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Michael Nirrengarten, G. Manatschal, J. Tugend, N. Kuszniir, D. Sauter

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## Tectonics

### RESEARCH ARTICLE

10.1002/2017TC004495

#### Key Points:

- Prebreakup and early seafloor spreading kinematic model of the southern North Atlantic
- Segmentation of the Iberia-Newfoundland rift system occurred during thinning
- Mantle exhumation and seafloor spreading occurred via V-shape propagation

#### Supporting Information:

- Supporting Information S1
- Table S1
- Figure S1
- Figure S2
- Movie S1
- Figure S3
- Data Set S1

#### Correspondence to:

M. Nirrengarten,  
m.nirrengarten@unistra.fr

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## Kinematic Evolution of the Southern North Atlantic: Implications for the Formation of Hyperextended Rift Systems

M. Nirrengarten<sup>1,2</sup> , G. Manatschal<sup>1</sup> , J. Tugend<sup>1</sup>, N. Kuszniir<sup>3</sup>, and D. Sauter<sup>1</sup> 

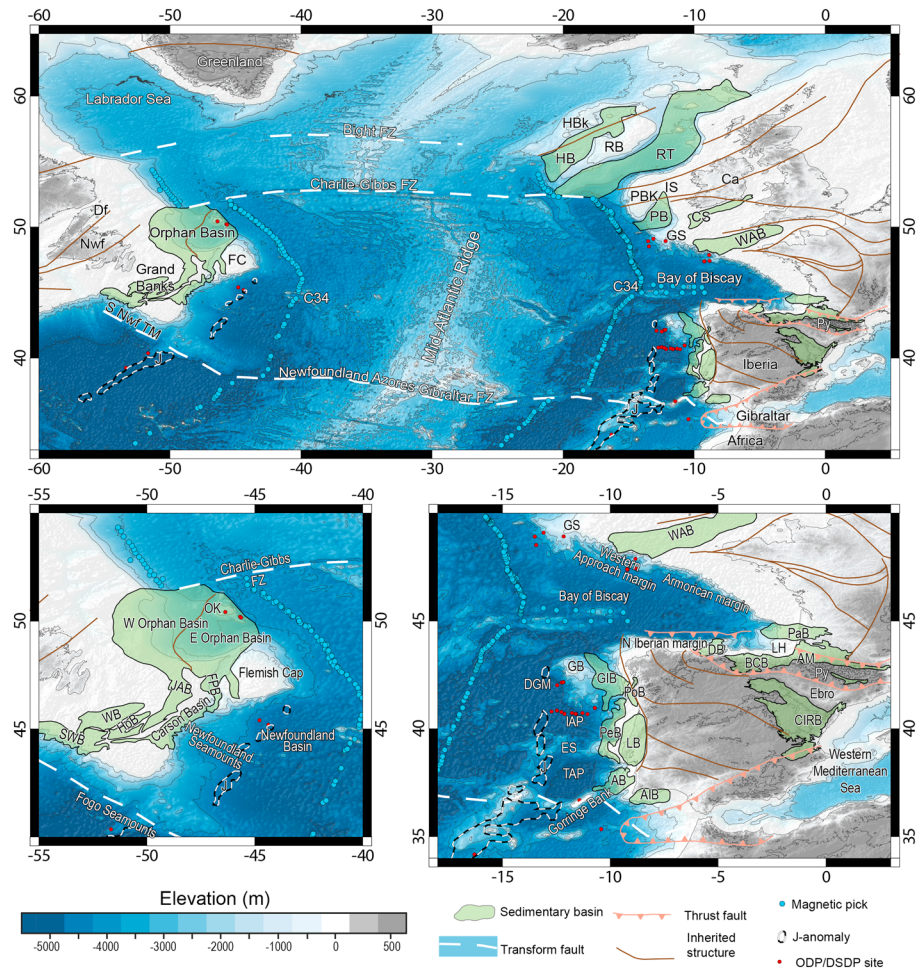
<sup>1</sup>Institut de Physique du Globe de Strasbourg, CNRS UMR 7516, Strasbourg, France, <sup>2</sup>Now at GEC, Université de Cergy-Pontoise, Cergy, France, <sup>3</sup>Earth, Ocean and Ecological Sciences, University of Liverpool, Liverpool, UK

**Abstract** We focus on the southern North Atlantic rifted margins to investigate the partitioning and propagation of deformation in hyperextended rift systems using plate kinematic modeling. The kinematic evolution of this area is well determined by oceanic magnetic anomalies after the Cretaceous normal polarity superchron. However, the rift and early seafloor spreading evolution (200–83 Ma) remains highly disputed due to contentious interpretations of the *J* magnetic anomaly on the Iberia-Newfoundland conjugate margins. Recent studies highlight that the *J* anomaly is probably polygenic, related to polyphased magmatic events, and therefore does not correspond to an isochron. We present a new palinspastic restoration without using the *J* magnetic anomaly as the chron M0. We combine 3-D gravity inversion results with local structural, stratigraphic, and geochronological constraints on the rift deformation history. The restoration of the southern North Atlantic itself is not the primary aim of the study but rather is used as a method to investigate the spatiotemporal evolution of hyperextended rift systems. We include continental microblocks that enable the partitioning of the deformation between different rift segments, which is of particular importance for the evolution of the Iberia-Eurasia plate boundary. Our modeling highlights the following: (1) the segmentation of the Iberia-Newfoundland rift system during continental crust thinning, (2) the northward V-shape propagation of mantle exhumation and seafloor spreading, (3) the complex partitioning of deformation along the Iberia-Eurasia plate boundary, and (4) a three-plate propagation model which implies transtension.

### 1. Introduction

Since Wegener (1915) and the pioneering studies of plate tectonics (Bullard et al., 1965; Heirtzler et al., 1968; Le Pichon, 1968; Morgan, 1968), many models have been proposed to restore the Atlantic Ocean by analyzing relative motions among Africa, America, and Eurasia (Bullard et al., 1965; Klitgord & Schouten, 1986; Le Pichon et al., 1977; Pitman & Talwani, 1972; Rowley & Lottes, 1988). Oceanic magnetic anomalies and oceanic fracture zones have provided very strong constraints on the evolution of the oceanic crust and have enabled the creation of global plate kinematic models (e.g., Müller et al., 2016; Seton et al., 2012). However, hyperextended domains that formed prior to seafloor spreading or oceanic domains without oceanic magnetic anomalies (formed during magnetic quiet zones) cannot be restored using classical methods of plate reconstruction and their restorations are therefore uncertain and often disputed (Barnett-Moore, Hosseinpour, et al., 2016; Neres et al., 2013; Vissers et al., 2016).

Hyperextended rift systems can accommodate significant extension prior to lithospheric breakup (up to 400 km (Brune et al., 2014; Sutra et al., 2013)), but their kinematic evolution is not well constrained as time markers are lacking. They are characterized by thin continental crust (often less than 10 km thick) (Sutra et al., 2013) sometimes accompanied by exhumed mantle domains and variable amounts of magmatic addition. The understanding of these hyperextended rift systems has been improved thanks to numerous studies based on seismic interpretations (e.g., Autin et al., 2010; Franke et al., 2014; Péron-Pinvidic et al., 2007; Ranero & Pérez-Gussinyé, 2010; Reston & McDermott, 2011), field observations (e.g., Jammes et al., 2009; Manatschal, 2004; Mohn et al., 2012), and numerical and analogue modeling (e.g., Brun & Beslier, 1996; Brune et al., 2014; Huisman & Beaumont, 2014; Lavier & Manatschal, 2006). Nevertheless, their integration in kinematic plate models remains complex. Previous studies proposed palinspastic restorations (e.g., Heine et al., 2013; Hosseinpour et al., 2013; Williams et al., 2011). However, they only use two rotation poles, one at full fit and the second at the first magnetic anomaly, assuming therefore a continuous evolution during rifting. Intermediate rotation poles are sometimes proposed for rift systems interrupted by



**Figure 1.** Topographic and bathymetric maps of the southern North Atlantic and zoom on the Newfoundland rifted margins and west European rifted margins. Magnetic picks are from the repository for magnetic anomaly (Seton et al., 2014). AB, Alentejo Basin; AIB, Algarve Basin; AM, Arzacq Mauléon Basins; BCB Basque Cantabrian Basin; CIRB, Central Iberian rift basins; Ca, Caledonian suture; CS, Celtic Sea; DB, Le Danois Basin; Df, Dover fault; DGM, Deep Galicia Margin; ES, Estremadura Spur; FC, Flemish Cap; FPB, Flemish Pass Basin; GB, Galicia Bank; GS, Goban Spur; HB, Hatton Basin; HBk, Hatton Bank; HoB, Horseshoe Basin; IAP, Iberia Abyssal Plain; IS, Irish shelf; JAB, Jeanne d’Arc Basin; LB, Lusitanian Basin; LH, Landes High; Nwf, Newfoundland; OK, Orphan Knoll; PaB, Parentis Basin; PB, Porcupine Basin; PBK, Porcupine Bank; PeB, Peniche Basin; PoB, Porto Basin; Py, Pyrenees; RB, Rockall Bank; RT, Rockall Trough; S Nwf TM, South Newfoundland Transform Margin; SWB, South Whale Basin; TAP, Tagus Abyssal Plain; WAB, Western Approach Basin; WB, Whale Basin. Basin outlines adapted from Funck et al. (2003), Alves et al. (2009), Lundin and Dore (2011), and Tugend et al. (2014). Inherited structures adapted from the compilations of Pollock et al. (2012), Welford et al. (2012), and Ballevre et al. (2014).

tectonic quiescence (Barnett-Moore, Müller, et al., 2016). Methodological improvements on the determination of spatiotemporal markers are needed to model both the rifting and the early stages of seafloor spreading.

Here we focus on the hyperextended rift systems of the southern North Atlantic between the Newfoundland Azores Gibraltar fracture zone (NAGFZ) and the Charlie Gibbs fracture zone (Figure 1). The study area includes the Iberia-Newfoundland rift system, the only place where scientific deep water drilling has penetrated hyperextended continental crust and exhumed mantle (Boillot et al., 1987; Sawyer, 1994; Tucholke & Sibuet, 2006; Whitmarsh & Wallace, 2001). Numerous seismic reflection and refraction data from this area are also available (e.g., Afilhado et al., 2008; Dean et al., 2000). This region includes the Pyrenean domain, which recorded an Albian to Cenomanian hyperextension phase (Clerc & Lagabrielle, 2014; Masini et al., 2014; Tugend et al., 2014; Vacherat et al., 2016), the oceanic Bay of Biscay (Montadert et al., 1979; Thion

et al., 2003), and the hyperextended rift basins on the Newfoundland and Irish margins (Welford et al., 2012). The first-order kinematic of the rifting, breakup, and seafloor spreading of the southern North Atlantic is related to the westward motion of the North American Plate relative to the Iberian/Eurasian Plate from Jurassic time to present. However, the Iberian Plate motion, before and during the Cretaceous normal polarity superchron, has often been questioned (Barnett-Moore, Hosseinpour, & Maus, 2016; Le Pichon & Sibuet, 1971; Olivet, 1996; Rosenbaum et al., 2002; Sibuet, 2004; Vissers & Meijer, 2012b). These controversies are mostly related to different interpretations and restorations of the *J* magnetic anomaly offshore the Iberia and Newfoundland rifted margins (Bronner et al., 2011; Tucholke & Sibuet, 2012). Based on a compilation of seismic and geochronological data, Nirrengarten et al. (2017) showed that the *J* anomaly does not correspond to an oceanic magnetic anomaly and therefore should not be used as such for kinematic restorations.

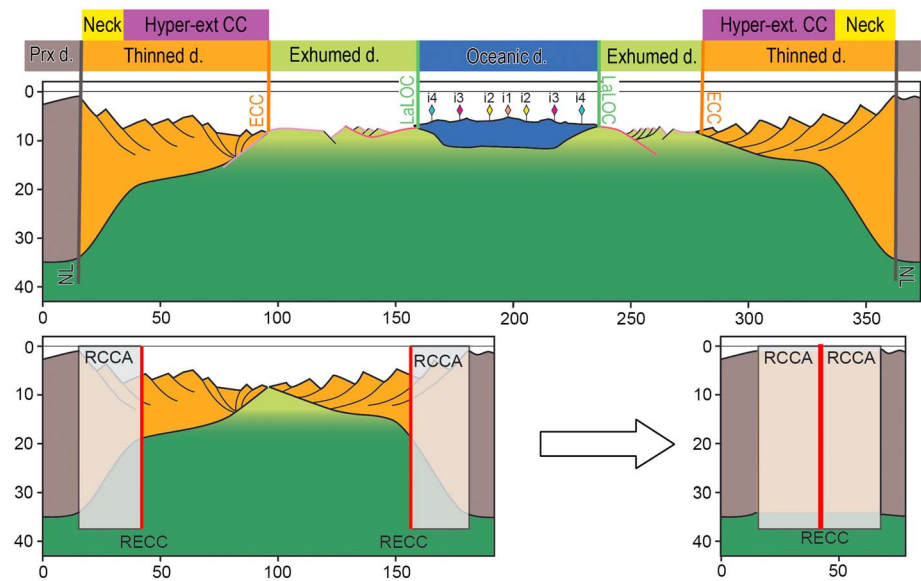
These uncertainties motivated the development of a methodology to improve the kinematic restoration of hyperextended rifted margins and early seafloor spreading stages in the southern North Atlantic. Our approach is to perform a palinspastic restoration of the southern North Atlantic that integrates the deformation of each hyperextended rift systems. In this approach continental microblocks (continental ribbons) (e.g., Lister et al., 1986; Peron-Pinvidic & Manatschal, 2010), the direction of the transfer zones, and key geological time constraints are introduced. A major input of this work is provided by a detailed mapping of the rift system based on seismic reflection and refraction interpretations and crustal thinning from gravity inversion coupled with key stratigraphic and geochronological data (e.g., Alves et al., 2009; Jagoutz et al., 2007). We perform areal balancing of gravity inversion-based crustal thickness cross sections to determine the predeformation dimensions of continental polygons that are used for plate modeling. Well-constrained areas (Iberia-Newfoundland margins and Pyrenees) are considered as inputs in the model, whereas areas that remain unconstrained by data (e.g., hyperextended basins along the Newfoundland and Irish shelves) represent outputs of the model. Although we propose a new plate model of the southern North Atlantic, it is important to note that the restored model itself is not the primary goal of this study but rather is a tool to examine and discuss the partitioning of the deformation in hyperextended rift systems. The model presented in this paper is one solution that can fit the sparse available data sets, but the existence of other solutions that do not contradict existing geological knowledge cannot be excluded.

## 2. The North Atlantic Ocean: Geography, Inheritance, and Kinematic Framework

The Atlantic Ocean is surrounded by passive continental margins and is only locally invaded by subduction zones (Duarte et al., 2013), providing an uninterrupted record from continental rifting to current oceanic accretion. Our study is limited to a segment of the North Atlantic located between the NAGFZ and the Charlie Gibbs fracture zone (Figure 1). As we aim to integrate the evolution of our study area in a global kinematic framework (Seton et al., 2012), we also investigate the neighboring segments, the central Atlantic south of the NAGFZ and the North Atlantic north of the Charlie Gibbs fracture zones.

The Atlantic Ocean resulted from the breakup of the supercontinent Pangea (e.g., Frizon de Lamotte et al., 2015), which was created by the progressive amalgamation of mainly continental blocks during the Carboniferous (Stampfli et al., 2013). The Late Triassic and Early Jurassic divergence of North America and Africa triggered the opening of the central Atlantic. Rifting history, interpretation of high-amplitude magnetic anomalies (East Coast magnetic anomaly and West African Coast magnetic anomaly), and initiation of seafloor spreading in the central Atlantic are still debated (Kneller et al., 2012; Labails et al., 2010; Sahabi et al., 2004; Schettino & Turco, 2009). However, the kinematic of the central Atlantic is relatively well defined by oceanic magnetic anomalies and fracture zones after the Jurassic Quiet Zone (post-154 Ma). The left-lateral movement of northwest Africa relative to Eurasia induced the rifting and formation of the Alpine Tethys ocean east of the Iberian Plate and the formation of several hyperextended/exhumed/oceanic domains in the Alpine realm (Meliata-Vardar and Piemonte-Liguria-Valais) (Handy et al., 2010; Schettino & Turco, 2011). The central Atlantic and the Alpine Tethys are connected by the Gibraltar transform zone, which probably experienced oblique rifting and highly segmented seafloor spreading during Late Jurassic (Frizon de Lamotte et al., 2011; Sallarès et al., 2011). However, the tectonic evolution east and south of the Iberian Plate is difficult to constrain quantitatively because of the Late Cretaceous and Cenozoic Alpine compression and subduction of parts of the system.





**Figure 2.** Schematic cross section highlighting the different rift domains determined by seismic interpretations in the southern North Atlantic (e.g., Gillard et al., 2016; Reston, 2009; Sutra et al., 2013; Tugend et al., 2014; Welford, Hall, et al., 2010). Step by step restoration of the rifted domains. Area balancing is used to determine the RECC (reconstructed edge of the continental crust). ECC, edge of the continental crust; i, isochrons oceanic magnetic anomalies; LaLOC, Landward Limit of Oceanic Crust; NL, necking line; RCCA, Restored Continental Crust Area.

Extensional deformation in the North Atlantic realm started during Permo-Triassic time or even before. However, the main crustal thinning occurred during the Late Jurassic–Early Cretaceous related to a crustal extension of 50–70 km (Skogseid et al., 2000). The magma-rich breakup between Greenland and Eurasia happened at about 55 Ma (Eldholm et al., 1989; Holbrook et al., 2001; White & Smith, 2009) after Late Cretaceous–Paleocene rifting. Eurasian Plate motion is constrained after chron C33n (Gaina et al., 2002). The motion of Greenland relative to North America is determined after chron C27n (~62 Ma) by the marine magnetic anomalies of the Labrador Sea and Baffin Bay (Oakey & Chalmers, 2012).

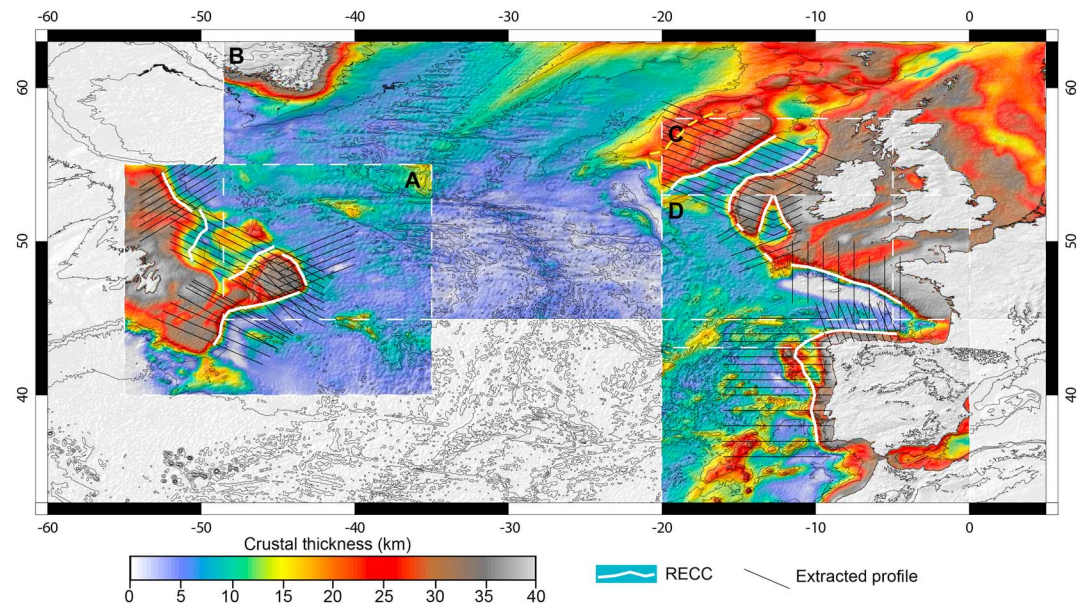
This study focuses on the tectonic and kinematic evolution of the southern North Atlantic segment between the earliest Jurassic (200 Ma) and magnetic chron C34 (83 Ma) (Gradstein, et al., 2012). Therefore, we model the plate motions during rifting and early stages of seafloor spreading.

### 3. Hyperextended Rifted Margins: A Kinematic Description

#### 3.1. Rift Domains and Domain Boundaries

Despite some variations on their crustal, sedimentary, and magmatic evolution, Atlantic magma-poor rifted margins share comparable first-order architectures (Peron-Pinvidic et al., 2013; Reston, 2009). These rifted margins are characterized from continent to ocean by a proximal domain, a thinned domain, and sometimes include an exhumed mantle domain before entering the oceanic domain (e.g., Sutra et al., 2013; Tugend, Manatschal, Kuszniir, & Masini, 2015) (Figure 2). The thinned domain includes the necking domain and the hyperextended continental crust. Boundaries between each domain are of particular importance for plate kinematic, but one question that needs to be addressed is whether a domain boundary can be considered to be a spatiotemporal marker. In this study, we defined and mapped the necking line (NL), the oceanward edge of the continental crust (ECC), and the continentward limit of first oceanic crust, referred to as the Landward Limit of Oceanic Crust (LaLOC) (Heine et al., 2013).

To map these boundaries, we determine variation of basement architecture and nature in 2-D sections (Sutra et al., 2013; Tugend, Manatschal, Kuszniir, & Masini, 2015), which then can be extrapolated laterally using potential field methods (Stanton et al., 2016; Tugend et al., 2014). We use seismic reflection and refraction data (Dean et al., 2000; Funck et al., 2003; Gerlings et al., 2011; Hopper et al., 2004; Shillington et al., 2006; Watremez et al., 2015), borehole data (e.g., Tucholke et al., 2007), gravity (Sandwell et al., 2014) and

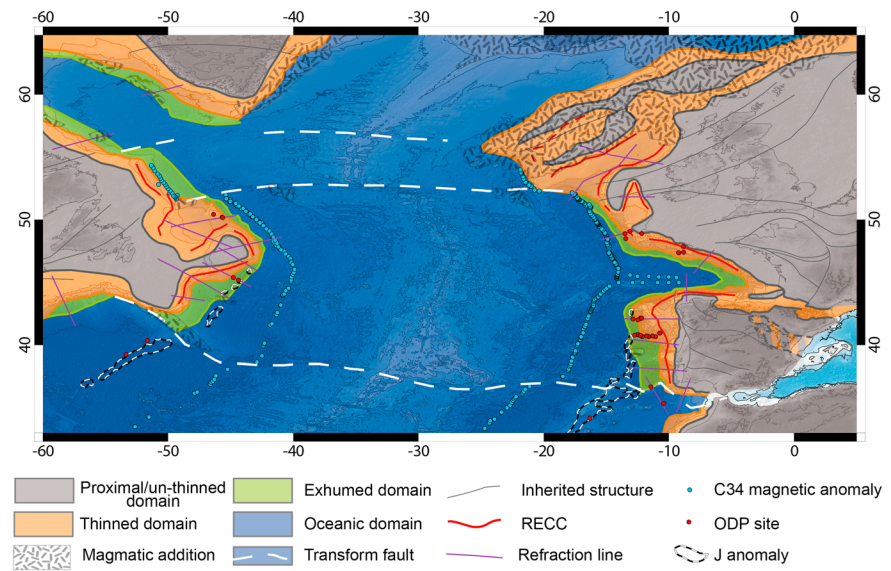


**Figure 3.** Crustal thickness maps resulting from gravity inversion. Parameters used for the gravity inversion and a sample of extracted crustal profiles are in supporting information S1 and S2. RECC, reconstructed edge of the continental crust.

magnetic maps (Meyer et al., 2016), published rift domain maps (Kneller et al., 2012; Nirrengarten et al., 2017; Peron-Pinvidic et al., 2013; Tugend et al., 2014; Welford, Hall, et al., 2010; Welford, Smith, et al., 2010) and crustal thickness maps (especially where seismic data are lacking). These latter (Figure 3) were determined by performing 3-D spectral domain gravity inversion. The detailed methodology is described in Chappell and Kusznir (2008) and includes a correction for the lithosphere thermal gravity anomaly and a parameterization of decompression melting to predict magmatic additions. The input data of the gravity inversion are public domain data (e.g., free-air gravity anomaly (Sandwell et al., 2014), bathymetry (Smith & Sandwell, 1997), sediment thickness map NOAA (Divins, 2003), and oceanic isochrons (Müller et al., 2016)).

The thermal reequilibration (cooling) time of oceanic lithosphere is given by ocean isochron ages (Müller et al., 2016); the cooling time of continental margin lithosphere is given by the margin breakup age. Gravity inversion results (Moho depth, crustal basement thickness, and thinning factor) are calibrated with the oceanic crustal thickness determined from seismic refraction data (Dean et al., 2000; Funck et al., 2003; Lau et al., 2006; Zelt et al., 2003). We parametrize the following: (1) the reference Moho depth, related to the long-wavelength component of the Earth's gravity field arising from deep mantle structures and processes and current dynamic topography; (2) the magnitude of magmatic additions from decompression melting corresponding to the thickness of the first oceanic crust; (3) the critical lithosphere thinning factor at which decompression melting commences; and (4) the continental breakup age, which provides the thermal reequilibration time of the continental margin lithosphere (Chappell & Kusznir, 2008). We carried out four different gravity inversions to parametrize as accurately as possible the areas of the Iberia, Newfoundland, Eurasian, Greenland, and Hatton rifted margins (Figure 3) (Parameter values and examples of calibrated profiles are given in the supporting information S1 and S2).

Proximal domains (brownish in Figure 2) are characterized by a weak thinning (thinning factor  $\gamma < 0.2$ ) of the continental crust/lithosphere other than under fault bounded half-graben basins. The necking line (NL Figure 2) corresponds to the continentward boundary of the thinned continental crust (orange in Figure 2). This line marks the transition from a weakly to a substantially thinned continental crust, where the Moho shallows and where the top basement deepens oceanward (Figure 2). The oceanward edge of the continental crust (ECC Figure 2) marks a change in material composition from continental crust to oceanic crust or exhumed subcontinental mantle (in green Figure 2). This boundary is best estimated by the location where a strong reflector at the base of the crust (*S*-reflection) (Reston et al., 2007) intersects



**Figure 4.** Map of the rift domains of the southern North Atlantic. The domain boundaries are derived from seismic interpretation, local mapping (Kneller et al., 2012; Nirrengarten et al., 2017; Tugend et al., 2014; Welford, Hall, et al., 2010), and crustal thickness maps (Figure 3). In parts where seismic data are sparse, the boundaries are an interpolation based on gravity inversion crustal thickness maps (see supporting information S3). RECC, reconstructed edge of the continental crust. The Central Iberian rift basins are poorly constrained and presented with dashed lines. Magmatic additions after the compilation of Chenin et al. (2015) and inherited structures adapted from the compilation of Welford et al. (2012), Pollock et al. (2012), and Balleve et al. (2014). Magnetic picks are from the repository for magnetic anomaly (Seton et al., 2014).

with top basement (for details see Sutra et al., 2013). In magma-poor margins, seismic facies often change from reflective (thinned continental crust) to transparent (serpentinized mantle) across the ECC. The LaLOC limits the first oceanic crust and typically coincides with the continentward termination of indisputable oceanic crust, characterized by parallel top basement and Moho reflections. Moreover, the oceanic top basement is typically observed near 8 s (two-way time, TWT) and characterized by a sedimentary passive infill. The LaLOC is often marked by a ramp morphology of the top basement from 9 to 10 s (top of exhumed mantle with few magmatic additions) to the typical 8 s two-way traveltime of the top of thermally equilibrated Penrose oceanic crust (for more details and definitions of rift domains see Tugend, Manatschal, & Kuszniir, 2015).

The resulting rift domain map (Figure 4) displays some important highlights: (1) the width of the thinned domain (measured in the extension direction) is highly variable and ranges from ~50 km to more than 300 km, (2) the limits between the domains are not necessarily parallel at the same margin, and (3) the distance between the first oceanic magnetic anomaly (C34) and the LaLOC decreases northward.

### 3.2. Restoration Methodology

Without any time constraints the restoration of an ideal 2-D cross section (Figure 2) can be achieved by removing the oceanic domain and, if present, also the exhumed mantle domain because both domains are newly formed materials defining a new top basement. The remaining cross section is formed by the thinned continental crust of the two conjugate margins. The extension of this deformed continental crust has to be taken into account for palinspastic restorations (e.g., Aslanian & Moulin, 2012; Dunbar & Sawyer, 1989; Heine et al., 2013; Williams et al., 2011). As magma-poor margins do not have a large amount of magmatic additions, we assume that the surface of thinned continental crust is equal to the initial crustal surface (no gain or loss of continental crust). We also assume that the material does not laterally flow along strike and that the restored section is oriented parallel to the extension direction (Figure 3). On a 2-D cross section, the surface of thinned continental crust between the mapped necking line (NL) and the edge of the continental crust (ECC) is extracted from the crustal thickness maps determined from gravity inversion. This area is divided by the initial crustal thickness to determine the reconstructed edge of the continental crust (RECC Figures 2–4) which is different to the RCOB (restored continent ocean boundary)

from Williams et al. (2011) because an exhumed mantle domain is often present in the southern North Atlantic between the ECC and LaLOC. The initial crustal thickness may differ depending on the prerift inheritance of each continental block. In this study we choose 37.5 km as the initial crustal thickness for all the system, which corresponds to the maximum crustal thickness modeled by refraction data at the Grand Banks of Newfoundland (Lau et al., 2006). This method is, however, not valid for magma-rich margins (e.g., Hatton Bank margin) since they present numerous magmatic additions, corresponding to new crust formation and consequently to a gain in crustal volume. In magma-rich margins restorations can only be undertaken if the volume of the magmatic additions can be estimated to determine a pseudo RECC (Barnett-Moore, Müller, et al., 2016).

Oceanic magnetic anomalies are the best spatiotemporal constraints for plate kinematic modeling. Some magnetic anomalies also exist on exhumed mantle domains (Russell & Whitmarsh, 2003; Sibuet, Srivastava, & Manatschal, 2007). However, the source of these anomalies is unclear. Serpentinized peridotites do not carry a sufficiently stable remanent magnetization to produce regionally coherent patterns of oceanic magnetic anomalies at mid-oceanic ridges (Bronner et al., 2014; Maffione et al., 2014). Moreover, since the emplacement process of exhumed mantle surfaces is not necessarily symmetric, these anomalies do not necessarily have a conjugate or a precise age (Gillard et al., 2015). The LaLOC can be dated by the first sedimentary layer downlapping onto the first oceanic crust, but this method does not work for exhumed mantle domains where complex in and out of sequence detachment faults may occur (Gillard et al., 2016). In thinned domains, deformation has been shown to migrate oceanward through time prior to mantle exhumation (Masini et al., 2013; Péron-Pinvidic & Manatschal, 2009; Ranero & Pérez-Gussinyé, 2010). Hence, sediments sealing the syntectonic wedges of a hyperextended continental domain have the same age as the initiation of mantle exhumation (e.g., Péron-Pinvidic et al., 2007).

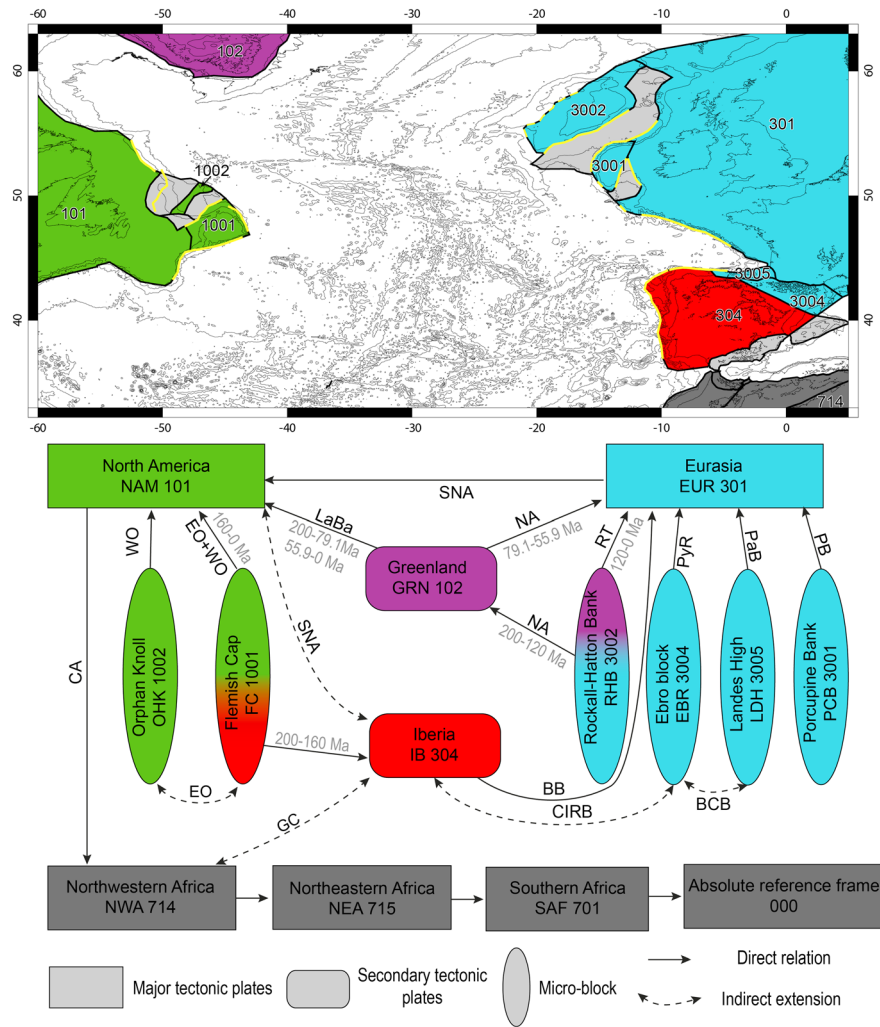
To determine the RECC on a map and to perform a full-fit restoration, we extracted profiles spaced at  $1^\circ$  to  $0.5^\circ$  intervals from the crustal thickness grid along all rifted margins and basins (Figures 3 and 4). Although Williams et al. (2011) showed that profile orientation does not change much the position of the RECC, we chose to extract them perpendicularly to the necking corresponding to the extension direction during maximum continental thinning. The RECC is used to determine the shape of each polygon defined in the plate modeling software GPlates (Boyden et al., 2011). It is important to note that the RECC defines the smallest possible polygon. The aim is to reduce the overlap even if it creates gaps elsewhere. The polygons can be subdivided into the following: (1) major tectonic plates, (2) secondary tectonic plates, and (3) microblocks (Figure 5). Major tectonic plates are currently limited by active plate boundaries. Secondary tectonic plates correspond to large continental units separated from the major plates by transient plate boundaries and are now integrated in a major tectonic plate. Microblocks are smaller and not separated from the major tectonic plates by active, well-determined plate boundaries (e.g., continental ribbons) (Peron-Pinvidic & Manatschal, 2010). Microblocks correspond to slightly thinned continental crust and may also have experienced a differential motion compared with the main plate which needs to be implemented in the kinematic model.

Transfer zones accommodate structural changes along strike roughly in the direction of extension in the exhumed domain and thinned continental crust. However, transfer faults affecting only continental crust can often be related to inherited structures and the deformation can be oriented obliquely to the main extension direction (Bellahsen et al., 2013), making them inappropriate for plate restoration. Transfer faults only represent good markers of the direction of extension if they penetrate newly created domains such as exhumed mantle or oceanic domains. The full rift motion cannot be determined from transfer zones only. The full-fit position is determined by the RECC fit of the conjugate margins and corresponds to the prehyper-extension setting of the southern North Atlantic.

### 3.3. Kinematic Significance of Domain Boundaries, Kinematic Markers, and Model Inputs

Mapped and restored domain boundaries (Figure 4) do not share the same geometries and implications for plate restoration. Therefore, establishing their significance as time markers is crucial for our study. It is meaningless to superpose two conjugate NLs because the deformed continental crust in between has to be restored and therefore the RECC has been delineated (Williams et al., 2011). However, there is no general observation to determine whether the necking starts simultaneously along the entire margin, along one segment or if it propagates. The ECC and the LaLOC are not necessarily synchronous along the entire





**Figure 5.** Polygons and plate circuit used in the modeling associated with their geographical positions. Yellow lines on the map correspond to the determined RECC. Ages below arrow correspond to the duration of motion of one block respectively to the other one. Abbreviations and the associated number are used in the GPlates rotation file provided in the supporting information. BB: Bay of Biscay, BCB: Basque Cantabrian Basin, CA: central Atlantic, CIRB: Central Iberian rift basins, EO: East Orphan Basin; GC: Gulf of Cadix; LaBa: Labrador Sea and Baffin Bay; Na: North Atlantic, PaB: Parentis Basin; PB: Porcupine Basin, PyR: Pyrenean rift systems, RT: Rockall Trough, SNA: Southern North Atlantic, WO: West Orphan Basin.

margin and may be the result of a propagation (Gaina et al., 1998; Heine et al., 2013), especially if well-defined marine magnetic anomalies are not parallel to these boundaries (e.g., South China Sea and Gulf of Aden) (Barckhausen et al., 2014; Fournier et al., 2010). Hence, the LaLOC and ECC, independently of the quality of the mapping, cannot be used as markers for plate kinematic restorations. Eagles et al. (2015) showed, for instance, that the LaLOCs can never be superposed. Transfer faults cutting the thinned and exhumed mantle domains provide the direction of motion during a stage of deformation, but they do not provide time indications.

#### 4. Rifted Margins and Basins

The southern North Atlantic system comprises three tectonic plates, Africa, North America, and Eurasia, two secondary plates, Greenland and Iberia, and six microblocks each separated by rift systems (Figure 5). In the next section, we review the timing, motion, and deformation mode of each rift system (Table 1).

**Table 1**  
Published Information on Rift and Oceanic Systems of the Southern North Atlantic

Rift system/basin	Events	Input of the model
Central Atlantic	Onset of rifting 240 Ma (Kneller et al., 2012), 230 Ma (Schettino & Turco, 2009; Sahabi et al., 2004), and 203 Ma (Labails et al., 2010)	Yes
	Onset of seafloor spreading 200 Ma (Sahabi et al., 2004) and 190 Ma (Labails et al., 2010; Schettino & Turco, 2009)	Yes
Tagus Abyssal Plain-Southern Newfoundland	First undisputed magnetic anomaly CM25 154 Ma (Klitgord & Schouten, 1986)	Yes
	Triassic to Middle-Jurassic rifting 200–161 Ma (Rasmussen et al., 1998)	Yes
	Gabbro intruded in peridotite 143 Ma (Féraud et al., 1986) marking possible onset of exhumation	Yes
	Onset of seafloor spreading 150 Ma (Mauffret et al., 1989; Srivastava et al., 2000), 133 Ma (Pinheiro et al., 1992), and 112 Ma (Tucholke, Sawyer, & Sibuet, 2007)	No
Iberia Abyssal Plain-Newfoundland	First undisputed magnetic anomaly C34 83 Ma (Srivastava & Roest, 1989)	Yes
	Triassic to Middle Jurassic rifting 200–161 Ma (Rasmussen et al., 1998; Murillas et al., 1990)	Yes
	Necking of continental crust 150–133 Ma (Mohn et al., 2015)	Yes
	Onset of mantle exhumation 133 Ma (Jagoutz et al., 2007; Mohn et al., 2015)	Yes
Galicia-Flemish Cap	Onset of seafloor spreading 145 Ma (Srivastava et al., 2000), 132 Ma (Schettino & Turco, 2009), 120 Ma (Bronner et al., 2011; Olivet, 1996), and 112 Ma (Tucholke, Sawyer, & Sibuet, 2007)	No
	Triassic to Late Jurassic rifting 200–145 Ma (Murillas et al., 1990)	Yes
	Necking of continental crust 150–133 Ma (Mohn et al., 2015)	Yes
	Onset of mantle exhumation 122 Ma (Féraud et al., 1986; Mohn et al., 2015; Schärer et al., 2000)	Yes
Orphan Basin	Onset of seafloor spreading 121 Ma Vissers & Meijer, 2012a and 112 Ma (Bronner et al., 2011; Tucholke, Sawyer, & Sibuet, 2007)	No
	Extension related to two main phases: Late Jurassic and Early Cretaceous (Enachescu, 2006; Dafoe et al., 2017)	No
Bay of Biscay	Ages of necking and possible mantle exhumation are not constraint	No
	Onset of rifting 156 Ma (Sibuet et al., 2004) and 145 Ma (Tugend et al., 2014)	No
	Age of Exhumation 125 Ma (Thinon, 2002; Tugend et al., 2014)	No
	Onset of seafloor spreading 124 Ma (Sibuet et al., 2004) and 112 Ma (Montadert et al., 1979; Thinon, 2002)	No
Parentis	End of seafloor spreading 83 Ma (Sibuet & Collette, 1991)	Yes
	Onset of rifting 130 Ma (Tugend, Manatschal, Kuszniir, & Masini, 2015)	Yes
Pyrenean Rift Systems	End of extension 106 Ma (Tugend, Manatschal, Kuszniir, & Masini, 2015)	yes
	Transtensional motion (Olivet, 1996) Onset of rifting dated as Late Jurassic (Desegaulx & Brunet, 1990)	Yes
	Hyperextension starts in Albian (Tugend, Manatschal, Kuszniir, & Masini, 2015)	
Central Iberian Rift System	Onset of convergence constraint as 83 Ma (Capote et al., 2002)	
	Onset of rifting 155 Ma (Salas & Casas, 1993)	No
Porcupine Basin	End of extension 106 Ma (Salas & Casas, 1993)	no
	Extension from Permian to Early Cretaceous (Tate et al., 1993)	No
Rockall Trough NE Newfoundland-Irish Margin	Main rift phase 161–145 Ma (Tate et al., 1993)	No
	Late Jurassic and Early Cretaceous extension (Naylor & Shannon, 2005)	Yes
	Onset of rifting 130 Ma (De Graciansky et al., 1985)	No
Labrador Sea	Onset of exhumation 112 Ma (De Graciansky et al., 1985)	No
	Onset of seafloor spreading	No
	First magnetic anomaly C34 83 Ma (Klitgord & Schouten, 1986)	
	First magnetic anomaly C27 61 Ma (Oakey & Chalmers, 2012)	No
	End of seafloor spreading 33 Ma (Oakey & Chalmers, 2012)	No

Note. The right column indicates if the information is used as an input of the plate modeling.

#### 4.1. Iberia-Newfoundland

The Iberia-Newfoundland conjugate margins result from the divergence between Iberia and North America. A first extension phase forming several sedimentary rift basins located over the Newfoundland Grand Banks and the Iberian Peninsula occurred from Triassic to Middle Jurassic (e.g., Lusitanian, Jeanne d'Arc, Porto, Horseshoe, and Whale basins, Figure 1) (Rasmussen et al., 1998; Tankard et al., 1989). These sedimentary basins are typically asymmetric half grabens, filled by Triassic evaporates and Jurassic carbonates

(Murillas et al., 1990; Pereira & Alves, 2011). Two-dimensional seismic interpretation studies across these conjugate margins were performed to determine structural domains (e.g., Péron-Pinvidic & Manatschal, 2009; Sutra et al., 2013; Welford, Smith, et al., 2010) (Figure 2) that were mapped (Figure 4) and used in the kinematic modeling. These margins are divided from south to north into three segments (Alves et al., 2009; Pinheiro et al., 1992; Welford, Smith, et al., 2010): the southern Newfoundland Basin-Tagus Abyssal Plain segment, the Iberia Abyssal Plain-northern Newfoundland Basin segment, and the Flemish Cap Galicia Bank segment (Figure 1). The deeper parts of the two northern segments have been drilled constraining the age of necking and breakup based on stratigraphic unconformities and deepening of depositional environments. In contrast, the distal part of the southern segment remains only approximated by indirect seismic correlations with borehole located on the proximal margin (Alves et al., 2009). Tithonian and Berriasian sediments from the Iberia Abyssal Plain indicate deep marine environments, whereas sediments of the same age on the Galicia Bank are shallow water carbonates. This observation argues for younger ages of the crustal necking northward (Jeannot et al., 2016; Mohn et al., 2015). The sedimentary record retrieved by Ocean Drilling Program (ODP) cruises indicates that necking had to occur in Tithonian-Berriasian time in the Iberia Abyssal Plain and Valanginian time in the Galicia margin. Exhumed serpentinitized mantle has been sampled in the two northern segments at ODP Sites 637, 897, 898, 1068, and 1070 (Boillot et al., 1987; Sawyer, 1994; Whitmarsh & Wallace, 2001). Seismic refraction data also suggest exhumed mantle on the Gorringe Bank in the Tagus Abyssal Plain (Sallarès et al., 2013) (Figures 1 and 4). Zircons obtained from gabbros included in serpentinitized peridotites have been dated at 137.5 Ma on the Gorringe Bank (Schärer et al., 2000), and  $^{40}\text{Ar}/^{39}\text{Ar}$  dating of hornblende in the same gabbros gave  $145 \pm 1$  Ma (Féraud et al., 1986). In the Iberia Abyssal Plain ages of plagioclase from deformed gabbros, dating the cooling of these rocks below 200 to 150°C, yield plateau ages of  $136.4 \pm 0.3$  Ma at ODP Site 900 (Féraud et al., 1996) and  $133.1 \pm 0.1$  Ma at ODP Site 1068 and  $141.8 \pm 0.4$  Ma at ODP Site 1067 (Jagoutz et al., 2007). This indicates that lower crustal rocks already had to be exhumed to shallower levels at that time; hence, mantle exhumation had to occur during Valanginian to Hauterivian time (~139–130 Ma). At ODP Site 637 at the Galicia margin, a diorite dyke has been dated by Ar/Ar on amphiboles at  $122 \pm 0.6$  Ma (Féraud et al., 1996). Zircons from metagabbros in the exhumed domain dredged at the Galicia margin (GAL-32-07) yield an emplacement age of  $121 \pm 0.4$  Ma (Schärer et al., 2000). Mantle exhumation at the seafloor did not start synchronously along the Iberia-Newfoundland margins, initiating in Valanginian time at the Gorringe Bank and in Aptian time farther north. A major sedimentary unconformity at the Aptian-Albian boundary has been interpreted as marking the complete separation of the lithosphere and the onset of stable oceanic accretion (Tucholke et al., 2007).

#### 4.2. Orphan Basin

The Orphan Basin is located northeast of the Newfoundland Island and is surrounded by the Flemish Cap to the southeast, the Grand Banks to the south, the northeastern Newfoundland margin to the west, and the Atlantic Ocean to the north (Figure 1). The crustal structure of this basin is divided in two subbasins, locally overlying less than 15 km thick continental crust (Watremez et al., 2015) in the East and West Orphan Basins, separated by a structural high (Figures 1–4). This high extends northward to the Orphan Knoll. The age of crustal extension and necking in these subbasins is ill constrained, but stratigraphic studies and well calibration provide some insight (Dafoe et al., 2017; Gouiza et al., 2016). After Late Triassic-Early Jurassic rifting in the East Orphan and the Jeanne d'Arc Basin (Enachescu, 2006) (Figure 1), two main extension phases have been suggested. The first one, during the Late Jurassic, is supposed to mainly affect the eastern subbasin, whereas the second phase, during Early Cretaceous, was likely more intense in the western subbasin (Dafoe et al., 2017; Enachescu, 2006; Gouiza et al., 2016). Seismic velocities from seismic refraction studies suggest a 5–8 km thick continental crust in the thinnest part of the subbasins and do not present evidence for large areas of exhumed mantle (Chian et al. 2001; Watremez et al., 2015).

#### 4.3. Bay of Biscay

The Bay of Biscay is a V-shaped oceanic basin limited to the northeast by the Western Approaches and the Armorican rifted margins and to the south by the North Iberian (or Cantabrian) margin (Figure 1). The Bay of Biscay is connected to the west to the southern North Atlantic and terminates to the east in the Parentis Basin. Several sedimentary basins in the Pyrenean realm may be linked kinematically to the opening of the Bay of Biscay (Roca et al., 2011; Tugend, Manatschal, & Kuszniir, 2015). The main rifting phase occurred

from latest Jurassic to Early Cretaceous (Berriasian to Barremian) leading to necking and extreme thinning of the crust (Montadert et al., 1979; Tugend et al., 2014). Mantle exhumation is inferred to occur during the Aptian (Thinon, 2002; Tugend et al., 2014), prior to the beginning of oceanic accretion in latest Aptian-earliest Albian (Montadert et al., 1979; Thinon et al., 2003). These ages are, however, only constrained by correlations from the few drill holes located on the Western Approaches margin (e.g., Deep Sea Drilling Project, DSDP Leg 48) (Montadert & Roberts, 1979) (Figure 4). Marine magnetic anomaly C34 is identified in the oceanic domain (Verhoef et al., 1986; Williams, 1975) and highlights the presence of a triple point between the Bay of Biscay and North Atlantic oceanic ridges (Sibuet & Collette, 1991). Seafloor spreading ceased in the Bay of Biscay between anomalies C34 and C33 (83 to 79 Ma) (Sibuet & Collette, 1991). The northern Iberian margin experienced compression from Late Cretaceous to Oligocene (Capot et al., 2002), whereas the Western Approaches and Armorican margins were only weakly reactivated (Thinon et al., 2003).

#### 4.4. Pyrenees, Parentis, Basque-Cantabrian, and Central Iberian Rift Systems

The Pyrenean mountain belt (in a broad sense) represents the remnant of a former transient plate boundary between Eurasia and Iberia (Canérot, 2008). Distributed strike slip and transtensional motion occurred from the Late Jurassic to Early Aptian (Canérot, 2008; Desegaulx & Brunet, 1990; Jammes et al., 2009; Olivet, 1996; Tugend, Manatschal, & Kuszniir, 2015) associated with a complex partitioning of the deformation between several oblique rift systems (Central Iberian, Bay of Biscay-Parentis, and Pyrenean-Basque-Cantabrian rift systems) (Figure 1). These rift systems are separated by crustal blocks (Landes High and Ebro Block; Figure 5) interpreted as continental ribbons (Tugend, Manatschal, & Kuszniir, 2015). These blocks are delimited by Cretaceous rift-related structures (Capote et al., 2002; Roca et al., 2011). The Central Iberian rift is assumed to be of Late Jurassic to Albian age (Salas & Casas, 1993). Basins from the Pyrenean domain (Basque-Cantabrian and Arzacq-Mauléon) are highly segmented (Tugend, Manatschal, Kuszniir, & Masini, 2015) and evolved from Late Jurassic to probably Cenomanian/Turonian time with mantle exhumation starting at Albian time (Jammes et al., 2009; Lagabrielle & Bodinier, 2008; Masini et al., 2014). The NNE-SSW segmentation of the Pyrenean-Basque-Cantabrian rift system is still observable at a crustal scale (e.g., Pamplona and Toulouse transfer zone) and seems to constrain the first-order direction of extension from Late Aptian onward (Chevrot et al., 2015; Jammes et al., 2009). The cessation of rift activity coincides with the onset of convergence in the Pyrenean domain in Santonian to Campanian time (Capote et al., 2002), whereas in the Central Iberian rift system the main convergence seems to occur during the Eocene (Salas & Casas, 1993). The convergence is related to the northward motion of Africa and led to the final collision stage of the Pyrenees during Oligocene time (Capote et al., 2002).

#### 4.5. Porcupine Basin, Rockall Trough, and Hatton Basin

The Porcupine Basin is a north trending V-shaped, hyperextended basin (Reston et al., 2004) located on the Irish shelf, north of the Goban Spur, bounded eastward by the Irish mainland platform and westward by a continental ribbon, referred to as the Porcupine Bank (Figure 1). The basin is narrow (<200 km wide) but has a large amount of crustal thinning (thinning factor  $\gamma > 0.8$ ), which may have resulted in local mantle exhumation and serpentinization (O'Reilly et al., 2006; Reston et al., 2004). Rifting occurred in several phases during Permo-Triassic, Jurassic, and earliest Cretaceous times with a major thinning phase during Middle to Late Jurassic (Tate et al., 1993).

The Rockall Trough (Figure 1) is located between the Irish mainland platform and the Porcupine Bank to the east, by the Rockall Bank to the west, and by the Charlie Gibbs fracture zone to the south. The basin is 1,100 km long and narrows from south to north (Figure 1). The crust outside the basin is about 30 km thick (Figure 3). The southern Rockall Basin is up to 200 km wide and floored by a crust that can be as thin as 10 to 8 km. There are different suggestions concerning the nature of the basement that floors the Rockall Basin: oceanic crust, serpentinized mantle, or hyperextended continental crust (O'Reilly et al., 1996). Permo-Triassic, Late Jurassic, and Early Cretaceous extension phases have been proposed (Naylor & Shannon, 2005), but a Middle to Late Cretaceous extensional phase is suggested by plate reconstructions (Cole & Peachey, 1999). However, the precise age of the main rift event remains unconstrained.

The Hatton Basin is oriented NE-SW and separates Rockall Bank on its eastern side and Hatton Bank on its western side. The basin stratigraphy is poorly understood due to the lack of well data and the numerous



magmatic additions; however, the main extension seems to be related to the breakup of the North Atlantic in Paleocene time (Hitchen, 2004).

#### 4.6. Northeastern Newfoundland Margin and Irish Shelf

The Irish shelf and the northeastern Newfoundland conjugate margins (Figure 1) are complex because they cut obliquely highly thinned continental crust and basins as well as unthinned continental crust. The Irish shelf margin is oriented NW-SE and continues from the Goban Spur at the northern termination of the Western Approaches margin to the southwest of the Rockall Bank. It follows the edge of the Porcupine Basin, the Porcupine Bank, and the Rockall Trough (Figure 1). The northeastern Newfoundland margin is composed of the northeastern Flemish Cap margin and the northern edge of the eastern and western Orphan subbasins to the Charlie Gibbs fracture zone. Only few studies are available on this area, and most of the data are from the Flemish Cap and Goban Spur conjugate margins. These margins are highly asymmetric (Gerlings et al., 2012). The Goban Spur shows a progressively necked crustal structure, with crust thinning from ~20 km to ~6 km over a distance of ~40 km and a zone of exhumed mantle that is up to 70 km wide (Bullock & Minshull, 2005). In contrast, the Flemish Cap margin presents a sharp necking, with continental crust thinning from 32 km to less than 10 km over only 30 km (Gerlings et al., 2011). The thin continental crust extends over 100 km oceanward and is bounded by about 20 km wide domain of exhumed mantle (Welford, Hall, et al., 2010). This structure is partly related to several NW-SE transfer zones affecting the margin during Late Jurassic-Early Cretaceous synchronously to extension in the Orphan Basin prior to the NE-SW extension between Ireland and Newfoundland. A small amount of Permo-Triassic extension could have affected Goban Spur; however, the oldest synrift strata recovered at DSDP Site 549 (Figure 1) on the Goban Spur are lower Barremian (De Graciansky et al., 1985). The unconformity between synrift and postrift sediments in continental half grabens is dated at the Aptian-Albian boundary (De Graciansky et al., 1985). However, this timing is only valid for the southernmost part of this area (Flemish Cap-Goban Spur rifted margins).

#### 4.7. Labrador Sea and Baffin Bay

The Labrador Sea and Baffin Bay are two oceanic basins separating Greenland from North America (Figure 1). These basins are not the focus of this study, but the position of Greenland has to be constrained to determine the position of Eurasia relative to North America. For this, we rely on the recent restoration of Barnett-Moore, Müller, et al. (2016). The oceanic evolution of these basins started in Maastrichtian or Paleocene time prior to Chron C27 (61 Ma) and ended at the Eocene-Oligocene boundary (33 Ma) (Oakey & Chalmers, 2012). The Labrador and SW Greenland margins experienced an important magma-poor rifting episode leading to the exhumation of lithospheric mantle (Chian et al., 1995). The tectonostratigraphic evolution of the proximal domain of the Labrador margin (Dickie et al., 2011) suggests two phases of rifting. A first one, Valanginian to Albian in age, is best documented on the proximal domain, whereas the second one, Cenomanian to Maastrichtian in age, is better recorded in more distal domains. We used existing kinematic models proposed for the northern North Atlantic (Barnett-Moore, Müller, et al., 2016; Hosseinpour et al., 2013) to locate the initial position of Greenland relative to North America at 200 Ma.

### 5. Tectonic Evolution of the Southern North Atlantic Rift Domains

As highlighted in the previous section, data sets and knowledge on each rift segment of the southern North Atlantic are very different. In some domains the necking phase, exhumation phase and onset of seafloor spreading are constrained by stratigraphy, geochronology, or dated magnetic anomalies, while in others only one or none of these constraints are known. Therefore, our restoration methodology is to implement the kinematic evolution of well-known domains and to test how less constrained areas evolve. Hence, the input data of the model are limited to areas from which data are reliable (Table 1). The temporal inputs to the model rely on the study of deep borehole data, outcrops, and their correlation to seismic data. Using the data and methods explained above, we have developed a new plate kinematic model of the southern North Atlantic to visualize the tectonic evolution of the rift systems. The kinematic model is composed of a set of finite rotation poles (Table 2). The solution presented in our model is nonunique and is used as a visualization tool (see supporting information) to examine the processes of rift propagation. North America (NAM) is kept fixed during all the investigated period, whereas the other plates are included in a plate circuit (Figure 5), highlighting the interaction between each plate through time.

**Table 2**  
Finite Rotation Poles Used in This Study

Total reconstruction poles										
Age	Latitude	Longitude	Angle	Fix plate	Ref	Age	Latitude	Longitude	Angle	Fix plate
Eurasia						Ebro block				
200.0	71.41	152.60	-23.68	NAM	<sup>a</sup>	200.0	51.59	2.80	-30.22	EUR
120.0	68.01	153.59	-21.01	NAM	<sup>a</sup>	160.0	51.59	2.80	-30.22	EUR
120.0	48.16	121.62	-12.71	GRN	<sup>a</sup>	150.0	51.69	2.68	-28.85	EUR
79.1	48.16	121.62	-12.71	GRN	<sup>a</sup>	140.0	51.93	2.76	-26.58	EUR
79.1	63.40	147.75	-18.48	NAM	<sup>b</sup>	120.0	49.91	1.49	-15.09	EUR
Iberia						112.0	47.87	-2.39	-10.17	EUR
200.0	50.43	1.53	-37.52	EUR		105.0	45.59	-5.96	-6.44	EUR
161.0	50.95	1.91	-37.60	EUR		86.0	27.56	-25.13	2.39	EUR
140.0	51.64	2.59	-35.62	EUR		83.0	27.56	-25.13	-2.39	EUR
120.0	-50.06	-177.68	22.83	EUR		Landes High				
112.0	-46.62	178.49	22.92	EUR		200.0	30.56	-1.68	-7.34	EBR
105.0	-44.99	176.89	19.01	EUR		122.0	30.56	-1.68	-7.34	EBR
86.0	-40.18	170.57	10.07	EUR		122.0	44.29	-0.79	-23.27	EUR
83.0	-39.36	170.34	9.08	EUR	<sup>c</sup>	100.0	43.99	-1.75	-6.53	EUR
Greenland						86.0	0.00	0.00	0.00	EUR
200.0	61.76	-132.95	-10.98	NAM	<sup>a</sup>	Porcupine Bank				
120.0	61.76	-132.95	-10.98	NAM	<sup>a</sup>	200.0	47.29	-13.48	12.65	RHB
83.0	61.42	-132.01	-9.76	NAM	<sup>a</sup>	140.0	73.96	45.49	-0.58	RHB
79.1	57.73	-138.96	-8.00	NAM	<sup>a</sup>	140.0	0.00	0.00	0.00	EUR
Flemish Cap						Rockall Bank				
200.0	63.94	-2.84	69.15	IB		200.0	50.23	122.38	-13.03	GRN
160.0	63.94	-2.84	69.15	IB		120.0	50.17	120.80	-12.88	GRN
160.0	44.65	-54.79	18.83	NAM		120.0	0.00	0.00	0.00	EUR
140.0	45.28	-53.47	20.30	NAM						
112.0	0.00	0.00	0.00	NAM						
Orphan Knoll										
200.0	42.38	-54.00	12.56	NAM						
140.0	42.38	-54.00	12.56	NAM						
130.0	44.97	-52.57	13.20	NAM						
112.0	0.00	0.00	0.00	NAM						

Note. The absence of symbol indicates that the rotation poles are from the presented model. Abbreviations are the same as Figure 5.

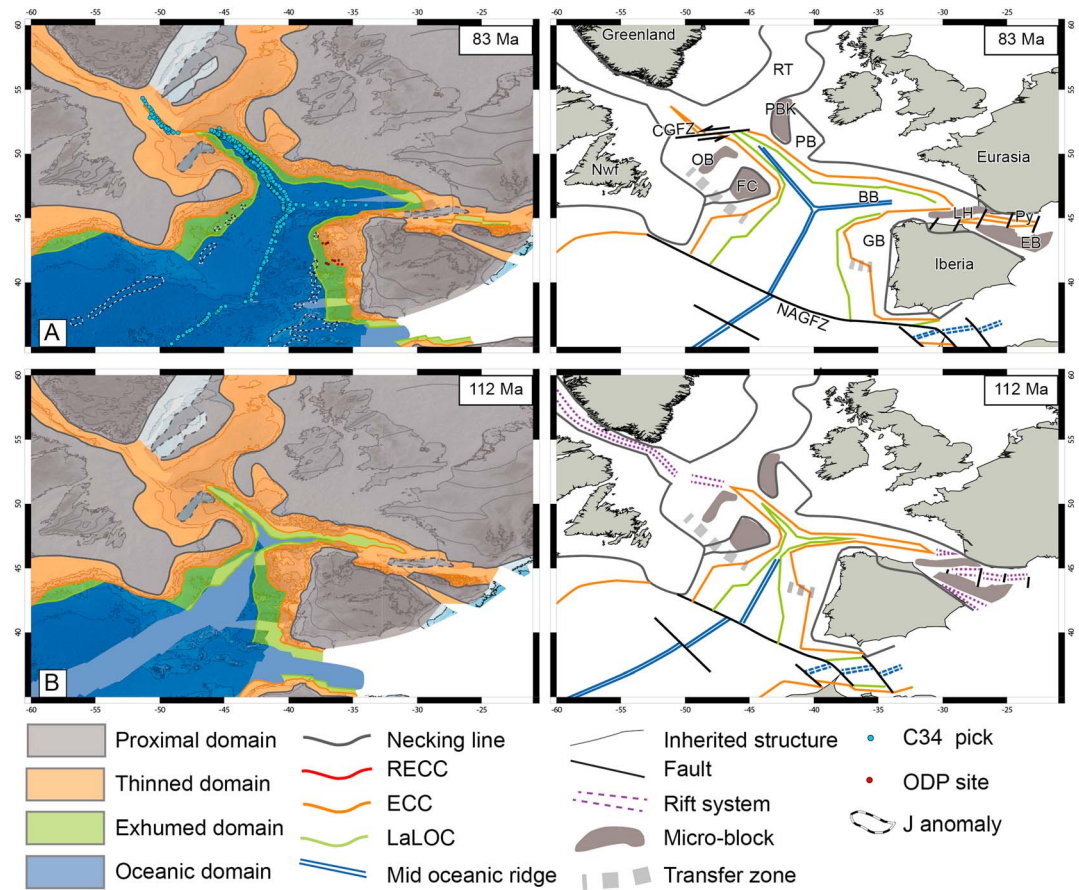
<sup>a</sup>Barnett-Moore, Müller, et al. (2016). <sup>b</sup>Gaina, Roest, and Müller (2002). <sup>c</sup>Vissers and Meijer (2012b).

### 5.1. Constrained Positions: Campanian (83 Ma) and Early Jurassic (200 Ma)

The Mesozoic evolution of the southern North Atlantic is highly debated; however, there is a general agreement on the plate configuration at the Triassic-Jurassic boundary and at the first magnetic anomaly after the Cretaceous normal polarity superchron C34, which is Campanian in age (83 Ma) (Barnett-Moore, Müller, et al., 2016; Vissers & Meijer, 2012a, 2012b). Widespread diffuse Permo-Triassic extensional events occurred in the southern North Atlantic realm, but the associated extension is limited (Leleu & Hartley, 2010; Stolfova & Shannon, 2009); therefore, we approximate the Triassic-Jurassic boundary restoration as the full fit.

#### 5.1.1. Campanian Position: First Constrained Oceanic Magnetic Anomaly (C34)

The post C34 evolution of the Iberian Plate has been recently modified (Vissers & Meijer, 2012a) to unify different studies with a consistent timescale, but most of the rotation poles are from previous studies (Gaina et al., 2002; Srivastava & Roest, 1989; Srivastava & Tapscott, 1986). Hence, continents are confidently positioned (Figure 6a) after C34. The rift propagates north of the Charlie Gibbs fracture zone between the northeastern Newfoundland margin and the southwest Rockall Bank margin and northward through the Labrador Sea, which experienced rift deformation until Maastrichtian time (Dickie et al., 2011). The Charlie Gibbs fracture zone laterally offsets the Mid-Atlantic Ridge and the Labrador rift system by 200 km (Figure 6a). The Pyrenean rifted domain is composed of thinned continental crust and local exhumed mantle. Its width is estimated to about 100 km (for more details see Tugend, Manatschal, & Kuszniir, 2015). This value may be a lower bound and is below the interpreted 140 km of total shortening (50 km subducted and 90 km



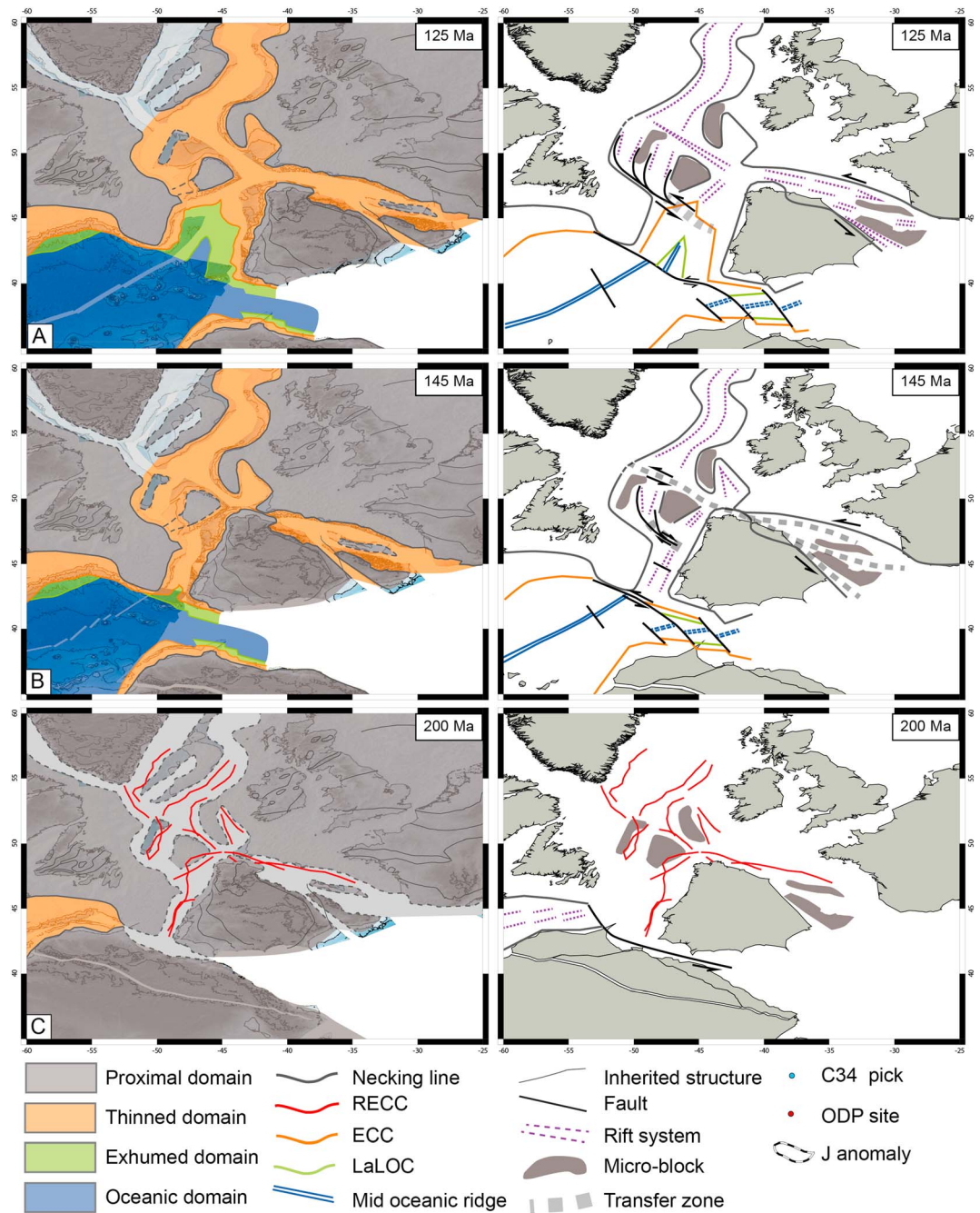
**Figure 6.** Restoration of the southern North Atlantic, with North America fixed on present-day coordinates. (left column) The restoration of the rifted domains and (right column) the plate restoration with the interpreted rift structures at each stage. Re d.: reconstructed domain. (For model animation, see supporting information S5.)

of crustal shortening) proposed by Mouthereau et al. (2014) and Wang et al. (2016) for the Campanian to Late Oligocene convergent phase (Vergés et al., 2002).

### 5.1.2. Triassic-Jurassic Boundary (200 Ma): The Full Fit

Most of the models (Jammes et al., 2009; Olivet, 1996; Sibuet, 2004; Vissers & Meijer, 2012a) agree on the closure of the southern North Atlantic and on the initial Triassic plate configuration. Strictly speaking, the Triassic-Jurassic boundary does not correspond to the onset of extensional deformation because Triassic rift basins are widespread across the central and North Atlantic realm (e.g., Jeanne d’Arc Basin, Celtic Sea, Lusitanian Basin) (see Leleu & Hartley, 2010; Rasmussen et al., 1998; Stolfova & Shannon, 2009; Tankard et al., 1989). Although the Triassic extension phase created sedimentary basins (see Leleu et al., 2016, for a review), the overall related crustal extension is ill constrained but seems relatively limited considering the present-day crustal thickness beneath the sedimentary basins that are not significantly reactivated during Jurassic to Cretaceous time (e.g., Celtic Sea and Western Approaches, Figure 3).

We visually fitted the polygons, limited by the RECC which corresponds to the minimum size of the predeformation plate. The position of northwestern Africa and Greenland relative to North America (Figure 7c) is taken from previous palinspastic restorations (Barnett-Moore, Müller, et al., 2016; Hosseinpour et al., 2013; Labails et al., 2010), but the position of Eurasia and Iberia is not constrained by such studies or by magnetic anomalies prior to C34 (Nirrengarten et al., 2017). The Iberian Plate is first fitted with respect to the North American and African Plates by closing of the Orphan Basin with the Flemish Cap. We restored the Porcupine Basin, Hatton Basin, and Rockall Trough and fitted Eurasia with the other plates from southeastern Greenland to the northern Galician margin. Our Triassic-Jurassic boundary position is very similar to previous



**Figure 7.** Restoration of the southern North Atlantic, with North America fixed on present-day coordinates. (left column) The restoration of the rift domains and (right column) the plate restoration with the interpreted rift structures at each stage. (For model animation, see supporting information S5.)

restorations (Rowley & Lottes, 1988; Barnett-Moore, Müller, et al., 2016). The gap in the future Bay of Biscay is filled by the Ebro Block and Landes High (Tugend, Manatschal, & Kuszniir, 2015). Even if the offshore continuation of Avalonian and Caledonian sutures is approximated, these structures present an offset through the proto-Atlantic (Pollock et al., 2012; Welford et al., 2012) (supporting information S4). Variscan domains show in previous reconstruction a bending structure between Western Europe and Iberia (Ballevre et al., 2014; Martínez Catalán, 2011; Matte, 2001); the correlation can be still valid even if Iberia is located westward compared to these restorations.



### 5.2. Jurassic Time (200–145 Ma): Distributed Deformation and Intracontinental Rifts

Seafloor spreading in the central Atlantic is proposed to have begun at around 190 Ma (Labails et al., 2010). In the plate model, with the North American Plate fixed, the eastward motion of NW Africa was accommodated by strike-slip motion south of the Iberian Plate. During Middle Jurassic oblique rifting is observed in the onshore Algarve Basin (Ramos et al., 2015) (Figure 1) and related to the motion of Iberia relative to NW Africa. Mantle exhumation and formation of oceanic crust in the Gulf of Cadiz are suggested to bridge the Alpine and Atlantic oceanic domains south of the Iberian Plate (Frizon de Lamotte et al., 2011; Martínez-Loriente et al., 2014; Sallarès et al., 2011). Our model suggests a maximum width of 300 km between south Iberia and North Africa at 145 Ma (Figure 7b).

Early Jurassic rifting follows the Triassic extension phase, which is recorded in several basins located over Newfoundland Grand Banks, the Iberia Peninsula, and on the Eurasia continent (e.g., Lusitanian, Jeanne d'Arc, Porto, Horseshoe, Whale, Porcupine, and Rockall basins) (Rasmussen et al., 1998; Shannon, 1991; Tankard et al., 1989). These sedimentary basins are typically grabens and asymmetric half grabens filled by subareal to shallow marine sediments (Murillas et al., 1990; Pereira & Alves, 2011). These basins are mostly oriented NNE–SSW following the trend of the later central Atlantic mid-oceanic ridge. The Jurassic extension is recorded through several rift basins that are discontinuous in time and space. It is implemented in our model by a slow eastward motion ( $<0.2$  cm/yr) of Eurasia and Iberia relative to North America from 200 Ma to 140 Ma. This motion is located between Iberia and the Grand Banks, in the East Orphan Basin and in the Rockall Trough (Skogseid, 2010). Strike-slip motion is modeled north of the Iberian Plate in the Pyrenean and Central Iberian rift systems. The Porcupine Basin is obliquely oriented compared to the Iberia–Newfoundland rift system, and its main rift event is suggested to be Middle to Late Jurassic (Shannon, 1991; Tate et al., 1993), which creates a sinistral transfer zone on the northeastern margin of Flemish Cap. The formation of the Porcupine basin is modeled by a V-shape opening; however, it causes compression between the Porcupine Bank and the Irish Shelf. Therefore, we suggest that the Porcupine Basin is likely to be highly segmented, enabling a differential extension without a rotation of the Porcupine Bank. However, a transtensional opening cannot be excluded.

The second extension phase between Iberia and Newfoundland starts in the Late Jurassic. This phase is diachronous, starting to the south in the Alentejo Basin at Callovian–Oxfordian time and in Berriasian time at the Galicia Margin (Alves et al., 2009). It initiated the motion of the Ebro Block and Landes High, leading to extensional and transtensional structures in the Bay of Biscay–Pyrenean corridor (Tavani & Muñoz, 2012; Tugend, Manatschal, & Kuszniir, 2015) and in the Central Iberian rift system (Salas & Casas, 1993).

### 5.3. Tithonian to Barremian (145–125 Ma): Necking, Hyperextension, and Exhumation

The snapshot (Figure 7b) at the Jurassic–Cretaceous boundary (145 Ma) presents an important change in the kinematic framework with the cessation of seafloor spreading between Iberia and North Africa (Schettino & Turco, 2011). It also corresponds to the time when the Mid-Atlantic Ridge propagated northward across the south Newfoundland transform margin. This phase induces in our model an increase of the eastward motion of Iberia relative to Eurasia at 140 Ma.

At 145 Ma the Iberia–Newfoundland rift system is still intracontinental. During Early Cretaceous the N–S segmentation of the deformation in the Iberia–Newfoundland rift system is well developed (Alves et al., 2009; Welford, Smith, et al., 2010). This is best observed within the sedimentary record between the Galicia–Flemish Cap system and the Iberia Abyssal Plain. In the first system, based on the deepening of the sedimentary depositional environments in ODP boreholes (e.g., Boillot et al., 1987; Tucholke et al., 2007) and subsidence analysis, the necking occurs in Valanginian to Hauterivian (140–130 Ma), whereas farther south it occurred during Tithonian to Berriasian times (150–140 Ma) (Mohn et al., 2015). This delay in the necking phase is interpreted by a differential extension, accommodated by the opening of the East Orphan subbasin (Enachescu, 2006) along a transfer zone located in the Flemish Pass Basin and south of the Galicia Bank (Figure 7b; see supporting information). On the same two segments, the phase of hyperextension is also decoupled and is younger northward (Mohn et al., 2015). In the model, extension in the East Orphan subbasin progressively decreased and migrated into the West Orphan subbasin. However, the structural and time relations between those two subbasins remain unclear due to the lack of public drill hole data. Mantle exhumation initiates at the southernmost part of the Iberia–Newfoundland system around 140 Ma, which corresponds to the crystallization age of gabbros in host peridotites on the Gorrington Bank

(145 Ma to 137.5 Ma) (Féraud et al., 1986; Schärer et al., 2000). In our model, mantle exhumation is quasi-synchronous at 140 Ma from the NAGFZ to latitude 45.5° (Figure 7b). Farther north, the model shows that mantle exhumation may be as young as 120 Ma (Figure 7a). This is due to the rotation and eastward movement of the Flemish Cap as a consequence of the extension in the Orphan Basin. The extension in East and West Orphan compensates the extension south of the Flemish Cap in the Iberia system. One of the major outputs of our model is that at 129 Ma our mapped LaLOC crosses north of the NAGFZ. We interpret this as the initiation of the V-shape propagation of seafloor spreading (Figure 7a). In our model, the Bay of Biscay-Pyrenean domain encompassed a strike-slip regime and accommodated the 1.6 cm/yr of left-lateral movement relative to Eurasia. Extension between North America and Eurasia is similar in magnitude and direction since the Late Jurassic phase and located mostly in the Rockall Trough. Interpretations of seismic data of the northeastern Flemish Cap margin suggest transfer fault structures (Welford, Hall, et al., 2010), which are explained by strike-slip motions between the Flemish Cap and the Porcupine Bank. In our model at the Jurassic-Cretaceous boundary the East Orphan Basin is aligned with Rockall Trough and the Orphan Knoll is located next to the Rockall Bank.

#### 5.4. Aptian (125–112 Ma): Seafloor Spreading Propagation

The Aptian evolution (Figures 7a and 6b) is marked by the northward propagation of seafloor spreading in the Iberia-Newfoundland conjugate system. Our model suggests that the tip of the propagator reached Galicia Bank at 113 Ma, in agreement with the 115 Ma old mid-ocean ridge basalt dyke drilled at ODP Site 1277 (Eddy et al., 2017) located ~100 km to the south. Indeed, this means that the early seafloor spreading on the southern part of the system is compensated by hyperextension in the Orphan Basin and mantle exhumation in the Flemish Cap Galicia Bank system. The left-lateral motion of Iberia is accommodated in the Bay of Biscay by transtension. The Parentis Basin and the Central Iberian rift are developing at this stage (Salas & Casas, 1993; Tugend, Manatschal, & Kuszniir, 2015), which inhibit extension between Eurasia and the Ebro Block in the Pyrenean domain. The model suggests a transtensional setting in the Pyrenean rift system. The North Galicia margin rifted away from the Goban Spur margin and experienced a strike-slip regime with the motion of the Flemish Cap. This motion is consistent with the highly asymmetric conjugate margins between Goban Spur and Flemish Cap (Gerlings et al., 2012). Rifting continued in the Rockall Trough until the Barremian/Aptian boundary and also between Newfoundland and Ireland where extension was first accommodated by crustal oblique deformation and later by mantle exhumation.

#### 5.5. Albian-Santonian (112 Ma–83 Ma): Northward Propagation and Pyrenean Hyperextension

This stage (Figures 6a and 6b) presents the propagation of seafloor spreading in the Bay of Biscay and between Newfoundland and Ireland resulting in stable simultaneous accretion at three oceanic ridges (Roest & Srivastava, 1991). Our model highlights a transtensional extension between Eurasia and the Ebro Block. Albian to Cenomanian time should correspond to hyperextension and mantle exhumation in the Pyrenean rift system (Jammes et al., 2009; Lagabrielle & Bodinier, 2008; Lagabrielle et al., 2010; Masini et al., 2014). This period coincides with the start of movement along the Charlie Gibbs fracture zone (Figure 6a), which separates the oceanic domain between Orphan and Porcupine Bank and the ongoing rift system between the northeastern Newfoundland margin and Rockall Bank. This fracture zone created a narrow transform margin between southern Rockall Trough and the West Orphan Basin with significant magmatism present on both sides (Keen et al., 2014). The Labrador rift system began its activity and accommodated the slow extension between Eurasia, North America, and Greenland, which was part of Eurasia at this stage. Extension in the Labrador rift system is concomitant to the cessation of extension in the hyperextended basins along the future Norwegian-Greenland margins (Gabrielsen et al., 1999).

## 6. Discussion

### 6.1. Method and Limitations

We propose a kinematic model to describe the rift evolution of the southern North Atlantic from the Triassic-Jurassic boundary to the first undisputed oceanic magnetic anomaly C34 (83 Ma, Campanian/Santonian, Figure 6a). Restoring rift systems is less accurate than reconstructing oceanic domains. However, the integration of critical data from rifted margins in a kinematic model can provide spatial and time constraints on the kinematics of the prebreakup stage of the future margins. Our method is based on the knowledge of small areas and their integration in a much larger kinematic framework including different tectonic plates and

microblocks. The first limitation is the variable quality and resolution of the spatiotemporal input data available in the southern North Atlantic. As a consequence, the model is not everywhere constrained by data with the same resolution. However, while the model needs to fit constrained areas (e.g., Iberia margin), it can be used as a predictive tool for unconstrained areas (e.g., Orphan Basin) and the model predictions may be tested by future scientific work or existing and new proprietary industry data.

Other issues are related to the use of “ages” and the definition of limits of polygons. As the southern North Atlantic does not have well-defined oceanic magnetic anomalies prior to C34 (Klitgord & Schouten, 1986; Nirrengarten et al., 2017) and as most of the time constraints used are from stratigraphy, we choose to follow the stratigraphic time chart of Gradstein et al. (2012). Moreover, the integration of geochronological data is not straightforward. It corresponds either to the time of crystallization or to the cooling of a mineral, and the interpretation of these ages has, therefore, important implications on the kinematic model. Hence, the ages and the way they are geologically interpreted and correlated remain sensitive to interpretation and therefore to the input data used in a model.

The determination of the shape and size of the polygons (Williams et al., 2011; Barnett-Moore, Müller, et al., 2016) relies on the areal balancing of the crustal thickness calculated from 3-D gravity inversion. In this study, we chose a rather high initial thickness of 37.5 km (Figure 2). This inversion is calibrated and is highly influenced by the accuracy of the sediment thickness grid NOAA (Divins, 2003). As 37.5 km is the maximum crustal thickness of the proximal domains determined by refraction studies (Lau et al., 2006), the polygon size determined in our work has the tendency to be underestimated and serves to minimize the overlap during the visual fitting. With an initial crustal thickness of 30 km, which is the thinnest estimation for non-rifted continental crust, the distance between the NL and the RECC would have increased by 20%. The domain mapping can further induce a bias in the determination of polygon shapes. Indeed, the necking line has to coincide with the Moho slope break; otherwise, a part of the thinning is omitted or added. Crustal thickness maps and profile extractions from these maps highlight clearly the location of this structural marker. Even if most of the extension and thinning occur between the two conjugate necking lines, the southern North Atlantic experienced extensional deformation since Late Carboniferous-Permian time (Stolfova & Shannon, 2009). Although the amount of extension is difficult to quantify accurately, we consider that it falls within the error bars/uncertainties of our model. In contrast to the necking line, the ECC is more difficult to determine with gravity inversion methods (Cowie et al., 2015) and access to high-quality seismic sections imaging deeper than 10 s TWT is needed. However, a mislocation of the ECC has a minor impact on the determination of the RECC. Since the continental crust is very thin close to the ECC, the surface/volume to restore is small as well as the induced shift in the RECC location. For instance, a miss-location by 50 km of the ECC only modifies the position of the RECC by 8 km assuming an average crustal thickness of 6 km for the distal domain. The RECC position is relatively insensitive to the orientation of extracted profiles along flow lines (Williams et al., 2011) unless the flow lines are completely oblique (Barnett-Moore, Müller, et al., 2016). To minimize this problem, we extracted “pseudo flow lines” perpendicular to the necking line (which are not necessarily the flow lines extracted from our model) along which the crustal thickness is restored.

## 6.2. Implications for the Southern North Atlantic

### 6.2.1. Partitioning of the Deformation Along the Iberia-Eurasia Plate Boundary

Previous kinematic models using the  $J$  anomaly as the magnetic anomaly M0 (Sibuet, 2004; Srivastava et al., 2000; Vissers & Meijer, 2012b) are inconsistent with the well-constrained Albian extension in the Pyrenean domain (Jammes et al., 2009; Lagabrielle & Bodinier, 2008; Masini et al., 2014; Vacherat et al., 2016) and instead propose an oceanic subduction. Paleomagnetic data support a counterclockwise rotation of Iberia of 35° (Gong et al., 2008; Van der Voo, 1969) but present inconsistencies with the global apparent polar wander path during Cretaceous time (Neres et al., 2013). Moreover, because the Iberian Plate is subdivided into different microblocks (Landes High and Ebro Block), it is difficult to evaluate the relative rotations of the subdomains and therefore to appraise the validity of the paleomagnetic data in the northeastern part of Iberia. Hence, following Barnett-Moore, Müller, et al. (2016), we prefer not to take into account paleomagnetic data. Olivet (1996) proposed a restoration of the Iberian Plate without using the M0 ( $J$ ) anomaly, as he already recognized the weakness of the interpretation of the  $J$  anomaly between Iberia and Newfoundland. Recent studies in the Pyrenees (Jammes et al., 2009; Masini et al., 2014; Tugend et al., 2014) highlight that transfer

faults are directed SSW-NNE almost perpendicular to the strike-slip motion, which means that some of the N-S extension has to be integrated in kinematic models (Jammes et al., 2009). A recent review of kinematic models of the Mesozoic evolution of the Iberian Plate by Barnett-Moore, Müller, et al. (2016) concludes that none of these models are consistent with the available geological and geophysical data. Nirrengarten et al. (2017) were able to demonstrate that the *J* magnetic anomaly is polygenic and does not correspond to an isochron, which disqualifies this anomaly as a kinematic indicator. The interpreted orthogonal extension suggested along the SSW-NNE oriented Pampolona fault (Jammes et al., 2009) was one of the targets of the modeling. The eastward motion of the Iberian Plate relative to the Eurasian Plate is due to the continuous seafloor spreading in the southern North Atlantic. This motion is partly compensated in the Central Iberian rift system and partly in the Pyrenean system creating transtensional corridors prior to the Late Aptian (Tugend, Manatschal, & Kuszniir, 2015). During Aptian (Figure 7a) the Central Iberian rift potentially reached its maximum extension (~75 km), which was able to compensate the opening of the Bay of Biscay and to keep a constant narrow rift system in the Pyrenean system (Lagabrielle & Bodinier, 2008). In our model the Central Iberian rift experienced strike-slip motion (Capote et al., 2002) (Figures 6a and 6b) and the Pyrenean system is a transtensional rift during Albian time. The flow line for the Pyrenean system (Figure 8) indicates that there is more strike-slip motion than orthogonal N-S extension (Jammes et al., 2009). Kinematic markers interpreted from field studies are hardly explained by plate kinematic modeling, a situation potentially due to rift localization and inherited structures. The modeled extension in the Pyrenean system is ~100 km, constant from west to east, which could be consistent with the presence of exhumed mantle (Tugend, Manatschal, & Kuszniir, 2015). Further studies are needed to better constrain the location and kinematics along the boundaries of the Ebro Block in order to refine its initial position and quantify the strike-slip motion predicted by our model and to also assess its evolution during the convergence phase. Our modeled flow lines in the Bay of Biscay indicate an oblique opening (Figure 8). The only alternative model which suggests a north-south seafloor spreading (Figure 8) implies simultaneous compression in the Pyrenees (Vissers & Meijer, 2012b).

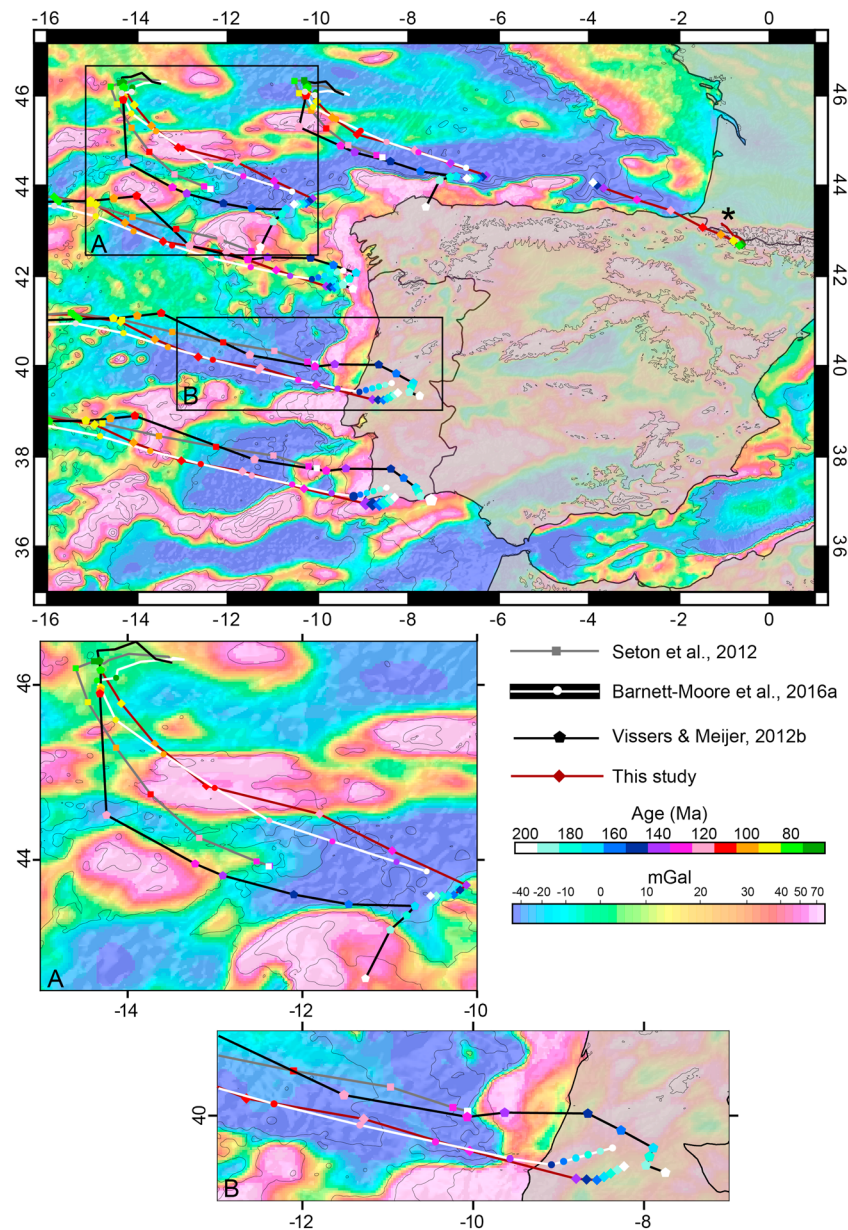
### 6.2.2. Partitioning of the Deformation in the Southern North Atlantic Rift System

The relatively well-constrained data from the Iberia Abyssal Plain, the Deep Galicia, and the northern Newfoundland margins are used as input data in our model. At the Iberia margin the kink of the flow line (Figure 8b) highlights the transition from necking to hyperextension, which is marked by an increase of the extension rate. To compensate for the time lag of extension observed between the deep Galicia margin and the Iberia Abyssal Plain (Mohn et al., 2015), we transfer extension from the southern Iberia-Newfoundland rift system into the Orphan Basin system (Figures 7a and 7b). Extension initiated in the east and progressively migrated westward, resulting in a younger West Orphan subbasin (Enachescu, 2006). The extension in the Orphan Basin is associated with the rotation of the Flemish Cap out of the Grand Banks (Sibuet, Srivastava, Enachescu, et al., 2007) (Figures 7a and 7b). This induced strike-slip motion between Goban-Spur and the Flemish Cap has been suggested by seismic interpretation (Welford, Hall, et al., 2010). Late Jurassic-Early to Cretaceous deformation is suggested to be accommodated also in the Rockall Basin, which is, at that time, located to the north of the Orphan Basin. The evolution of the Porcupine Basin is more difficult to constrain within our model as its orientation is not perpendicular to the main extension trend. Extension between North America and Eurasia localized after mid-Aptian time (120 Ma) in the Labrador rift system, concomitant with the cessation of rift activity in the Rockall Basin and the West Orphan subbasin at the Aptian-Albian boundary (112 Ma). The Aptian-Albian boundary (Figure 6b) is recorded in the Iberia Newfoundland conjugate margins by a prominent regional stratigraphic unconformity (e.g., Soares et al., 2012). An alkaline magmatic vein drilled at ODP Site 1277 yields an age of  $113.2 \pm 2.1$  Ma (Jagoutz et al., 2007) and was interpreted to be related to an excess magmatic event (Bronner et al., 2011). Further evidence for this unconformity and magmatic activity are found in the Bay of Biscay (Montadert et al., 1979), Goban Spur margin (De Graciansky et al., 1985), Orphan Basin (Dafoe et al., 2017; Gouiza et al., 2016), Galicia margin (Grobe et al., 2014), and Porcupine Basin (Calvès et al., 2012). This time corresponds in our plate modeling to the initiation of a triple ridge junction between the Mid-Atlantic Ridge and the Bay of Biscay ridge. Indeed, interactions between triple junctions, magmatic activity, and uplift are observed on many examples (Burke & Dewey, 1973), but the relationship between those processes remains disputed.

### 6.3. Propagation and Distribution of Deformation Modes

Our modeling of the opening of the southern North Atlantic enables us to examine the propagation mode of each rift domain, the role of inheritance, segmentation, and the associated velocities. In the following





**Figure 8.** A comparison of age-coded tectonic flow lines around Iberia extracted from different plate kinematic models plotted on free air gravity map (Sandwell et al., 2014). The flow line with the cross corresponds to the motion of the Ebro Block. (a) Tectonic flow lines of Iberia relative to Eurasia and (b) tectonic flow lines of Iberia relative to North America.

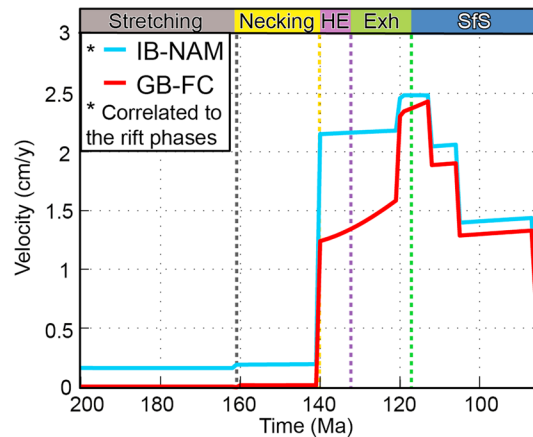
discussion, we describe the extension velocities (full rate) of each deformation phase using the timing of deformation of the Iberia Abyssal Plain and its conjugate northern Newfoundland margin.

**6.3.1. Distributed Rift Deformation**

During the stretching phase (Figure 10cl), extension is not localized and the deformation is characterized by a wide rift mode (Buck, 1991) in which several half-graben basins formed simultaneously. In the modeling, the velocity of this phase (200–160 Ma) is about 2 mm/yr. Proximal basins are mostly oriented perpendicular to the main extension (e.g., Jeanne d’Arc and Lusitanian Basins). However, crustal inheritance can influence the orientation and location of these basins (e.g., Kivu Basin in the East African Rift (Smets et al., 2016)).

**6.3.2. Necking and Hyperextension of the Continental Crust**

The necking process corresponds to the thinning of the continental crust and the excision of midcrustal ductile levels (Mohn et al., 2012). It also corresponds to the localization of the deformation into a narrower zone



**Figure 9.** Extensional velocity of Iberia relative to North America and to the Flemish Cap microblock. The correlation between rifting phases and velocity can only be made on the blue line annotated with the star, which corresponds to the velocities in the Iberia Abyssal Plain. HE: Hyperextension; Exh: mantle exhumation; SfS: seafloor spreading.

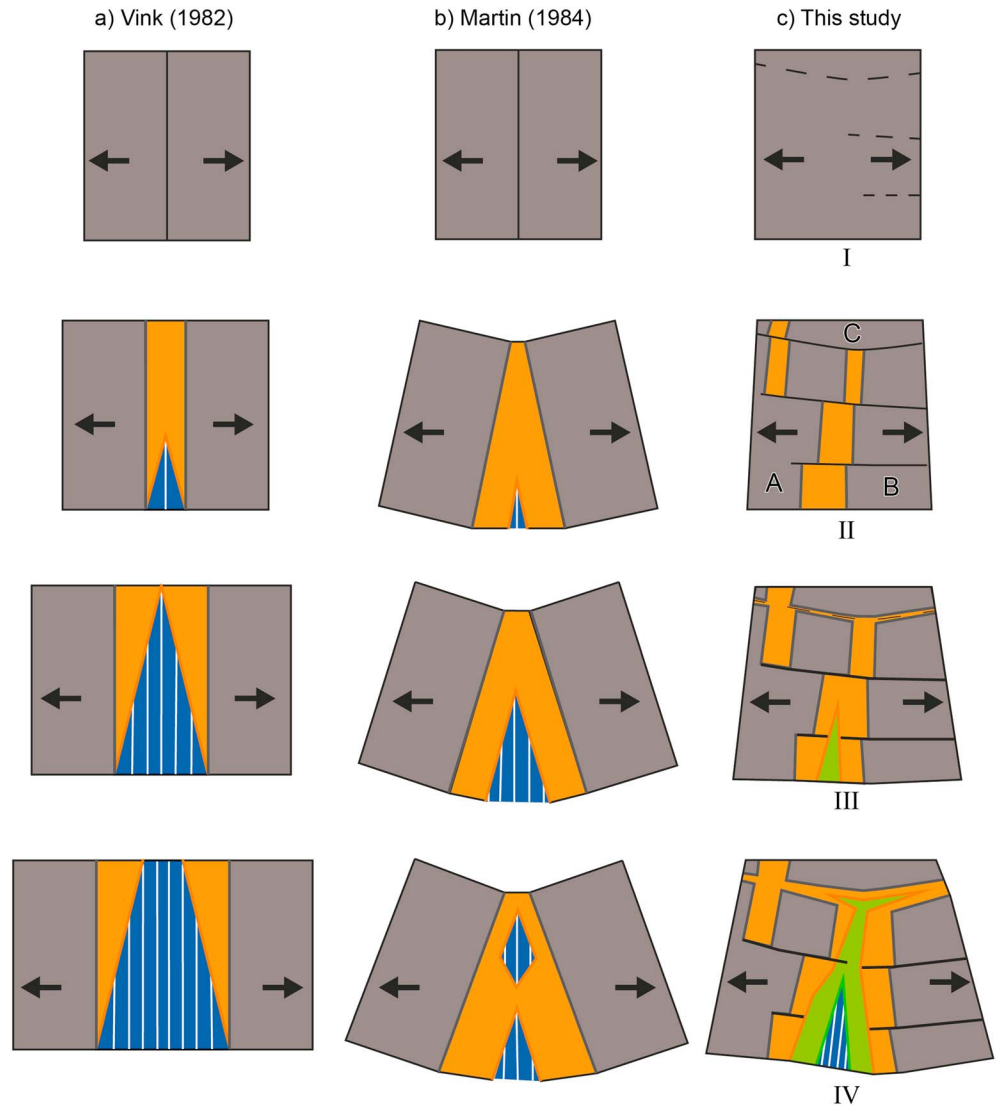
2009; Mohn et al., 2015) within several segments (Welford, Smith, et al., 2010). However, rigid plate kinematic modeling cannot well represent this propagation and segmentation using a single rotation pole for Iberia. Each segment is limited by transfer faults, which seem, on the Iberia side, preferentially located on Variscan domain boundaries (e.g., Balleve et al., 2014). The only difference in the velocities of Iberia relative to North America is the northward decreasing velocity due to the increasing proximity to the rotation pole. Nevertheless, it should be noted that during this phase, deformation is not only accommodated on a single straight rift system (Figure 10cII). The deformation south of the Galicia Bank is accommodated to the north in two different rift systems (Orphan Basin and the Flemish Cap Galicia Bank rift, Figures 7a and 7b), which caused a major transfer zone south of the Flemish Cap (through the Flemish Pass Basin) potentially linked to an inherited structure.

The transition from the necking phase to the hyperextension phase corresponds to a change from a decoupled ductile-dominated crustal deformation to a fully brittle/coupled deformation at the scale of the crust (Pérez-Gussinyé & Reston, 2001; Sutra & Manatschal, 2012). As a consequence, the main parameters controlling the crustal deformation during final rifting will be the frictional properties of the hyperextended crust (Nirrengarten et al., 2016). This event occurred around Berriasian time in the Iberia Abyssal plain and corresponds in our kinematic model to the major acceleration to 2.2 cm/yr, which is within the range of slow-spreading oceanic crust. This acceleration seems simultaneous along the entire margin based on the velocity extracted from the kinematic model (Figure 9). However, as highlighted by subsidence analysis (Mohn et al., 2015), hyperextension is not simultaneous along strike and the simultaneous step in the velocity graph is due to the use of a unique rotation pole for the Iberian Plate. The acceleration is meaningful, observed on many rift systems and proposed to be related to the decay of the strength of the lithosphere approaching breakup (Brune et al., 2016). In other words, this acceleration and the transition to hyperextension coincided with the mechanical and/or thermal/magmatic weakening of the lithosphere leading to lithospheric breakup. This acceleration is also present in the Orphan Basin, which in our model is kinematically linked to the extension in the Iberia-Newfoundland system.

### 6.3.3. Mantle Exhumation

The onset of mantle exhumation (Figure 7a) corresponds to the creation of a new basement surface and a change in the mode of extension, mainly controlled by exhumation faults associated with serpentization. In our model, mantle exhumation is not linked to a change in extension velocities. South of the Flemish Cap and Galicia Bank the exhumed mantle propagates northward in a very narrow V-shaped domain (the separation is almost simultaneous from the NAGFZ to south of the Flemish Cap/ Galicia Bank), whereas the V-shape propagation has a larger angle between Galicia and Flemish Cap. This V-shape opening (Figure 10cII) is compensated by extension in the Bay of Biscay and between the Irish Shelf and the north-eastern Newfoundland margin, which also undergo a V-shape exhumation during Aptian and Early Albian

(H-block (Lavier & Manatschal, 2006)), which will be later deformed during the hyperextension phase. Therefore, the volume of rocks between the two conjugate necking lines greatly impacts the final width of the thinned continental domain (necking domain and hyperextended continental crust). During this deformation phase our model shows only a slight increase of velocity from 2 mm/yr to 3 mm/yr, which may not be significant for rift evolution. The duration of the necking phase is related to the strain rate and the thickness of the ductile levels in the crust (Sutra et al., 2013). Seismic and stratigraphic studies clearly present that the onset of necking gets younger northward (Alves et al.,



**Figure 10.** (a) Rift propagation model from Vink (1982), (b) rift propagation model from Martin (1984), and (c) propagation model of rift and seafloor spreading of a three-plate system. Note that the rotation of Figures 10a and 10b does not imply compression at the tip of the propagation.

time. The nonsegmented propagation may indicate that inherited structures do not influence anymore the partitioning of the deformation. The exhumed serpentinized mantle deformation is brittle and is only related to the frictional properties, the tectonic regime, and the volume and location of magmatic additions. The transfer zone south of the Flemish Cap is still active during mantle exhumation, due to the partitioning of deformation between the Iberian-Newfoundland and Orphan systems.

**6.3.4. Oceanic Seafloor Spreading**

The onset of steady state seafloor spreading is related to the emplacement of a stable (Cannat et al., 2009) localized spreading ridge (Gillard et al., 2016). The transition from crustal extension or mantle exhumation to seafloor spreading is associated with an increase of the magmatic budget without an acceleration of extension rate. Seafloor spreading propagates northward in a V-shape (Figures 7a and 10cIV) with a velocity of 3.8 cm/yr of the propagation tip. This V-shaped seafloor spreading propagation initiated at the Aptian/Albian boundary in the Bay of Biscay and between the Irish and the northeast Newfoundland margins creating a triple ridge junction. If well-defined magnetic anomalies could be observed (note that this propagation occurred during a magnetic superchron), they might have been oblique to the LaLOCs as observed in the Southern Atlantic (Rabinowitz & LaBrecque, 1979) or in the South China Sea (Barckhausen et al., 2014).

Oceanic propagation does not always correspond to a V-shape and can also be segmented following oceanic transfer zones (e.g., Gulf of Aden (Fournier et al., 2010), Woodlark Basin (Taylor et al., 2009), and South China Sea (Sibuet et al., 2016)).

### 6.3.5. Propagation Model

Onset of seafloor spreading is not instantaneous over a whole rift system and creates V-shape oceanic basins (e.g., South Atlantic, Gulf of Aden, Red Sea, and South China Sea). Some models propose a differential extension of the continental crust where rifted margins are wider toward the tip of the propagator (Vink, 1982, Figure 11a). However, it seems that the variation of the width of the margin is not systematic (Eagles et al., 2015). Another model proposes a V-shape propagation of rifts and breakup (Martin, 1984, Figure 10b), which implies a constant width of the margins along strike and compression at the tip of propagation. This compression is not seen in our case. We propose a modification of the V-shape propagation model based on observations and kinematic modeling of the southern North Atlantic (Figure 10c). We suggest that the initial rift phase is best described by a segmented propagation. Each segment has its own timing and is separated by transfer faults from its neighbor (Figure 10c). This segmentation may be related to crustal or lithospheric inheritance often reactivated in extensional settings (e.g., Sutra & Manatschal, 2012; Tugend et al., 2014). The motion and the area that formed during the necking phase are almost insignificant relative to large rigid plate kinematics (Brune et al., 2016); hence, the relative differential extension along the propagation can be accommodated by intracontinental transfer zones. The exhumed mantle and seafloor spreading propagates in a V-shape style through the rifted domain implying a rotation of the two surrounding plates. However, contrarily to the classical V-shape propagation model (Martin, 1984), the southern North Atlantic is defined by three tectonic plates during its opening. This plate system avoids compression at the front of the propagator by creating extensional domains obliquely oriented to the main propagation producing a triple junction. The location of this oblique extensional system remains intriguing and could be related to a preexisting weakness.

## 7. Conclusion

We reevaluated the kinematic evolution of the southern North Atlantic by incorporating data and knowledge from hyperextended rift systems. This study merges local constraints with large-scale plate kinematic modeling. The resolution of such a model cannot be as strongly constrained as restorations using unequivocal oceanic magnetic anomalies. However, the presented model corresponds to one possible solution that aims to reconcile the discrepancies raised on the Iberian Plate kinematics. Further detailed studies are nevertheless needed to better unravel the strain partitioning and kinematic evolution along the Iberia-Eurasia plate boundary. In addition to the proposed kinematic model of the southern North Atlantic, which provides the framework of this study, the main scientific questions addressed in this paper relate to the rift evolution and breakup propagation. The main conclusions that result from our study are as follows.

1. The propagation of rift deformation during the necking phase is observable but accommodated within several segments bounded by transfer zones (e.g., Iberia-Newfoundland and Orphan rift systems).
2. Mantle exhumation and seafloor spreading propagates in the southern North Atlantic in a V-shape style.
3. The V-shape propagation implies a rotation of the adjacent plate and compression ahead of the rotation pole. We suggest that compression can be avoided by transtensional opening of rift systems obliquely oriented to the spreading ridge in front of the propagation tip. The observation of a V-shape propagator and the transtensional corridor ultimately results in an oceanic triple junction between the Mid-Atlantic Ridge and the Bay of Biscay ridge at the Aptian/Albian boundary.
4. In a regional transtensional system the partitioning of the deformation between several microblocks may enable the creation of hyperextended rift systems (e.g., Pyrenees).

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