicrefractionlines, we developed a set of plate kinematic reconstructions of the South, Central and North Atlantic using the GPlates Software (Müller et al., 2018) and including the position of published wide-angle seismic

profiles. Reconstructions at complete closure as well as reconstructions at early opening stages of oceanic basins (Figure 2) were made using the rotational poles from Matthews et al. (2016). However, for most regions more detailed studies exist (Eagles and Koenig, 2008; Moulin et al., 2010; Heine et al., 2013; Barnett-Moore et al., 2018; Thompson et al., 2019) and they are appropriately cited in the text. The location of fracture zones (Matthews, 2015), magmatic provinces, hotspots (Whittaker et al., 2015), and mid-oceanic ridges (Müller et al., 2016) were used to define segment boundaries and those parts of the margins influenced by volcanism. Existing wide-angle seismic profiles were subsequently digitized and converted to a common color scheme to allow direct comparison of deep seismic profiles. However, the initial interpretations of these profiles remained unchanged as they often represent a more complete work than just modeling seismic velocities, including interpretation of additional data, such as gravity modeling, seismicreflectioninterpretations or geological constraints from the field.

One of the main difficulties of this study is the difference in the quality of the seismic data. While older profiles were often acquired with instruments sparsely deployed at 10 km intervals, modern studies include dense receivers spacings of only a few kilometres. While the error of the Moho depth is typically less than a kilometer in modern studies (Biari et al., 2015), it might reach 2-3 kilometer in older work (LASE Study Group, 1986). However, all profiles selected provide sufficient detail to characterize the nature of the crust and the geometry of the crustal layers in the scope of this manuscript. A third point is the possible obliquity of rifting, which might lead to structural features imaged in an oblique angle by perpendicular profiles. The plate reconstructions themselves assume rigid plates, which probably was not the case during the opening of most of the Atlantic Margins. It is difficult to estimate the error inherent in plate kinematic reconstructions, but using different rotational poles published in the literature indicates differences of less than 15-20 km. Again, this is sufficient for the analysis and comparison presented in this study. For consistency we used the rotational poles of Matthews et al. (2016) throughout this work. To have exact details on error calculations and additional constraints the reader is referred to the original publications.

2 Central Atlantic

The Central Atlantic is defined in this study within the bounds of the Pico and Gloria fracture zones to the north and the 15°20' N, and Guinea fracture zones to the south (Figure 4). Rifting in the Atlantic started first in the central Atlantic Ocean (CAO) which encompasses the area between the conjugate northeast America (Nova Scotia and US Atlantic margin) and the northwest Africa margins (from Morocco in the north to the Guinea fracture zone in the south) (Figure 4). Rifting occurred in a single phase with the break-up of the supercontinent Gondwana in the latest Triassic-earliest Jurassic (Le Roy and Piqué, 2001; Michard et al., 2008; Frizon de Lamotte et al., 2015) (Figure 2). Sea floor spreading and the formation of the first oceanic crust began at about 175 Ma during the Middle Jurassic (Klitgord and Schouten, 1986) or around 195 Ma during Sinemurian times (Sahabi et al., 2004).

The opening of the Atlantic Ocean was associated with the establishment of a large igneous province known as the Central Atlantic Magmatic Province (CAMP). It is defined by tholeiitic basalts outcropping in South and North America, Africa, and Europe (Marzoli et al., 1999 and 2004, McHone et al., 2000 and 2003) and dated between 197 and 203 Ma (Hames et al., 2000; Knight et al., 2004; Marzoli et al., 2004). The emplacement of the CAMP volcanic products was associated with the Triassic–Jurassic extinction event, which is considered as one of Earth's major extinction events (Courillot et al., 1996; Blackburn et al., 2013).

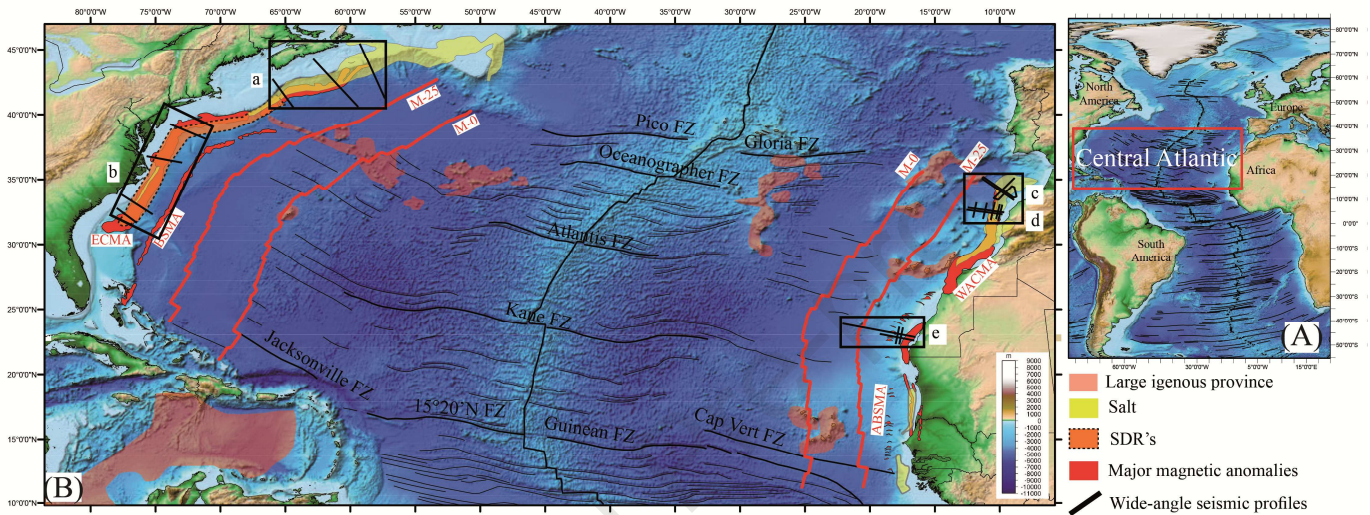


Figure 4: A) Topographic and bathymetric map of the Atlantic Ocean from ETOPO1 bathymetry grids (Amante and Eakins, 2009). B) Overview of the central Atlantic Ocean located between the Pico and Gloria fracture zones to the north and the 15°20'N and Guinea fracture zones to the south (thick black lines), bathymetry and topography from ETOPO1 bathymetry grids (Amante and Eakins, 2009). The position of the salt (light yellow) is taken from Sahabi et al. (2004), Davison (2005), Tari et al. (2012). Major magnetic anomalies are from Sahabi et al. (2004) and Labails et al. (2010). Offshore CAMP volcanism is from Coffin and Eldholm (1994) and Ernst (2014). Going from north to south. Black lines along the North American margin represent wide-angle seismic profiles of the SMART-1, -2, and -3 (a) (Funck et al., 2004; Wu et al., 2006; Loudon et al., 2010), LASE, EDGE-801, USGS-32, and BAa-6 (b) (LASE Study Group, 1986; Tréhu et al., 1989; Austin et al., 1990; Holbrook et al., 1994a, 1994b; Lizarralde et al., 1994; Lizarralde and Holbrook, 1997). Black lines along the North African margin represent wide-angle seismic profiles of the Sismar (c), Mirror (d) and Dakhla (e) cruises (Contrucci et al., 2004b; Klingelhoefer et al., 2009; Biari et al., 2015).

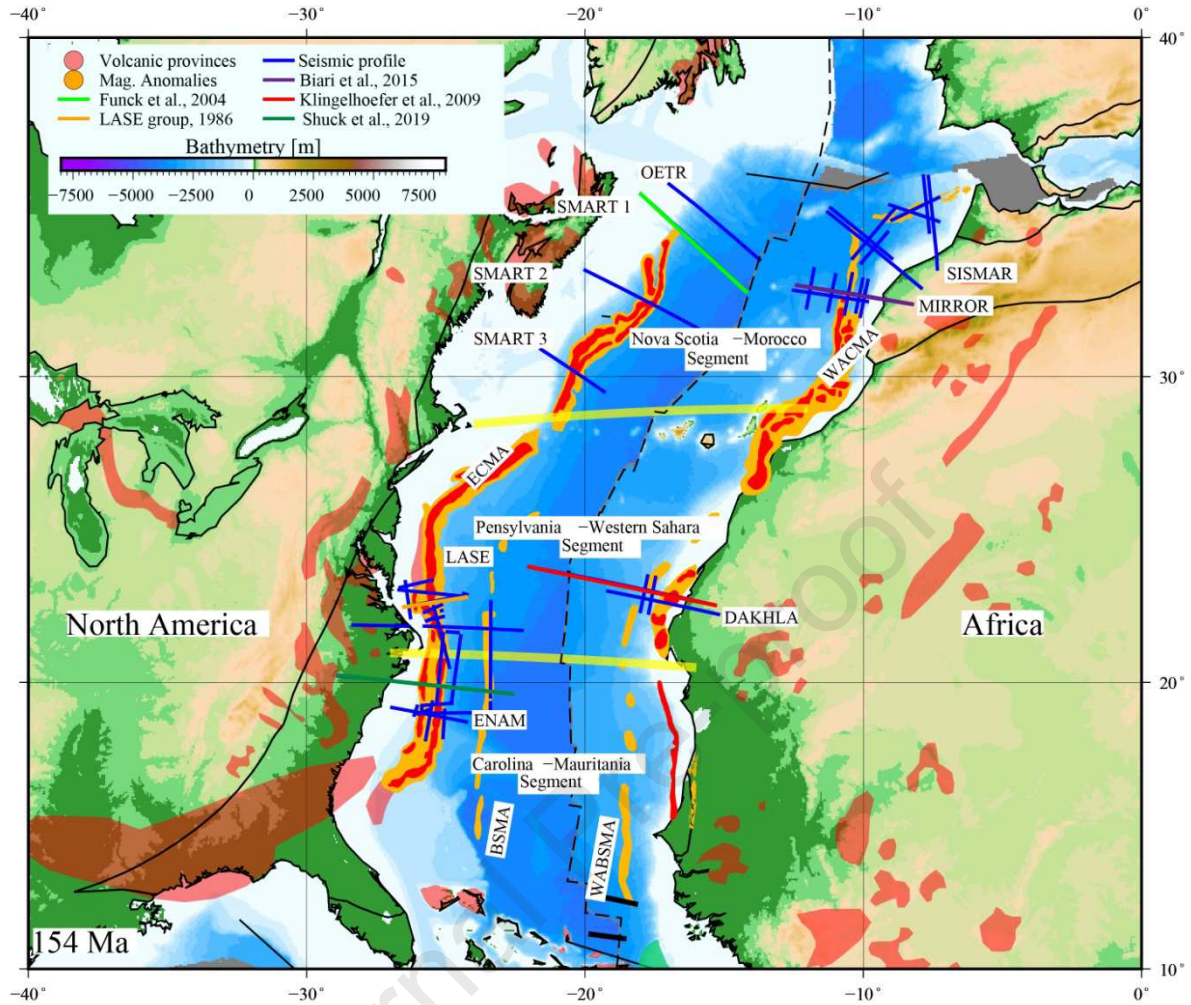


Figure 5: Reconstruction of the Central Atlantic at 154 Ma (Chron M25) with rotational poles from Matthews et al., 2016. The fixed plate is Africa. Locations of wide-angle seismic profiles are marked by solid lines. Orange transparent polygons represent the location of volcanic provinces. Yellow lines depict segment boundaries.

The northeast American and northwest African margins are characterized by the presence of four magnetic anomalies named East Coast Magnetic Anomaly (ECMA) and its conjugate West African Coast Magnetic Anomaly (WACMA), the Blake Spur Magnetic Anomaly (BSMA) and its possible conjugate the African Blake Spur Magnetic Anomaly (ABSMA) (Figure 5) (for the East Canadian margin see Drake et al., 1959; Dehler, 2011; Funck et al., 2004; for the African margin see Hinz et al., 1982; Roeser, 1982; Roeser et al., 2002; Sahabi, 2004). ECMA and WACMA are thought to mark the end of rifting and onset of oceanization (Hinz et al., 1982; Roeser, 1982; Roeser et al., 2002; Sahabi, 2004; Biari et al., 2017), however the origin of the anomalies is still debated and they may originate from seaward-dipping reflector sequences or magmatic underplating (Holbrook et al., 1994a; Keen and Potter 1995 a,b). The BSMA was dated by drilling at DSDP site 534 and proposed to be Callovian (165 Ma) (Sheridan et al., 1982; Gradstein et al., 2004).

Along the eastern United States margin, thick layers of SDR and magmatic underplate were imaged (LASE Study Group, 1986; Tréhu et al., 1989; Austin et al., 1990; Holbrook et al., 1994a, 1994b). Volcanism decreases to the north towards the Nova Scotian margin (e. g. Funck et al. 2004, Loudon et al., 2010). Along the conjugate African margin, only few volcanic structures interpreted as SDR and some seamounts were imaged (Roeser et al., 2002), so the northwestern African margin was defined as magma poor (Contrucci et al., 2004b; Labails and Olivet, 2009; Klingelhoefer et al., 2016; Benabdellouahed et al., 2017). During rifting four major conjugate salt basins were formed along the Central Atlantic margins: the Moroccan Atlantic salt basin and its conjugate salt basin off Nova Scotia in the northern Central Atlantic margin and the Mauritanian salt basin conjugate to the salt basin of the Carolina Trough, offshore the US margin in the southern Central Atlantic.

Klitgord and Schouten (1986) proposed an evolution model for the Central Atlantic considering the ECMA and WACMA as conjugates, including an eastward ridge jump south of the Canary Islands at the time of BSMA and an onset of seafloor spreading at 175 Ma. However, other authors suggest that it corresponds to a boundary between oceanic crust with a "rough" morphology in the west and a domain with a smoother morphology in the east related to an acceleration of seafloor spreading (Markl and Bryan, 1983).

Based on the identification of magnetic anomalies, salt basin location and the orientation of fracture zones, the models of Olivet et al. (1984), Sahabi et al. (2004) and Labails et al., (2010) describe the evolution of the North and Central Atlantic from the early stages of oceanic opening to the present. In their reconstruction both the ECMA and WACMA coincide with the limit of the salt basins. Sahabi et al., (2004) propose an age of initial opening at Sinemurian times (195 Ma), about 20 Ma earlier than proposed by Klitgord and Schouten (1986).

Based on deep seismic data, Shuck et al. (2019) propose that full lithospheric break-up between North America and Africa did not happen during the time of volcanism that produced the ECMA, and that normal seafloor spreading started during the creation of the BSMA under a moderately high mantle potential temperature.

For comparison of the deep structure of the conjugate margins, two roughly conjugate profile pairs were selected and are presented in the following subchapters, the SMART and MIRROR profiles (Funck et al., 2004 and Biari et al., 2015, respectively) and the LASE and DAKHLA profiles (LASE Study Group, 1986 and Klingelhoefer et al., 2009, respectively).

2.1 The Nova Scotia and Moroccan conjugate margin pair

Both margins have been sampled by wide-angle seismic cruises, the SMART (Funck et al., 2004; Wu et al., 2006) and OETR (Lau et al., 2018) cruises on the North American side and the SISMAR (Contrucci et al., 2004) and MIRROR (Biari et al., 2015) cruises on the African side (Figure 5). The three SMART profiles were acquired in regions where the ECMA shows different amplitudes to study the origin of this magnetic anomaly and the transition

from volcanic to magma-poor rifting. While the SMART 1 profile in the north shows a 150 km broad region with serpentinized upper mantle material exhumed to the seafloor and very thin oceanic crust, the central profile, SMART 2, is characterized by high lower crustal velocities, probably due to a mixture of serpentinized upper mantle material with gabbroic layer 3 (Funck et al., 2004; Wu et al., 2006). The southernmost profile, SMART 3, images a 100km-wide region of thin crust underlain by a 5 to 10km-thick body interpreted to be magmatic underplate neighboring normal oceanic crust of 7 km thickness (Louden et al., 2010). The OETR profile is located about 100 km north of the northernmost SMART profile and shows structures similar to the SMART 1 profile, including a zone of partially serpentinized upper mantle and a thin oceanic crust (Lau et al., 2018) indicating that rifting was magma-poor in this region. Seismic profiles from the MIRROR cruise on the NW-African margin and located conjugate to the SMART 1 profile show no indications for the existence of serpentinized upper mantle material (Biari et al., 2015) and image an oceanic crust of 7-8 km thickness, typical of oceanic crust. The crustal structure of unthinned continental crust is similar along both profiles, including three distinct layers and a thickness of 35-38 km, thinning over a distance between 150-200 km. Therefore, a symmetric rifting phase followed by asymmetric oceanic crustal accretion with accretion of mainly serpentinized mantle material towards the Canadian side and magmatic crust towards the African side has been proposed (Biari et al., 2017) (Figure 6). The difference in oceanic crustal thickness may be explained by the passage of the Canary Hotspot in the region around 60 Ma (Holik et al., 1991). Alternatively, a 16 km eastward mid-oceanic rift jump has been proposed to explain this asymmetry (Lau et al., 2018).

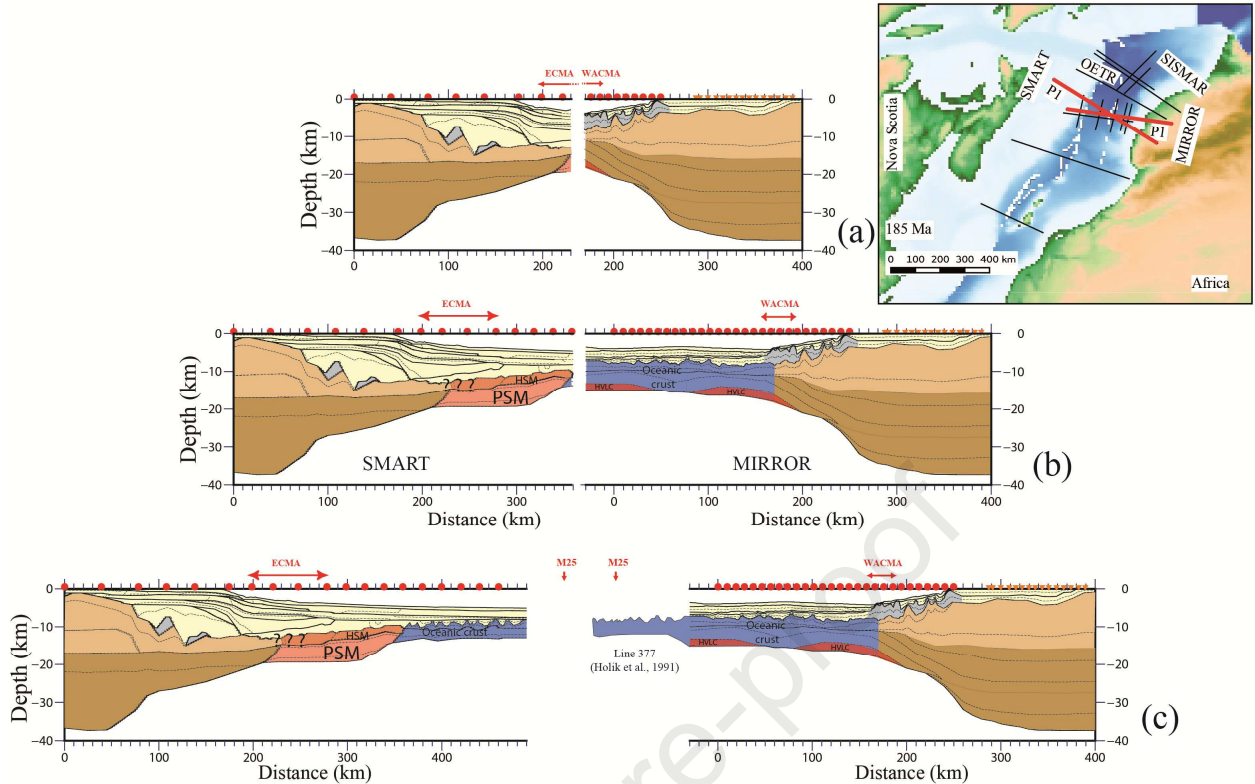


Figure 6. Reconstruction of the Nova-Scotia-Moroccan margin pair. (a) After initial extension, (b) asymmetric seafloor spreading, with exhumation to the west and magmatic accretion to the east, and (c) accretion of thin magmatic oceanic crust (after Biari et al., 2017). Abbreviations are HSM, highly serpentinised mantle; PSM, serpentinised mantle.

2.2 The US and NW-African margin pair

A second conjugate seismic profile pair is located offshore NW-Africa (Klingelhoefer et al., 2009) and the US continental coast (LASE Study Group, 1986) (Figure 7). Here, the American margin is highly magmatic, with a thick layer of SDR and other volcanic products. Along its conjugate no SDR were identified, which might explain the lower magnitude of the magnetic anomalies in this region. Some volcanic intrusions are emplaced into the lower crust along the Dakhla profile (Klingelhoefer et al., 2009). One major difference is the location of the first identifiable magnetic anomaly, located at thinned continental crust on the African margin and in the oceanic domain at its conjugate. A second difference is the thickness of the pre-rift continental crust, which is 27 km at the African side compared to 40 km along the US continental margin. The fact that the crustal thickness is less along the DAKHLA profiles when compared to those to the north and on the conjugate margin can be explained by the presence of the Reguibat Ridge, a large basement high which forms an uplifted part of the West African Craton, (Klingelhoefer et al., 2009; Labails and Olivet, 2009; Benabdellouahed et al., 2017). This Ridge remained stable during the Hercynian orogeny and the Central Atlantic opening and led to the formation of several NE-SW oriented basins (Tarfaya-Laayoune and Mauritanian-Senegal BASINS). Therefore, its crustal thickness is not a direct result of rifting only. The oceanic part of both profiles images a 7 to 10 km-thick crust with velocities typical for oceanic crust.

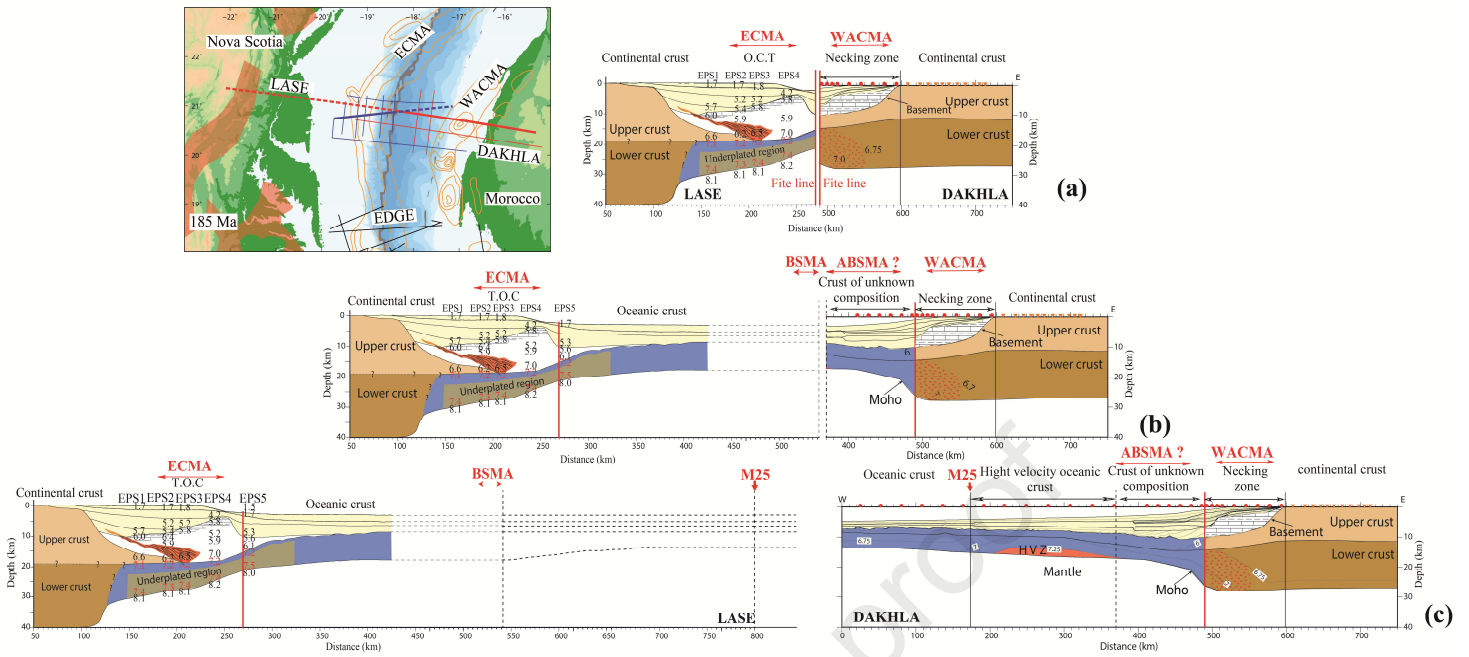


Figure 7: Reconstruction of the U.S.-Mauritania margin pair. (a) After initial extension, (b) accretion of highly intruded and underplated oceanic crust and seaward dipping reflector sequences on the western margin and slightly intruded oceanic crust on the eastern margin, and (c) accretion of thin normal oceanic crust (Biari et al., 2017).

3 Southern Atlantic

Several studies have been published to improve the Bullard (1965) original opening model of the South Atlantic, by using other geological, geophysical, magnetic and gravity data, but until now the opening mechanism of the South Atlantic Ocean has remained a matter of debate (Le Pichon and Hayes; 1971; Dietz, 1973; Mascle and Sibuet, 1974; Smith and Briden, 1977; Sibuet and Mascle, 1978; Curie, 1984; Unternehr et al., 1988; Eagles, 2007; Moulin et al., 2010). During the opening of the South Atlantic, several conjugate pairs of rift basins were formed, Argentina – Namibia BASINS, Santos, Campos - Espirito Santo – Angola basins and Bahia-Sergipe-Gabon BASINS. However, even if they show almost the same tectono-sedimentary evolution, they display differences from very wide marginal basins with the presence of a thick layer of Aptian salt to very narrow basins with no Aptian salt (Davison 1997, Mohriak et al., 2008). From analyses of an extensive magnetic dataset it was proposed that a significant portion of the SDR was produced subaerially (Collier et al., 2017).

Relative plate motions between the South American and African plates are still debated due to the lack of identifiable magnetic anomalies during most of the continental break-up period. The Cretaceous Magnetic Quiet Zone (CMQZ) is related to a period during which there were no geomagnetic reversals (Rabinowitz and LaBrecque, 1979; Malinverno et al., 2012; Granot and Dymant, 2015; Collier et al., 2017). Due to the lack of magnetic data, weak magnetic anomalies close to the equator and a lack of magnetic reversals, several kinematic

reconstructions for the Equatorial segment have been suggested based on fracture zone positions (e.g Olivet et al., 1984; Campan, 1995; Fidalgo, 2001). Along the Central and Austral segments, the thickness of sedimentary layers along the continental domain mask and complicate the gravity response, and only a few gravimetric anomalies identified in the deep ocean can be extended towards the continents (Moulin et al., 2010).

After the first work of Rabinowitz and LaBrecque (1979), identifying magnetic signatures back to M11 on the African side off South Africa and back to M4 on the conjugate South American plate off Argentina, the authors identified a high amplitude magnetic anomaly, named the G Anomaly interpreted to mark the ocean-continent transition. As it coincides with thick wedges of seaward-dipping reflectors (SDR) (Gladczenko et al., 1997; Bray et al., 1998) these reflectors have been suggested to be its cause (Bauer et al., 2000; Corner et al., 2002). Later, Moulin et al.; 2010 proposed that the M11 and M9 anomalies do not exist on the South American side and interpret a part of the G anomaly in the PelotasBASIN as M4 anomaly, suggesting to merge them into a single anomaly called Large Marginal Anomaly (LMA) (Moulin et al.; 2010).

By constructing a modern evolution model of the South Atlantic Ocean taking into account the direction and timing of the opening as well as the best continental fits, Moulin et al., (2010) proposes 4 phases of opening : (1) Opening of the Austral Segment between LMA (~133 Ma) and M4 (130 Ma) (2) Movement of the Santos Block westward and start of extensional movement in the Central Segment (3) Opening of the Central Segment between Late Aptian/Early Albian accompanied by an eastward ridge jump, very close to the African coast, allowing the formation of a first continuous mid-ocean ridge (4) Opening of the Equatorial Atlantic Ocean during Albian times, and definitive water connection between the Central and South Atlantic Oceans. Opening of the Falkland segment preceded these phases. However, it is linked closely to the movement of the Antarctic plate and therefore partly disconnected from the other segments (Moulin et al., 2010).

In the following figures we present deep seismic transects from the Falkland, Austral, and Central South Atlantic segments (Figure 8 and 9).

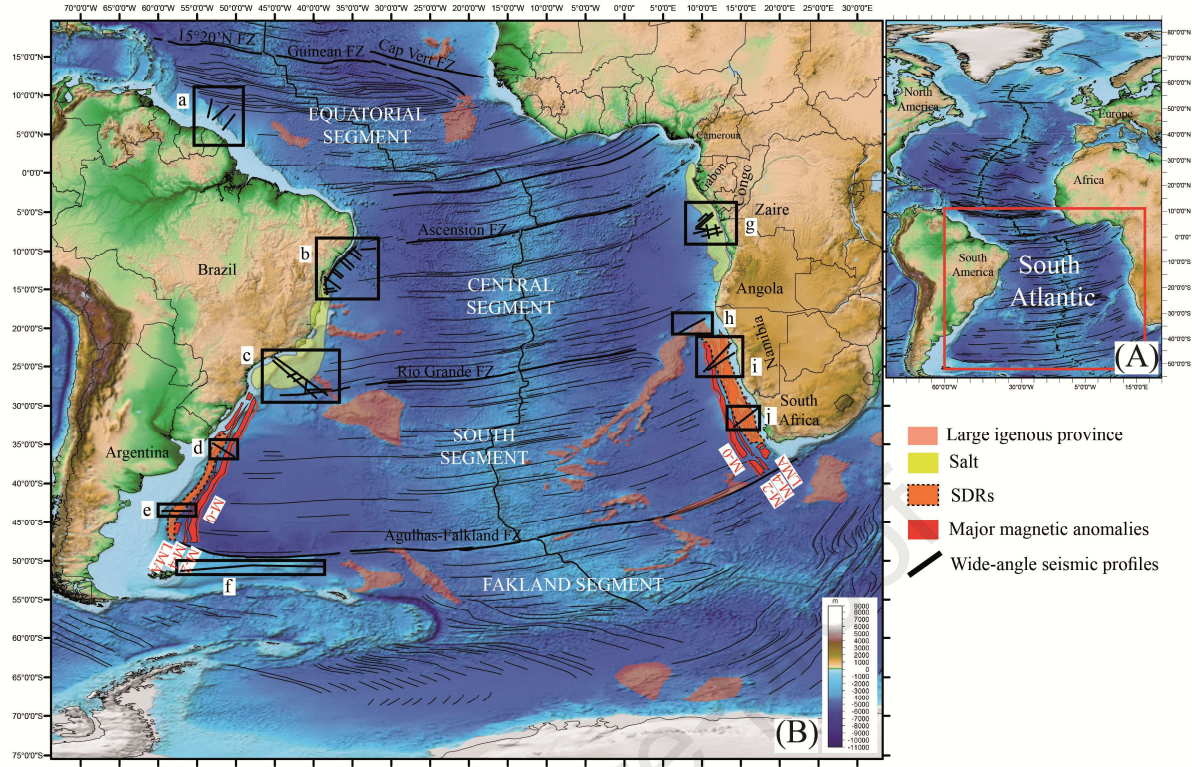


Figure 8: A) Topographic and bathymetric map of the Atlantic Ocean from ETOPO1 bathymetry grids (Amante and Eakins, 2009). (B) Overview of the South Atlantic Ocean divided into four segments from the north to the south, the Equatorial segment, the Central segment, the South Segment and the Falkland Segment (e.g Popoff, 1988, Moulin et al. 2005). Bathymetry and topography from ETOPO1 bathymetry grids (Amante and Eakins., 2009). The Aptian salt as well as SDR structures are from Torsvik et al. (2009). The large igneous province volcanics are from Coffin and Eldholm (1994). Magnetic anomalies are from Moulin et al., 2005. Black lines along the South American margin represent wide-angle seismic profiles of the Amazon Cone Experiment and the MARGATS cruise (a) (Greenroyd et al., 2008, Museur et al., 2020); SalSa experiment (b) Loureiro et al., 2018; SanBa experiment (c) (Klingelhoefer et al., 2015; Evain et al., 2015); BGR04-REFR01(d) and BGR98-REFR2 (e) are from Becker et al., 2014; The AWI-20130010 (f) wide-angle seismic profile running from East Falkland to the Georgia BASIN is after Schimschal and Jokat, 2019. On the African conjugate margin (g) represents wide-angle seismic profiles of the ZaiAngo cruise (Contrucci et al., 2004); (h) P2 profile along the Walvis Ridge after Fromm et al., 2015; (i) Profile 2 from the MAMBA experiments after Bauer et al., 2000 and (j) Springbok line after Schinkel et al., 2007.

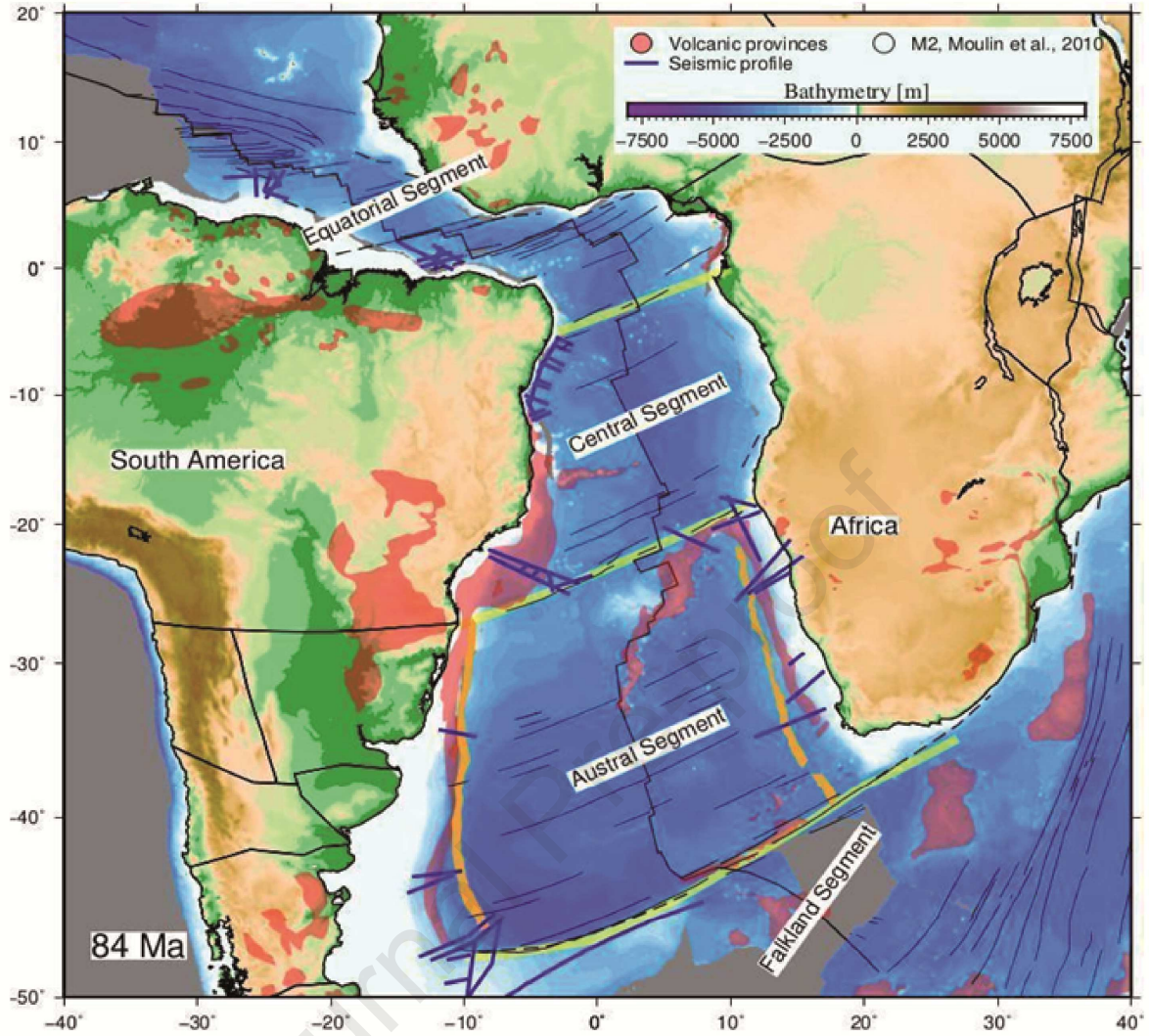


Figure 9: Reconstruction of the South Atlantic at 84 Ma (Chron 34) with rotational poles from Matthews et al., 2016. The fixed plate is Africa. Locations of wide-angle seismic profiles are marked by solid lines. Orange transparent polygons represent the location of volcanic provinces.

3.1 Falkland Segment

Opening of the South Atlantic started in the south during the Early Cretaceous (Figure 10) (Smith and Hallam, 1970; Tarling, 1972). There may be a link to the earlier break-up of the southwestern Weddell Sea that probably started around 160 Ma and was preceded by the emplacement of the Karoo/Ferrar mafic Large Igneous Province (LIP) (Lawver et al., 1991; Jourdan et al., 2005). Break-up of East Gondwana and the initiation of seafloor spreading in the Riiser-Larsen Sea and Mozambique BASIN and West Somali BASINS took place at similar times by 160 Ma (Roeser et al., 1996; Eagles and Koenig, 2008; Leinweber and Jokat, 2012). Torsvik and Cocks (2013) link the break-up of Gondwana to its location over hot mantle over the large African low shear-wave velocity province. The exact motions during break-up are still debated and many detailed studies taking into account plate motions and overlaps, paleo

stress, sedimentary basins and seismic studies have recently been published (Leinweber and Jokat, 2012; Gaina et al., 2013; Schimschal and Jokat, 2019; Thompson et al., 2019).

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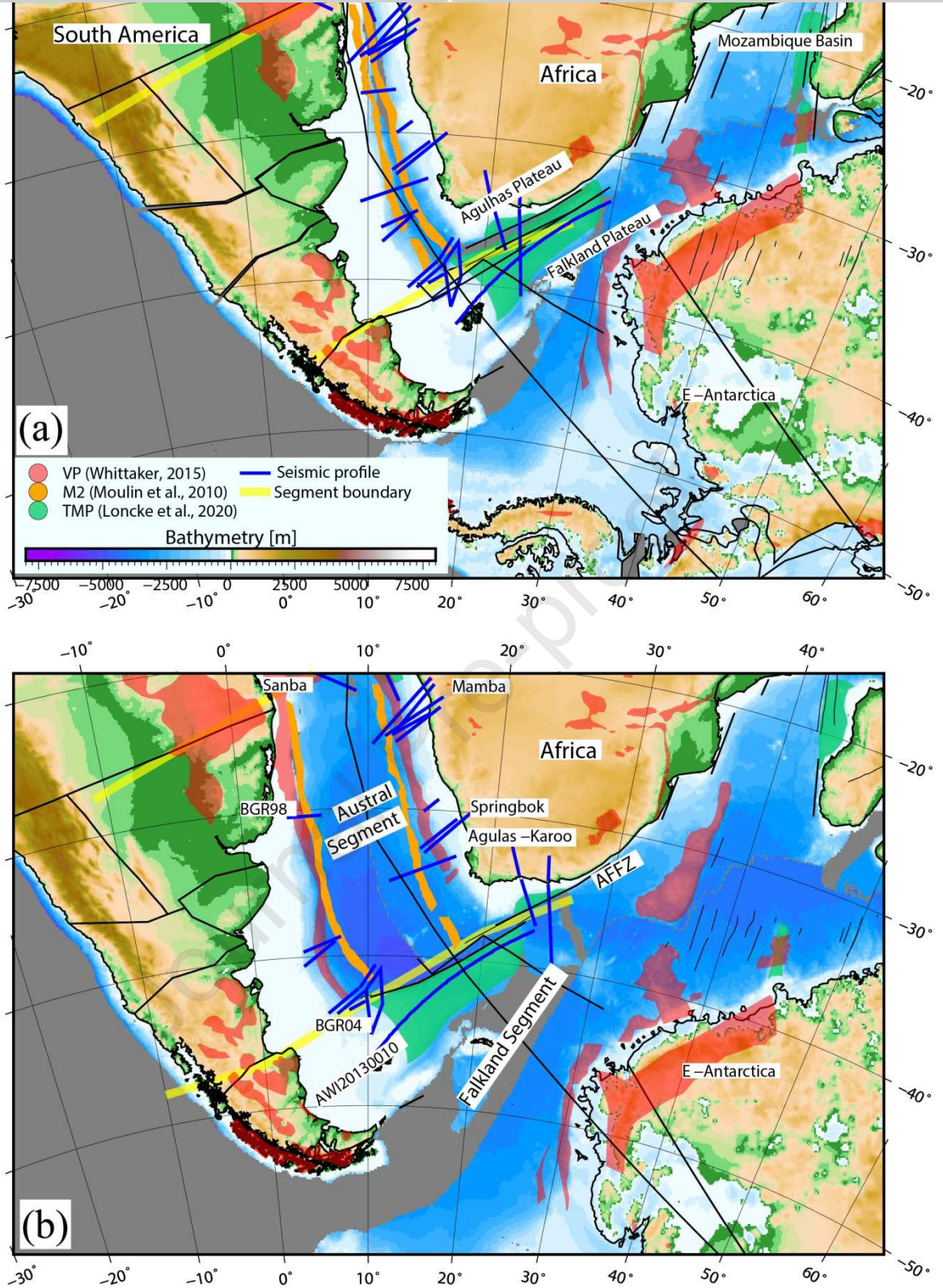


Figure 10: Reconstruction of the South Atlantic at 130 Ma (a) and at 110 Ma (b) with rotational poles from Matthews et al, 2016. The fixed plate is Africa. Locations of wide-angle seismic profiles are marked by solid lines. AFFZ = Agulhas-Falkland Fracture zone.

In the southern South Atlantic, the Falkland/Malvinas TMP and its conjugate, the Agulhas TMP, are two of the most prominent morphological features (Figure 10). They are located south and north of the Agulhas-Falkland/Malvinas Fracture zone (AFFZ), respectively. First studies of the Scotia Sea and Falkland PLATEAUS using sonobuoy data led to the suggestion

of a continental nature of the Falkland PLATEAU (Ewing et al., 1971). Later analysis of magnetic anomalies and seismic data led to the hypothesis that the eastern boundary of the plateau corresponds to a volcanic margin (Barker, 1999). Recent studies show that it comprises three regions: the Falkland/Malvinas islands, underlain by a 35 km crust of continental origin; the Falkland PLATEAU BASIN, underlain by thick oceanic crust; and the Maurice Ewing Bank, a 28 km thick continental fragment which is heavily overprinted by magmatic intrusions (Schimschal and Jokat, 2018; Schimschal and Jokat, 2019) (Figure 11).

The southern margin of the African margin, including the conjugate Agulhas PLATEAU were studied in the scope of the Inkabaye Africa project. The plateau underwent two major stretching phases during the Early Jurassic with the initial Gondwana break-up and during the Early Cretaceous due to shear tension along the AFFZ (Parsiegla et al, 2007 and 2009). Along the profile FRA the Moho rises from 40 km inland to 30 km at the coast and then the crust is further thinning toward the AFFZ, where the ocean-continent transition is located (Stankiewicz et al., 2008). On the subparallel profile GRA, lava flows are identified from seismicreflectiondata. The middle crust is found to be thickened by intrusions and a lower crustal body with velocities between 7.0 km/s and 7.6 km/s makes up about half of the total crustal thickness (Parsiegla et al., 2008) (Figure 12).

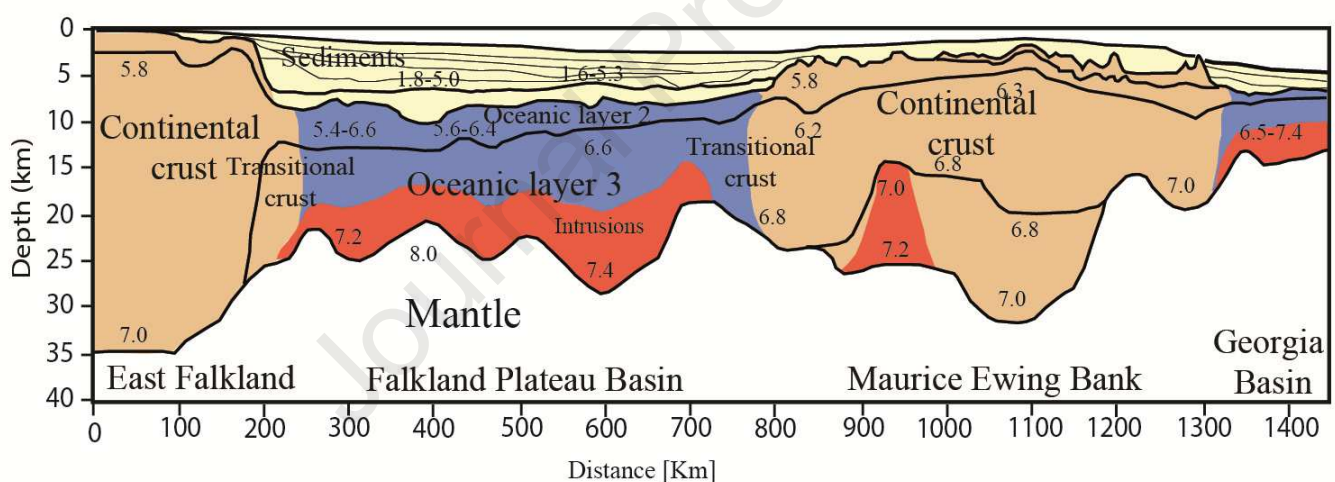


Figure 11: Deep structure of the Falkland PLATEAU and the Maurice Ewing Bank along profile AWI20130010 (after Schimschal and Jokat, 2019). For location see Figure 10.

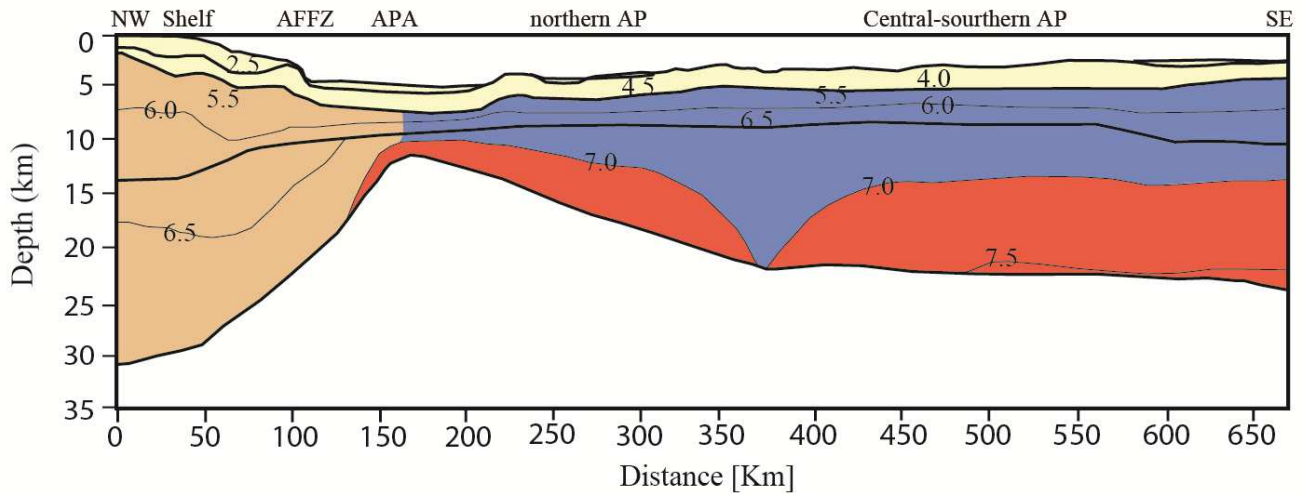


Figure 12: Deep structure of the Agulhas PLATEAU from profile AWI-20050200. Abbreviations are AFFZ, Agulhas-Falkland Fracture Zone; AP, Agulhas PLATEAU; APA, Agulhas Passage (after Parsieglä et al., 2008).

3.2 Austral Segment

From the Falkland segment, rifting propagated to the north into the Austral segment (Figure 13) about 133 Ma ago (Becker et al., 2012). The South Atlantic opened in the Early Cretaceous with suggested opening ages between 126 Ma and 137 Ma (Rabinowitz and LaBrecque 1979; Unternehr et al. 1988; Nürnberg and Müller 1991; Gladchenko et al. 1997; Jokat et al. 2003), and it is commonly proposed that this process progressed from south to north (Rabinowitz & LaBrecque 1979; Austin and Uchupi 1982; Uchupi 1989; Jackson et al. 2000). The opening of this segment of the Atlantic Ocean may thus be described as a northward unzipping of the continent.

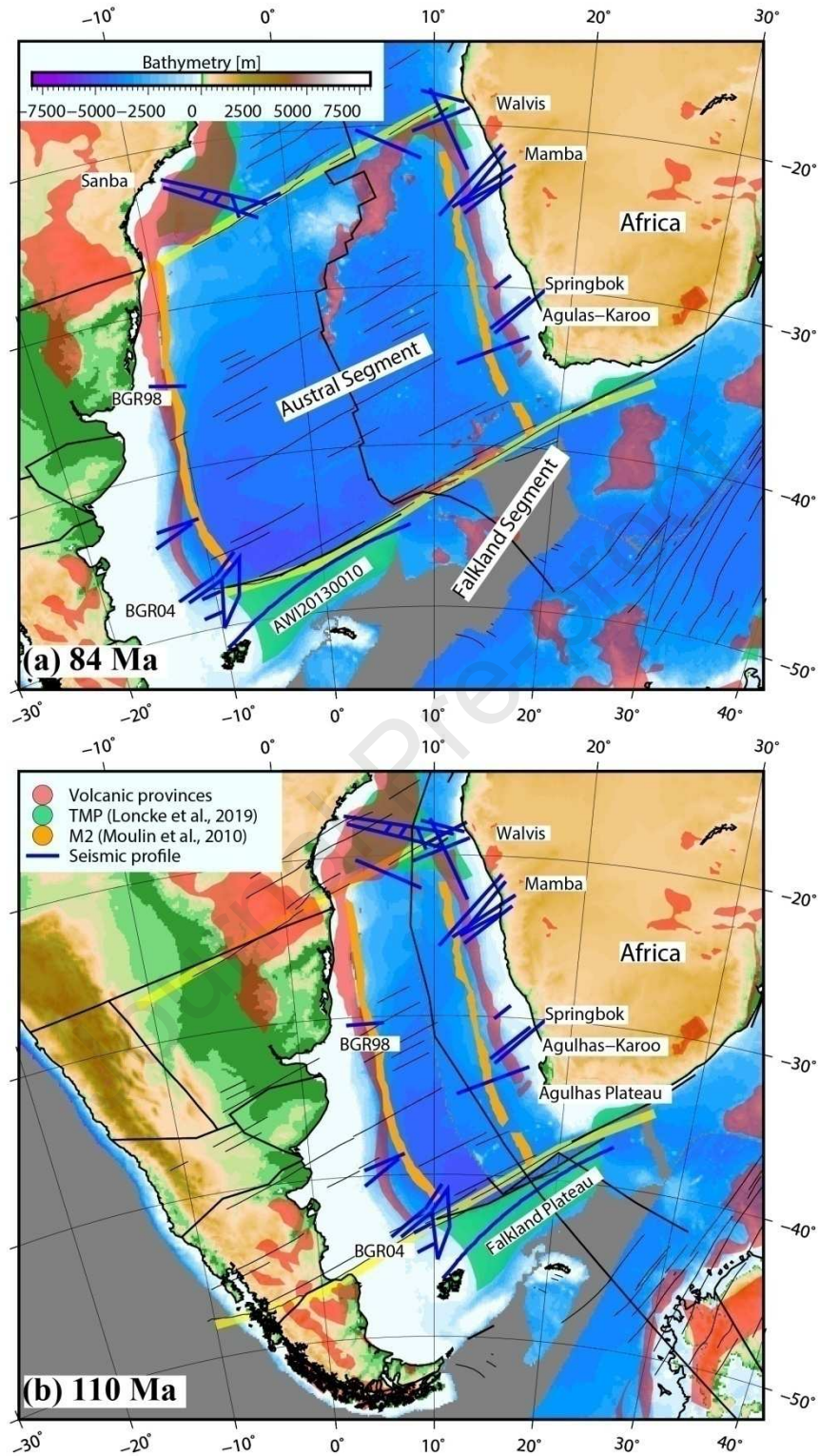


Figure 13: (A) Reconstruction of the southern or Austral South Atlantic at 84 Ma with rotational poles from Matthews et al, 2016. The fixed plate is Africa. Locations of wide-angle seismic profiles are marked by solid blue lines. Orange transparent polygons represent the location of volcanic provinces. (B) Reconstruction at 110 Ma.

Due to volcanism during rifting, the Austral segment is characterized by volcanic products in the form of layered volcanic sequences and mafic intrusions/underplate at the base of the crust (Bauer et al., 2000; Schnabel et al., 2008; Blaich et al., 2011; Becker et al., 2014). However, the rifted margins are not continuously of the volcanic type since the southernmost 200 to 300 kilometers are magma-poor (Koopmann et al., 2014b; Koopmann et al., 2016; Franke et al., 2010). The change from a non-volcanic to volcanic margin occurs abruptly at both conjugate margins.

The deep structure of the conjugate margins has been imaged over several surveys, across the Colorado BASIN offshore Argentina (Franke et al., 2006), at the north Falkland BASIN (Becker et al., 2012), the Namibian margin (Bauer et al., 2000), and the southern margin of the African continent (Stankiewicz et al., 2008). In addition, seismic wide-angle studies, seismic reflection data and gravity data have been used (Franke et al., 2007; Blaich et al., 2009; Franke et al., 2010). The relationship between SDR and the COB (continent ocean boundary) is not homogeneous along the South Atlantic margins and most likely variable on either conjugate section (Koopmann et al., 2016). According to Koopmann et al., 2016, there is no reason to doubt the existence of mainly mafic (oceanic) crust seaward of the outer SDR due to magnetic lineations and seismic character, and there is little doubt that thinned continental crust underlies the feather edge of the SDR.

Paton et al.; 2017a; 2017b identified that the Colorado-Orange BASIN rift phase formed a tripartite system with the Namibian Gariep Belt (for location see Figure 33), which they called the Garies Triple Junction. All three rift branches were active during the break-up of Gondwana, but during the Atlantic rift phase the Colorado BASIN failed while the other branches continued to rift, defining the present-day location of the South Atlantic. This clearly contradicts the suggestion that rifting follows zones of crustal heterogeneity (Paton et al.; 2017a and b).

Two wide-angle seismic profiles acquired perpendicular to the Namibian continental margin cross two wedges of seaward dipping reflectors (inner and outer SDR) at 100-150 km and 300 km offshore (Bauer et al., 2000) (Figure 14). Normal oceanic crust is imaged oceanward of the western SDR wedge, while the continent-ocean transition zone is located at the landward end of the eastern inner SDRs (Bauer et al., 2000). For the northern Namibian margin, Bauer et al. (2000) show that the landward-most SDR complex overlies continental crust of about 30 km thickness, and the seaward SDR overlies a high-velocity lower crustal body (HVLC) of 15–20 km thickness with seismic velocities exceeding 7.0 km/s. On the conjugate South American margin, Schnabel et al. (2008) propose values of 20–25 km for the thickness of the (partly thinned) continental crust underlain by a first high velocity lower crustal body up to 7.3 km/s below the SDR and a further thinning crust seaward of the SDR. The entire transition zone is underlain by up to 20 km of high velocity material, which the authors interpret to be exclusively igneous material and not intruded continental crust. Koopmann et al., 2016 conclude that the SDR occur at the very end of the rifting process prior to continental break-up of crust of varying thickness that thins quite dramatically towards the ‘seaward-most’ SDR.

Further south, several wide-angle seismic profiles were acquired offshore South Africa (Schinkel et al., 2006; Becker et al., 2014). The profiles show a single thick SDR complex and up to 10 km thick underplate layers at the transition zone. On the conjugate Argentinian and Uruguay margin three wide-angle seismic transects were acquired (Becker et al., 2014). Although this margin also shows SDR packages and underplate layers, the amount of volcanic products is much lower here, only about 25% of that on the conjugate African margin. According to the authors, the difference in the amount of volcanic products can be explained by asymmetrical rifting in simple shear mode, similar to structures observed in the conjugate East Greenland/Hatton Bank margins. The results also question a simple causal relationship between SDR and high velocity lower layers (Becker et al., 2014). In addition, the location of the underplate with respect to the SDR sequences varies from south to north along the margins of the Austral South Atlantic. In the south, close to the Falkland–Agulhas Fracture Zone, a thin high velocity layer is present, without volcanic extrusives like SDR. In the center of the margin segment, the underplate layer is located beneath the inner wedges of the SDR packages and in the north close to the Rio Grande Rise–Walvis Ridge this layer reaches seaward of the SDR domain (Becker et al., 2014).

The only timing information on the SDR in the South Atlantic is from the Kudu wells at the border between Namibia and South Africa. Sediments directly overlying the drilled basalts are dated as Late Barremian (in the time-scale used here c. 122–123 Ma (Gradstein et al., 2004). Sparse microfauna in the lowermost interval interbedded with volcanics may indicate an age no older than Valanginian (Erlank et al. 1990). From the successive disappearance of magnetic anomalies in the region of SDR, Koopmann et al., 2016 derive a minimum age estimate for the seaward-most volcanic flows and suggest that the volcanism related to the formation of the South Atlantic propagated from south to north along with the opening of the oceanic basin itself (Koopmann et al., 2016).

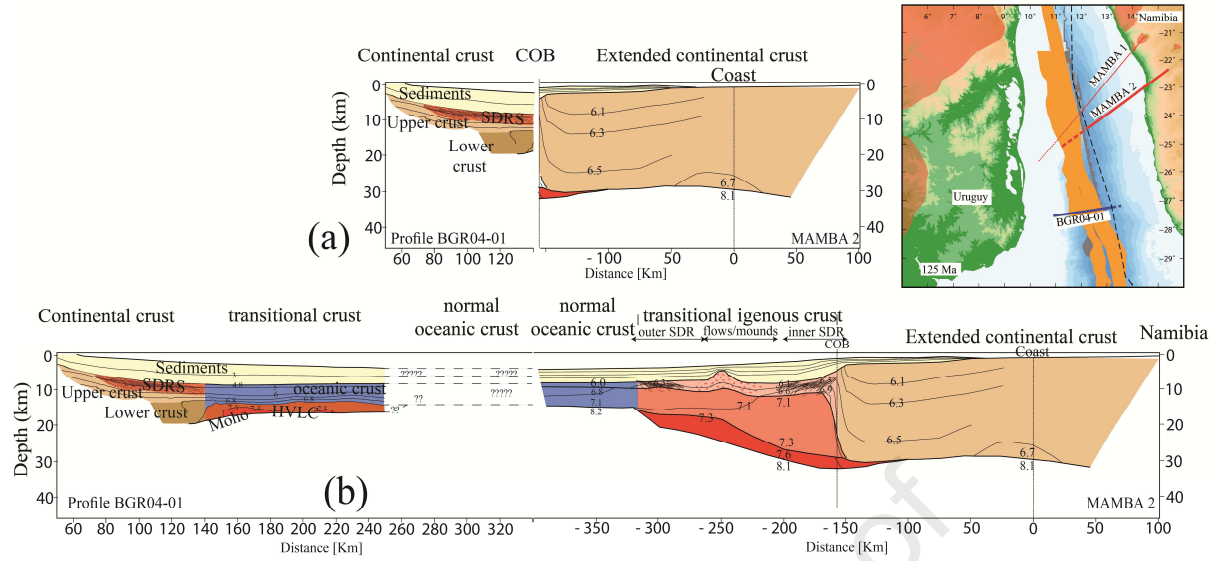


Figure 14: Deep structure of the Uruguay continental margin and the Namibia continental margin (after Bauer et al., 2000 and Becker et al., 2014). (a) reconstruction for break-up (COB), (b) accretion of oceanic crust.

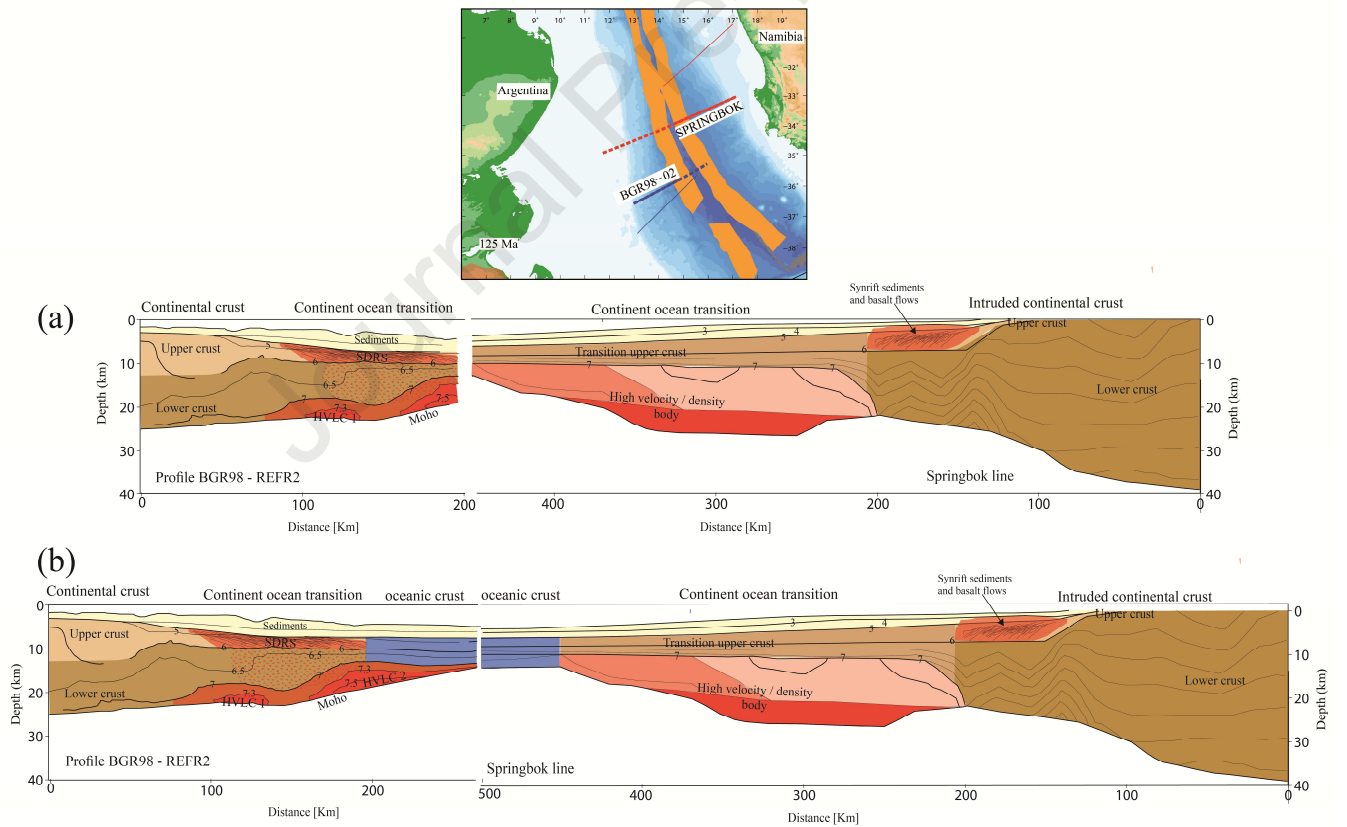


Figure 15: Deep structure of the Argentinian and Namibian continental margins (after Schinkel et al., 2007 and Becker et al., 2014). (a) reconstruction for break-up (COB), (b) accretion of oceanic crust.

3.3 Central segment

The extension propagated further to the North and break-up in the central segment took place in the late Early Cretaceous between the Eastern Brazilian/Western African continental margins (Rabinowitz and LaBrecque, 1979; Changet al., 1992; Heilbron et al., 2000; Mohriak et al., 2008; Blaich et al., 2011) (Figure 16).

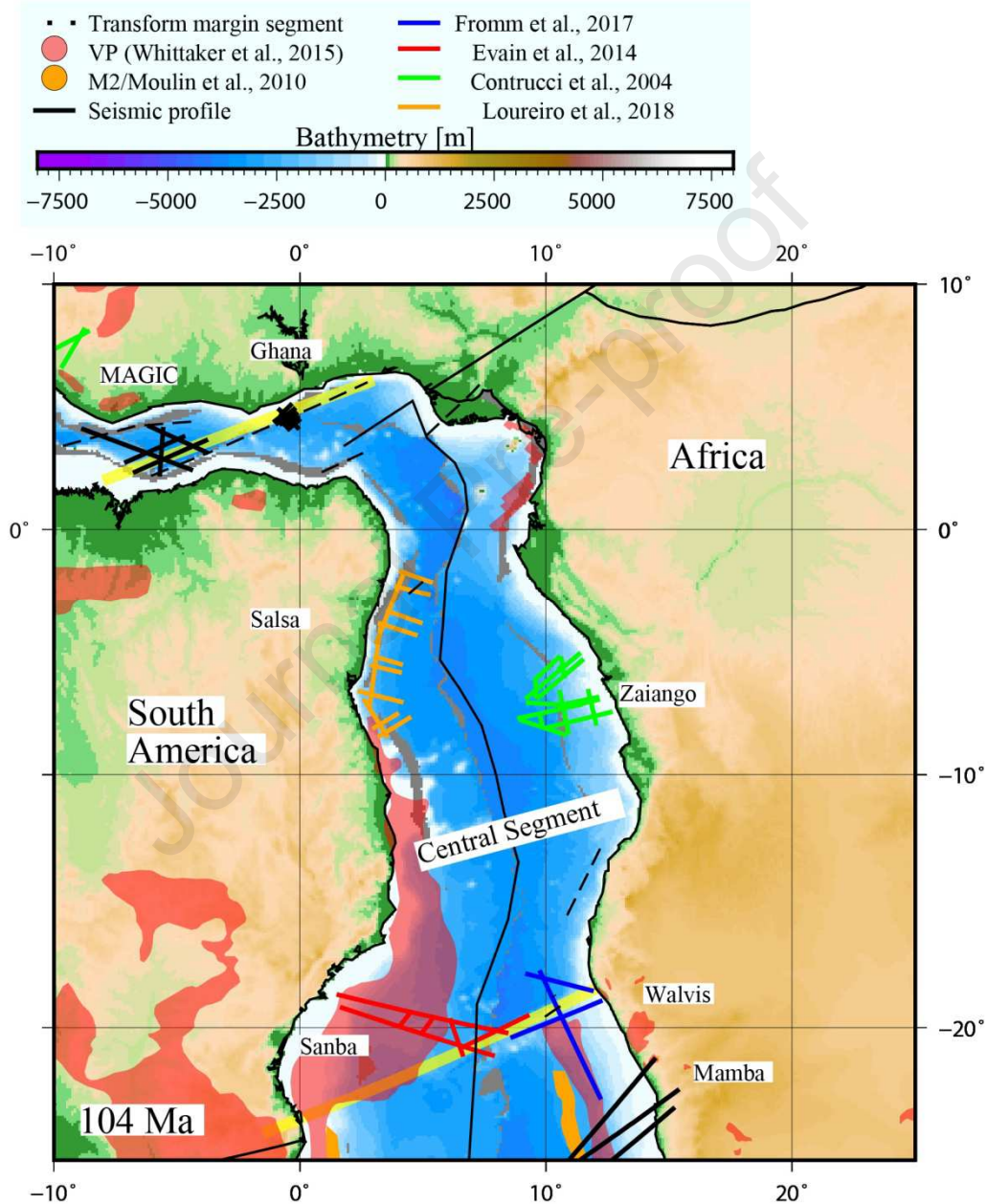


Figure 16: Reconstruction of the South Atlantic at 104 Ma with rotational poles from Matthews et al, 2016. The fixed plate is Africa. Locations of wide-angle seismic profiles are marked by solid lines.

3.3.1 Brazilian–Zaire Angolan Margin pair

The north of the central segment of the South Atlantic was imaged by two wide-angle seismic surveys, the SALSA survey in 2014 (Loureiro et al., 2018; Pinheiro et al., 2018) on the Brazilian margin at the Sergipe-Alagoas and Jequitinhonha BASINS and the ZAIANGO survey in 2000 in the Congo BASIN (Contrucci et al., 2004; Moulin et al 2005). In the reconstruction (Figure 17), the Zaiango 14 profile is located about 150 km south of the Salsa 11 profile, during opening times.

Spanning the Angola-Zairemargin, Profile 14 is 270 km long (Figure 17b), from the coast to the abyssal plain and crosses the Zaire canyon (Contrucci et al., 2004). It is located nearly conjugate to the SALSA 11 profile acquired across the Brazilian Campos margin (Loureiro et al., 2018). The Zaire-Angolan margin is characterized by the presence of a deep subsalt basin which is underlain by an intermediary or transitional type crust neither purely oceanic nor continental (Contrucci et al., 2004). This transitional zone is bounded by a basement ridge with high seismic velocities (6.6 km/s to 6.8 km/s). A relatively thin (2-3 km) high velocity layer at the base of the crust has been suggested to consist of either serpentinizedmantle material or lower crustal material (Contrucci et al., 2004) (Figure 17). The neighboring oceanic crust, despite having a typical thickness of oceanic crust, is unusual in that the upper layer is 5 km thick with laterally highly variable velocities. It has been proposed to consist of an alternation of gabbros, basalt and serpentine, overlying a lower layer of only 2 km thickness with a seismic velocity of 7.5 km/s possibly consisting of serpentinite. The up-to-6 km-thick layer of evaporites can be explained by either an elevated position of the basin at opening (Moulin et al., 2005) or a pre-existing depression, repetitively refilled and closed off (Crosby et al., 2011).

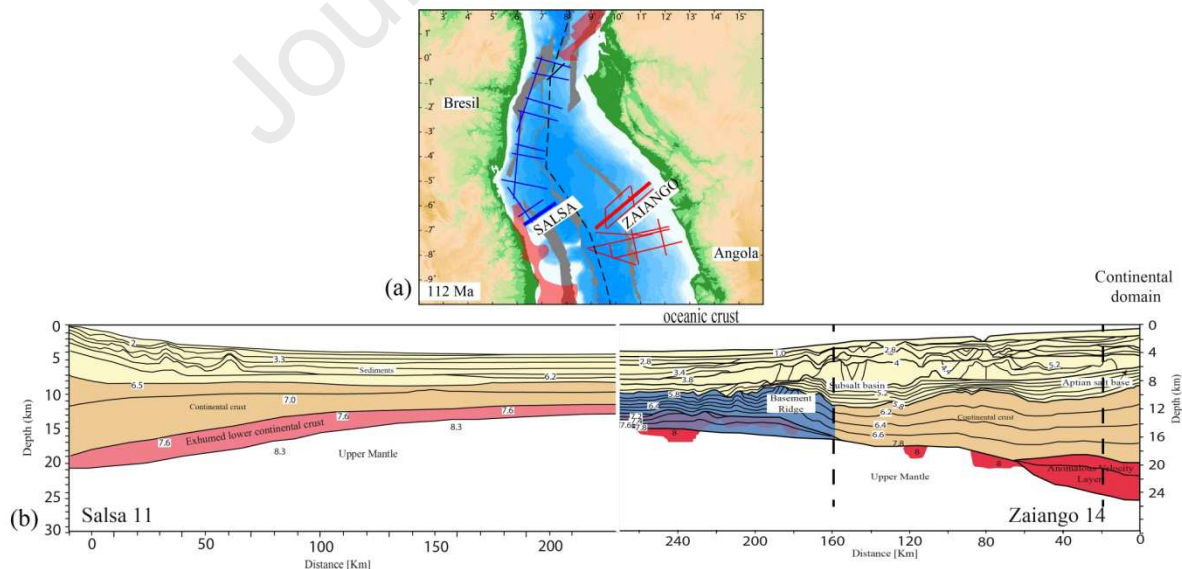


Figure 17: Deep structure of the Brazilian and Zaire continental margin (after Loureiro et al., 2018 and Contrucci et al., 2004).

Along the Jequitinhonha BASIN the SALSA 11 profile images a continental crust of about 20 km thickness thinning to less than 5 km in the distal basin. Here, it is characterized by high seismic velocities and gradients and therefore cannot, according to the authors, be classified as typical oceanic crust, such that typical oceanic crust was never observed along the profile. Although the authors suggest that the crust consists of exhumed lower continental crust which has crept towards the basin (Loureiro et al., 2018) (Figure 17), the seismic velocities (> 7.6 km/s) of this layer are significantly higher than those measured in typical lower continental crust elsewhere. In fact, the velocities are much closer to those of the high velocity layer above the Moho identified for example at the Nova Scotia margin, where it is proposed to consist of serpentinized mantle material from seawater descending through thin oceanic crust into the upper mantle (Wu et al., 2006).

3.3.2 Santos BASIN –Namibia BASIN pair

South of these profiles, another near conjugate wide-angle seismic profile pair consists of the SanBa experiment in the Santos BASIN offshore Brazil and an experiment in the Walvis ridge area. However, the two experiments are not located in the same segment as the SanBa profiles are situated north and the Walvis ridge profiles south of the Florianopolis fracture zone, so a direct comparison is not possible. The nature of the conjugate margins changes across the fracture zone from moderate to low magmatic in the north to heavily magmatic in the south.

The profiles, located in the vicinity of the Walvis ridge and a proposed hotspot trail (Duncan, 1984; Perez-Diaz and Eagles, 2014), show high lower crustal velocities of up to 7.5 km/s. In contrast to the neighboring profiles at the southwestern African margin where these high velocities do not extend underneath the continent (Bauer et al., 2000; Schinkel, 2007; Hirsch et al., 2009), here, they extend several tens of kilometers underneath the African continent, and might therefore be connected to a mantle thermal anomaly or hot spot (Fromm et al., 2015; Fromm et al., 2017). However, the authors find it unlikely that this thermal anomaly triggered the initial opening of the South Atlantic Ocean due to the missing large-scale intrusions within the continental crust in the region.

During one of the first deep seismic studies along the Santos BASIN, a combined reflection and expanding-spread-profiles (ESP) study was conducted (Figure 18) to roughly image the COB in the Santos BASIN (Leyden et al. 1971). The interpretation of 12 000 km of deep sounding seismic reflection data, coupled with gravimetric and magnetic modelling, provided a first 3-D view of the deep structure of the Santos, Campos and Espirito Santo BASINS (Zalán et al., 2011). The authors interpreted a large area of proposed exhumed and serpentinized upper mantle material. Based on gravity modelling, serpentinization was deduced to occur down to several kilometers depth.

The SanBa profiles (Evain et al., 2015) along the Brazilian continental margin spanning the Santos BASIN-São Paulo PLATEAU System, show a very abrupt thinning from a 40 km thick continental crust to less than 15 km. Adjacent is a 350 km-wide basin underlain mostly by thinned continental crust (Evain et al., 2015). Parts of this basin show anomalous velocities for thinned continental crust and they are proposed to be either atypical oceanic crust or, exhumed lower crust overlying altered mantle, or, alternatively, upper continental crust intruded by mafic material, overlying intruded lower continental crust (Evain et al., 2015). A triangular -shaped region in the SE of the Santos BASIN-São Paulo PLATEAU System is proposed to consist of either serpentinized exhumed upper mantle material or proto oceanic crust including basaltic flows on top of a layer of serpentinized upper mantle material (Klingelhoefer et al., 2015).

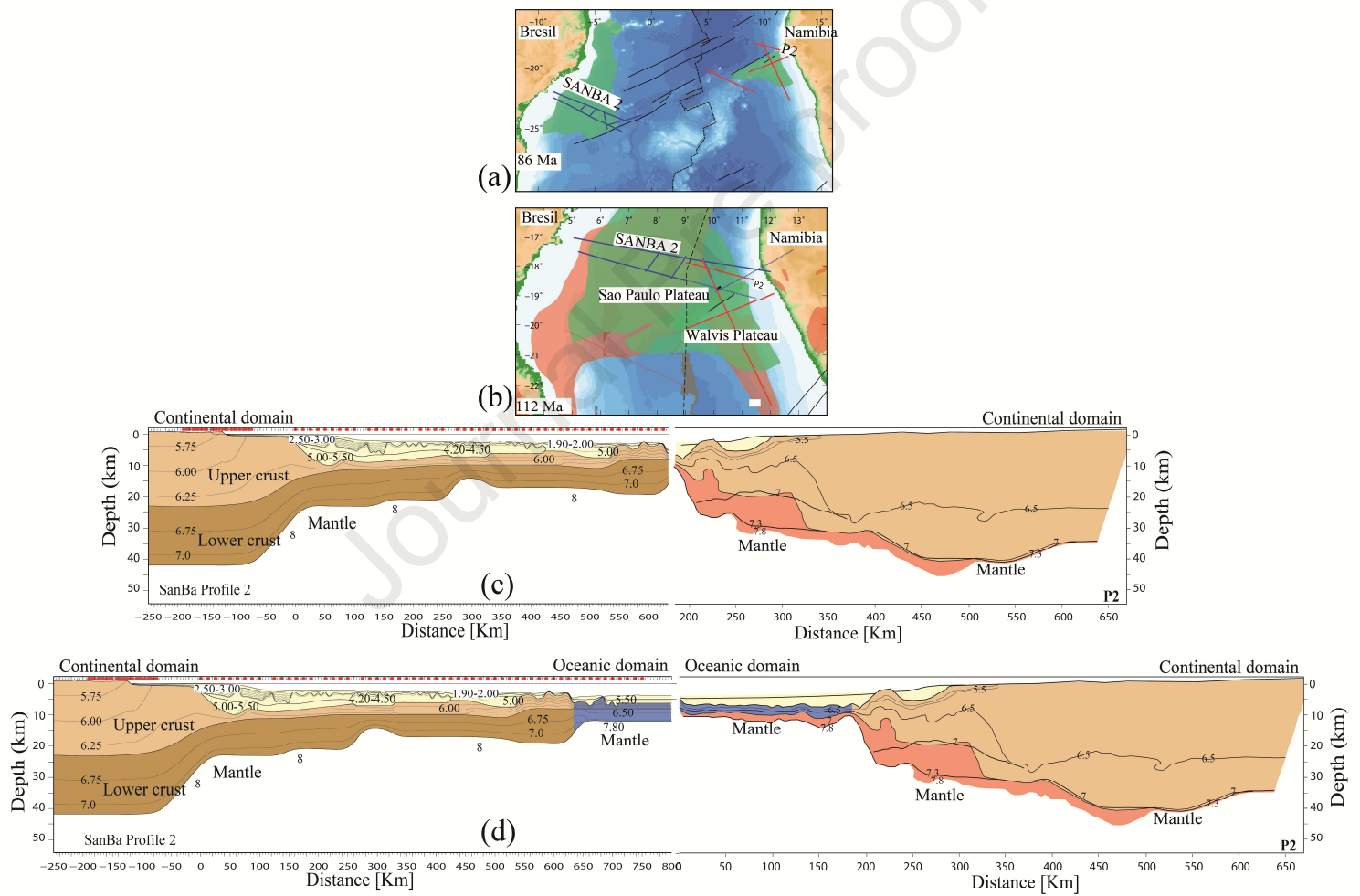


Figure 18: Deep structure of the Santos BASIN and Namibia margin (Evain et al., 2015 and Fromm et al., 2015). (a) Reconstruction at 86 Ma, (b) Reconstruction at 112 Ma, (c) reconstruction for break-up (COB), (d) accretion of oceanic crust. Red polygons mark volcanism (Whittaker et al., 2015) and green polygons mark transform marginal plateaus (Loncke et al., 2020).

Two transform marginal plateaus (TMP) are located along this segment, the Sao Paulo PLATEAU north of the Florianopolis Fracture Zone and the Walvis PLATEAU south of it. The Walvis PLATEAU consists of a 26 to 35 km-thick heavily intruded crust, probably of continental origin with a more-than-300 km-wide HVLC and the western part is proposed to consist of “Icelandic” type crust (Fromm et al., 2017; Planert et al., 2017). The Sao Paulo PLATEAU was found to consist of thinned continental crust; however, small scale volcanic activity also took place during its construction (Evain et al., 2015, Klingelhoefer et al., 2015).

3.4 Equatorial Segment

The last and northernmost segment of the Southern Atlantic between South America and Africa started rifting in a predominantly transform mode in the Early Cretaceous (Burke and Dewey, 1973; Sibuet and Mascle, 1978; Unternehr et al., 1988) (Figure 19). Rifting here was strike-slip or highly oblique in nature and controlled by large-scale transform faults, (from south to north the Chain, Romanche and Saint Paul fracture zones) with intermittent short segments of divergent margins (Nürnberg and Müller, 1991; Edwards et al., 1997; Basile et al., 2005; Heine et al., 2013). A transtensional shear corridor with a dextral sense of displacement developed at the northern continental margin of Brazil (Azevedo, 2011) (Figure 19). From numeric modeling, a break-up age of 118 Ma is predicted (Heine et al., 2013) which is in good agreement with other studies (Karner et al., 1997; Dupré et al., 2011). Numerical modeling also shows that oblique rifting facilitates the rifting process (Brune et al., 2012), the obliqueness of a rift can accelerate its evolution and final opening as compared to orthogonal rifts, thereby explaining the failing of the central and western African rifts (Heine and Brune, 2014) around 110 Ma. Linked to this failure is the onset of transpression along the Côte d’Ivoire-Ghana Ridge leading to a regional uplift event and the formation of a regional unconformity in mid-Albian times (Pletsch et al., 2001; Basile et al., 2005). The break-up between the African and South American continental lithospheres occurred at 104 Ma with the oceanic spreading center passing the Côte d’Ivoire/Ghana Marginal Ridge by 100–99 Ma (Clift et al., 1997; Pletsch et al., 2001; Basile et al., 2005; Antobreh et al., 2009). The rift BASIN entered a phase of thermal subsidence with a minor reactivation in post-Early Cretaceous times (e.g. Janssen et al., 1995). The first deepwater connection between the Central and South Atlantic became established in the late Albian to middle Coniacian (Pletsch et al., 2001). This separation had far-reaching implications for the definitive separation of African and South American fauna and flora, oceanic currents and the sedimentary and geochemical evolution in the neighboring ocean basins (e.g. Wegener, 1928; Pletsch et al., 2001).

While the Central Atlantic creates margins with long rifted segments separated by a few transform faults which offset the mid-ocean ridge, the rifted margins along the Equatorial Atlantic are generally short and associated with complex transform fault tectonics (Azevedo, 1991).

The African side of this margin was studied during the Charles Darwin Cruise 55 (Peirce et al., 1996) as well as part of the Equasis programs of 1983 and 1990 (Edwards et al., 1997a and b) on a NW-SE profile offshore Ghana, spanning the complete transform margin. It is

extremely narrow here thinning the crust from 23 km to 10 km over a distance of only 15 km. The neighboring oceanic crust is thin (~4-5 km) and the transition zone includes crust of high velocities but no sign of volcanic underplating (Peirce et al., 1996; Edwards et al., 1997a and b). Along the Suriname and French Guyana margin and the Brazilian margin, seismic data were acquired in the scope of the ACE study (Amazon Cone Experiment) (Greenroyd et al., 2007, 2008a and 2008b) and offshore French Guiana and Suriname during the MARGATS deep seismic cruise (Museum et al., 2020) (Figure 8 and 19). Profile A (Figure 20) images a 38 km-thick crust interpreted to be of continental origin at the margin of French Guiana (Greenroyd et al., 2007). The zone of crustal thinning is only 50 km wide, narrower than normal rifted margins, but wider than most transform margins. The continent-ocean transition is about 30 km wide and oceanwards, a 4-5 km thin oceanic crust was modeled, comparable to that on the African side, despite the fact that these profiles are not conjugate (Figure 19). The profile is interpreted to reflect a “leaky” transform margin formed through transtensional extension (Greenroyd et al., 2008b) or a hyper-oblique margin (Sapin et al., 2016).

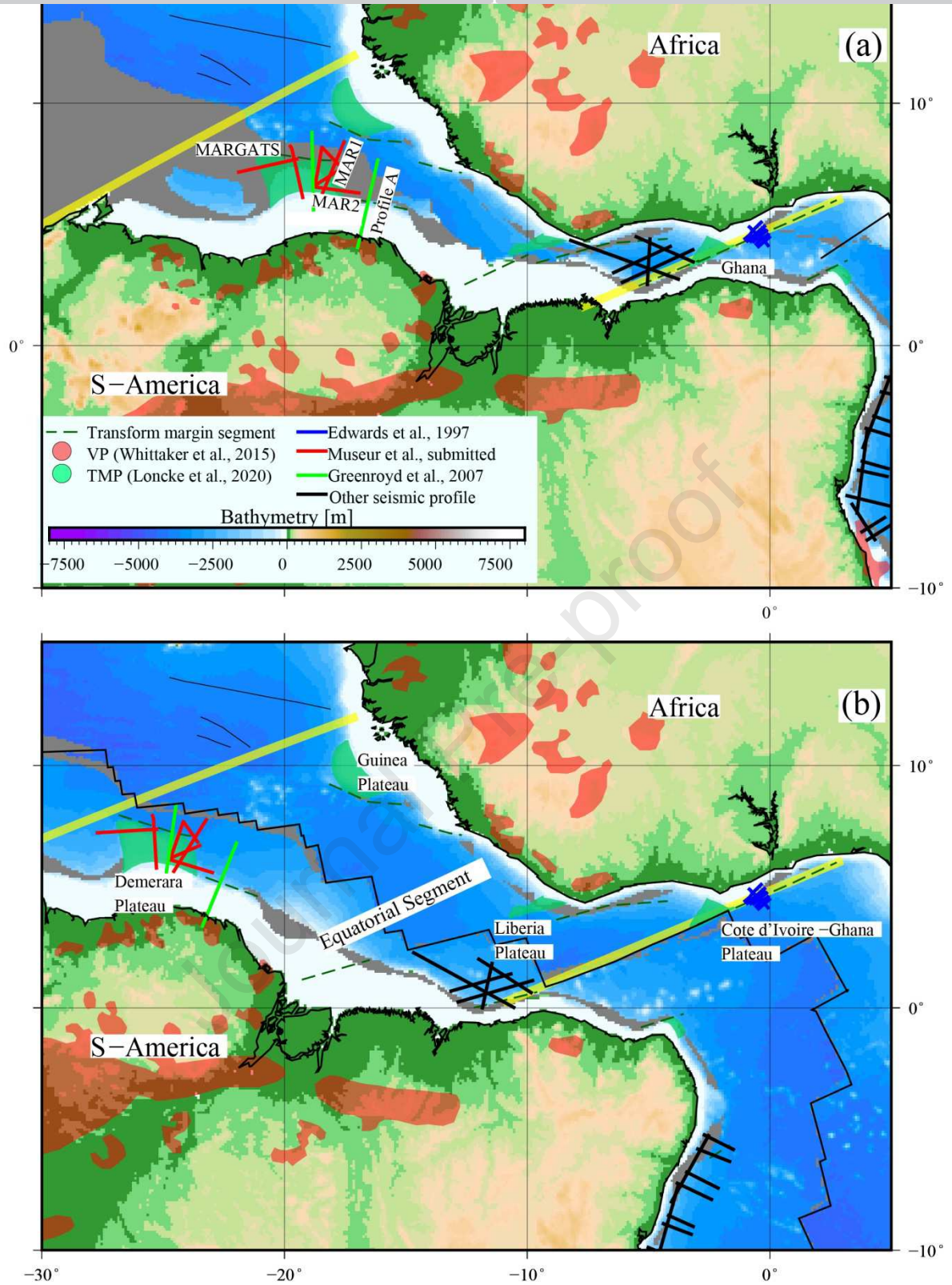


Figure 19: (a) Reconstruction of the equatorial Atlantic at 106 Ma with rotational poles from Matthews et al., 2016. The fixed plate is Africa. (b) Reconstruction of the equatorial Atlantic at 85 Ma. Locations of wide-angle seismic profiles are marked by solid lines (red lines: MARGATS from Museur et al., 2020; Green lines: lines from Greenroyd et al., 2007 and 2008). Orange transparent polygons represent the location of volcanic provinces (Whittaker et al., 2015) and green transparent areas TMP (Loncke et al., 2020).

Four TMPs are located along this segment: the conjugate Demerara and Guinea PLATEAUS, the Liberia PLATEAU, and the Côte d'Ivoire-Ghana PLATEAU. Based on a tight network of industrial deep seismic reflection profiles, Reuber et al. (2016) propose the presence of thick SDR sequences possibly related to the Bahamas hotspot constituting a large part of the Demerara PLATEAU. These SDR could originate from the presence of the Bahamas HotSpot in the region during the initial opening of the Central Atlantic (Reuber et al., 2016). These SDR were confirmed later by two wide-angle seismic velocity models from the MARGATS cruise seismic experiment (Museum et al., 2020) (Figure 20). Results of Museum et al., 2020, show a 30 km-thick crust, with an upper SDR crust and a middle crust, where velocities vary from 4.50 km/s to 7.00 km/s; the deepest layer was modeled by high velocities between 7.20 km/s and 7.60 km/s, and are interpreted as intruded magmatic material. For the authors, the Demerara PLATEAU would, therefore, be an inherited Jurassic volcanic margin boarding the Central Atlantic (Museum et al., 2020). The presence of SDR sequences and HVLC at the Demerara PLATEAU led to suggesting a polyphased origin of the Plateau including a first phase creating the volcanic western margin and, during the Cretaceous, a second phase creating its northern (transform or highly oblique) and eastern (divergent) margins (Basile et al., 2013; Reuber et al., 2016; Loncke et al., 2020; Museum et al., 2020). Although much-less explored, thick SDR sequences were also identified along the conjugate Guinea PLATEAU (Mercier de Lepinay et al., 2016). In a later phase (Aptian-Albian), the Guinea PLATEAU was overprinted by volcanic extrusions (Olyphant et al., 2017). At the neighboring Sierra Leone Rise, magma feeder dikes for these volcanics are proposed to have been guided by transform-related lithospheric discontinuities (Jones et al., 1991). While little data exist on the Liberia TMP, the Côte d'Ivoire Ghana PLATEAU has been extensively studied and consists of stretched continental crust at the intersection between a divergent and a transform margin (Sage et al., 2000).

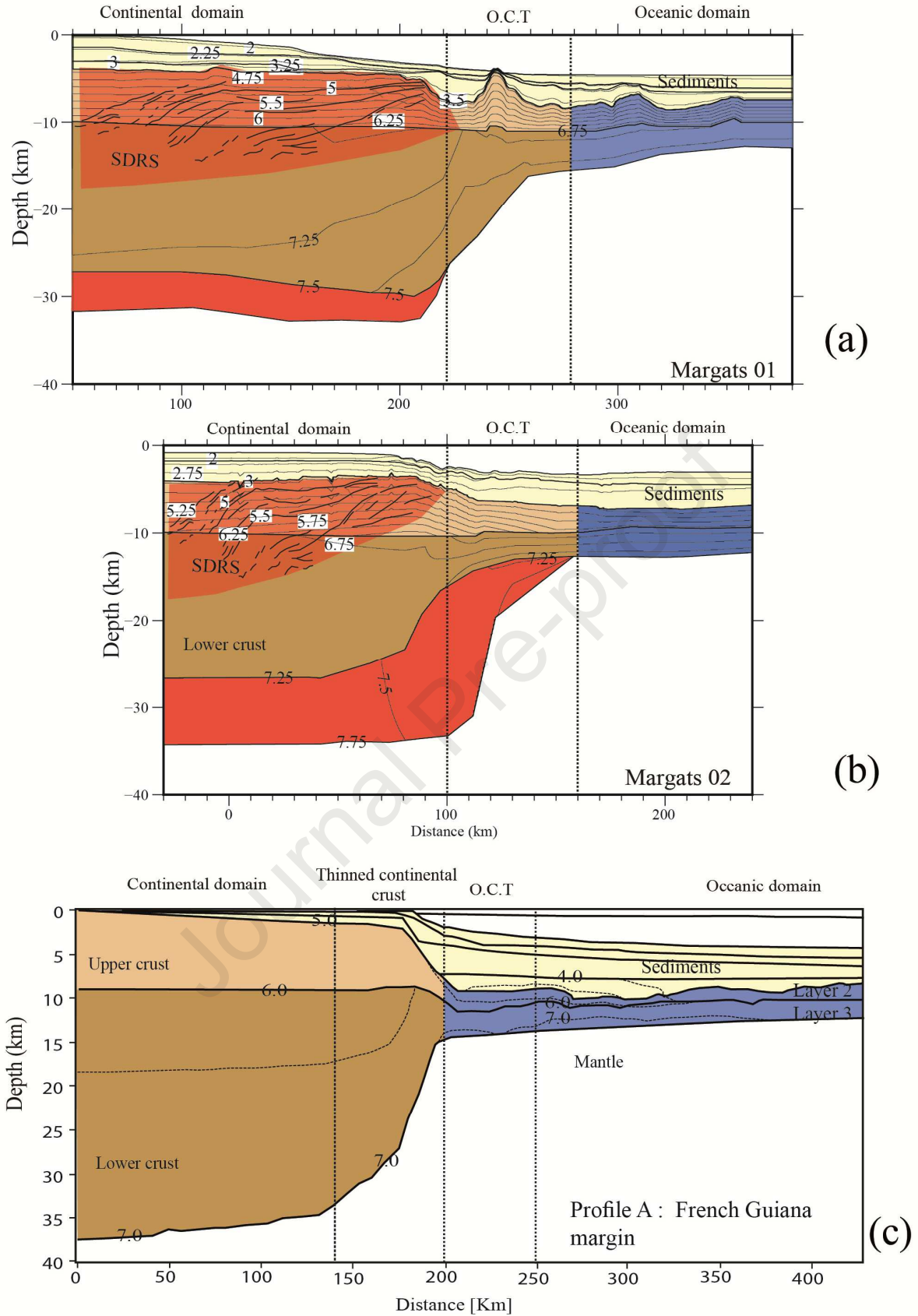


Figure 20: Deep structure of the Demerara Plateau and the French Guiana continental margin (Greenroyd et al, 2007; Greenroyd et al., 2008; Museur et al., 2020).

4 Northern Atlantic

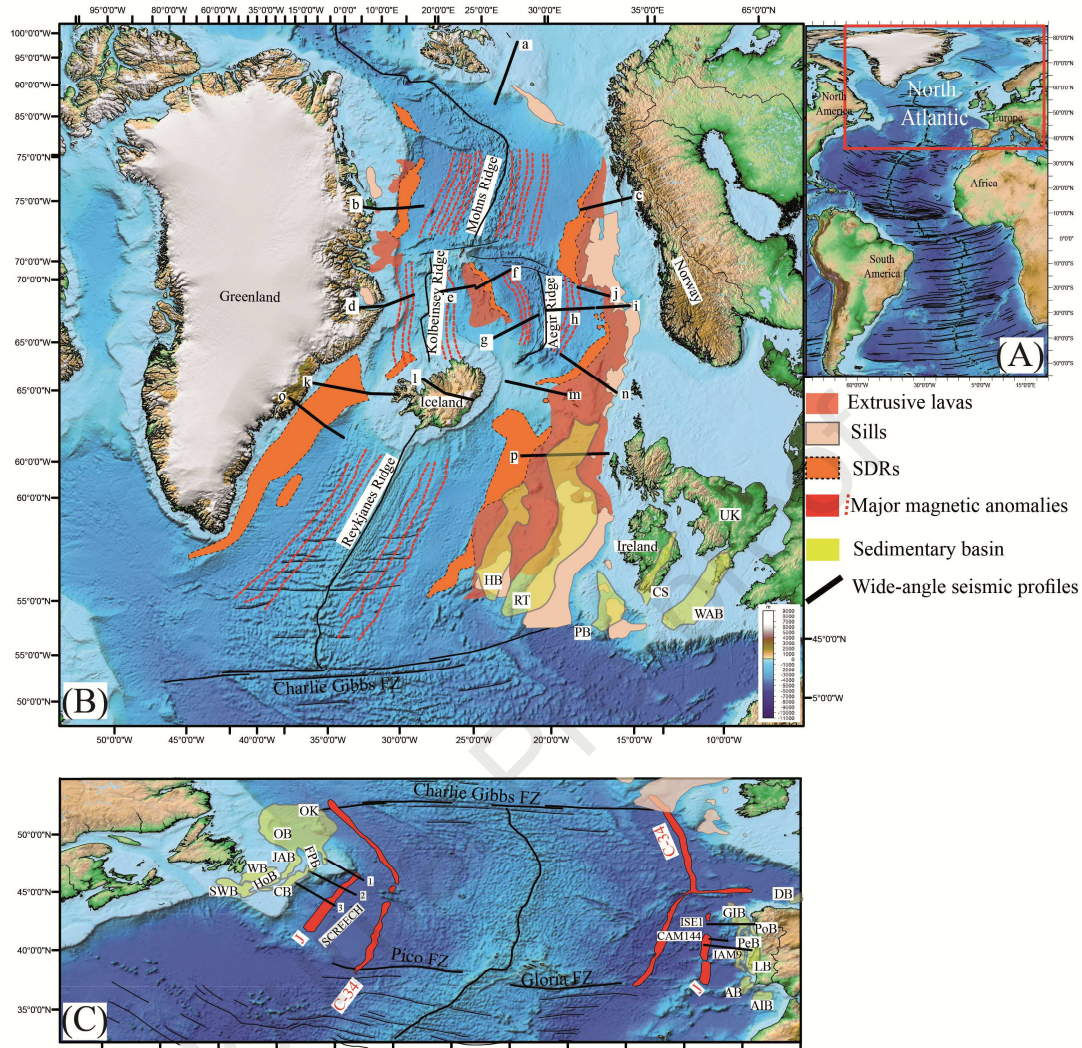


Figure 21: (A) Topographic and bathymetric map of the Atlantic Ocean from ETOPO1 bathymetry grids (Amante and Eakins, 2009). (B) and (C) Overview of the North Atlantic Ocean. SDR, extrusive lavas and sills are from Horni et al., 2014, 2017. Magnetic anomalies are from Bronner et al., (2011) and Nirrengarten et al., (2018). Sedimentary basins from Funck et al. (2014), Alves et al. (2009), Lundin and Doré (2011), and Tugend et al. (2014): HB (Hatton BASIN); RT (Rockall Trough); PB (Porcupine BASIN); CS (Celtic Sea); WAB (Western Approach BASIN); OK (Orphan Knoll); OB (OrphinBASIN); FPB (Flemish Pass BASIN); JAB (Jeanne d 'Arc BASIN); WB (Whale BASIN); HoB (Horseshoe BASIN); SWB (South Whale BASIN); HoB (Horseshoe BASIN); CB (Carson BASIN); DB (Le DanoisBASIN); PoB (Porto BASIN); PeB (PenicheBASIN); LB (Lusitanian BASIN); AB (AlentejoBASIN); ALB (Algarve BASIN). Black lines along the NE Atlantic margin represent wide-angle seismic profiles of the (a) Barents 98 line 3E (after Breivik et al. 2003) (b) AWI 20030200 (after Voss et al., 2009); (c) LVMT (after Tsikalas et al., 2005); (d) ARK 1988-3 (after Weigel et al. 1995); (e) JMKR 95-3 (after Kodaira et al., 1998a); (f) OBS2000-8 (after Mjelde et al., 2008); (g) KRISE 7 (after Brandsdottir et al., 2006); (h) OBS 2000-1 (after Breivik et al., 2006); (i) Vøring 96 Line 8A (after Raum, 2002); (j) Vøring 96 Line 8B (after

Raum et al., 2002); (k) SIGMA 1 (after Holbrook et al. 2001); (l) ICEMELT Line (after Darbyshire et al. 1998); (m) FIRE offshore (after Richardson et al. 1998) ; (n) iSIMM Faroe line (after Roberts et al. 2009); (o) SIGMA 2 (Korenaga et al. 2000); (p) AMP Line E (after Klingelhofer et al. 2005). Black lines along offshore West Iberia margin, represent wide-angle seismic profiles of the ISE1, CAM144, and IAM9 (Bullock and Minshull., 2005; Zelt et al., 2003; Chian et al., 1999; Dean et al., 2000) and wide-angle seismic profiles Screech1,2 and 3 along the Newfoundland margin (Funck et al., 2003; Van Avendonk et al., 2006; Lau et al., 2006).

4.1 Southern North Atlantic

Some studies consider the North Atlantic Ocean to be completely closed at Early Jurassic (e.g. Schettino and Turco, 2011; Vissers and Meijer, 2012; Barnett-Moore et al., 2016), while other studies indicate that rifting started already in the Late Triassic to Early Jurassic south of Iberia due to the north-westward propagation of CAO extension (e.g. Mohriak and Leroy., 2013). During this time, rifting is accompanied by only minor crustal thinning (Pérez-Gussinyé et al., 2006; Tucholke and Sibuet, 2007; Mohn et al., 2015 and references therein). Break-up started first in Late Jurassic-Early Cretaceous along the southern Iberia margin and then propagated northward to the Galicia Margin. During this phase, the extension led to the formation of a series of rift basins with typically asymmetric half grabens (e.g. Orphan and Rockall BASINS, Bay of Biscay, Jeanne d'Arc and Whale BASINS) filled by Triassic evaporites and Jurassic carbonates (Murillas et al., 1990; Enachescu, 2006; Tucholke et al., 2007; Tucholke and Sibuet, 2007; Pereira and Alves, 2011). Seafloor spreading started during the mid-Cretaceous (Aptian-Albian) (Tucholke and Sibuet, 2007). The full oceanic connection between the Central and North Atlantic occurred during Albian–Turonian times (Koutsoukos and Dias-Brito 1987; Koutsoukos et al. 1991).

In the southern North Atlantic, the conjugate Iberian and Newfoundland margins are magma-poor rifted margins, with no SDRs or other volcanic structures observed (Manatschal and Bernoulli 1998; Péron-Pinvidic and Manatschal, 2008) (Figures 21 and 22).

To the south, the J anomaly was imaged over exhumed mantle and thinned continental crust. It is a large and high amplitude magnetic anomaly (> 800 nT) that characterizes the Iberia–Newfoundland conjugate margin pair (Srivastava et al., 2000; Russell et al., 2003, Funck et al., 2003). These conjugate anomalies were used to identify the OCT zone. For Srivastava et al. (2000), the J anomaly represents the beginning of the M sequence of seafloor-spreading magnetic anomalies (M0–M3). The nature of the J-magnetic anomaly is debated. Many authors propose magmatic activity as the cause, including magmatic intrusions (Russell et al., 2003) or a pulse of magmatism that accompanied the final break-up of the Iberian–Newfoundland rift systems (Bronner et al., 2011). However, Sibuet et al. (2007) propose that here, exhumed and serpentinized subcontinental mantle material might form weak magnetic anomalies similar to oceanic anomalies, with an alignment of magnetite grains during the serpentinization process. Furthermore, it has been argued that the J anomaly might be polygenetic, i.e. originating from several magmatic events and not suitable for reconstructing

the opening and seafloor spreading (Bronner et al., 2011; Nirrengarten et al., 2018), an hypothesis which was contested by Tucholke and Sibuet (2012).

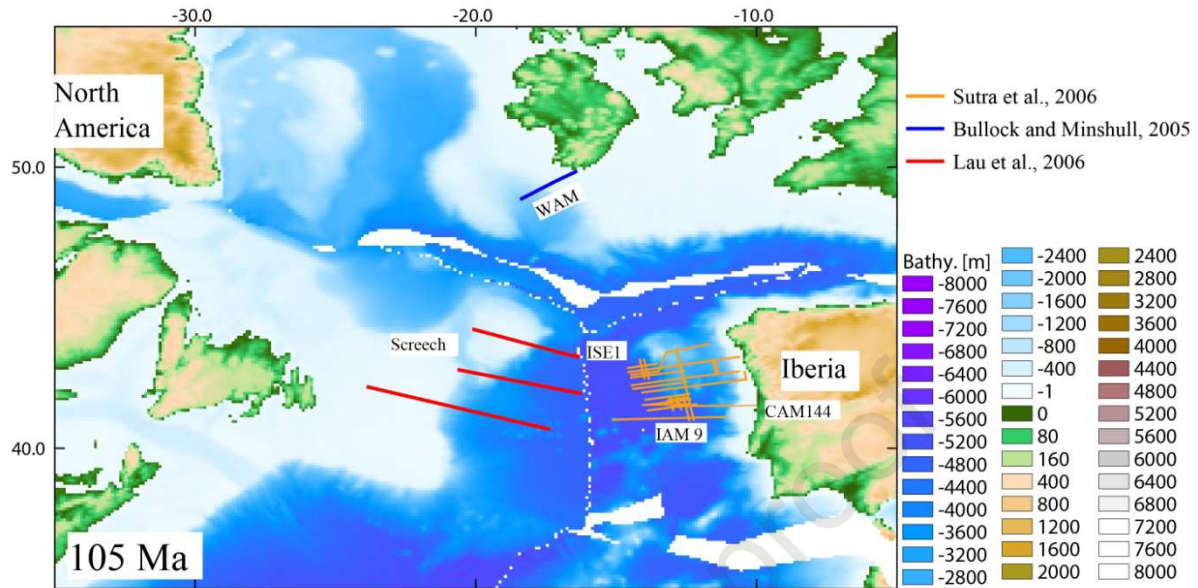


Figure 22: Reconstruction of the Southern North Atlantic at 105 Ma with rotational poles from Matthews et al. (2016). The fixed plate is North America. Locations of wide-angle seismic profiles are marked by solid lines.

The southern North Atlantic margins are among the best-studied margins on Earth with a wealth of marine geophysical surveys including reflection and wide-angle seismic data. In the scope of the IODP and ODP drilling program, several wells have been drilled (Boillot et al., 1988; Sawyer, 1994; Whitmarsh and Wallace, 2001; ODP Leg 173 Shipboard Scientific Party, 1998; Tucholke and Sibuet, 2007; Bayrakci et al., 2016). However, magnetic anomalies that are key for kinematic reconstructions are unclear in this region and do not necessarily have a conjugate or a precise age (Gillard et al., 2015; Causer et al., 2020). Based on a compilation of seismic and geochronological data, Nirrengarten et al. (2017) showed that the J anomaly, which was used for many kinematic models, does not correspond to an oceanic magnetic anomaly and therefore should not be used as such for kinematic restorations. As a result, the rift and early seafloor spreading evolution as well as the rotation poles remain still highly debated (Le Pichon and Sibuet, 1971; Olivet, 1996; Rosenbaum et al., 2002; Sibuet et al., 2004; Vissers and Meijer, 2012; Barnett-Moore et al., 2016). Taking these factors into account, the opening of the southern North Atlantic most likely started with an extension phase leading to the formation of half-graben type basins (Peron-Pinvidic and Manatschal, 2009), followed by increased thinning in Tithonian to Barremian times (Boillot et al., 1980). Numerous interpretations have been made for the break-up age. Based on structural and stratigraphic data, Wilson et al. (2001) propose an early Valanginian age (138 Ma). In contrast, Whitmarsh and Miles (1995) and Russell and Whitmarsh (2003) suggest a Barremian age (anomaly M5 to M3, ~130 Ma) based on the modeling of magnetic anomalies. Péron-Pinvidic et al. (2007) propose a break-up age of 112 Ma at the Aptian-Albian boundary. These differences in the interpretation of the break-up age can originate

from the fact that serpentinized upper mantle material with magmatic additions can form regions characterized by a magnetic anomaly pattern similar to the pattern over true oceanic crust (Bronner et al., 2011; Causer et al., 2020).

In terms of deep crustal structure, the Newfoundland margin was studied as part of the SCREECH seismic experiment. Three wide-angle seismicrefractionprofiles were acquired across the margins of the Grand Banks of Newfoundland and Flemish Cap in the direction of extension, with the objective to image the deep structure of the margin from the continental platform onto the abyssal plain. The conjugate Iberian margin was studied by lines ISE 1 (Zelt et al.; 2003), CAM 144 (Chian et al., 1999), and IAM9 (Dean et al., 2000, Davy et al., 2016). We divide the Iberian – Newfoundland conjugate rifted margins into northern, central and southern segments.

4.1.1 Northern segment

SCREECH Line 1 is a 320-km-long profile acquired on the southeastern Flemish Cap continental margin (Figure 23). Along this line, Funck et al. (2003) found a 30-km-thick continental crust, with velocities ranging from 6.0 km/s to 6.7 km/s. This continental crust thins to just 1.5 km, over a distance of 90 km. A 3km-thick oceanic crust that thins locally to 1 km is imaged with velocities of 4.7 km/s and 6.9 km/s. Over an 80-km-wide zone, a layer with velocities of 7.6 km/s and 8.0 km/s underlying both the thinned continental crust and the thin oceanic crust is imaged and interpreted as partially serpentinized mantle with velocities between 7.6 km/s and 8.0 km/s (Figure 20). Adjacent to the zone with serpentinized mantle, a 6-km-thick oceanic crust is encountered.

The western margin of Galicia Bank was imaged along the ISE 1 line (Whitmarsh et al., 1996). The continental crust at Galicia Bank is divided into only two layers with velocities between 5.2 km/s and 6.9 km/s. The velocity model shows a 60km-wide zone of serpentinized mantle with velocities between 7.2 km/s and 7.6 km/s underlying both thin continental and thin oceanic crust. A peridotite ridge exposed at the seafloor was interpreted as the border of the margin that separates continental from oceanic crust. It has been modeled with velocities of 3.5 km/s. A 7km-thick oceanic crust displays velocities from 4.6 km/s to 4.8 km/s at the top of layer 2 and 7.0 km/s in layer 3. Where the oceanic crust overlies serpentinized mantle, the thickness is reduced from 2.5 to 3.5 km (Figure 23).

Comparison between the two conjugate profiles suggests symmetric oceanic crustal accretion as both profiles show the same velocities and thickness for the oceanic part overlying the serpentinized mantle. However, the continental crust at Galicia Bank is divided into only two layers with velocities between 5.2 km/s and 6.9 km/s, whereas the Flemish Cap crust exhibits three layers with velocities ranging from 6.0 km/s to 6.7 km/s. Funck et al. (2003) relate this to differences in the internal structure or composition. In addition, the crustal thinning at Galicia Bank extends over a wider zone compared to the conjugate margin surveyed with SCREECH line 1. Some basement highs imaged along the West Iberia margin were, by comparison with the drilled peridotite ridge, proposed to also consist mainly of exhumed

continental serpentinite (Dean et al., 2015). These ridges are similar to oceanic core complexes consisting of exhumed mantle material sampled at the Mid-Atlantic Ridge, implying the existence of a gradual transition between continental extension, accretion of amagmatic seafloor and magmatic spreading (Jagoutz et al., 2007; Dean et al., 2015). The first magmatically accreted oceanic crust located 20 km from the peridotite ridge, has a thickness of only 0.5 km (Davy et al., 2016).

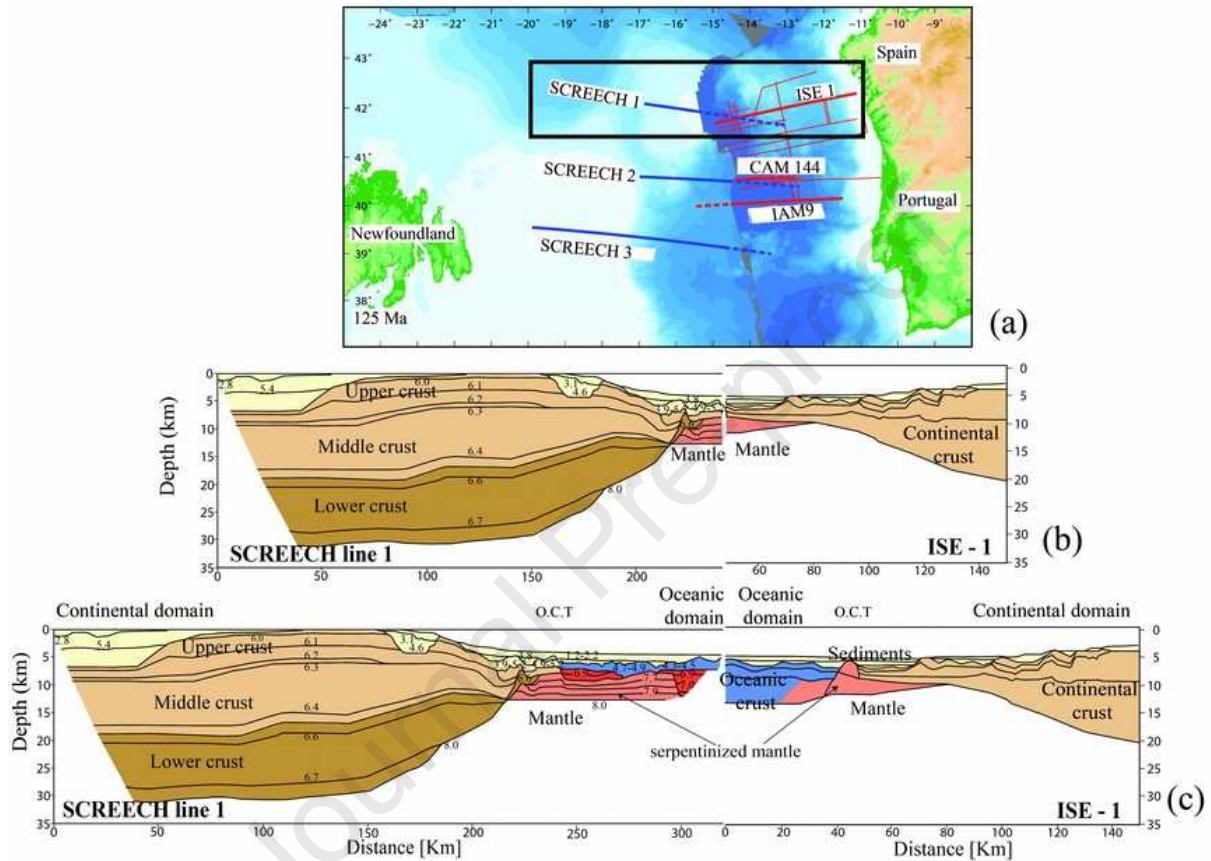


Figure 23: Deep structure of the Newfoundland and Iberian margin (after Funck et al., 2003 and Whitmarsh et al., 1996). (a) Reconstruction of the Southern North Atlantic at 125 Ma (b) reconstructed for the initiation of seafloor spreading and (c) at the time of magnetic anomaly $M0$ (118 Ma).

4.1.2 Central segment

The 368-km-long seismicrefractionline SCREECH 2 (Van Avendonk et al., 2006; 2009) extends from the eastern Grand Banks into the NewfoundlandBASIN (Figure 24). The tomographic inversion indicates that the continental crust has a maximum thickness of 27 km, with velocities varying from 5.5 km/s to 6.5 km/s. The crustal thickness thins to 6 km beneath the continental slope of Grand Banks and coincides with anomalously-high seismic velocities ($7.0\text{--}7.2\text{ km}\cdot\text{s}^{-1}$) in the lower crust of the shelf edge. Further seaward, a 25-km-wide zone is interpreted to consist of exhumed and serpentinized continental mantle with velocities from 6.3 km/s to 7.7 km/s. Unroofed mantle and very thin oceanic crust were imaged seaward, with

seismic velocities ranging from 5.0 km/s to 7.0 km/s. The mantle layer was modeled with velocities between 7.8 km/s and 8.2 km/s (Figure 24).

At the conjugate Iberia Abyssal Plain south of Galicia Bank, the seismic velocity structure along line CAM 144 shows 2-5 km thick upper continental crust with several fault blocks and velocities of 5.0 km/s to 6.6 km/s. This crust is underlain by a high-velocity layer (7.3 km/s to 7.9 km/s) of poorly serpentinized peridotite (0-25%). There was no evidence for the existence of a basement with characteristics of oceanic crust along this profile (Chian et al., 1999).

Thick continental crust is not modeled on either conjugate profiles, however, thin continental crust is imaged on both sides of the rift with a different observed nature (Van Avendonk et al., 2006; 2009). Exhumed mantle that was serpentinized to a great depth is exposed on both margins. No prerift sediments were imaged within on the transition zone along the SCREECH line 2, which lead Van Avendonk et al., 2006 to suggest that the Newfoundland BASIN was formed by exhumation of midcrustal or lower crustal rocks. However, on the conjugate margin along Cam144, continental crust blocks appear to be cut by seaward-dipping normal faults. Therefore, compared to the transition zone of SCREECH line 2, the basement relief along Cam144 is higher and the continental crust is thinner (Chian et al., 1999; Van Avendonk et al., 2006)

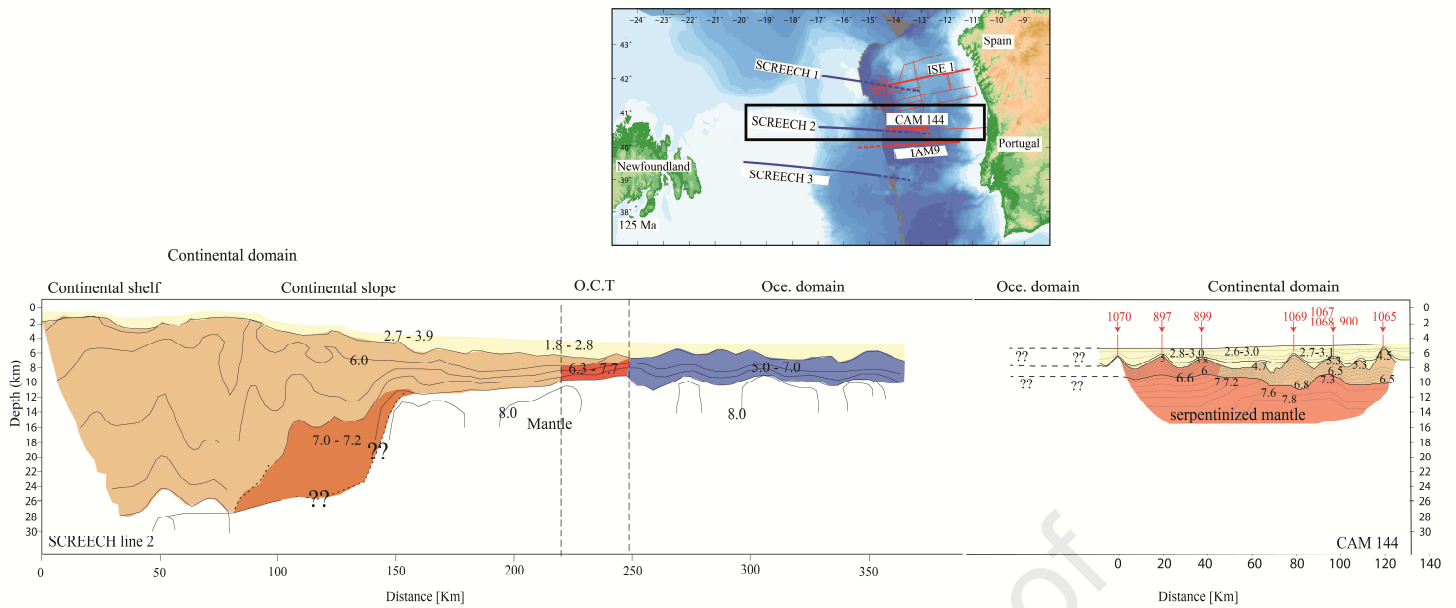


Figure 24: Deep structure of the Newfoundland and Iberian margin (after van Avendonk et al., 2006; 2009 and Chian et al., 1999).

4.1.3 Southern segment

SCREECH line 3 is 565 km long and is the southernmost profile acquired across the Grand Banks-Newfoundland BASIN (Lau et al., 2006) (Figure 25). The velocity model of this part of the margin shows a 36km-thick continental crust with P wave velocities of 5.8 km/s to 6.9 km/s. Crustal thinning occurs over a 170km-wide zone and is characterized by the presence of several large tilted blocks with a width of ~20 km that decreases oceanward. An 80km-wide zone of transitional crust is imaged, with a thickness of 5 to 6 km and velocities ranging from 4.4 to 7.8 km/s and interpreted as exhumed serpentinized mantle. A 200-km-wide zone with reduced mantle velocities (7.6 km/s to 7.9 km/s) is observed under the transitional zone and thin continental crust and interpreted as partially (<10 percent) serpentinized mantle. Oceanwards, a 2 to 6km-thick oceanic crust is imaged and interpreted as a result of slow seafloor spreading (Lau et al., 2006) (Figure 25).

The conjugate line IAM 9 is 320 km long and was acquired in the southern Iberian Abyssal Plain (Dean et al., 2000). The velocity model shows a 28km-thick continental crust that thins to 7 km over an 80-km-wide zone and is characterized by the presence of several large tilted blocks. A 170-km-wide zone of transitional crust is imaged with a thickness of 4.5 km and velocities ranging from 7.3 km/s to 7.9 km/s. This zone is interpreted as a peridotite ridge and has been directly sampled (Dean et al., 2000).

Even if mantle exhumation and serpentinization appears to have occurred on both margins before oceanic crust was formed, comparison between the two conjugate profiles suggests asymmetrical margins. This strong asymmetry shows that continental break-up was concentrated on the Iberian side and a wider zone of extended continental crust has been left on the Newfoundland side. Lau et al., (2006) suggest that such significant asymmetry may be related to the large width (270 km) of the rift. Moreover, the S reflector, a reflector

which slopes slightly towards the continent, is not observed on the Newfoundland margin, which contradicts Wernicke's simple shear model (Wernicke, 1985).

Figure 25: Deep structure of the Newfoundland and Iberian margins (after Lau et al., 2006 and Dean et al., 2000). (a) Reconstruction of the Southern North Atlantic at 125 Ma (b) reconstructed for the initiation of seafloor spreading and (c) at accretion of oceanic crust.

4.2 Northeast Atlantic

During the Cretaceous, EW extension between the North American and Eurasian plates led to rifting and break-up in the Labrador Sea and Baffin Bay. The start of seafloor spreading is disputed but lies between magnetic Chron C31 (69 Ma, e.g. Roest and Srivastava, 1989; Keen et al., 2018) and C27 (62 Ma, e.g. Oakey and Chalmers, 2012; Hosseinpour et al., 2013) and

ceased before Chron C13 (33 Ma, Chalmers and Pulvertaft, 2001; Delescluse et al., 2015; Peace et al., 2016). East of Greenland, break-up occurred at 55.5 Ma in the Early Eocene (e.g. Storey et al. 2007, Seton et al. 2012; Gaina et al., 2017) resulting in simultaneous oceanic crustal accretion east and west of Greenland until the spreading ceased in the Labrador Sea and Baffin Bay.

The NE Atlantic is bounded by the Senja Fracture Zone in the north and the Charlie Gibbs Fracture Zone in the south. It opened after several rifting phases spanning more than 300 Ma, from the Late Palaeozoic to the Mesozoic/Early Cenozoic (e.g. Fossen, 2010 and references therein). However, it is important to note that in the NE Atlantic there were multiple phases of rifting in contrast to e.g. the Newfoundland-Iberia margin that saw multiphase rifting (Reston, 2007; Peron-Pinvindic et al., 2013).

The North Atlantic Igneous Province (NAIP) formed prior to, during and after break-up (Coffin and Eldholm, 1994; Saunders et al., 1997; A Horni et al., 2017). Basaltic rocks of this large igneous province extend over a surface area of at least 1.3×10^6 km² (Eldholm and Grue, 1994). Two main periods of magmatism have been proposed. The first at 62–57 Ma and a second major pulse of magmatism at c. 56–54 Ma (Saunders et al., 1997; Storey et al., 2007). However, Wilkinson et al. (2017) propose that there was no time gap in volcanism.

The associated volcanic activity was extensive and produced thick seaward dipping reflector (SDR) sequences along the continental margins (Parkin et al., 2008). The highest amount of volcanic/magmatic products is recorded in the vicinity of the Greenland-Iceland-Faroe Ridge (GIFR) (Holbrook et al., 2001). The transform margin in the north between Svalbard/SW Barent Sea and NE Greenland (Ritzman and Jokat, 2003; Ritzmann et al., 2002), is unaffected by the magmatism. After break-up east of Greenland, the Aegir Ridge spreading center and the proto-Reykjanes Ridge propagated through the North Atlantic Craton (Figure 21). Several models have been proposed to explain the shape of the NAIP, the volcanic Yermak PLATEAU and Morris Jesup Rise (e.g. White and McKenzie, 1989; Jones and White, 2003). Some continental fragments, such as the Jan Mayen Microplate north of Iceland were left behind after rifting and break-up (Gernigon et al., 2012). This continental fragment was the result of the separation of a part of Greenland through the Kolbeinsey Ridge after spreading ceased at the Aegir Ridge (Gaina et al., 2009). Many studies associate the 35 to 40-km-thick igneous crust along the Greenland-Iceland-Faroe Ridge (GIFR) with the magmatic activity of an Iceland hotspot or plume (Staples et al., 1997; Darbyshire et al., 1998). In a study involving receiver functions and numerical modeling, Foulger et al. (2019) propose that part of the lower crust of Iceland and the GIFR may contain heavily intruded continental crust (Foulger et al., 2019). Similarly, Torsvik et al. (2015) suggested the

existence of an up to 70 km wide zone of continental crust beneath southeastern Iceland based on isotope data.

Between Greenland and NW Europe, the NE Atlantic Ocean contains patches of ocean crust with a series of symmetric magnetic anomalies parallel to the oceanic ridges. Chron 24 (52.9 Ma) is the oldest magnetic anomaly identified in the NE Atlantic Ocean. However, the origin of some of these magnetic anomalies remains disputed. For example, anomalies between the Jan Mayen Ridge and the Vøring Marginal High originally thought to be part of the 24 chron system (Skogseid and Eldholm, 1987) were later interpreted to result from recent magmatic activity on Jan Mayen Island rather than by seafloor spreading, as indicated by aeromagnetic data and geological sampling (Gernigon et al., 2009; Torsvik et al., 2015).

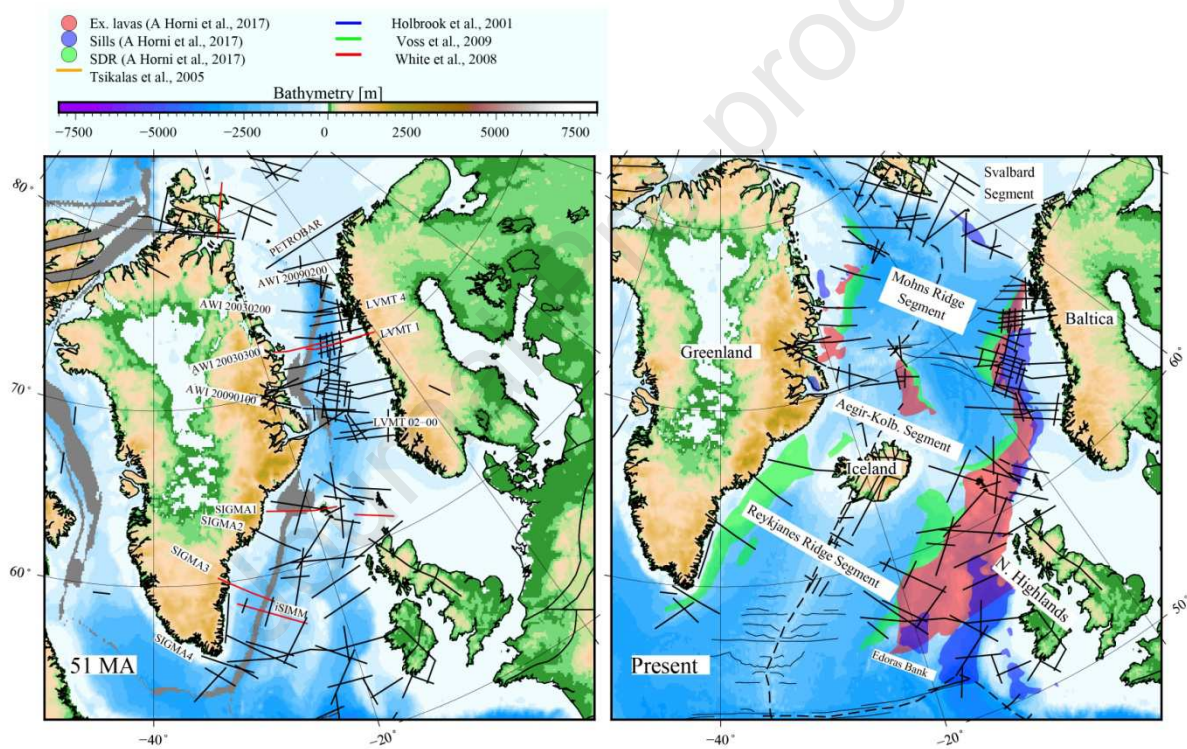


Figure 26: (a) Reconstruction of the North Atlantic at 51 Ma with rotational poles from Matthews et al. (2016). The fixed plate is North America. Locations of wide-angle seismic profiles are marked by solid lines. (b) Present day bathymetry with extruded lavas (red), sills (blue) and SDR (green) from Á Horni et al. (2014; 2017).

The NE Atlantic can be subdivided into four regions: the Svalbard transform zone, the northern region around the Mohs Ridge, the central region around the Aegir and Kolbeinsey ridges together with the Greenland-Iceland-Faroe Ridge, and the southern region around Reykjanes Ridge segment. A synthesis of existing wide-angle seismic lines in the NE Atlantic is given in Funck et al. (2017a; b).

4.2.1 Svalbard transform segment

Based on wide-angle seismic data of Breivik et al. (2003), the northern transform margin west of Svalbard has a narrow continent-ocean transition zone typical of transform margins, thinning from 25 km to 10 km thickness over a distance of only 50 km (Figure 27). Two transform marginal plateaus are located along this margin, the Morris Jesup Rise and the YermakPLATEAU at the border between the NE Atlantic and Arctic Ocean (Loncke et al., 2020). However, the crustal nature of both plateaus is still debated to be either continental slivers overprinted by volcanism or Eocene oceanic plateaus (Jackson et al., 1984; Engen et al., 2008; Faleide et al., 2008). The YermakPLATEAU experienced volcanic influence during its construction, and it might have recently been rejuvenated as indicated by high heat-flow values (Okay and Crane, 1992). Using seismic reflection data, Berglar et al. (2016) identified rotated fault blocks at the NE YermakPLATEAU margin, bounded by major listric normal faults, and half-grabens filled with syn-rift sediments. Based on these observations, Berglar et al. (2016) interpret this region as a rifted continental margin, implying that the NE YermakPLATEAU is of continental origin. While part of the Yermak PLATEAU might be continental, the northern section and the Morris Jesup Rise may have formed a continuous volcanic massif when the propagating rift arrived at the Spitsbergen or de Geer fault zone (Okay and Crane, 1992).

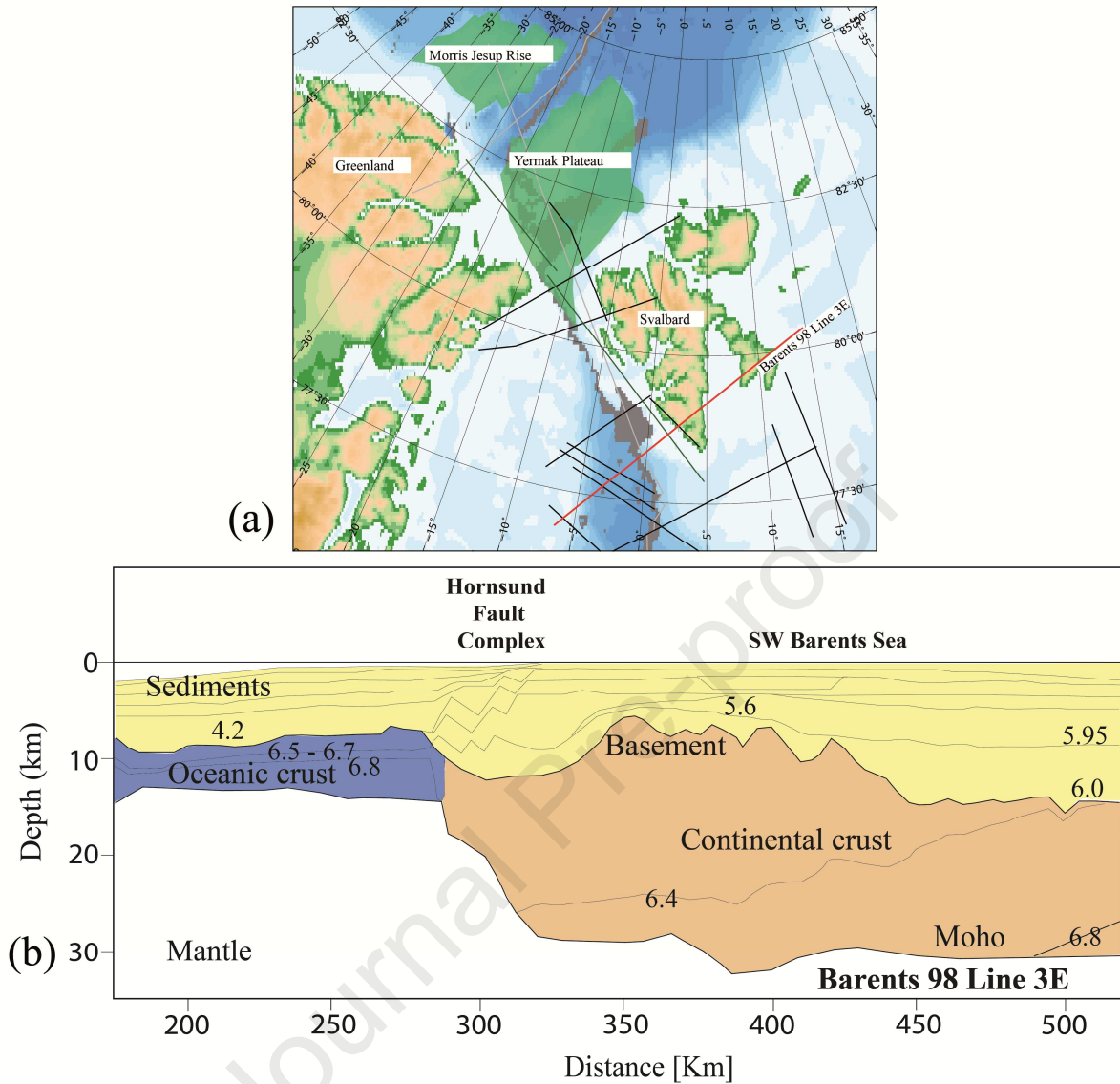


Figure 27: Deep structure of the SW Barents Sea margin along profile Barents 98-line 3E (after Breivik et al. 2003).

4.2.2 The northern segment around Mohns Ridge

The Greenland side of this segment has been studied through four wide-angle seismic profiles: AWI-20030200 to AWI20030500 (Figures 21 & 28) (Schlindwein et al., 1999a and 1999b; Voss and Jokat, 2007; Voss et al., 2009). The northernmost profile (AWI-20030200) samples a margin with only small amounts of magma and an oceanic crust thinning over a very short distance (120 km) until magnetic anomaly C21 (47.1 Ma). The southern seismic transect (AWI-20030500) is associated with a sedimentary basin of up to 7 km depth beneath flood basalts (Voss et al., 2009). The continental region of about 30 km thickness shows volcanic edifices and some magmatic underplating landward of the COB magmatic underplate. Initial oceanic crustal thickness decreases from 9–13 km in the south to 5 km in

the north (Voss et al., 2009). The thickness of the HVLC increases towards the south up to a maximum of 15 km. In comparison, along the conjugate profiles LVMT4 and LVMT1 offshore the Lofoten, the oceanic crust is 1-2 km thicker while continental crustal thickness is similar (Tsikalas et al., 2005) (Figure 28). Along both margins a HVLC extends from the C23 anomaly to underneath the continent, with its maximum thickness of 5-8 km at the COT (Voss et al., 2009; Tsikalas et al., 2005). Along this part of the Greenland margin, break-up related SDR packages decrease in volume both northward and southward from a central point around 75°N (Franke et al., 2019).

The north-south differences in the distribution of the HVLC along the basin are proposed to result from either a transfer zone that acted as a barrier to northward magma flow, or a second magmatic event associated with the separation of the Jan Mayen microcontinent from Greenland (Voss et al., 2009).

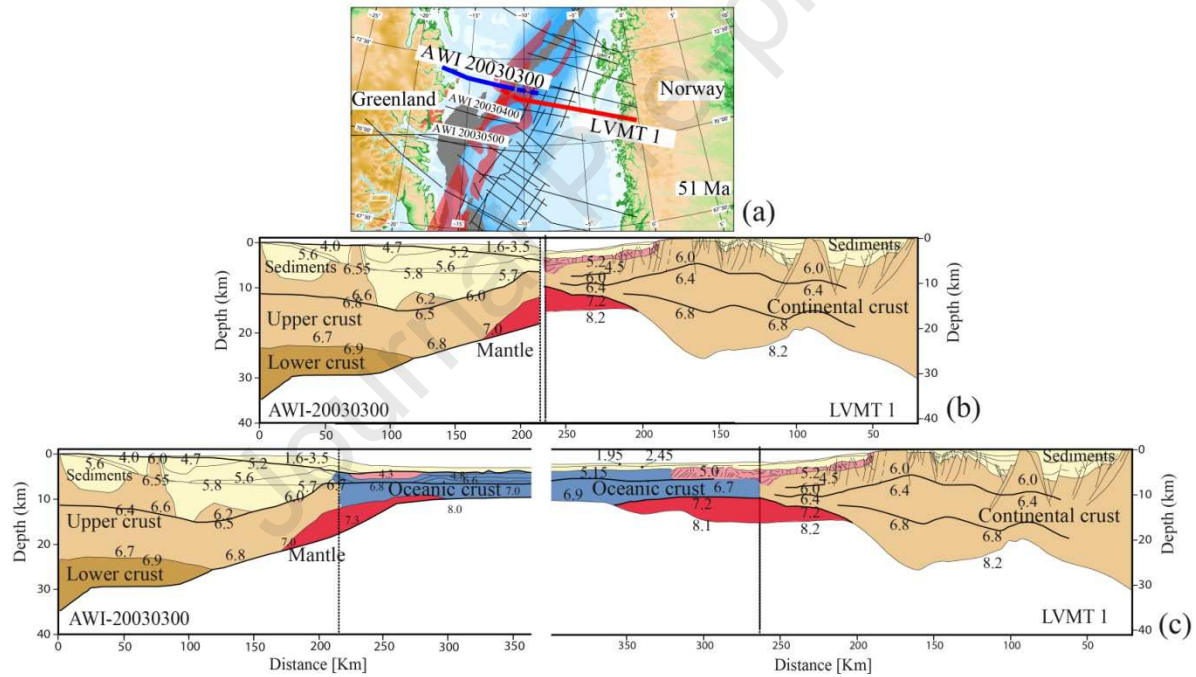


Figure 28: Deep structure of the Norwegian and Greenland continental margins (after Korenaga et al., 2000; Holbrook et al., 2001; Voss et al., 2009; Tsikalas et al., 2005). (a) Reconstruction of the North Atlantic at 51 Ma (b) reconstructed for the initiation of seafloor spreading and (c) at accretion of oceanic crust.

The Vøring Transform Marginal PLATEAU is situated north of the East Jan Mayen FZ (Loncke et al., 2020). Wide-angle seismic data indicate a magmatically overprinted continental crust for the plateau (Mjelde et al., 1997, Berndt et al., 2000). A high velocity layer interpreted to be mafic intrusions/underplate is underlying the plateau with a highly

variable thickness of up to 8 km (Mjelde et al. 1997). The flexed continental crust is dilated by numerous feeder-dikes of the thick overlying SDR sequences (Abdelmalak et al., 2015). The conjugate of the VøringPLATEAU would be situated at the NE-Greenland margin (Loncke et al., 2019).

4.2.3 The central segment around the Aegir and Kolbeinsey ridges and the Greenland-Iceland-Faroe Ridge

This segment was the object of several wide-angle seismic surveys mostly concentrating on the origin of the Icelandic magmatic province and the Jan Mayen Microcontinent (e.g. ARK - Weigel et al., 1995, Vøring 96 - Raum, 2000; FIRE - Richardson et al., 1998; iSIMM Faroe line - Roberts et al., 2009) (Figure 29 and 30). North of the Greenland-Iceland Faroe Ridge (GIFR), the ARK 1988-3 line (Weigel et al., 1995) shows 35 km thick continental crust beneath the Greenland margin (Figure 30). Still onshore, the 15km-deep Jameson Land BASIN is crossed with the basement high forming Liverpool Land to the east of it. From there, the continental crust thins seaward and is underlain by a HVLC that Weigel et al. (1995) propose to result from magmatism. The adjacent oceanic crust is 6-7 km thick (Weigel et al., 1995). Conjugate to this line is profile JKMR-95 L3 crossing the western margin of the Jan Mayen Microcontinent (Figure 29). The oceanic crust on this line is 9-10 km thick in the west but thins to 5 km towards the Jan Mayen Microcontinent (Kodaira et al., 1998) (Figure 30). Line JKMR-95L3 connects eastward with OBS 2000 Line 8, showing a 17km-thick crust beneath the Jan Mayen Microcontinent (Mjelde et al., 2008). There is no evidence for a HVLC, which is why Kodaira et al. (1998) proposed the western margin of the Jan Mayen Microcontinent to be non-volcanic. The exact outline of the Jan Mayen Microcontinent is still disputed (see the various proposed continent ocean boundaries in Funck et al., 2014), it extends from south of the island of Jan Mayen to somewhere north of Iceland (Gaina et al., 2009). The microcontinent has a crustal thickness of about 17 km and elevated lower crustal velocities and basaltic flows on its eastern side (Kodaira et al., 1998). Further to the east, profile Vøring-96 8B images 7-8 km-thick oceanic crust (Raum et al., 2002). In the south, KRISE line 7 is located at the continent-ocean transition zone of the Jan Mayen Microcontinent (Brandsdottir et al., 2006) linking with OBS 2000 Line 1 and in its extension Vøring-96 8A (Breivik et al., 2006; Raum., 2000; Raum et al., 2002) (Figure 30) at the mid-Norwegian continental margin. This 160km-wide margin is underlain by a HVLC of up to 4 km thickness.

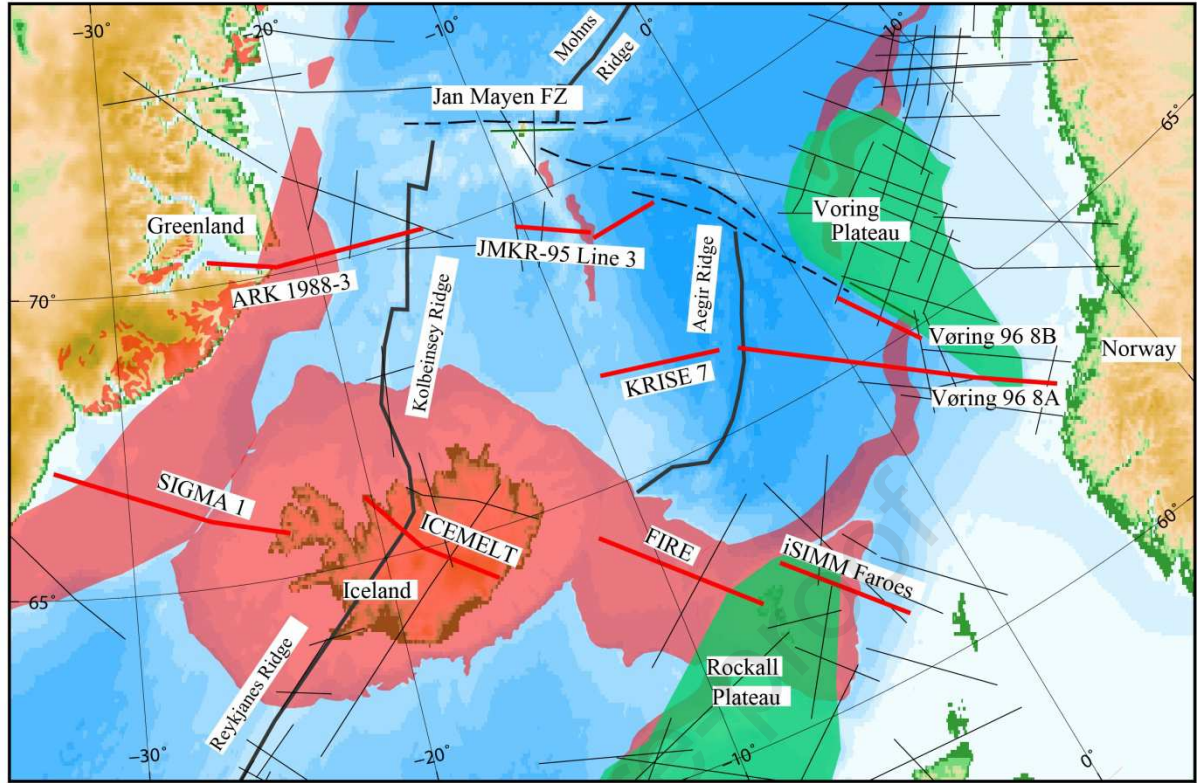


Figure 29: Location map of the wide-angle seismic profiles shown in Figures 30 and 31. Orange transparent areas represent the location of volcanic provinces (Whittaker et al., 2015) and green transparent areas represent transform marginal plateaus (Loncke et al., 2020).

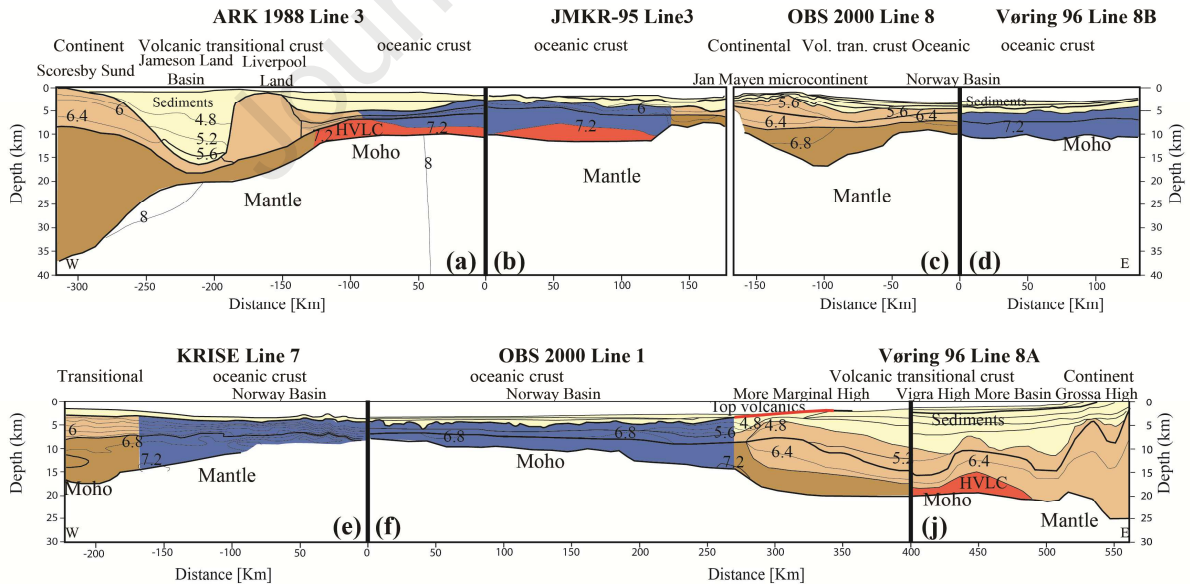


Figure 30: P-wave velocity models for conjugate profiles in the Jan Mayen region. (a) ARK 1988 Line 3 (after Weigel et al. 1995); (b) JMCR-95 Line 3 (after Kodaira et al., 1998a); (c) OBS 2000 Line 8 (after Mjelde et al., 2008); (d) Voring 96 Line 8B (after Raum, 2000); (e) KRISE Line 7 (after Brandsdottir et al., 2006); (f) OBS 2000 Line 1 (after Breivik et al., 2006); (j) Voring 96 Line 8A (after Raum, 2000; Raum et al., 2002).

A second transect chosen in this margin segment is located along the Greenland-Iceland-Faroe Ridge including Iceland itself (Figures 29 and 31). Located at the Greenland margin, SIGMA Line 1 extends onto the Greenland-Iceland Ridge and shows a crust exceeding 30 km in thickness with a HVLC with velocities up to 7.5 km/s extending at least for 150 km from the coast and possibly even farther towards Iceland (Holbrook et al., 2001). The COT zone is probably located landward of the western end of the profile. Spanning Iceland, the ICEMELT line indicates that the crustal thickness increases from 25 km in the north to 38-40 km in the southern central part (Darbyshire et al., 1998). In the young volcanic regions, the upper crust is only 5 km thick, but ridge jumps may have doubled the upper crustal layer thickness locally (Darbyshire et al., 1998). Lower crustal velocities, between 7.1 km/s and 7.35 km/s are higher than in normal oceanic crust, which can be explained by elevated temperatures in the mantle. East of Iceland, the transition between the continental Faroe block and the thickened oceanic crust of the Faroe-Iceland Ridge was mapped by the FIRE profile (Richardson et al., 1998). The lower part of the crust beneath the Faroe Islands exhibits high velocities that may relate to magmatic underplating or mafic intrusions (Richardson et al., 1998). To the north, the iSIMM Faroes profile was acquired with a very dense seismometer spacing of 2 to 5 km and was able constrain a narrow COT zone of only 50 km width and large amounts of intrusives and extrusives within the transition zone (Roberts et al., 2009).

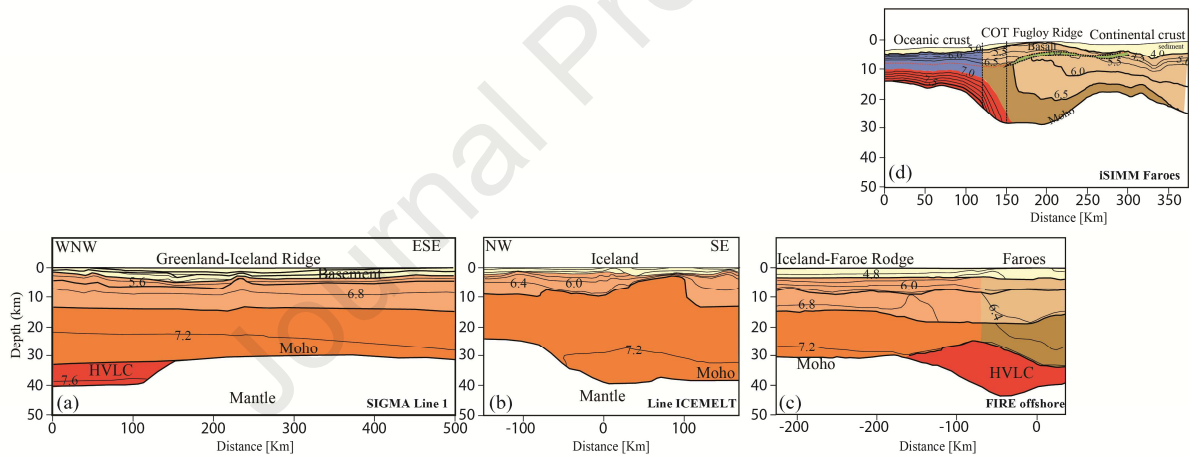


Figure 31: P-wave velocity models along the Greenland-Iceland-Faroe Ridge and the Faroe-Shetland BASIN (FSB). (a) SIGMA line 1 (after Holbrook et al. 2001); (b) ICEMELT line (after Darbyshire et al. 1998); (c) FIRE offshore line (after Richardson et al. 1998); (d) iSIMM Faroe line (after Roberts et al. 2009). Orange colors depict Icelandic type crust including a high degree of volcanic material possibly mixed with continental crustal fragments.

4.2.4 The southern segment along the Reykjanes Ridge

This segment was imaged by three wide-angle seismic profiles of the SIGMA experiment (Korenaga et al., 2000; Holbrook et al., 2001; Hopper et al., 2003) showing the magma-rich nature of the margin (Figure 32). SIGMA 2 is located closest to the Iceland Hot Spot and shows 27-km-thick crust at the continent-ocean boundary (COB), from where the oceanic crust thins to about 10 km over a distance of 150 km (Holbrook et al., 2001). The profile SIGMA 3 is located further away from Iceland showing an initial oceanic crustal thickness of 16 km at the COB thinning gradually to 8 km over a distance of 170 km. The southernmost profile SIGMA 4 also has an initial oceanic crustal thickness of 16 km but here the thinning to 8-9 km occurs over only 75 km, reflecting the increasing distance to the Iceland plume (Holbrook et al. 2001).

The conjugate margin of the Faroe-Hatton-Rockall PLATEAU was sampled along several profiles (e. g. Klingelhoefer et al., 2005; Parkin and White, 2008; Vogt et al. 1998; Shannon et al. 1999). Figure 32 shows a conjugate reconstruction for SIGMA line 3 (Hopper et al. 2003) and the iSIMM Hatton dip line (White and Smith 2009). Although the original crustal thickness might have been over 50 km, the crustal thickness at the landward limit of the profiles is about 32 km on the Greenland side and only 24 km on the Hatton Bank side, probably due to previous rifting events (Ziegler 1988; Doré et al. 1999; Funck et al., 2017). Spreading rates along the margins are asymmetric in the oceanic crust, where only 65 km of crust was accreted between Chron C22 and the COB on the Hatton margin and up to 140 km during the same period at the SE Greenland margin, when using the interpretation of Funck et al. (2014) for the position of the COB. The high amount of flood basalts and thick oceanic crustal thickness of up to 18 km can be explained by either high mantle temperature potential (White and McKenzie, 1989) or the pre-break-up rifting in the neighboring Hatton BASIN (Armitage et al., 2010). In that case, hot mantle material from upwelling during the previous rifting event can spread laterally to already slightly extended lithosphere before the break-up (Armitage et al., 2010).

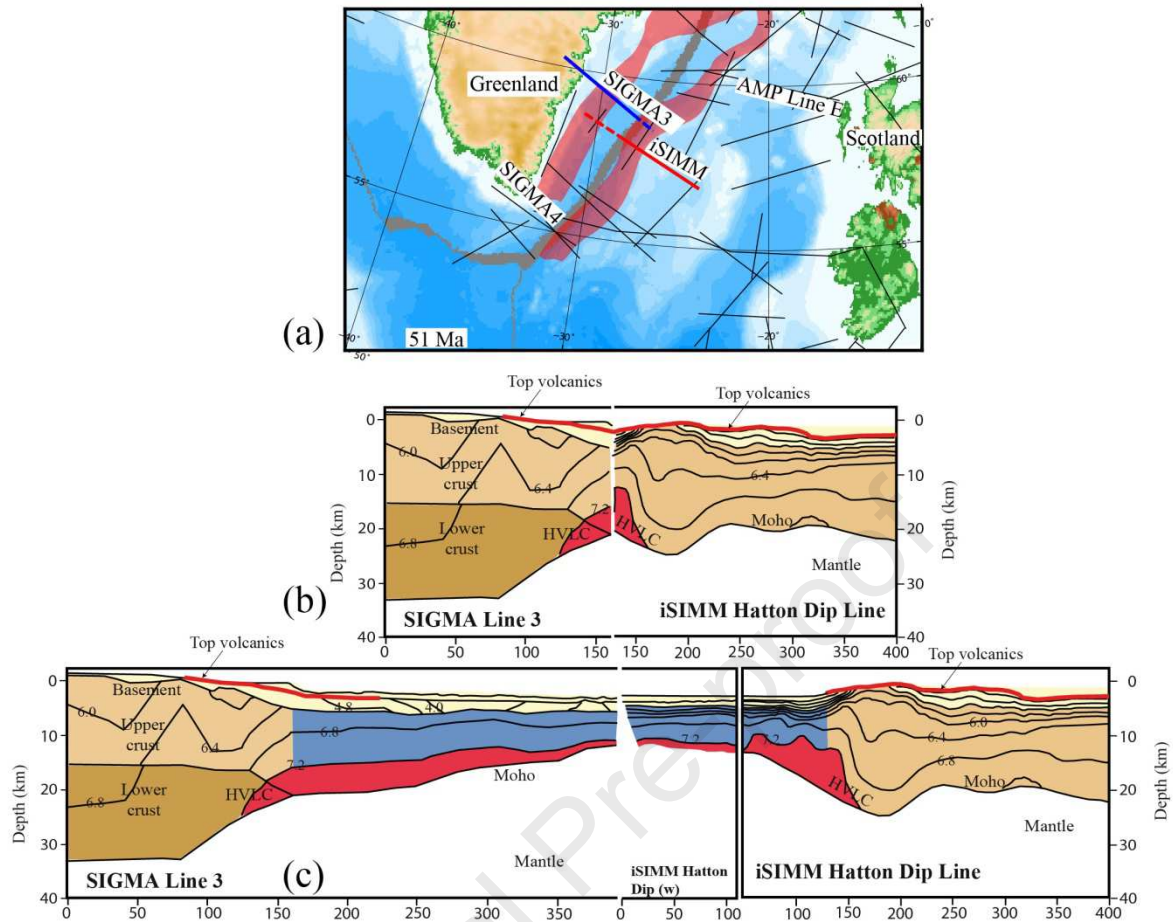


Figure 32: Deep structure of the Greenland margin and the Rockall Trough: SIGMA Line 3 after Hopper et al. (2003), iSIMM Hatton dip line (W) after Parkin and White (2008) and iSIMM Hatton dip line after White and Smith (2009). (a) Reconstruction of the North Atlantic at 51 Ma (b) reconstructed for the initiation of seafloor spreading and (c) at accretion of oceanic crust.

5 Discussion

This review of main conjugate deep seismic profiles along the Atlantic margins allows for discussion of the unresolved questions regarding rifting and continental margin formation identified in the introduction. As a summary and base for the discussion, the original crustal thickness, oceanic crustal thickness, and approximate beta factors are provided in Table 2.

1) What are the factors leading to the development of either volcanic or magma-poor margins and during which phase?

The continental margins of the Atlantic cover the full range of passive margins from highly volcanic to magma-poor margin types (Figure 33). Continental margins along the volcanic parts show a distinctively different structure to non-volcanic ones. First of all, they tend to be

narrower, e.g. the Argentine volcanic rifted margin reveals a scarcity of rift basins and abrupt thinning of the continental crust (Franke et al., 2007; Schnabel et al., 2008; Becker et al., 2014). Except for the locations of rift-perpendicular basins (e.g. Colorado, Salado), there is no evidence of significant crustal thinning prior to break-up along most of the Argentine margin. In our compilation the magma-poor Galicia margin and the Flemish Cap and Nova Scotia margins are wider (e. g. Dean et al., 2000; Funck et al., 2004). Second, zones of upper mantle exhumed at the seafloor and serpentized by water intruding through cracks and faults have been discovered so far only at magma-poor margins. These zones are characterized by seismic velocities higher than lower crustal velocities and lower than un-serpentized mantle. Deep reflections observed in wide-angle seismic data might originate from a serpentization front rather than from the Moho (Minshull et al., 1998; Minshull, 2009). These zones are clearly imaged along the margins of the southern North Atlantic (Dean et al., 2015) and northern parts of the American margin of the Central Atlantic (Funck et al., 2004; Wu et al., 2006). None of the wide-angle seismic models from volcanic margins along the Atlantic Ocean revisited in this study show indications for serpentized mantle material between the extrusive SDR sequences and HVLC bodies (e.g. Bauer et al., 2000; Korenaga et al., 2000; Planert et al., 2017).

The first oceanic crust accreted along magma-poor margins can be unusually thin, as for example at the Nova Scotia margin (Funck et al., 2004) and correspondingly it could be expected that oceanic crust accreted at volcanic margins is overthickened. However, this is complicated as layers of magmatic underplate might be present underneath the crust, as for example along the US continental margin (LASE study group, 1984). Our comparison shows thick oceanic crust is accreted in the N-E Atlantic, and thin oceanic crust in the western Central Atlantic.

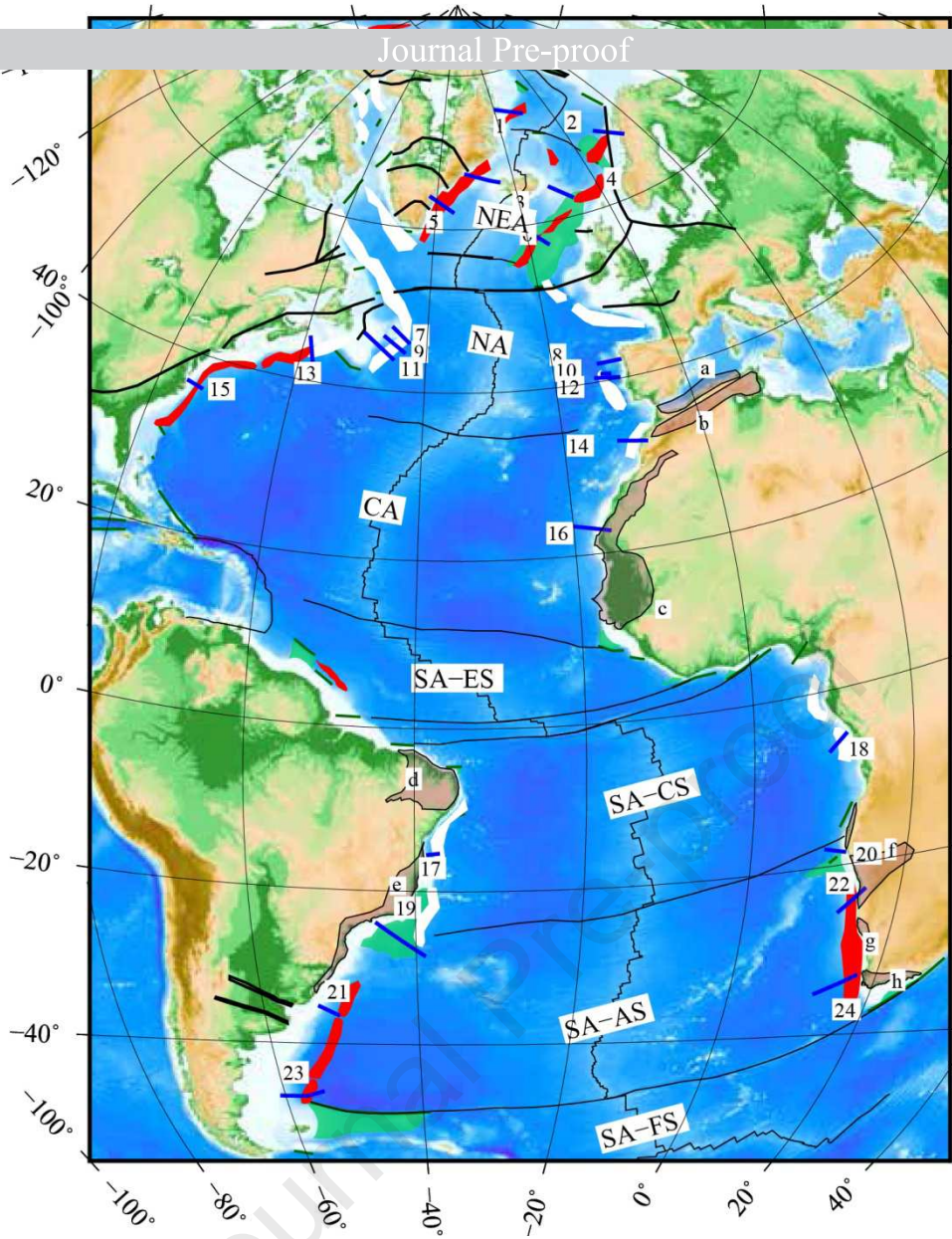


Figure 33: Location of the wide-angle seismic profiles reviewed in this study. Red and white polygons are volcanic margin parts (red) and magma-poor margins (white) Lundin et al., 2018. Green polygons depict TMP from Loncke et al., 2020. (a – Rif-Tell Fold Belt (Bumby and Guiraud, 2005); b – Atlas Fold Belt (Bumby and Guiraud, 2005); c – West African Fold Belt (Sougy, 1962); d – Borboruna Fold Belt (Assumpção et al., 2013); e – Dom Feliciano Fold Belt (Reuber and Mann, 2019); f – Damara Fold Belt (Reuber and Mann, 2019); g – Gariep Fold Belt (Reuber and Mann, 2019); h – Cape Fold Belt (Reuber and Mann, 2019)). Explanation of 1-24 is given in the caption of Table 2.

Profile W	(A)	(B)	Beta	(C)	Profile E	(A)	(B)	Beta	(C)	(D)	Ocean Basin	Magma poor /magmatic
1) AWI-20030300	43	8	5,37 5	5-6	2) LVMT1	43	10	4,3	6,7	parallel	NE-Atlantic	Magmatic
3) SIGMA 1	43	20	2,15	15	4) FIRE	43	15	2,8	15	craton	NE-Atlantic	Magmatic
5) SIGMA 3	43	15	2,86	8	6) iSIMM	46	13	3,53	9	craton	NE-Atlantic	Magmatic
7) SCREECH1	30	6,5	4,61	4,5	8) ISE1	30	3	10	7,5	-	Southern North Atlantic	Magma poor
9) SCREECH2	30	10	3	4	10) CAM144	30	3	10	?	-	Southern North Atlantic	Magma poor
11) SCREECH3	30	5	6	4	12) IAM9	30	8	3,75	8-10	-	Southern North Atlantic	Magma poor
13) SMART	37	5	7,4	3-4	14) MIRROR	37	9	4,11	7-8	parallel	Central Atlantic	Magma poor
15) LASE	40	20	2	10	16) DAKHLA	37	17	2,17	7-8	parallel	Central Atlantic	Magmatic
17) SALSA11	40	6	6,66	5/?	18) ZAIANGO14	40	5	8	5,5	parallel	South Atlantic/Central segment	Magma poor
19) SANBA2	42	12	3,5	6	20) P2 (Namibia)	40	15	2,66	6	parallel	South Atlantic/Central segment	Magma poor
21) BGR98-REF2	45	5	9	6,5	22) SPRINGBOK	45	10	4,5	7,5	orthogonal	South Atlantic/Austral segment	Magmatic
23) BGR04-01	45	12	3,75	6,5	24) MAMBA2	45	19	2,36	7,5	parallel/oblique	South Atlantic/Austral segment	Magmatic

(A) Original crustal thickness [km] ^(a)

(B) Thinned Continental crustal thickness [km]

(C) Oceanic crustal thickness [km]

(D) Angle to pre-existing fold belt or different crustal configuration

Table 2: Crustal thickness, beta-factor and oceanic thickness from the seismic profiles reviewed in this study. (1 - Voss et al., 2009; 2 - Tsikalas et al., 2008; 3 - Holbrook et al., 2001; 4 - Staples et al., 1997; 5 - Holbrook et al., 2001; 6 - White and Smith, 2009; 7 - Funck et al., 2003; 8 - Whitmarsh et al., 1996; 9 - Van Avendonk et al., 2006; 10 - Chian et al., 1999; 11 - Lau et al., 2006; 12 - Dean et al., 2000; 13 - Funck et al., 2004; 14 - Biari et al.,

2015; 15 - LASE Studygroup, 1984; 16 - Klingelhoefer et al., 2009; 17- Loureiro et al., 2018; 18 - Contrucci et al., 2004; 19 - Evain et al. 2015 ; 20 - Becker et al., 2014 ; 24 - Schinkel et al., 2007 ; 23 - Becker et al., 2014 ; 24 - Bauer et al., 2000). (a) From Szwillus et al., 2019

What is the influence of inheritance in margin break-up?

Our review highlights that the influence of inheritance on the position and style of rifting is still highly debated. While some oceanic basins developed preferentially along old sutures from previous Wilson Cycles (e. g. Piqué and Laville, 1996) other rifts broke through old cratons (Newman and White, 1999). A good example for the first is the northern Central Atlantic, which opened at least during two Wilson cycles along similar axes (Wilson, 1966; Pique and Laville, 1996; Whalen et al., 2015; Peace et al., 2019; Thomas, 2019). In some places, margins which formed along old suture zones, still show the presence of remaining subducted conjugate slabs, e.g. East Greenland (Schiffer et al., 2015).

Conversely, rifting broke through the Greenland Craton despite the existence of highly thinned continental crust at the nearby Hatton BASIN. Such behavior may be explained by the cooling and corresponding strengthening of the lithospheric mantle when rifting stops for a sufficient amount of time before the new rift develops (Newman and White al., 1999; England, 1986). When extension reoccurs, a new rift will develop next to the extinct one, preferring the then less competent lithosphere. Based on numerical modeling Armitage et al., 2010, propose that a first rifting episode directly influences the rifting style of following episodes. For example, during the inter-rifting period the thermal anomaly located at the older Hatton BASIN rift might have produced melt flowing into the NE-Atlantic rift producing volcanism and thick oceanic crust, and therefore could be a cause of magmatism during the break-up of the NE-Atlantic (Armitage et al., 2010).

The late Cretaceous/Early Cenozoic rifting in the NE Atlantic saw two configurations. In the north, it widely followed the generally N-S trend of the Caledonides, while in the south it rifted a craton and crossed the Nagssugtoqidian Orogen almost orthogonally. Similarly, in the Austral South Atlantic the Late Jurassic/Early Cretaceous rift crossed the Gondwana Fold Belt in Argentina/South Africa nearly orthogonally and went parallel to the Dom Feliciano Belt at the latitude of Brazil and Angola. The arms of the Proterozoic-Cambrian Pan-African Fold Belt (Damara F.B.-orthogonal, Kaoko, (Gariep), Ribeira F.B.-parallel) are arranged to form an orogenic triple junction at the point where the Walvis Ridge intersects with the Namibian shore (Clemson et al., 1999), i.e. at the position of the Etendeka lavas. At the intersection of propagating rifts with pre-existing suture zones, the production of magmatism probably increased (Koopmann et al., 2014a; Schiffer et al., 2015; Peace et al., 2017; Petersen et al., 2018). Where rifts stagnated at a barrier such as a suture zone, such excess magmatism may contribute to the formation of LIPs (Foulger et al., 2019, Peace, 2019).

Based on a concept from Courtilot, 1982, Reuber and Mann (2019) proposed that where orogenic fabric trends parallel to the north–south South Atlantic rift direction, narrow (55–90 km) rift zones with modest continental beta factors of 2.5–3.5 develop, while a basement fabric trending near-orthogonally to the north–south South Atlantic rift direction results in wider (185–220 km) rift zones with higher beta factors of 4.3–5. These authors explain that greater amounts of stretching are needed to rupture the orthogonal, orogenic, basement fabric, while smaller amounts of rifting are needed to stretch the weaker and parallel, orogenic, basement fabric (Figure 34). This hypothesis holds true for the South Atlantic where wider, locked zones in the South Atlantic like the rift-orthogonal Salado suture zone or Damara foldbelt, exhibit higher beta factors and narrower rift zones like the rift-parallel Dom Feliciano and Kaoko foldbelts which exhibit lower beta factors (Figures 32 and 33). The central Atlantic is characterized by the margin parallel West African Fold Belt and the Variscan and Iapetus sutures along the north American margin, which opened in a single phase (Sahabi et al., 2004). The NE Atlantic opened along pre-existing structures, however other factors such as changing stress directions were influential (Schiffer et al., 2019).

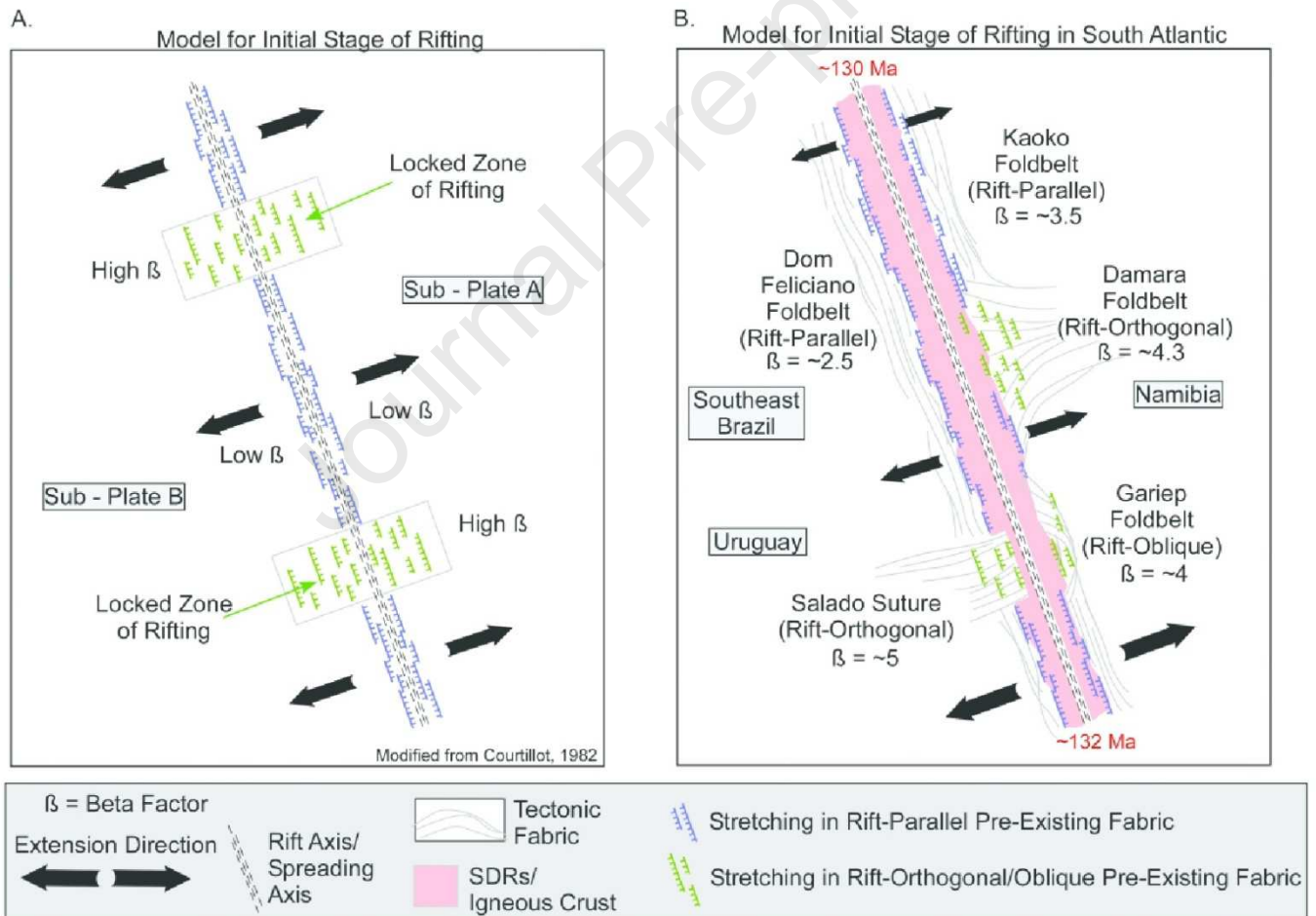


Figure 34: A) Modification of the schematic model from Courtilot (1982) illustrating how a propagating rift is affected by along-strike zones of wider, “locked” zones of continental extension between two plates. B) Model from Reuber and Mann. (2019) showing that the wider, locked zones correspond to areas of pre-rift, orogenic grains that are at right angles or oblique to the Cretaceous South Atlantic opening direction. Wider, locked zones in the

South Atlantic like the rift-orthogonal Salado suture zone or Damara foldbelt exhibit higher beta factors. Narrower rift zones like the rift-parallel Dom Feliciano and Kaoko foldbelts exhibit lower beta factors. Regional trends of pre-existing, orogenic basement fabrics are shown by the gray lines (After Reuber and Mann., 2019).

3) What are the origins and the mechanisms of formation of transform marginal plateaus?

TMPs are located at the boundary of two oceanic basins of different ages, for example the Demerara and Guinea PLATEAUS; the Walvis and Sao Paulo PLATEAUS and the Yermack and Morris-Jesup PLATEAUS, the Hatton-Rockall PLATEAU; the Falkland-Malvinas PLATEAU (Mercier de Lepinay et al., 2016; Loncke et al., 2020 and references therein). Although widespread, the architecture and layout of marine marginal plateaus have not yet been satisfactorily understood. By definition, they are associated with a transform margin. These transform segments might have formed at rift propagation barriers, or between two opening basins forming transfer zones during opening and break-up. Most of these plateaus have experienced volcanism at one stage of their evolution and are considered, at least for part of their surface area, as volcanic margins (Elliott et al., 2008; 2009; Berndt et al., 2001; Planert et al., 2016; Reuber et al., 2016; Abdelmalak et al., 2017; Schimschal and Jokat, 2019; Museur et al., 2020). However, a marginal ridge system is not always present on sheared edges. For example, oceanward the crust in front of SW Newfoundland thins from 20 to 6 km in thickness without creating a ridge (Reid and Jackson, 1997). Their genesis has been linked to a thermally induced uplift by heat transfer from the oceanic crust into the neighboring continental crust (Bird, 2001; Parsiegla et al., 2009) or to the uplift of crustal segments as a result of transpressive movements (Parsiegla et al., 2009). Also, as observed in the Red Sea, a compressional front can precede the tip of a propagating rift (Bonatti and Crane, 1982). When this front impacts against an oblique barrier such as a fracture zone, the crust along the barrier may be compressively uplifted (Crane and Bonatti, 1987) (Figure 35).

Numerical experiments and observations from the Red Sea indicate that spreading after continental rifting starts with sporadic seafloor spreading triggered by local asthenospheric upwelling in cells around 130 km apart (Bonatti et al., 1985; Corti et al., 2003). A contrary mechanism with unzipping from south to north has been proposed for the South Atlantic (Franke et al., 2007). Final rifting along the NE Greenland margin most likely also included opening along sporadic spreading cells (Franke et al., 2019). Systematic mapping of SDR sequences along the Argentine/Uruguay continental margins shows that the volume of magma emplaced is highest close to transfer zones and that the width of the SDR wedges decreases in the direction of rifting in each individual margin segment (Franke et al., 2007). A conceptual model from Koopmann et al., (2014a) proposes that once a rift arrives at a transfer zone rifting might come to a halt, volcanism temporarily diminishes, and heat accumulates in the mantle. When the next segment starts opening the accumulated heat may result in excess magmatism (Koopmann et al., 2014a). In this case, this accumulated heat might be responsible for the volcanism along TMPs but also in a top-down approach it might cause elevated heat regions in the mantle to develop at the position of rift barriers (Franke, 2013). This might explain earlier observations of the existence of mantle plumes close to

triple junctions observed by Burke and Dewey, 1973. The absence of TMP along some sheared margins might be explained by a different rifting style or the rift not halting for a longer period of time and thus not leading to the accumulation of heat in the mantle. However, some TMP seem to have developed at the terminus of a segment before reaching the transfer zone (e.g. Demerara/Guinea) and others after resuming opening when the transfer zone has been overcome (e.g. Hatton). This might be explained in some cases by rift barriers which allow the focusing of mantle melts and later the localization of transfer faults, and in other cases, the transform faults allow the compartmentalization of melt cells. The exact relationship between mantle melting cells, localization of TMP and transform faults will need to be studied in more detail in the future.

A number of recent studies also show that the transition from continental to oceanic crust on sheared continental margins is by no means as abrupt as shown in previous studies. The Ascension Fracture Zone in the Gulf of Guinea shows an approximately 70 km zone of a 'proto-oceanic' crust (Turner and Wilson, 2009). Along the Agulhas-Falkland Fracture Zone, there is a 50 km wide transition zone between the continental and oceanic crust (Parsiegla et al., 2009). This relatively broad transition zone observed along these margins can result from readjustments in the opening direction during oblique rifting for example due to a plate reorganisation (Acocella et al., 1999; Farangitakis et al., 2019).

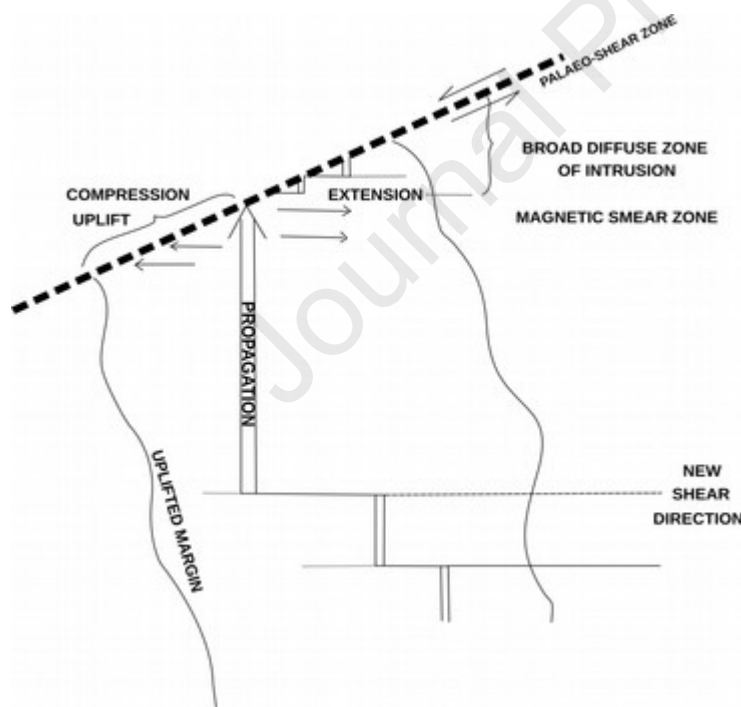


Figure 35: Schematic diagram of the deflection of a propagating rift along a continental shear zone (redrawn after Crane and Bonatti, 1987). One side of the intersection will be uplifted while the other side will be under oblique extension.

6 Conclusions and outlook

This study presents a compilation of deep seismic data along the Atlantic continental margins and selected conjugate or near-conjugate profiles based on plate kinematic reconstructions. It therefore allows comparison of asymmetry of rifting, influence of volcanism and the presence of serpentinized upper mantle material at the different margin segments. Some factors such as wide-angle seismic model quality and the plate reconstructional pole choice will influence our results, however as we are studying model features on a scale larger than 10-20 kilometers our results remain valid. The comparison of deep seismic models from the literature allows for the testing and validation of existing hypotheses about rifting and the opening of an oceanbasins. Our review shows that only magma-poor margins exhibit zones that have been qualified by the original authors of the studies as zones of serpentinized upper mantle material, exhumed during rifting. For magma rich margin segments the volume of volcanic material increases towards transfer zones, where rifting probably comes to a halt and leads to local heating of the upper mantle. Once the barrier was broken, rifting resumed with a lesser amount of volcanism. Highly oblique or shear margins are often associated with transform marginal plateaus and mostly include large quantities of volcanic material.

In the future more and denser spaced deep seismic profiles located along conjugate margins will be needed to explore rifting in true 3D geometry. Plate kinematic reconstructions taking plate deformation into account will be helpful to assess the finer degrees of asymmetry of margin pairs. Lastly, seismicreflectionprofiling will be needed to better account for the volume of volcanic material and associated variability along continental margins.

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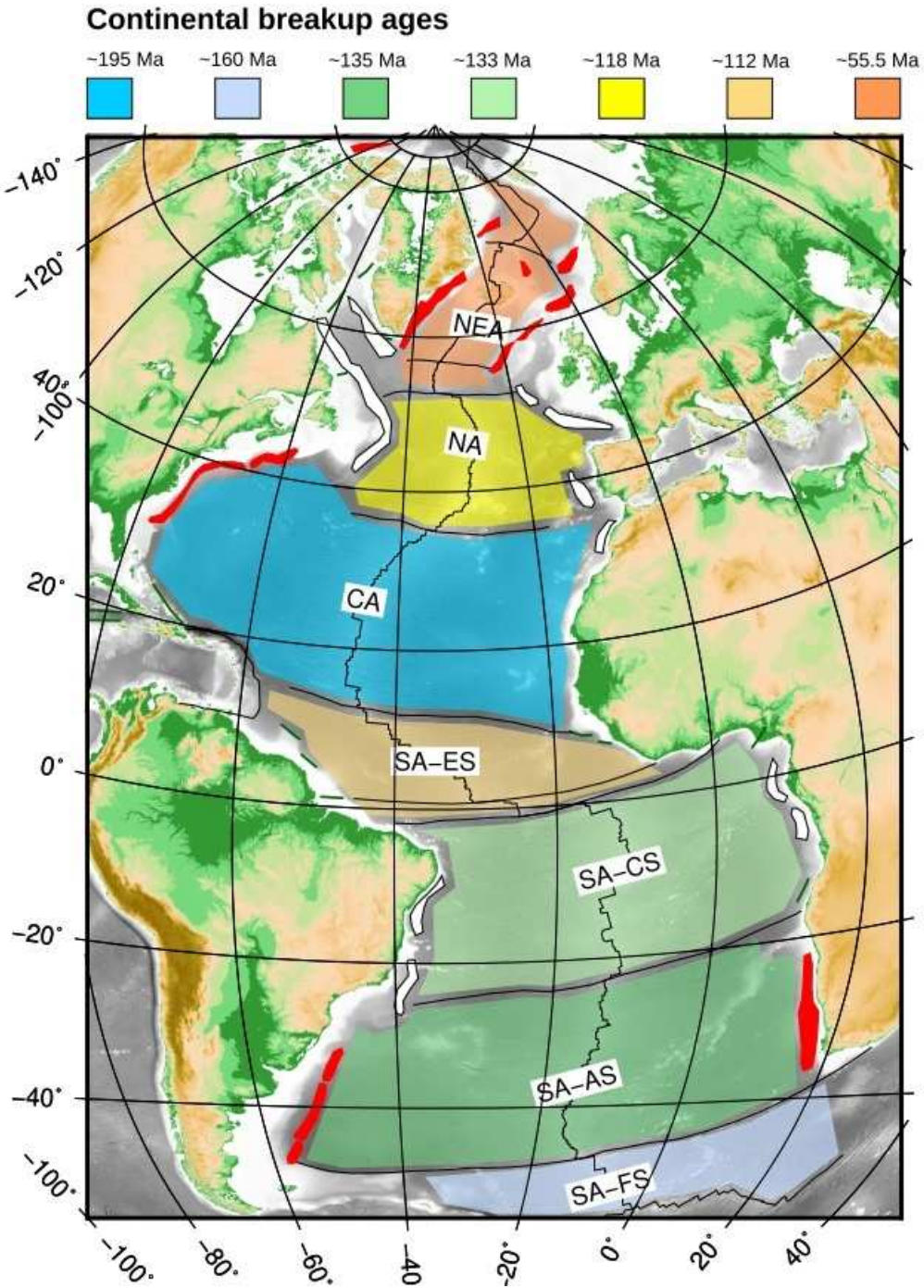
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Geographic domain	Age of opening	Full spreading rate at opening	Margin type	Salt basin	Rifting obliquity ⁽¹⁴⁾	Transform marginal Plateau ^{(15),(16)}	Associated magmatic province
Central Atlantic	~195 Ma ⁽¹⁾ ~175 Ma ⁽²⁾	0.8 cm/yr ⁽⁹⁾	North: Magma-poor South: Heavily magmatic (SDR, Underplate)	Am. : Sable, Shelburne, and Georges Bank basins Af. : Seine, Meseta, Mauritania, Casamance, Caroline basins	Divergent (0-20°)	-	CAMP
South Atlantic/Falkland segment	~160 Ma ⁽³⁾	1-2 cm/yr ⁽¹⁰⁾	Magmatic	Danmarks Havn Basin (NE Greenland)	Transform or highly oblique (60-90°)	Falkland-Malvinas Agulhas	KLIP
South Atlantic/Austral segment	~133 Ma ⁽⁴⁾	7 cm/yr ⁽¹⁰⁾	High magmatism (SDR, Underplate)	-	Divergent (0-20°)	Sao Paulo, Walvis	Paraná-Etendeka LIP
South Atlantic/Central segment	~118 Ma ⁽⁷⁾	3.8 cm/yr ⁽¹¹⁾ 4.7 cm/yr ⁽¹²⁾	Moderate magmatism.	Am. : Esperito Santo, Campos, Santos basins Af.: Gabon, Kwanza, Benguela basins	Divergent (0-20°)	-	CAMP
South Atlantic/Equatorial segment	~112 Ma ⁽⁵⁾	1 cm/yr ⁽¹⁰⁾	Magmatic	-	Transform or highly oblique (60-90°)	Demerara, Guinea, Liberia, Côte d'Ivoire-Ghana	CAMP
Southern North Atlantic	~133 Ma ⁽⁶⁾	1 cm/yr ⁽¹⁰⁾	Magma-poor	-	Divergent (0-30°)	-	NAIP
Northeast Atlantic	~55.5 Ma ⁽⁸⁾	1.5-2.5 cm/yr ⁽¹³⁾	Strong magmatism (SDR, Underplate)	- Paleozoic salt basin- Danmarkshavn Basin (NE Greenland)	Oblique (40-55°)	Morris Jesup Rise Yermack Plateau, NE Greenland, Vøring, Faroe-Rockall	NAIP



Highlights

- The comparison of published deep seismic velocity models, plate kinematic reconstructions and geological information allows to assess rifting styles and volcanic influence along the Atlantic Margins.
- Zones of upper mantle exhumed at the seafloor and serpentinized by water intruding through cracks and faults have been discovered only at magma-poor margins.
- At magma-rich margins the volume of magma is largest close to transfer zones and decreases in the direction of rifting in each individual margin segment.
- Sheared margins are often associated with transform marginal plateaus, which have undergone at least one volcanic phase.

Declaration of interests

☒ The authors declare that they have no known competing financial interests or personal relationships that could have appeared to influence the work reported in this paper.

☐ The authors declare the following financial interests/personal relationships which may be considered as potential competing interests: