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Quantification and restoration of extensional deformation along the Western Iberia and Newfoundland rifted margins

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[1] Many recent papers describe the structure of the Iberia and Newfoundland rifted margins; however, none of them propose kinematic restorations of the complete rift system to quantify the amount of extension necessary to exhume mantle and initiate seafloor spreading. In our study, we use two pairs of cross sections considered as conjugate lines: one across the Galicia Bank-Flemish Cap and the other across the Southern Iberia Abyssal Plain-Flemish Pass. Both transects have been imaged by reflection- and refraction-seismic methods and have been drilled during Ocean Drilling Program Legs 103, 149, 173, and 210. Drilling penetrated parts of the rift stratigraphy and the underlying basement. The cross sections can therefore be considered as the best-documented conjugate transects across present-day hyperextended, magma-poor rifted margins. The aim of this paper is threefold: (1) provide a detailed description of the crustal architecture of the two conjugate sections, (2) define the extensional structures and their ages, and (3) quantify the amount of strain and strain rate accommodated along these lines. This paper proposes a quantitative description of extension along the Iberia-Newfoundland rift system and discusses the limitations and problems in quantifying extensional deformation along hyperextended rifted margins.

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1. Introduction

[2] Numerical models show that extension leading to continental breakup occurs via a series of different tectonic styles. *Huismans and Beaumont* [2002] and *Lavier and Manatschal* [2006] suggested that these different styles reflect changes of the thermal structure and rheological evolution of the lithosphere. Although this idea is generally accepted, it is yet unclear how these different tectonic styles can be mapped in present-day margins and how and when the transition from one style into another occurs. Moreover, no study exists in which the extension accommodated by the different styles has been quantified for a conjugate rift system.

[3] Quantification of extension in rifted margins requires access to refraction- and reflection-seismic data, drill hole data, as well as good assessment of the prerift stage and the kinematic evolution of the rift system. In cases where magma has been added to the crust, erosion removed parts of the rift section or the kinematic transport direction changed during rifting, restoration models conserving mass are inappropriate and cannot be used to quantify extension. Thanks to the absence of main magmatic additions, magma-poor rifted margins are privileged sites to quantify extension leading to continental breakup.

[4] The Iberia-Newfoundland rift system is at present one of the rare examples that fulfills all these requirements and where restorations can be proposed using real data. In this paper, we describe structures and their ages and propose a restoration of these structures with the aim to quantify extension and extension rates accommodated along the Iberia-Newfoundland rift system. The fact that the strain evolution is polyphase, the mechanisms leading to crustal thinning are yet little understood, and reliable strain markers in the crust are difficult to define put some limitations on the precision of our restorations. Identifying and discussing the limitations of restorations is important in order to propose correct plate kinematic reconstructions and model rift evolution.

2. Crustal Structure of the Iberia-Newfoundland Rifted Margins

2.1. Geological Context

[5] The Iberia-Newfoundland rifted margins resulted from the superposition of two rift events, a first event dated as Late Triassic to Early Jurassic and a second, more important event dated as Late Jurassic to Early Cretaceous, leading to the breakup of the southern North Atlantic [*Tucholke et al*., 2007]. The relatively thin sedimentary cover, the absence of salt (at least in the northern and most distal part of the rift system) and the scarce synrift to postrift magmatism enabled to drill basement along this margin and to image the rift structures in the basement. The exceptional data set is at the origin of a considerable advancement in the understanding and development of new concepts for deep-water rifted margins [*Hopper et al*., 2004; *Shillington et al*., 2006; *Van Avendonk et al*., 2006; *Lavier and Manatschal*, 2006; *Reston*, 2009; *Ranero and Pérez-Guissinyé*, 2010; *Huismans and Beaumont*, 2011]. The most characteristic feature of the Iberia-Newfoundland rift system is the existence of a Zone of Exhumed Continental Mantle (ZECM) [*Whitmarsh et al*., 2001] that does neither show continental nor oceanic characteristics [*Péron-Pinvidic and Manatschal*, 2009]. Equally important but less discussed is the occurrence of hyperextended continental crust, $\leq$10 km thick and composed of brittle hydrated crust, and the existence of necking zones marking the transition between the hyperextended crust and the normal thick ($\sim$30 km) crust. The necking zones on both conjugate margins are well imaged in refraction-seismic sections [*Afilhado et al*., 2008; *Lau et al*., 2006; *Van Avendonk et al*., 2009].

2.2. Conjugate Sections

[6] For this study, two sets of conjugated sections were chosen across the Iberia-Newfoundland rifted margins, one referred to as the "Northern" section corresponding to the Study of Continental Rifting and Extension on the Eastern Canadian sHelf Iberia Seismic Experiment (SCREECH 1-ISE 1) transect and the other across the southern margin referred to as the "Southern" section, corresponding to the SCREECH 2-TGS/LG12 transect (Figure 2).

2.2.1. Northern Section

[7] Along the northern transect, the SCREECH 1 and ISE 1 lines were both imaged as reflection- and refraction-seismic lines, allowing to estimate the present-day crustal thickness as well as to determine the first-order crustal structure. The SCREECH 1 line was first published and discussed by *Funck et al*. [2003] and *Hopper et al*. [2004]. The ISE 1 line on the conjugate Galicia margin was acquired in 1997 and first published and discussed by *Zelt et al*. [2003] and *Henning et al*. [2004]. During Ocean Drilling Program (ODP) Leg 103 [*Sibuet et al*., 1979; *Boillot et al*., 1987a], Site 637 was

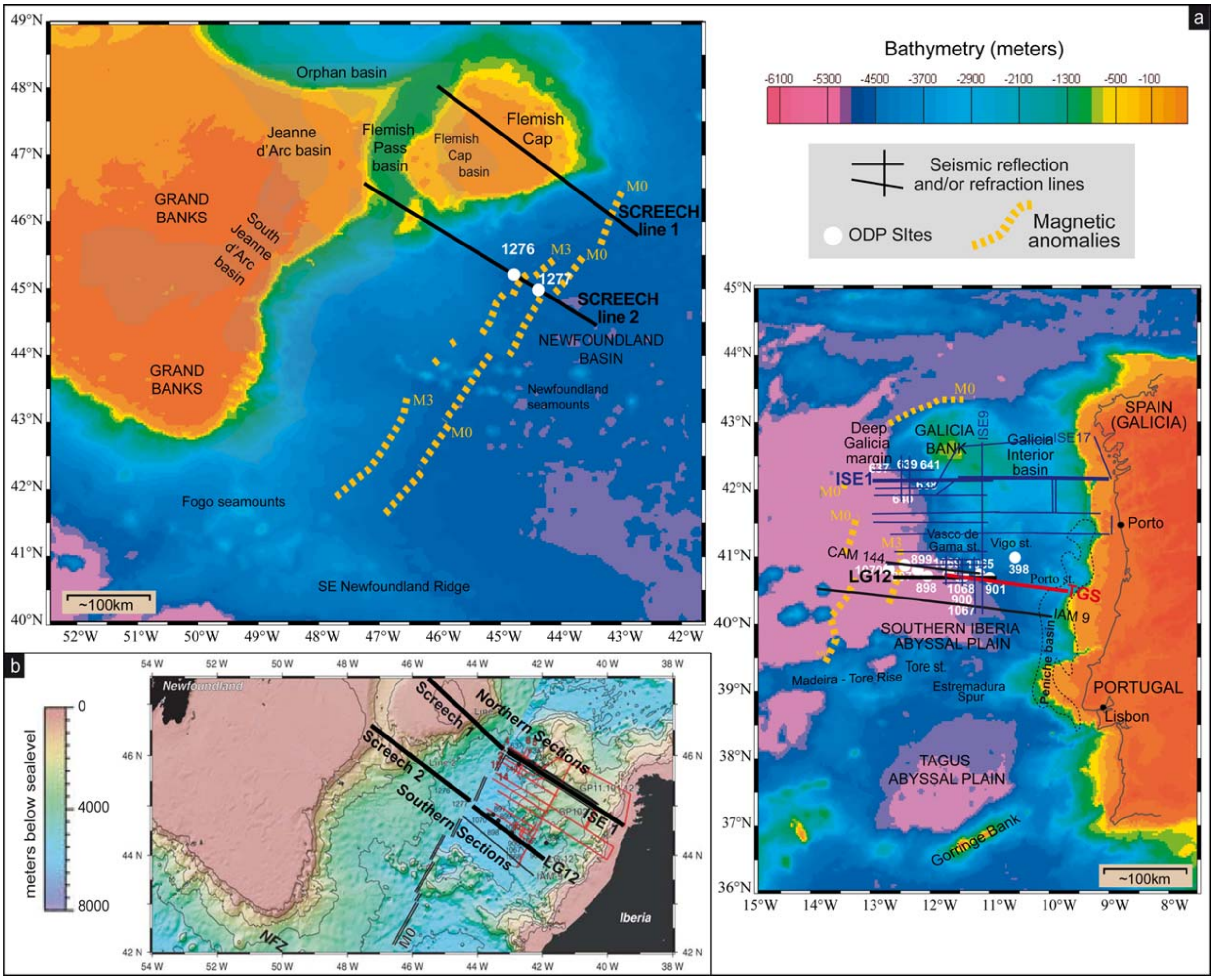

Figure 1. (a) Bathymetric maps of the Iberia-Newfoundland rifted margins with the location of the principal seismic reflection/refraction lines and drill hole data discussed in this paper. (b) Position of the seismic sections discussed in this paper within a bathymetric map of the Iberia-Newfoundland rift system restored at anomaly M0 (modified from *Hopper et al*. [2006]).

drilled exactly along the trace of the ISE 1 section while Sites 638, 639, 640, and 641 were drilled close to the section. In Figure 2, the location of all ODP sites are projected and shown on the ISE 1 section.

2.2.1. Southern Section

[8] Along the southern transect, the SCREECH 2 line was acquired during the same campaign as SCREECH 1 and was first published and described by *Shillington et al*. [2006]. This line connects the Flemish Pass Basin with the Newfoundland Basin. The corresponding velocity model used in this paper is by *Van Avendonk et al*. [2006]. The SCREECH 2 line was drilled at Sites 1276 and 1277 during ODP Leg 210 [*Tucholke et al*., 2004]. For the conjugate Iberian margin, we choose the LG12 and the TGS lines that image the Iberia Abyssal Plain and its continuation into the Peniche basin, the former imaging the ZECM and the most distal hyperextended continental crust, the later imaging the more proximal parts of the hyperextended crust and the necking zone. The LG12 reflection line was acquired in 1990 and described by *Beslier et al*. [1996] and *Krawczyk et al*. [1996]. The TGS line is an industrial acquisition that can be used here with the courtesy of TGS-NOPEC [*Sutra and Manatschal*, 2012]. Unfortunately, the TGS/LG12 sections do not have an equivalent refraction profile; however, a gravity model along this transect was realized by *Cunha* [2008] in the Iberia Abyssal Plain. This model was used to describe depth of Moho along the composite line. Several drill holes of ODP Legs 149 [*Sawyer et al*., 1994] and 173 [*Whitmarsh et al*., 1998] sampled basement rocks along the TGS-LG12 section. Sites 898, 900, 901, 1067, and 1068 were

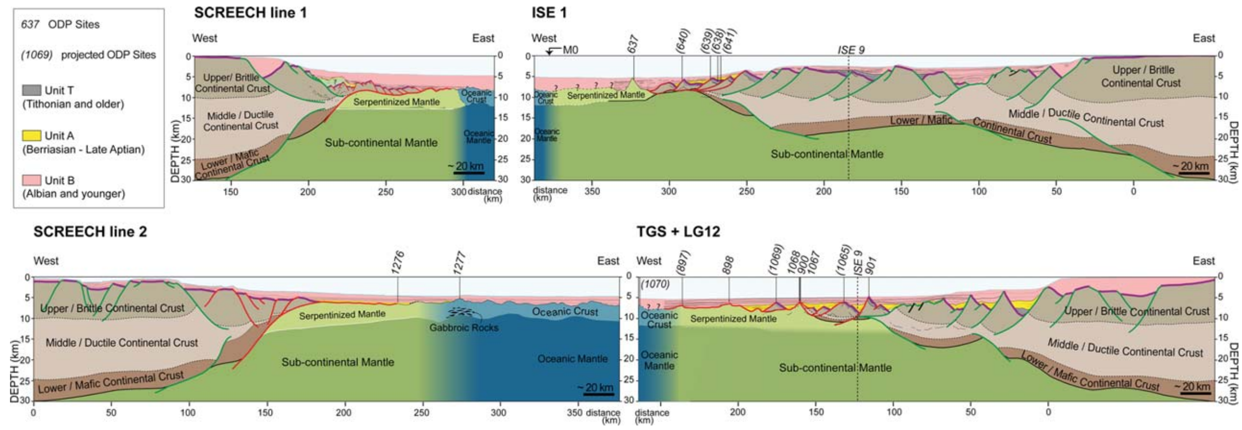

Figure 2. Interpreted sections across the two conjugate pairs of rifted margins: the northern transect corresponds to the SCREECH 1-ISE1 lines and the southern transect to the SCREECH 2-TGS/LG12 lines. The interpretations are based on all available geophysical (seismic reflection and refraction) and ODP drill hole data. For the sediments, see work of *Péron-Pinvidic et al.* [2007], for the subdivision into "upper brittle," "middle ductile," and "lower mafic" continental crust and "serpentinized," "subcontinental," and "oceanic" mantle see text). The continental limits of the lines are either cut or extended (using gravity modeling from *Cunha* [2008]) to the location where crustal thickness reaches about 30 km.

drilled exactly along the LG12 transect (for localization, see Figure 2), while Sites 897, 1065, and 1069 are projected on the line.

2.3. Definition of Stratigraphic and Lithological Units Shown in the Two Transects

[9] In the two conjugate sections across the Iberia-Newfoundland rifted margins, we distinguished among sedimentary sequences, seismic basement (continental and oceanic), and mantle rocks. The further subdivisions are based on geophysical and geological criteria described in the following sections.

[10] *The sediments* have been subdivided, following the results of ODP drilling [*Wilson et al.*, 2001; *Péron-Pinvidic et al,*. 2007; *Tucholke and Sibuet*, 2007] in three major stratigraphic levels that are: present-day to early Albian (postrift); Berriasian to Late Aptian (syntectonic in distal margin, posttectonic in proximal margin) and Late Jurassic (pretectonic in distal margin, syntectonic in proximal margin) (for an overview see *Péron-Pinvidic et al*. [2007]. Older sediments, if present, have been mapped as part of the seismic basement [e.g., *Hölker et al*., 2002]

[11] *The continental crust* has mainly been defined based on refraction-seismic data and, on the more distal parts, using ODP drill hole data. The base of the crust was determined as the transition into velocity 8 km/s, where the continental crust is thicker than 10 km. In the more distal parts, where the crust is <10 km thick, mantle rocks can be serpentinized and consequently show velocities <8 km/s. Therefore, in these domains, we used reflection seismic and drill hole data to determine the base of the continental crust (see later for details).

[12] To subdivide and map different parts of the continental crust, we used two different criteria, one based on the composition (velocity structure) and the other based on the bulk rheological behavior of the crust at the onset of rifting.

[13] *Composition of the crust*: Following *Christensen and Money* [1995] and *Rudnick and Gao* [2003], the continental crust is suggested to be stratified and can in many cases be subdivided into an upper part that has an average granodioritic composition (e.g., predominantly quartzofeldspatic rocks) with expected velocities of about 6 km/s and a lower part that is typically more mafic (e.g., feldspatic rocks) with expected velocities of about 7 km/s. Thus, we mapped all crustal rocks with velocities of ≥ 7 km/s as mafic lower crust and all crustal rocks with velocities of about 6 km/s as quartzofeldspatic upper (brittle) and middle (ductile) crust. We are aware that this is a crude simplification, since a continental crust is in reality much more complex. However, we believe that with this approach we can capture the first-order compositional structure of the prerift continental crust.

[14] *Bulk rheological behavior of the crust*: In the proximal, little affected parts of the rifted margins we distinguish, where possible, between brittle

and ductile layers in the prerift continental crust based on the distribution of decoupling horizons imaged in seismic sections [e.g., *Zalán et al.*, 2011]. Indeed, decoupling horizons should coincide with the location where "brittle" faults sole out in ductile layers and do not offset anymore pretectonic layers. Thus, the depth where large high-angle faults bounding major rift basins sole out in the crust should correspond to the brittle-ductile transition. A good example is imaged beneath the Jeanne d'Arc Basin [e.g., *Tankard et al.*, 1989]. It is commonly accepted that the brittle ductile transition depends on the composition of the crust and the thermal gradient. In the case of a predominantly quartzofeldspatic crust, i.e., domains showing velocities around 6 km/s, this transition is expected to correspond to the temperature at which quartz becomes ductile, which is at about 300°C [*Stipp et al.*, 2002]. In the case of a dominantly feldspatic crust, this transition is expected to occur at ≥500°C [*Passchier and Trouw*, 1996, and references therein]. Thus, in a crust with a velocity of about 6 km/s corresponding to a quartz dominated crust and a thermal gradient of 30°C/km, the brittle-ductile transition should be located at about 10 km depth. *Van Avendonk et al.* [2009, see Figure 12 and references therein] proposed that the thermal gradient is higher on the Iberia margin compared to the conjugate Newfoundland margin. This subdivision of the crust into upper brittle and middle ductile quartzofeldspatic and lower mafic crust is a strong simplification and applies only for the proximal, little affected parts of the rifted margins. In reality, the structure of the crust may be more complex and controlled by compositional and structural inheritance. In addition, the crustal composition and, in particular, the rheology of more extended parts of the crust are likely to be more complex. It is important to note that the rheological behavior of the crustal rocks changes as a function of rifting due to cooling and hydration during thinning and exhumation, which makes that the limits between brittle and "ductile" rocks change during rifting. Because the brittle-ductile transition cannot be considered as a material line, this limit cannot be used as a marker horizon to restore in cross sections. Moreover, it is important to note that the relative volume of brittle and ductile layers will change during rifting; thus, one should not expect that the initial volume of brittle material remains the same during rifting.

[15] *Mantle rocks*: The subdivision of mantle rocks into different subcontinental and oceanic mantle rocks [*Péron-Pinvidic and Manatschal*, 2009] is in reality complex. Therefore, in this work we propose to subdivide the mantle into serpentinized mantle and unserpentinized mantle. Serpentinized mantle rocks occur in the ZECM but may also be present in the hyperextended domain below the thin continental crust. Serpentinized mantle is characterized by gradual transition of *P* wave velocities from 8 km/s [*Minshull*, 2009, and references therein]. Based on refraction data [*Chian et al.*, 1999], the serpentinization front, i.e., where the velocity reaches 8 km/s, is suggested to occur at about 6 km below the top of the basement. However, as suggested by the study of *Bronner et al.* [2011], velocities between 7 and 8 km/s may also be explained by gabbro intrusions into mantle rocks near the place where lithospheric breakup occurred. We put "fresh" unserpentinized mantle where velocities of 8 km/s were imaged in seismic refraction sections.

2.4. Nature of Basement Reflections

[16] The interpretation of the nature of major basement reflections is, in particular in hyperextended domains, a matter of debate. In our interpretation, we define top-basement, top-mantle, and intra-basement reflections.

[17] *The top-basement reflection* is commonly well defined in the less extended proximal domain, where it corresponds to either a stratigraphic contact or to a high-angle normal fault showing a characteristic tilted block geometry. In hyperextended domains, the top-basement reflection is either a stratigraphic contact (e.g., ODP Sites 901, 1065, 1069 and 638, 639, 640, and 641) or an exhumed top basement detachment fault exhuming continental crust (e.g., ODP Sites 900 and 1068) or serpentinized mantle (e.g., ODP Sites 637, 1068, 889, 897, 1270, and 1277). Over oceanic crust, the top of the basement reflection is either the interface of sediments with mafic (accreted) or ultramafic (exhumed) rocks.

[18] *The top-mantle reflect*ion should, by definition, correspond to the transition to rocks with velocities of 8 km/s. In time sections, this reflection often corresponds to a reflection that occurs typically at about 10 s. In hyperextended domains, the interpretation of top mantle is more difficult, since the nature of this limit may be either a tectonic or a primary lithological contact, and the top of the mantle can be serpentinized [*Manatschal et al.*, 2006] and does therefore not correspond to seismic Moho (e.g., velocity 8 km/s). Indeed, in these domains, seismic Moho is within the mantle and does not correspond anymore to a petrologic

Moho (contact between ultramafic and crustal rocks). This is well documented in the ZECM in the Iberia Abyssal Plain, where the drilled mantle is heavily serpentinized showing velocities ranging from 5 to 8 km/s [*Minshull et al*., 2009]. In the two Iberia sections shown in Figure 2, the crust-mantle boundary has been interpreted to correspond to a strong reflection, the "H" reflection in the southern section and the "S" reflection in the northern section [*Krawczyk and Reston*, 1995; *Reston et al*., 1996; *Manatschal et al*., 2001]. Mapping the location where these reflections pinch out at the top of the basement therefore enables to define the oceanward termination of the continental crust. On the Newfoundland margin, exhumed mantle rocks have been drilled at Site 1277 along the southern section. Apart from this drill hole, we do not have further direct evidence for exhumed mantle rocks in the Newfoundland margin, however, in analogy with the Iberia margin, we interpret the strong continentward dipping reflections pinching out at the top of the basement as the boundary between serpentinized mantle and hyperextended continental crust. This interpretation is compatible with the refraction-seismic data presented by *Hopper et al*. [2006] and *Van Avendonk et al*. [2006]. In the ZECM, exhumed mantle can either be overlain by extensional allochthons of continental origin (e.g., block underlying ODP Site 1069 in Figure 2), sediments, or basalts of either tholeiitic or alkaline composition [*Manatschal*, 2004].

[19] *Intrabasement reflections* can be observed across the whole margin, but their position and characteristics change across the margin. In the proximal margin, where the crust often preserves its initial thickness, reflectivity in the crust, if present, is complex and likely related to prerift, inherited structures. In the necking zone, where the crust thins to less than 10 km, reflectivity in the middle crust is often well developed [*Zalán et al*., 2011]. In the hyperextended domain, intrabasement reflectivity is very prominent, and reflections such as "C," "L," "H," and "S" can be determined [*Krawczyk et al*., 1996; *Whitmarsh et al*., 2000]. These reflections are interpreted as either the contact between upper and lower crust (e.g., C reflection; [*Manatschal*, 2004]) or the contact between hyperextended crust and serpentinized mantle (e.g., "S-type reflection; [*Reston et al*. 1996]) or high-angle faults cutting into mantle (L and H reflections; [*Manatschal et al*., 2001]). Within the exhumed mantle domain, reflections are rare, except for seismic line IAM 9 (for localization, see Figure 2), where such reflections were interpreted as faults [*Reston and McDermott*, 2011]. In contrast, intrabasement subhorizontal reflectivity in the oceanic crust is widespread and commonly interpreted as related to lithological boundaries (e.g., three layered oceanic crust [*Penrose*, 1972]).

2.5. Crustal Architecture

[20] The crustal architecture of rifted margins is strongly variable along strike as well as between conjugated margins (see Figure 3) [*Sutra and Manatschal*, 2012]. The observation of characteristic building blocks, identified in each of the sections, can be used to define and map domains. In order to characterize these domains, we use accommodation space (water plus sediments), crustal thickness, stratigraphic and crustal architecture, geometry of rift structures, depth of penetration of faults and their relationship to intrabasement, and top-mantle reflections. For the latter, two cases can be envisaged: (1) faults soling out within the crust, indicating that upper crust and upper mantle are decoupled and that top-basement and Moho topography are unrelated or (2) faults transecting the whole crust indicating that upper crust and upper mantle are coupled and that top basement and top mantle are affected by one and the same fault. In high-quality long-offset reflection-seismic sections where top basement and top mantle are imaged, decoupled and coupled domains can be distinguished and used for the characterization of the margin architecture (Figure 2).

[21] *The decoupled domain*: This domain is characterized by fault-bounded basins (tilted blocks) in the brittle upper crust. The faults bounding these basins sole out at midcrustal levels and consequently do not affect Moho (e.g., Jeanne d'Arc basin) [*Tankard et al*., 1989]. This observation enables us to suggest that deformation is, on the scale of the crust, decoupled. Within the decoupled domain, two subdomains can be characterized:

[22] (1) A first subdomain, referred to as "stretched domain," corresponds to the part of the margin where brittle faulting can be mapped but no major thinning of the midcrustal ductile layers can be observed. In this subdomain, rift basins form over a continental crust maintaining its original (prerift) crustal thickness. As a consequence, top basement and top mantle are, on a larger scale, subparallel. Due to the lack of thinning, only little accommodation space is created across this domain during and after rifting outside the rift-basin.

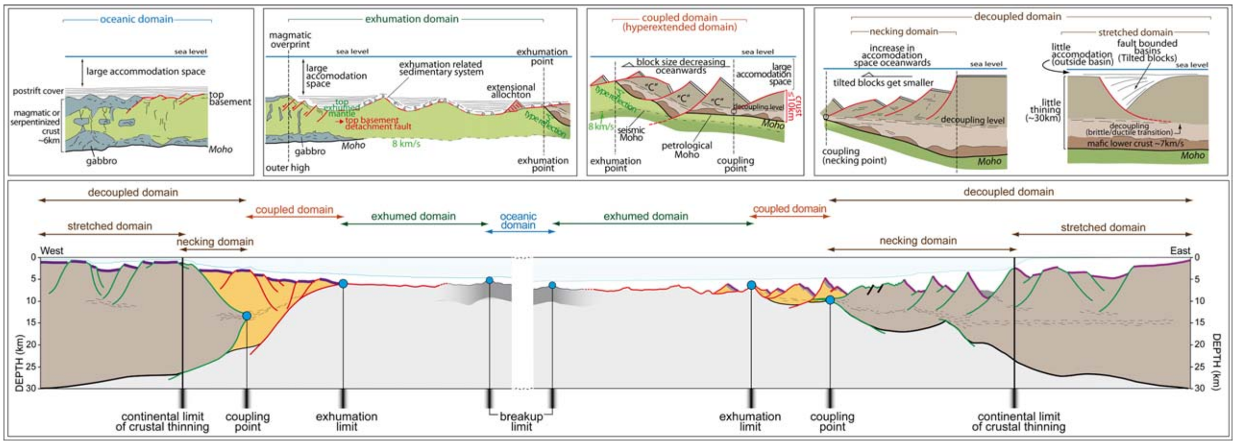

Figure 3. Schematic section across the southern transect (SCREECH 2-TGS/LG12) illustrating the limits and domains defined in this work. Brown domain is the decoupled domain, yellow the coupled domain, light gray the exhumed domain, and dark gray the oceanic crust. Sections above show the geological characteristics of each of the domains.

[23] (2) A second subdomain, referred to as "necking domain," corresponds to the zone where deformation is still decoupled on the scale of the crust, but in contrast to the stretched domain, top basement and top mantle are not parallel anymore but converge oceanward (e.g., taper geometry defined by *Osmundsen and Redfield* [2011]). In this domain, fault-related top basement topography decreases oceanward as a function of the thinning of the brittle upper crust [*Ranero and Pérez-Gussinyé*, 2010; *Sutra and Manatschal*, 2012]. The oceanward limit of this domain is defined by the first brittle faults transecting the crust and penetrating into mantle. This location, referred to in this paper as "coupling point" (e.g., "crustal embrittlement" in *Pérez-Gussinyé et al*. [2003]), limits the decoupled from the coupled domain and corresponds to the location where no ductile layers prevailed in the crust anymore. In many seismic sections, intracrustal reflectivity increases in intensity toward the location where top basement and top mantle (Moho) converge (e.g., TGS line in Figure 3). The continentward limit of the necking domain coincides with the crustal hinge zone, i.e., the location where the continental crust starts to thin considerably and the accommodation space starts to increase. Along the Iberia margin, shape and width of the necking domain changes along strike. In the southern section (e.g., TGS-LG12 in Figure 2), the necking domain shows a gradual narrow transition (single narrow neck), while in the northern section (e.g., ISE 1 in Figure 2), the necking domain is wide and controlled by two necks (double neck) (for more details see *Sutra and Manatschal* [2012]). On the Newfoundland margin, a sharp neck can be observed in the refraction-seismic lines. Comparing the shape and width of the necking domain on the conjugate margins, it appears that the necking domain is wider and more complex on the Iberia side; however, as discussed below, the necking geometry on the Newfoundland margin is less well defined and therefore difficult to compare.

[24] *The coupled domain*: This domain, also referred to as hyperextended domain, corresponds to the zone where brittle faults cut through the complete thinned continental crust and penetrate into the underlying mantle. No remnant ductile layers prevail in the crust, allowing faults to cut from the surface into mantle. This coupling is well illustrated on the TGS line and the ISE 1 lines (Figure 3). In both lines, tilted blocks in the hyperextended domain decrease in size oceanward. These fault-bounded basins are underlain by a strong reflection, referred to as H reflection on the TGS-LG12 lines and the S reflection on the ISE1 line [*Sutra and Manatschal*, 2012]. The hanging wall and footwall of the H reflection have been drilled at Hobby High (see ODP Sites 900, 1067, and 1068 in the TGS-LG12 section in Figure 2), indicating that the H reflection corresponds to the crust/mantle boundary (petrological Moho). Both H and S reflections (in this work grouped as S-type reflection) are pinching out at the seafloor (e.g., TGS-LG12 and ISE1 lines in Figure 2). Therefore, the location where this S-type reflection reaches the surface coincides with the wedging out of the continental crust and the exhumation of mantle at the seafloor. Further oceanward, continental crust may exist but only as extensional allochthons overlying exhumed mantle

[*Manatschal*, 2004] (see ODP Site 1069 in the TGS-LG12 section in Figure 2). It is important to note that shape and width of the coupled domain are very similar along the whole Iberia margin [*Sutra and Manatschal*, 2012]. In contrast, the structures found in the coupled domain on the conjugate Newfoundland margin are very different. In the SCREECH 1 and 2 lines oceanward of the necking zone, continent and oceanward dipping reflections can be observed. Although this basement has not been drilled, refraction-seismic studies [*Hopper et al*., 2006; *Van Avendonk et al*., 2006] suggest that this domain is made of continental crust, highlighting an abrupt thinning of the crust. The crustal block, located between necking zone and exhumation domain, also referred to as "residual H-block" [*Péron-Pinvidic and Manatschal*, 2010], does not have an equivalent on the conjugate Iberian margin. Therefore, it is important to note that the coupled domain is very differently expressed and less well defined on the Newfoundland margin compared to the Iberia margin. This observation points to a strong asymmetry of the two conjugate margins within the coupled domain.

[25] *The exhumation domain*: This domain shows in contrast to the coupled domain major differences. The top of the basement is not anymore formed by fault-bounded basins but either by extensional allochthons or basement highs made of exhumed footwall (see ODP Sites 897, 898, 1070 in the TGS-LG12 section in Figure 2). Moreover, the basement is devoid of intrabasement reflections, except for some very deep, intramantle reflections imaged in the IAM 9 line [*Dean et al*., 2000; *Pickup et al*., 1996; *Reston and McDermott*, 2011]. This domain, which is mainly formed by serpentinized mantle (ODP Sites 637, 897, 898, 1070, 1277), shows a characteristic velocity structure, highlighting a downward increase in velocity, commonly interpreted to reflect a downward degree in serpentinization [*Minshull*, 2009; *Reston*, 2009, 2010]. Magmatic rocks may, however, also occur, as indicated by the finding of basalt clasts in debris flows drilled at ODP Site 1068 [*Whitmarsh et al*., 1998]. Deeper crustal rocks can also be exhumed at the seafloor as shown by ODP drilling (ODP Sites 900 and 1067). Defining the domain of exhumed continental crust is, however, difficult without drill hole information. Therefore, we correlate the continentward limit of the exhumation domain with the pinching out of the S-type reflection at the seafloor.

[26] *The oceanic domain*: This domain is defined in reflection-seismic sections by strong reflections at the base of the lower crust parallel to top basement. The reflectivity in the lower crust, often at about 6 km below top basement, is characteristic and cannot be observed in the ZECM. These reflections may be interpreted as underplated gabbros. Where this reflectivity is missing, the base of the crust may consist of serpentinized mantle. In the case of the Iberia margin, the velocity structure of the oceanic crust is different from the adjacent exhumation domain [*Minshull*, 2009]. Moreover, the average depth of the top of oceanic crust is shallower than the top of the exhumed mantle domain. As a consequence, the transition between the exhumed mantle domain and the first oceanic crust is characterized by a ramp that is also referred to as the outer high [see *Péron-Pinvidic and Manatschal*, 2010, Figure 2]. In the past, the continentward limit of the oceanic domain was either defined by magnetic anomalies [*Dunbar and Sawyer*, 1989] or the seismic velocity structure [*Whitmarsh et al*., 1986, 1990]. Here we follow the work of *Bronner et al*. [2011] and define the limit to first oceanic crust as the outer high over which the "J" anomaly has been imaged. This outer high was drilled at ODP Sites 1070 and 1277. At Site 1277, a post- to synexhumation sedimentary sequence was drilled, which is tilted continentward and onlapped by Albian sediments [*Péron-Pinvidic et al*., 2007] (Figure 4). Moreover, drilling showed evidence for polyphase magmatic activity [*Jagoutz et al*., 2007]. Refraction-seismic data [*Van Avendonk et al*., 2006] show that the crust thickens beneath the outer high. *Bronner et al*. [2011] reinterpreted the J magnetic anomaly overlying the outer high as an edge effect of a latest Aptian-earliest Albian excess magmatic event that resulted in magmatic underplating and continental breakup. Thus, the J-anomaly does not correspond, in their model, to a classical oceanic anomaly but to the continentward limit of the oceanic domain. This has an important implication for the age and mechanism of lithospheric breakup.

3. Time Constraints and Rift Evolution

[27] *Tucholke and Sibuet* [2007] and *Péron-Pinvidic et al*. [2007] reviewed the timing of rift events along the Iberia-Newfoundland margins. These authors highlighted the change of the age of the syntectonic sequences across the margins, showing that the concept of a breakup unconformity is not

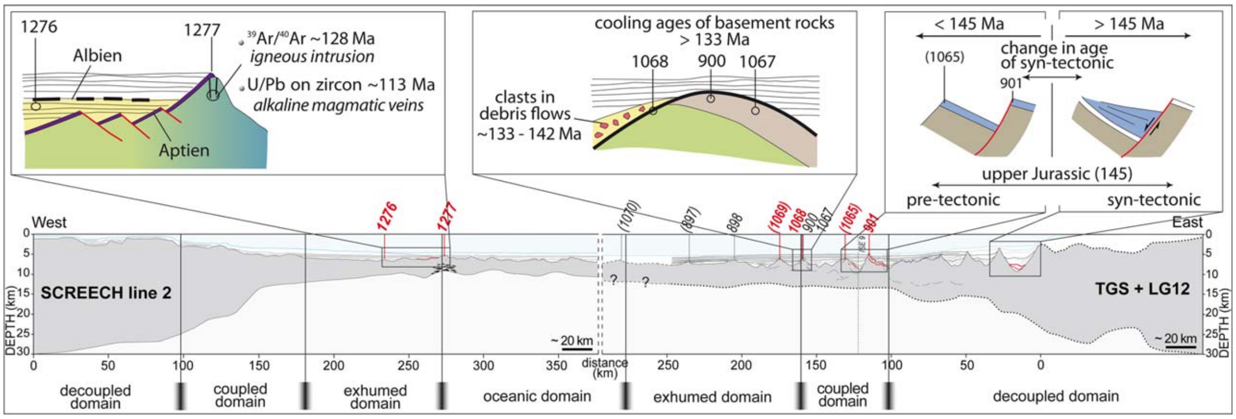

Figure 4. Key observations made across the southern transect (SCREECH 2-TGS/LG12) enables to define the time constraints for the major tectonic activity in each of the domains and to date the evolution of these domains during the formation of the margins. Migration of extension from the decoupled to the coupled domain is dated by the change from a syntectonic to a pretectonic geometry of the Upper Jurassic sediments. The exhumation is dated by cooling ages of rocks drilled at ODP Sites 1068, 900, and 1067, and the maximum age of the debris flows sealing the first exhume mantle at ODP Site 1068. The breakup has been determined using the unconformity between the Upper Aptian and Lower Albian sediments between ODP Sites 1276 and 1277 and the ages derived from ODP Site 1277.

applicable at the scale of the whole margin. *Bronner et al*. [2011] investigated the magnetic anomalies that were previously interpreted by *Sibuet et al*. [2007] as belonging to the M series. The new interpretation by *Bronner et al*. [2011] suggests that these anomalies do not correspond to seafloor-spreading anomalies and, consequently, cannot be interpreted as isochrons. Thus, the classical approach used to establish timing of rifting, based on the identification of the synrift sequence and the breakup unconformity in the proximal margin and the first magnetic anomaly over oceanic crust, does not work for the Iberia-Newfoundland rift system. In this study, we used mainly drill hole data to establish timing of rifting within the different domains of the margin. All ages referred to in the following will use the *Gradstein et al*. [2004] time scale.

[28] *Decoupled domain*: Age constraints for rift activity in the decoupled domain, including the stretching and necking domains, are based on the identification of syntectonic units (*Alves et al*. [2006] for the Iberia margin and *Tankard et al*. [1989] for the Newfoundland margin). In the proximal basins (e.g., Lusitanian basin), the youngest syntectonic sequence defined by growth structures is dated as Early Oxfordian to Berriasian (161.2 ± 4.0 to 140.2 ± 3.0 Ma) [*Alves et al*., 2006] (Figure 4). However, onset of rifting is ill defined, and it is not clear how this event is related to the Triassic and Early Jurassic rifting documented onshore. It is also important to note that along both conjugate margins, rifting stopped in the decoupled domain before onset of mantle exhumation further oceanward.

[29] *Coupled domain*: Timing of rifting in the coupled domain is constrained by ODP drilling in relation with reflection-seismic imaging [*Wilson et al*., 2001]. In both the northern and southern sections, the youngest sediments showing pretectonic relationships are Tithonian in age (150.8 ± 4.0 to 145.5 ± 4.0Ma). At ODP Site 639D in the northern section, the Tithonian sediments are made of biohermal dolomites and shallow water limestones that overlie directly Variscan basement. In the southern section, the Tithonian sequence was drilled at ODP Sites 901, 1065, and 1069 (Figure 4). Since in both the northern and southern sections the Tithonian sediments are tilted and consequently predate major extension and block rotation, we assume that brittle faulting had to initiate along both sections after Tithonian within the coupled domain. As a consequence, onset of block rotation in the future distal margin initiated at both sections in the Early Cretaceous, as indicated by the occurrence of Berriasian to Aptian sediments 145.5 ± 4.0 to 112.0 ± 1.0 Ma) showing growth structures [e.g. *Péron-Pinvidic et al*., 2007; *Tucholke and Sibuet*, 2007; *Whitmarsh et al*., 2001].

[30] *Exhumation domain*: The timing of exhumation is derived from the youngest sediments deposited on exhumed basement, providing a minimum age for this process. A maximum limit is given by

cooling ages obtained from the crustal rocks belonging to the footwall of the exhumation fault. The oldest dated sediments deposited over exhumed mantle in the Iberia margin are Barremian to Aptian in age (130.0 ± 1.5 to 112.0 ± 1.0 Ma) (Sites 897 and 1070). However, these sediments are drilled on highs within the exhumation domain; therefore, exhumation may have initiated earlier. The basal unit described at ODP Site 1276 along the SCREECH 2 line, dated as latest Aptian to earliest Albian (about 112.0 ± 1.0Ma) [*Tucholke et al.*, 2004], is onlapped by the Albian sequence on the Western side of the outer high [see *Bronner et al.*, 2011, Figure 2]. The Albian sequence is intruded by sills drilled at ODP Site 1276. These sills have been dated at 105 and 98 Ma [*Hart and Blutsztajn*, 2006], respectively.

[31] Accepting the idea that the Iberia margin was in a lower plate position relative to the Newfoundland margin during exhumation [*Péron-Pinvidic and Manatschal*, 2009], the oldest sediments related to exhumation should be expected on the Iberia margin. The sediments drilled at ODP Site 1068 are made of debris flows containing continent-derived basement clasts that yield Ar/Ar ages on Plagioclase, interpreted as cooling ages, ranging between 142 and 133 Ma [*Jagoutz et al.*, 2007] (Figure 4). These ages fall in the same age range like those drilled at ODP Sites 900 and 1067 within the continental basement [*Manatschal et al.*, 2001; *Jagoutz et al.*, 2007]. These results suggest that these crustal basement rocks had to be exhumed after 133 Ma. Furthermore, microfossils derived from the sedimentary matrix of the debris flows drilled at ODP Site 1068 were dated as Barremian [*Wilson et al.*, 2001]. Thus, Barremian (130.0 ± 1.5 to 125.0 ± 1.0Ma) represents a minimum age for the exhumation of these rocks at the seafloor. This is also in agreement with the observations of *Péron-Pinvidic et al.* [2007] along the CAM144 and LG12 seismic lines, showing that the basal syntectonic unit range in age from Valanginian to late Aptian (140.2 ± 3.0 to 112.0 ± 1.0 Ma). Thus, exhumation of crustal and mantle rocks may have started as early as Valanginian (140.2 ± 3.0 to 136.4 ± 2.0 Ma) [*Manatschal et al.*, 2001]. The basement topography observed in the ZECM was acquired after exhumation, most likely during latest Aptian time, but before early Albian (about 112.0 ± 1.0 Ma). This is indicated by early Albian to Cenomanian sediments (112.0 ± 1.0 to 93.5 ± 0.8 Ma) passively onlapping onto basement highs covered by Aptian sediments [*Péron-Pinvidic et al.* 2007; *Tucholke and Sibuet*, 2007].

[32] *Oceanic domain*: The age of breakup is still a matter of debate. In this study, we define the breakup as the moment when deformation and magmatic activity localized and remained localized in a steady-state spreading center. The ages proposed for breakup range from 112 [*Boillot et al.*, 1989] to 142 Ma [*Srivastava et al.*, 2000]. More recently, *Tucholke and Sibuet* [2007] reinterpreted the age of breakup based on mapping of a basinwide unconformity of latest Aptian to earliest Albian (about 112.0 ± 1.0 Ma) observed throughout the northern Iberia Abyssal Plain.

[33] The reevaluation of the J-magnetic anomaly lead to the interpretation that continental breakup occurred at latest Aptian (about 112.0 ± 1.0 Ma) as a result of an excess magmatic event, leading to continental breakup [*Bronner et al.*, 2011]. The high drilled at ODP Site 1277 on the Newfoundland margin corresponds to the location of the J anomaly. Drilling revealed the presence of alkaline magmatic veins dated by U/Pb on zircon at 113 ± 2 Ma [*Jagoutz et al.*, 2007]. At the same site, older gabbros were also drilled that give ages of 128 ±3 Ma ($^{39}Ar/^{40}Ar$ ages) [*Jagoutz et al.*, 2007]. This suggests that the serpentinized mantle rocks into which the gabbros were emplaced had to be exhumed before 113 ± 2 Ma, which is the age of onset of steady-state seafloor spreading [see *Bronner et al.*, 2011, Figure 4].

[34] *Age of the major events*: The data and observations presented above enable us to determine the age of each major event associated with the evolution from decoupled to coupled deformation to first exhumation of mantle rocks and lithospheric breakup within the Iberia-Newfoundland rift system. The age of coupling is reasonably well defined across the deep Iberia margin. It corresponds to the migration of deformation further outboard and is dated by the onset of tilting of blocks over thinned, brittle crust within the coupled domain. This event occurred at latest Tithonian to Berriasian time (about 145.5 ± 4.0 Ma) in both the southern and northern sections (Figure 4). This age is relatively well constrained.

[35] In contrary, the age of first exhumation is less well defined. Age of exhumation is based on the drill hole data obtained from Hobby High (ODP Sites 900, 1067, and 1068). The age relationships between exhumation (cooling ages of basement rocks and clasts in debris flows) and the depositional age of the debris flows over exhumed mantle show that mantle exhumation may have occurred

as early as Valanginian time (140.2 ± 3.0 to 136.4 ± 2.0 Ma).

[36] The age of lithospheric breakup is constrained by mapping of a basinwide unconformity [*Tucholke and Sibuet*, 2007] and the drilling results at ODP Sites 1070 and 1277 [*Bronner et al*., 2011], both suggesting that breakup occurred at the Aptian-Albian boundary (about 112.0 ± 1.0 Ma). This age is relatively well constrained. From correlation of ages derived from the proximal wells [*Alves et al*., 2006] and the ODP sites [*Wilson et al*., 2001], it looks as if the transition from decoupled to coupled deformation and lithospheric breakup occurred almost simultaneously along the northern Iberia-Newfoundland rift system. It is, however, unclear if first exhumation was simultaneous along the two studied sections.

4. Restoration and Quantification of Crustal Extension

4.1. Restoration of Rifted Margins

[37] At present, there is no published kinematic, balanced restoration of a conjugate pair of rifted margins that restores the polyphase rift evolution back to the prerift stage using drill hole and refraction and reflection-seismic data. Classical restoration methods quantify extension by either using an areal balancing or by measuring and restoring fault heaves. While the former provides only an estimate of the bulk extension of the crust, the latter only describes deformation in the brittle upper crust. None of the two techniques is able to take into account the partitioning between brittle and ductile deformation. This is mainly due the fact that the limit between brittle and ductile layers changes during rifting as a function of strain, access of fluid, and change in the thermal state. It is important to note that the limit between brittle and ductile crust does not correspond to a conservative boundary. As a consequence, the volume of brittle and ductile material within the crust changes during rifting. Other limitations of restorations are due to (1) error in the estimation of crustal thickness due to velocity and gravity modeling, (2) inaccuracy in defining top and base crust, (3) potential loss or gain of crustal material during extension due to erosion, addition of magma, alteration/modification of crust, and (4) restoring lines that are not parallel to the transport direction. All these potential complications do not only limit the accuracy of restorations but also make the estimations of error bars difficult. Thus, restorations cannot be proposed for margins from which complete data sets are missing and the kinematics are not constrained.

[38] The restorations presented in this study are done for one of the best-investigated and constrained pairs of conjugated rifted margins worldwide. The advantage of restoring the Iberia-Newfoundland rift system is mainly due to the existence of a complete data set including refraction, reflection, gravity, and drill whole data and the well-constrained prerift history and kinematic evolution of the rifted margin. Although we are aware of the limitations of kinematic restorations (see earlier section), we believe that at present the Iberia-Newfoundland rift system is the only conjugated rift system from which we can get results that are sufficiently well constrained.

4.2. Areal Balancing of Refraction-Seismic Sections

[39] The simplest way to restore and quantify extension accommodated across rifted margins is by areal balancing of seismic refraction sections that are parallel to the kinematic transport direction. Areal balancing of such sections is straightforward, if one can assume that the whole crustal section can be imaged accurately, initial crustal thickness is known, sections are perfectly imaged parallel to the extensional direction, no material can be removed (erosion) or added (magma) to the crust, and the crust-mantle boundary is imaged with sufficient precision.

[40] The method of areal balancing used in this work is described in *Williams et al*. [2011] and discussed in *Heine et al*. [2013]. The basic idea is to restore a crustal section across a conjugated pair of rifted margins back to its prerift shape, which can be considered as a rectangle. The vertical side represents the initial crustal thickness and the horizontal side the initial width of the rift system. The initial crustal thickness is assumed to be that observed at present at the unextended continentward edges of the rift system. Thus, by dividing the present-day crustal surface through the initial crustal thickness, the prerift width of rifted crust can be estimated. This restoration transfers the final shape of the crust back into its initial shape, without giving insights on when and where the deformation was accommodated. Thus, the results of this restoration are independent from the processes that deformed the crust. The major assumption is that material was neither lost nor gained during extension. Implicit in this method is that (1)

top prerift and top mantle can be imaged across the whole margin, (2) the crust was thermally equilibrated and had a constant thickness across the domain, (3) the observed crustal thickness variations are primarily related to extensional deformation, i.e., that neither significant amount of magma was added to the crust or erosion removed parts of the crust, and (4) kinematic transport direction remained constant and parallel to the restored section

[41] For the sections considered in this study, these assumptions are reasonably well respected: (1) top-basement and top-mantle are well imaged, thanks to seismic refraction and reflection data, (2) the lithosphere can be considered to be thermally equilibrated (last orogenic events predated the rift event by more than 100 Ma and major thickness variations of the prerift sediments cannot be observed within the prerift sediments across the margin [*Wilson et al*., 2001], (3) only very minor synrift magmatic activity has been found [*Grange et al*., 2008; *Müntener and Manatschal*, 2006] and no evidence for erosion of the crust exists, and (4) previous studies show that the kinematic transport direction of extension was orthogonal, as indicated by major E-W trending transform systems bounding the Iberia-Newfoundland rift segment to the North and the South [*Hopper et al*., 2006; *Sibuet et al*., 2007]. Therefore, we think that a conservative areal balancing restoration can be used to quantify extension along the Iberia-Newfoundland rift system. The results of the restoration are presented in Figure 5. We calculated the surface of the sections shown in Figure 5 using the program ImageJ (freely available on the Internet: http://rsbweb.nih.gov/ij/). The results show strong variations for the total extension of the continental crust along the two conjugated margins as well as along one and the same margin. The Iberia margin is more extended as its conjugate Newfoundland margin. The factor of global extension varies slightly along the same margin: from ß 1.95 to 2.27 for the ISE1 and TGS-LG12 lines and from ß 2.14 to 1.36 for the SCREECH 1 and SCREECH 2 lines, respectively. The total extension across the northern transect is in the order of 255.5 km (78.5 km in Newfoundland/177 km in Iberia) and in the southern transect in the order of 229 km (65 km in Newfoundland/164 km in Iberia). These results not only highlight an important difference in amount of crustal extension on the conjugated margins but also a slightly higher total extension in the northern transect comparing to the southern transect.

4.3. Kinematic Inversion of Crustal Scale Sections

[42] A kinematic inversion of the two sections across the Iberia-Newfoundland rifted margins (Figure 2) is shown in Figure 6 and discussed in this chapter. At present there is no software or restoration algorithms that are able to restore hyperextended, polyphase rifted margins, taking into account the partitioning between brittle and ductile deformation in rocks undergoing modifications of their rheology as a function of rifting. The restoration of the sections shown in Figure 6 follows the model published in *Mohn et al*. [2012, Figure 14]. This model allows for a strain partitioning between brittle and ductile layers during thinning and exhumation of the crustal

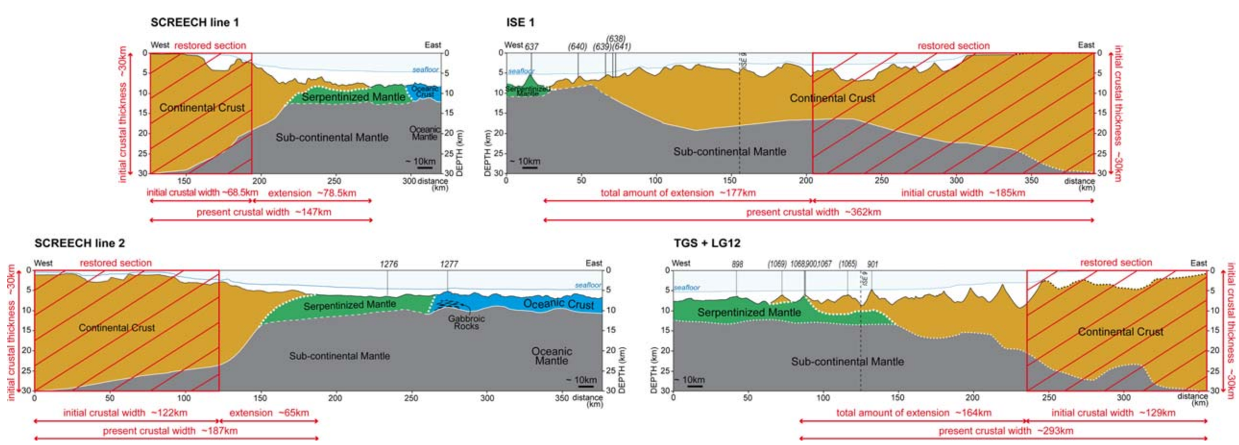

Figure 5. Schematic representation of the areal restoration method used to quantify the total extension of the continental crust. The orange color shows the continental crust, green the serpentinized mantle, and blue the oceanic crust. The red square shows that restoration of the continental crust (orange surface) back to its original crustal thickness, enabling to determine the initial width of the crust before onset of rifting.

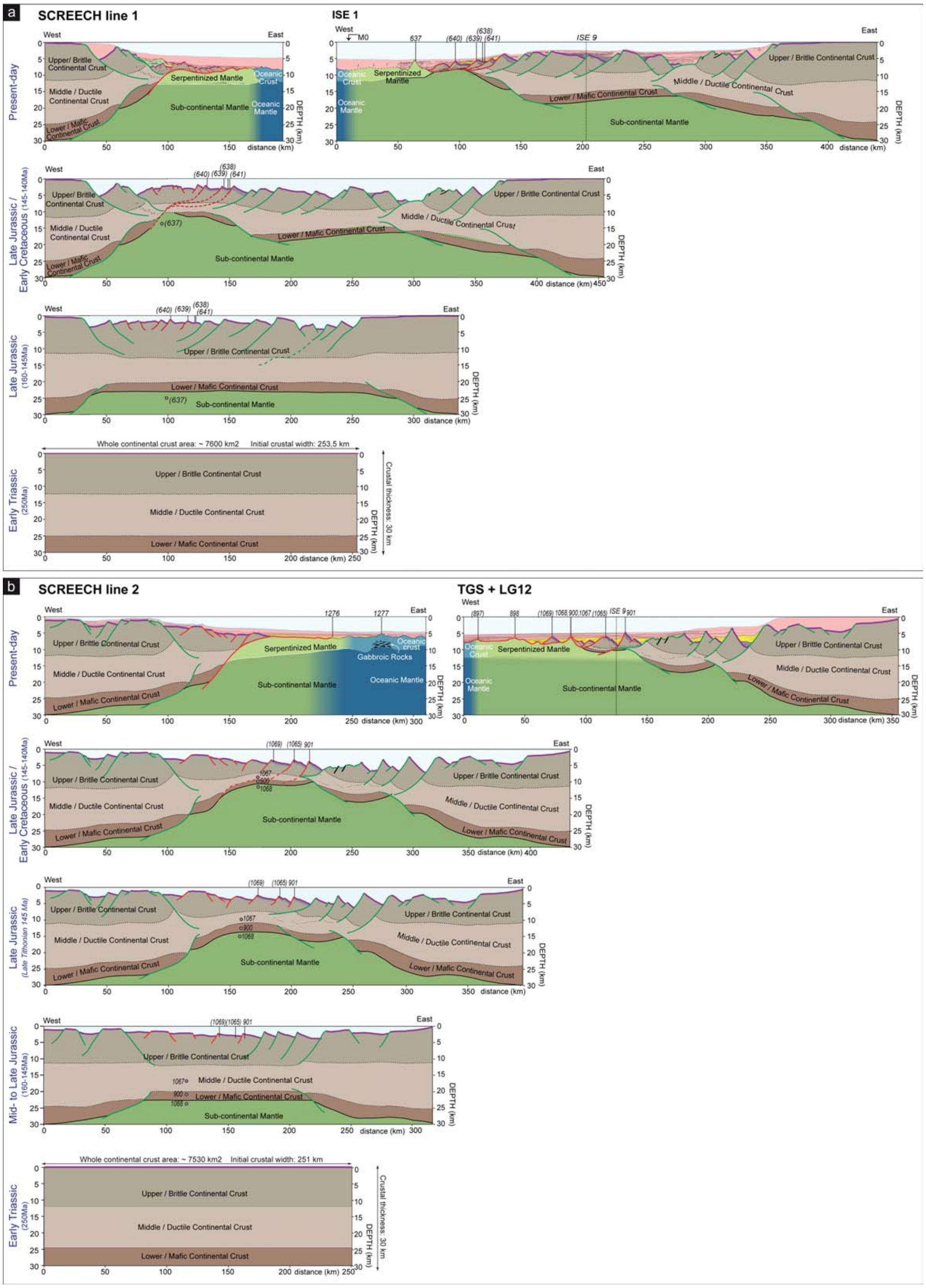

Figure 6. Restoration of the two pairs of conjugated sections: (a) the northern sections and (b) the southern sections. For location, see Figure 1. Amount of horizontal extension is based on the restoration of fault heaves and of the prerift layer back to a continuous horizontal section. Crustal restoration is based on the conceptual model published in *Mohn et al*. [2012], which describes the distribution of brittle and ductile deformation during thinning and exhumation. The restorations shown in Figures 6a and 6b show the distribution of initial "middle ductile" crust during rifting. Note that in reality the volume of brittle and ductile layers changes as a function of rifting and does not remain constant. Note that we do not restore the sedimentary evolution nor the subsidence history.

lithosphere. At the onset of rifting, deformation of the upper crust is brittle while deeper crustal levels behave ductile. With increasing thinning of the crust, the ductile levels are "consumed," and the crust eventually loses its brittle layers. At this stage, deformation in the crust will be mechanically coupled, and brittle faults can start to exhume deeper crustal and mantle layers. The initial distribution of brittle and ductile layers depends on the composition and thermal gradient, however, with ongoing extension, the thermal structure changes as a function of lithospheric thinning and the embrittlement of the crust enables fluids to move into the crust and to hydrate the rocks, leading to more semiductile, hydrated material. Thus, in reality, the volume of brittle and ductile layers may change during rifting and consequently simple areal balances of brittle upper crust are not possible. In our restoration shown in Figure 6, we tried to restore the different domains of the margin step by step, following the ideas developed in the model of *Mohn et al*. [2012]. To quantify the extension, we used a prerift marker horizon. We defined across the conjugate rifted margins the prerift sedimentary unit, and we restored it back to a continuous horizontal section. We also calculated the surface of the sections for each deformation step shown in the restoration in Figure 6 using the program ImageJ (freely available on the internet: http://rsbweb.nih.gov/ij/) in order to balance the surface of the continental crust throughout the restoration.

[43] *The assumptions* we made in our restoration are that (1) the domains defined in the sections formed in sequence, i.e., that they formed as the result of a progressive migration and localization of rifting toward the area of breakup; (2) the limits between the domains, i.e., the coupling point, exhumation point, and breakup point correspond to time events that can be localized in sections of the two conjugate margins; and (3) the prerift or pretectonic sequence can be defined accurately and used to restore the initial width of the margin.

[44] The assumption that the different domains formed as a sequence of events is supported by the observation that rift basins at the proximal margin are older than those at the distal margin. The idea that deformation develops from decoupled to coupled is compatible with previous observations made along the Iberia margin [*Pérez-Gussinyé et al*., 2003]. Both assumptions imply an oceanward migration of deformation and a younging of the syntectonic sequence in the same direction. One of the big unknowns is how much distal margin was thinned and extended already during the initial rifting stage. Drilling in the distal margin at ODP Site 639 in the northern section and ODP Site 1069 in the southern section indicate that the Upper Jurassic sediments were deposited onto Paleozoic rocks, suggesting that the former distal margin was not strongly affected by extension during initial Triassic to Early Jurassic rifting.

[45] The two restored sections shown in Figure 6 include intermediate steps of the rift evolution, which are (1) the stretching stage, (2) the thinning stage, (3) the coupling stage, and (4) the final stage (exhumation and breakup stage). Restoration is mainly based on the definition of the coupling point and the exhumation point (edge of the continent) and the definition of pretectonic marker horizon (in the decoupled domain we used the Lower Jurassic-Triassic and in the coupled domain the Late Jurassic. The overall horizontal restoration (extension) is based on the assumption that all pretectonic layers were horizontal and continuous before rifting. In the exhumation domain, the amount of extension is equal to that of the horizontal distance between the point of first exhumation (edge of the continent) and the location where breakup occurred, which coincides with the J anomaly. Based on these assumptions, we were able to calculate how much horizontal movement was necessary to restore the present-day sections back to their prerift stage. It is important to note that this restoration depends on the recognition of prerift sequences and the determination of fault heaves.

[46] The results obtained by this restoration are shown in Figure 7. The amount of extension accommodated in the crust is quite similar in the two transects; 255.5 km in the northern transect and 228.8 km in the southern transect. In contrast, the extension accommodated in the ZECM is quite different; 86 km in the north and 180 km in the south. Another observation is that the Iberia margin accommodated more extension prior to mantle exhumation (177 km in the north and 164 km in the south) compared to the Newfoundland margin (78.5 km in the north and 64.8 km in the south). The most interesting observation is that the amount of extension in the coupled domain is very constant across all four sections and ranges between 23 and 35 km, while the deformation in the decoupled domain is more heterogeneous, ranging from 29.8 to 145 km.

[47] If we look at ß values for the bulk extension of the crust, we have ß = 2.01 in the north and ß = 1.91 in the southern section. ß values are smaller for the SCREECH 1 and 2 sections (2.14

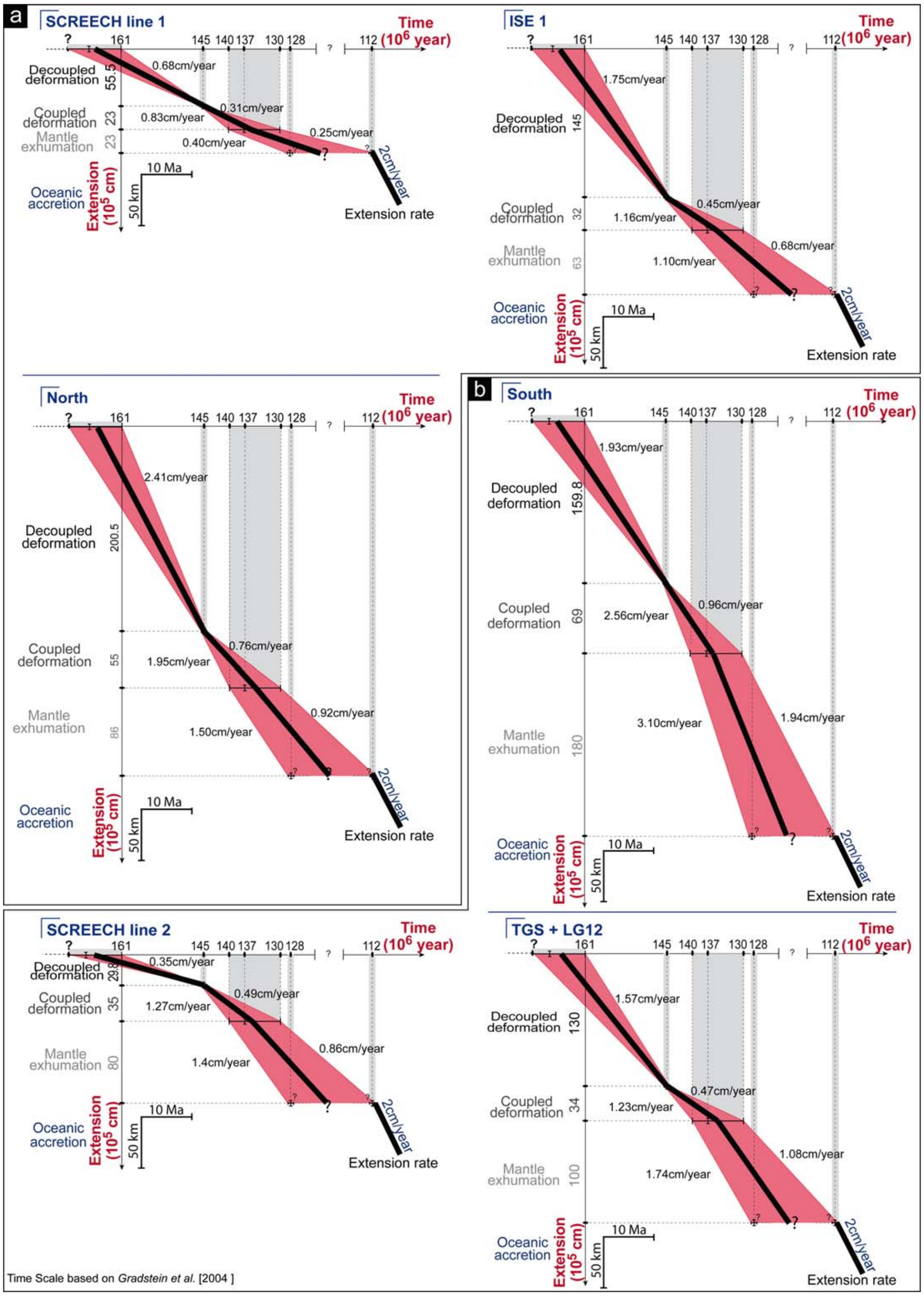

Figure 7. Diagrams presenting the results of the quantification of the extension along the different domains along (a) the northern transect and (b) the southern transect. Red surface represents estimated error bars, and thick black line estimated average extension rate. Rates used for first seafloor spreading are from *Minshull et al*. [2001].

in the north and 1.36 in the south) and larger for the ISE 1 and TGS-LG12 sections (1.95 in the north and 2.27 in the south). If we compare the coupled and decoupled domains, we see that decoupled domains show values ranging from 1.44 to 2.82, while the cumulative coupled domains show values in the order of 1.60 and 1.85).

[48] *Marrett and Allmendinger* [1992] demonstrated that by using fractal scaling relationships small-scale faults, invisible in seismic sections, can account from 25% to 60% of extension. Thus, we suggest that there is a tendency to underestimate the amount of extension in seismic interpretations and that the estimates derived from the areal restorations are more precise.

4.4. Determination of Strain Rates for the Whole Rift Event and for Distinct Phases

[49] In the previous two sections, we determined the timing and duration of distinct rift events (t) and the amount of extension (x) accommodated in each of these rift events. Using the simple linear function $x = vt$, where v corresponds to the strain rate, enable us to determine v, which corresponds to the slope of the linear function. Thus, by plotting x against t, we can determine the strain rate for each of the events. It is important to note that the calculated value corresponds to a full rate (not half rate as usually referred to in oceanic domains). Moreover, maximum and minimum strain rates can be determined by integrating the error bars related to the determination of the timing and amount of extension. It is important to note that the values strongly depend on the error bars that we need to include in our estimations. These error bars are difficult to quantify. For the determination of the amount of extension, we can assume that the real values for the extension based on the restoration of fault heaves and the prerift section underestimate the amount of extension and therefore also the rates of extension [*Marrett and Allmendinger*, 1992].

[50] Figure 7 shows the rates that we calculated as well as maximum and minimum rates for the different stages of rifting within the Iberia-Newfoundland rift-system. Rates for the decoupled domain are difficult to determine since onset of rifting is ill defined, and it is not clear how the Triassic rifting is linked to the Late Jurassic/Early Cretaceous rift event that led to continental breakup. Assuming that the rift event leading to continental breakup initiated in Oxfordian time (161 Ma), as suggested by *Alves et al*. [2006], rates of 2.41 cm/yr are obtained for the northern transect and 1.93cm/yr for the southern section. For onset of rifting initiating in Triassic time (about 220 Ma), much lower values would be obtained. Comparing the values obtained from the Iberia and Newfoundland margins, the values for the former are between 1.57 and 1.75 cm/yr. Those from the conjugate Newfoundland margin are between 0.35 and 0.68 cm/yr.

[51] Rates obtained for the coupled domain range between 0.76 and 1.95 cm/yr in the northern transect and between 0.96 and 2.56 cm/yr in the southern transect. Comparing the values obtained from the Iberia and Newfoundland margins, the average values for the former are larger (0.80–0.85 cm/yr), compared with those from the conjugate Newfoundland margin (0.57–0.88 cm/yr). However, it is important to note that the coupling point for the Newfoundland margin is ill defined, and therefore the values obtained for the Newfoundland margin are not very accurate.

[52] Rates obtained for mantle exhumation are much faster in the southern transect (1.94–3.10 cm/yr) compared to the northern transect (0.92–1.50 cm/yr). It is important to note that these values are assuming and that exhumation and onset of seafloor spreading started on both sections almost simultaneously [e.g., *Wilson et al*., 2001; *Bronner et al*., 2011]. Comparing the values obtained from the Iberia and Newfoundland margins, the average values for the Iberia margin are larger (0.89–1.41 cm/yr) compared with those from the conjugate Newfoundland margin (0.33–1.13 cm/yr).

5. Discussion

5.1. Architecture of Magma-Poor Rifted Margins

[53] The structural variability observed across magma-poor rifted margins enables to define different domains (Figure 3). These domains can be characterized by crustal thickness, geometry of top-basement and top-mantle reflections, accommodation space, stratigraphic and crustal architecture, geometry of the structures creating accommodation space, and their relation to decoupling horizons in the crust or the underlying mantle.

[54] Based on these characteristic mappable features, four domains and two subdomains (Figure 3) can be defined. The first-order control on the development of these domains is bulk rheology of the extending lithosphere, which changes as a function

of extension, access of fluids, and formation of magma. We describe furthermore "limits" that define the boundaries between the domains. These limits correspond to particular "events" in the time-space evolution of a rift system (Figure 8). These limits correspond to a location and an instant (time point) at which (1) coupling between residual brittle layers (usually upper crust and upper lithospheric mantle) occurs, 2) first exhumation of subcontinental mantle takes places, and (3) seafloor spreading starts. In reality these limits may not necessarily correspond to sharp boundaries but rather to zones in which case neither an exact age nor an exact location can be determined (see for instance, the coupling point along the Newfoundland margin). In an idealized rift system, strain localizes

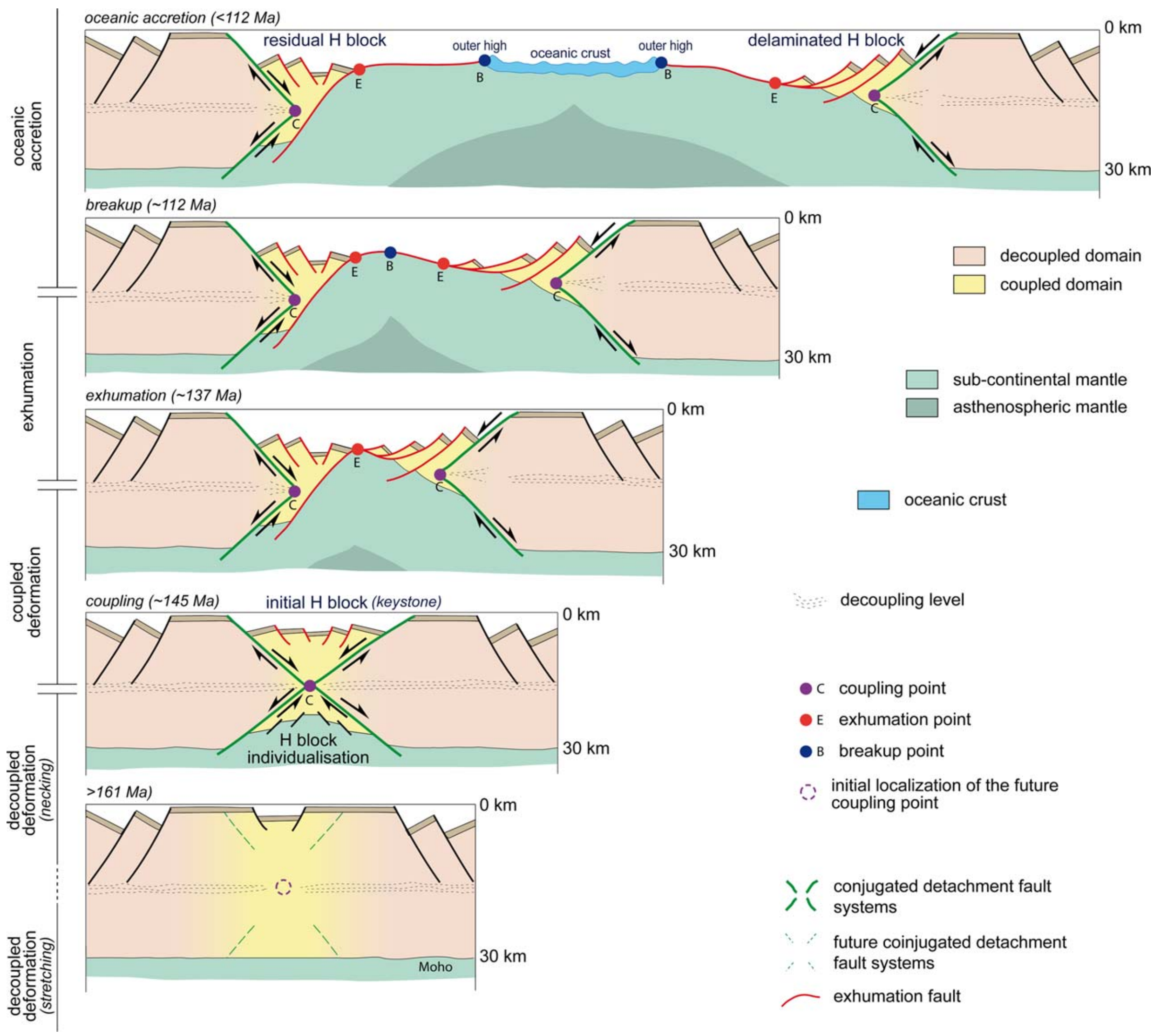

Figure 8. Conceptual restoration showing the time-space evolution of the Iberia-Newfoundland rift system. Note that the limits correspond to major events but also to mappable locations along margins limiting domains that show different evolutions of rifting. Note also that the limits defined on one margin have time equivalents on the conjugate margin. Another important point illustrated in Figure 8 is the evolution from symmetric to asymmetric rifting that depends on the capacity of the crust to either decouple or couple deformation. The evolution of the coupled domain shows an evolution from a "symmetric" keystone (H-Block) to delaminated blocks that will eventually be split up between exhumed domains on the two future conjugate margins. In the final architecture of the margin, one can define a residual H-Block in the upper plate margin and a delaminated H-Block in the lower plate margin within the coupled domain separated by an exhumed and an oceanic domain.

through time, brittle-ductile transitions are simple and well-defined surfaces, and onset of seafloor spreading is instantaneous and controlled by asthenospheric processes. In such an idealized rift system, the limits between the different domains may be considered to be sharp, in which case timing and location of an event can be determined with accuracy. In such an idealized case, changes in the mode of deformation may be abrupt and may correspond to major events in the lifetime of a rift system. These major changes in the mode of extension may explain the different ways of accommodating strain and creating accommodation space previously described in the dynamic models proposed by *Huismans and Beaumont* [2002] and *Lavier and Manatschal* [2006]. The change in the mode of extension from stretching to thinning to exhumation and symmetric seafloor spreading may correspond to an in-sequence evolution of a rift system undergoing strain localization. We propose that in such a sequential evolution of rifting, each of the domains and limits found on one-rifted margin has a time-equivalent domain and limit on its conjugate-rifted margin (see conceptual model shown in Figure 8).

5.2. Symmetric Versus Asymmetric Rifting

[55] A major debate that persisted in the rifted margin community over more than two decades was related to the question if rifted margins are symmetric or asymmetric. *Huismans and Beaumont* [2008] showed in their experiments that rifted margins can be either symmetric or asymmetric, depending on strain rate and bulk rheology of the crust. Of major importance for the structural evolution of a rifted margin is whether the strain is coupled or decoupled on the scale of the crust. In this study, we use the transition from decoupled to coupled deformation as a criteria to distinguish between two major domains in the margin, which are the decoupled and coupled domains, respectively (Figure 3). The transition from decoupled to coupled deformation corresponds to a change from symmetric (pure shear) to asymmetric (simple shear) deformation on the scale of the crust. On the scale of the lithosphere, rifting may remain symmetric throughout rifting. While the structures in the decoupled domain are very similar on both margins, those in the coupled domain are more complex and asymmetric. This is well shown in the two conjugate sections in Figure 2. In these sections, a distal upper plate margin (residual hanging wall block (H-Block)) can be distinguished from a distal lower plate margin (delaminated H-Block) (Figures 2 and 8) [*Lavier and Manatschal*, 2006; *Péron-Pinvidic and Manatschal*, 2010]. The evolution from decoupled to coupled rifting is shown in Figure 8. In this model, rifting initiated as distributed, symmetric, decoupled deformation. Localization of deformation in the crust, probably controlled by the boudinage of the underlying lithospheric mantle [*Lavier and Manatschal*, 2006; *Mohn et al.*, 2012], results in localization and thinning of the crust at the edges of a keystone (e.g., H-Block of *Lavier and Manatschal* [2006]). This H-Block will be affected by brittle faulting during the subsequent coupled deformation, resulting in two asymmetric blocks, a residual H-Block in the upper plate margin and a delaminated H-Block, formed by extremely thinned tilted blocks and extensional allochthons on the lower-plate margin. These two domains are separated by an exhumation domain, and after breakup, by the future oceanic domain. Thus, the coupled, hyperextended domain is the most complex and asymmetric part of the margin.

[56] Whether the rifted margins evolve as symmetric or asymmetric consequently depends mainly on the rheological evolution of the rift system. In the case of Iberia-Newfoundland, the transition from symmetric to asymmetric rifting on the scale of the crust occurred as a consequence of the elimination of ductile layers during extreme crustal thinning. The change back to pure-shear extension only occurred when the arrival of magma was able to rupture the lithosphere, leading to the onset of seafloor spreading. Rift systems where abundant magma can be produced before embrittlement of the crust (e.g., magma-rich systems) may develop as symmetric-rifted margins. Excess magma during rifting may also force the lithosphere to break, in which case breakup may occur before exhumation started. Thus, in magma-rich rifted margins, the exhumation mode may be suppressed. It is also important to note that in the case of magma-rich margins, the crust can be thickened by magma additions, which makes that restorations are only possible if the volume of added magma corresponding to new crust can be determined.

5.3. Restoration Techniques and Strain Rates

[57] In this study, we used two different restoration techniques. A first one is based on an areal restoration and a second one on the restoration of fault heaves and the prerift layer back to a

continuous horizontal section. Although the results of the two restorations are different, the values obtained for the two transects show that in order to thin the crust to zero and exhume mantle rocks at the seafloor, an average ß value of 2 is required. The fact that the restoration of fault heaves gives values that are constantly lower comparing with the results obtained from the areal restoration may be explained by small-scale faults, invisible in seismic sections, which are estimated to account for 25%–60% of the extension [*Marrett and Allmendinger*, 1992]. Therefore, we suggest that the total amount of extension in the crust is better approximated by areal restoration. More precise restoration methods would need to combine an areal restoration by adding reliable maker horizons within the crust. Such well-defined, reliable strain markers in the crust are difficult to determine. Many authors use the brittle-ductile transition that forms the base of the brittle crust, as a strain marker; however, this boundary does not behave as a "conservative" boundary, since the rocks can change their rheology during rifting. Ductile material can become brittle as a consequence of cooling during thinning and exhumation. On the other hand, brittle material may behave as semiductile (or macroductile) as a consequence of the breakdown and hydration of olivine and feldspar and the formation of low friction minerals such as serpentine and clay minerals [*Jammes et al*., 2010]. The formation of such low-friction minerals may also result in the development of a network of low frictional shear zones in the brittle domain [*Chester et al*., 1985], which may accommodate a component of flattening (pure shear) within hyperextended domains.

[58] One of the key parameters commonly suggested to control rifting are extension rates. A prerequisite to determine extension rates is to be able to estimate duration of extension and total amount of extension. Since rates may change during rifting, it is important to be able to define these values for different stages of rifting. As discussed in the previous sections, such restorations are difficult to obtain and depend on many assumptions that are in most rifted margins unconstrained. In our study, we determined rates of extension by measuring the total amount of extension of each domain accommodated during the time of its formation. Although we need to add error bars to the values (see Figure 7 and related discussion), it looks as if extension rates increase during rifting and approach during exhumation those of seafloor spreading. This increase in strain rates may be explained by either an overall increase of strain rates in the southern North Atlantic rift system or by increasing localization of rifting within the Iberia-Newfoundland rift system, while competing rift systems such as the Orphan rift stopped. In this context, it is important to note that more mantle has been exhumed in the southern transect compared with the northern transect. Assuming that the timing of mantle exhumation is the same, this would suggest that rates decrease northward. This may be explained by a rotation of Flamish Cap at this stage, resulting in the formation of V-shaped basins on both sides of Flemish Cap [*Sibuet et al*., 2007]. An alternative interpretation is that mantle exhumation was controlled by a polar rotation during progressive northward migration, leading to the unzipping of the Iberia-Newfoundland rifted margins.

6. Conclusions

[59] The aim of this paper was to characterize the rift architecture of two transects across the Iberia Newfoundland rift system, to restore the two conjugate transects, and to quantify the extensional deformation and determine the extension rates for single stages of rifting for these two transects. Although this exercise seems to be simple, in reality, it is complex. Indeed, true kinematic inversions, including areal balancing and strain rate determinations for single rift stages, are not yet reported from rifted margins. The nonexistence of such studies is not due to missing software or the restoration alogorithms, but due to the lack of data and the understanding of how crustal sections really deform. A requirement to determine extension rates is to determine the amount of extension and its duration for each stages of rifting. Although extension rates are reported in the literature for most rifted margins, most of the numbers are crude estimates and are not based on dated, and most of the restoration algorithms used for the restoration only take into account simple block tilting [*Ranero and Pérez-Gussinyé*, 2010]. In our study, we show that the quality of restorations depends on the quality of the data as well as the possibility to map structures within sections and date the age of these structures. This asks to have access to high-quality reflection-seismic data imaging down to >10 s combined with refraction-seismic data. A second requirement is that the rift evolution is relatively simple, which means that there is no out-of-sequence rifting or oblique extension or major additions of magma during

rifting or deposition of salt that could mask some of the rift structures. All these requirements are only satisfied in few rift systems, one of which is the Iberia-Newfoundland conjugate rift system.

[60] For two conjugate pairs across the Iberia-Newfoundland rift system, we propose a subdivision of the margins in four domains, referred to as decoupled, coupled, exhumed, and oceanic domains. We use a number of characteristics such as crustal thickness, geometry of top-basement and top-mantle reflections, accommodation space, stratigraphic and crustal architecture, geometry of the structures creating accommodation space, and their relation to decoupling horizons. This enables us to define limits separating these domains that juxtapose parts of the margin that show different isostatic, structural, magmatic, and sedimentary evolutions during rifting. These limits not only correspond to geographical locations that can be mapped in a rifted margin but also represent instants in the evolution of the margins that record major changes in the rheological, mechanical, and magmatic evolution of the rift system. These changes are expressed by a change in the mode of extension from stretching to thinning, to exhumation to onset of seafloor spreading. Since these changes correspond to a sequential evolution associated with strain localization, we assume that these limits have time equivalents on both conjugate margins and occur, in the conjugate margin, simultaneously (Figure 8). The recognition of these domains and their time relationships enabled us to propose a restoration of two rift sections and to quantify the total amount of extension and to define the strain rates for the two transects across the Iberia-Newfoundland rift system. The major result is that bulk crustal extension leading to mantle exhumation is in the order of ß = 2, corresponding to a total extension that is in the order of about 200 km. The amount of extension depends, however, on the deformation mode. Extension in the decoupled domain is strongly variable across the rift system and depends on the inherited ratio between ductile and brittle layers. In contrast, coupled domains show a very homogeneous strain distribution across the whole rift system. Exhumed mantle domains show strong lateral variations, suggesting that these domains formed as V-shaped basins. How these lateral variations in strain and strain rate can be explained is yet unclear. Two end-member interpretations are possible: either that these changes reflect overall changes in total amount of strain and strain rates in the overall North Atlantic rift system or that these changes are related to strain distribution and localization within different simultaneous rift systems.

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References

Afilhado, A., L. Matias, H. Shiobara, A. Hirn, L. Mendes-Victor, and H. Shimamura (2008), From unthinned continent to ocean: The deep structure of the West Iberia passive continental margin at 38°N, *Tectonophysics*, *458*, 9–50, doi:10.1016/j.tecto.2008.03.002.

Alves, T. M., C. Moita, F. Sandnes, T. Cunha, J. H. Monteiro, and L. M. Pinheiro (2006), Mesozoic–Cenozoic evolution of North Atlantic continental-slope basins: The Peniche basin, western Iberian margin, *Am. Assoc. Pet. Geol. Bull.*, *90*, 31–60, doi:10.1306/08110504138.

Beslier, M. O. (1996), Data report: Seismic line LG12 in the Iberia Abyssal Plain, *Proc. ODP Sci. Results*, *149*, 737–739.

Boillot, G., E. L. Winterer, A. W. Meyer, et al. (1987a), *Proceedings of the ODP Initial Reports, 103*, Ocean Drilling Program, College Station, Tex.

Boillot, G., M. Recq, and Scientific party ODP LEG 103 (1987b), Tectonic denudation of the upper mantle along passive margins: a model based on drilling results (ODP leg 103, western Galicia margin, Spain), *Tectonophysics*, *132*, 335–342.

Boillot, G., J. Girardeau, and J. Kornprobst (1989), Rifting of the West Galicia continental margin: A review, *Bull. Soci. Géol. France*, *4*, 393–400.

Bronner, A., D. Sauter, G. Manatschal, G. Péron-Pinvidic, and M. Munschy (2011), Magmatic breakup as an explanation for magnetic anomalies at magma-poor rifted margins, *Nature Geosci.*, *4*, 549–553, doi:10.1038/NGEO1201.

Chester, F. M., M. Friedman, and J. M. Logan (1985), Foliated cataclasites, *Tectonophysics*, *111*, 139–146.

Chian, D., K. Louden, T. A. Minshull, and R. B. Whitmarsh (1999), Deep structure of the ocean-continent transition in the southern Iberia Abyssal Plain from seismic refraction profiles: Ocean Drilling Program (Legs 149 and 173) transect, *J. Geophys. Res.*, *104*(B4), 7443–7462.

Clark, S. A., D. Sawyer, J. A. Austin, G. L. Christeson, and Y. Nakamura (2007), Characterizing the Galicia Bank-Southern Iberia Abyssal Plain rifted margin segment boundary using multichannel seismic and ocean bottom seismometer data, *J. Geophys. Res.*, *112*, B03408, doi:10.1029/2006JB004581.

Christensen, N. I., and W. D. Money (1995), Seismic velocity and composition of the continental crust: A global view, *J. Geophys. Res.*, *100*, 9761–9788.

Cunha, T. (2008), Gravity anomalies, flexure, and the thermal and mechanical evolution of the West Iberia Margin and its conjugate of Newfoundland, Ph.D. thesis, Department of Earth Sciences, Oxford University, Oxford, UK.

Dean, S. M., T. A. Minshull, R. B. Whitmarsh, and K. E. Louden (2000), Deep structure of the ocean-continent transition in the Southern Iberia Abyssal Plain from seismic refraction profiles: The IAM-9 transect at 40°20′N, *J. Geophys. Res.*, *105*, 5859–5885.

Discovery 215 Working Group (1998), Deep structure in the vicinity of the ocean-continent transition zone under the Southern Iberia Abyssal Plain, *Geology*, *26*, 743–746.

Dunbar, J. A., and D. S. Sawyer (1989), How preexisting weaknesses control the style of continental breakup, *J. Geophys. Res.*, *94*, 7278–7292.

Funck, T., J. R. Hopper, H. G. Larsen, K. E. Louden, B. E. Tucholke, and W. S. Holbrooket (2003), Crystal structure of the ocean-continent transition at Flemish Cap: Seismic refraction results, *J. Geophys. Res.*, *108*(B11), 2531, doi:10.1029/2003JB002434.

Gradstein, F. M., J. G. Ogg, and A. G. Smith (2004), *A Geologic Time Scale 2004*, Cambridge University Press, Cambridge, UK, 610 pp.

Grange, M., U. Schärer, G. Cornen, and J. Girardeau (2008), First alkaline magmatism during Iberia–Newfoundland rifting, *Terra Nova*, *20*, 494–503.

Galice, G. (1979), The continental margin off Galicia and Portugal: Acoustical stratigraphy, dredge stratigraphy, and structural evolution, *Initial Rep. Deep Sea Drill. Proj.*, *47*, 633–662.

Hart, S. R., and J. Blusztajn (2006), Age and geochemistry of the Mafic sills, ODP Site 1276, Newfoundland Margin, *Chem. Geol.*, *235*, 222–237.

Heine, C., J. Zoethout, and R. D. Muller (2013), Kinematics of the South Atlantic rift, *Solid Earth Discuss.*, *5*, 41–116, doi:10.5194/sed-5-41-2013.

Henning, A. T., D. S. Sawyer, and D. S. Templeton (2004), Exhumed upper mantle within the ocean-continent transition of the Northern West Iberia margin: Evidence from prestack depth migration and total tectonic subsidence analyses, *J. Geophys. Res.*, *109*, B05103, doi:10.1029/2003JB002526.

Hölker, A. B., K. Holliger, G. Manatschal, and F. Anselmetti (2002), Seismic reflectivity of detachment faults of the Iberian and Tethyan distal continental margins based on geological and petrophysical data, *Tectonophysics*, *350*, 127–156.

Hopper, J. R., T. Funck, B. E. Tucholke, H. G. Larsen, W. S. Holbrook, K. E. Louden, D. Shillington, and H. Lau (2004), Continental breakup and the onset of ultraslow seafloor spreading off Flemish Cap on the Newfoundland rifted margin, *Geology*, *32*, 93–96, doi:10.1130/G19694.1.

Hopper, J. R., T. Funck, B. E. Tucholke, K. E. Louden, W. S. Holbrook, and H. C. Larsen (2006), A deep seismic investigation of the Flemish Cap margin: Implications for the origin of deep reflectivity and evidence for asymmetric break-up between Newfoundland and Iberia, *Geophys. J. Int.*, *164*, 501–515, doi:10.1111/j.1365-246X.2006.02800.x.

Huismans, R. S., and C. Beaumont (2008), Complex rifted continental margins explained by dynamical models of depth-dependent lithospheric extension, *Geology*, *36*, 163–166, doi:10.1130/G24231A.1.

Huismans, R. S., and C. Beaumont (2011), Depth-dependent extension, two-stage breakup and cratonic underplating at rifted margins, *Nature*, *473*, 74–78, doi:10.1038/nature09988.

Jagoutz, O., O. Müntener, G. Manatschal, D. Rubatto, G. Péron-Pinvidic, D. B. Turrin, and I. M. Villa (2007), The rift-to-drift transition in the North Atlantic: A stuttering start of the MORB machine?, *Geology*, *35*, 1087–1090, doi:10.1130/G23613A.1.

Jammes, S., L. Lavier, and G. Manatschal (2010), Extreme crustal thinning in the Bay of Biscay and the Western Pyrenees: from observation to modeling, *Geochem. Geophys. Geosyst.*, *11*, Q10016, doi:10.1029/2010GC003218.

Krawczyk, C. M., and T. J. Reston (1995), Detachment faulting and continental break-up: The S reflector offshore Galicia, in *Rifted Ocean-Continent Boundaries*, vol. 463, edited by E. Banda, M. Torne, and M. Talwani, pp. 231–246, Kluwer, Amsterdam.

Krawczyk, C. M., T. J. Reston, M. O. Beslier, and G. Boillot (1996), Evidence for Detachment Tectonics on the Iberia Abyssal Plain rifted margin, in *Proceedings of the Ocean Drilling Program, Scientific Results*, vol. 149, edited by R. B. Whitmarsh et al., pp. 603–615.

Lau, K. W. H., K. E. Louden, T. Funck, B. E. Tucholke, W. S. Holbrook, J. R. Hopper, and H. C. Larsen (2006), Crystal structure across the Grand Banks–Newfoundland Basin Continental Margin. I. Results from a seismic refraction profile, *Geophys. J. Int.*, *167*, 127–156, doi:10.1111/j.1365-246X.2006.02988.x.

Lavier, L., and G. Manatschal (2006), A mechanism to thin the continental lithosphere at magma-poor margins, *Nature*, *440*, 324–328, doi:10.1038/nature04608.

Manatschal, G. (2004), New models for evolution of magma-poor rifted margins based on a review of data and concepts from West Iberia and the Alps, *Int. J. Earth Sci.*, *93*, 432–466, doi:10.1007/s00531-004-0394-7.

Manatschal, G., N. Froitzheim, M. Rubenach, and B. D. Turrin (2001), The role of detachment faulting in the formation of an ocean-continent transition: Insights from the Iberia Abyssal Plain, in *Non-Volcanic Rifting of Continental Margins: A Comparison of Evidence From Land and Sea*, vol. 187, edited by R. C. L. Wilson et al., pp. 405–428, Geol. Soc. of London, Special Publications.

Marrett, R., and R. W. Allmendinger (1992), Amount of extension on small faults—An example from the Viking Graben, *Geology*, *20*(1), 47–50.

Minshull, T. A. (2009), Geophysical characterisation of the ocean-continent transition at magma-poor rifted margins, *C. R. Geosci.*, *41*, 382–393, doi:10.1016/j.crte.2008.09.003.

Minshull, T. A., S. M. Dean, R. S. White, and R. B. Whitmarsh (2001), Anomalous melt production after continental break-up in the southern Iberia Abyssal Plain, in *Non-Volcanic Rifting of Continental Margins: A Comparison of Evidence from Land and Sea*., vol. 187, edited by R. C. L. Wilson et al., pp. 537–550, Geol. Soc. of London, Special Publications.

Mohn, G., G. Manatschal, M. Beltrando, E. Masini, and N. Kusznir (2012), Necking of continental crust in magma-poor rifted margins: Evidence from the fossil Alpine Tethys margins, *Tectonics*, *31*, TC1012, doi:10.1029/2011TC002961.

Montadert, L., D. G. Roberts, O. De Charpal, and P. Guennoc (1979), Rifting and subsidence of the northern continental margin of the bay of Biscay, *Texas A&M Univ.*, Ocean Drilling Program, College Station, Tex.

Müntener, O., and G. Manatschal (2006), High degrees of melt extraction recorded by spinel harzburgite of the Newfoundland margin: The role of inheritance and consequences for the evolution of the southern North Atlantic, *Earth Planet. Sci. Lett.*, *252*, 437–452, doi:10.1016/j.epsl.2006.10.009.

Osmundsen, P. T., and T. F. Redfield (2011), Crustal taper and topography at passive continental margins, *Terra Nova*, *23*, 349–361, doi:10.1111/j.1365-3121.2011.01014.x.

Passchier, C. W., and R. A. J. Trouw (1996), *Microtectonics*, 289 pp., Springer, Berlin.

Penrose (1972), Penrose field conference on ophiolites, *Geotimes*, *17*, 24–25.

Pérez-Gussinyé, M., C. R. Ranero, and T. J. Reston (2003), Mechanisms of extension at nonvolcanic margins: Evidence from the Galicia interior basin, west of Iberia, *J. Geophys. Res.*, *108*(B5), 2245, doi:10.1029/2001JB000901.

Peron-Pinvidic, G., and G. Manatschal (2009), The final rifting evolution at deep magma-poor passive margins from Iberia-Newfoundland: A new point of view, *Int J. Earth Sci.*, *98*, 1581–1597, doi:10.1007/s00531-008-0337-9.

Péron-Pinvidic, G., G. Manatschal, T. A. Minshull, and S. Dean (2007), Tectonosedimentary evolution of the deep Iberia–Newfoundland margins: Evidence for a complex breakup history, *Tectonics*, *26*, TC2011, doi:10.1029/2006TC001970.

Péron-Pinvidic, G., and G. Manatschal (2010), From microcontinents to extensional allochthons: Witnesses of how continents rift and beak apart?, *Pet. Geosci.*, *16*, 189–197, doi: 10.1144/1354-079309-903.

Pickup, S. L. B., R. B. Whitmarsh, C. M. R. Fowler, and T. J. Reston (1996), Insight into the nature of the ocean-continent transition off West Iberia from a deep multichanel seismic reflection profile, *Geology*, *24*, 1079–1082.

Ranero C. R., and M. Pérez-Gussinyé (2010), Sequential faulting explains the asymmetry and extension discrepancy of conjugate margins, *Nature*, *468*, 294–297, doi:10.1038/nature09520.

Reston, T. J. (2009), The structure, evolution and symmetry of the magma-poor rifted margins of the North and Central Atlantic: A synthesis, *Tectonophysics*, *468*, 6–27, doi:10.1016/j.tecto.2008.09.002.

Reston, T. J. (2010), The opening of the central segment of the South Atlantic: symmetry and the extension discrepancy, *Pet. Geosci.*, *16*, 199–206, doi:10.1144/1354-079309-907.

Reston, T. J., and K. G. McDermott (2011), Successive detachment faults and mantle unroofing at magma-poor rifted margins, *Geology*, *39*, 1071–1074, doi:10.1130/G32428.1.

Reston, T. J., C. M. Krawczyk, and D. Klaeschen (1996), The S reflector west of Galicia (Spain): Evidence from prestack depth migration for detachment faulting during continental breakup, *J. Geophys. Res.*, *101*, 8075–8091.

Rudnick, R. L., and S. Gao (2003), Composition of the continental crust, *Treat. Geochem.*, *3*, 1–64.

Sawyer, D. S., R. B. Whitmarsh, A. Klaus, et al. (1994), *Proceedings of the ODP Initial Reports*, *149*, Ocean Drilling Program, College Station, Tex.

Shillington, D. J., W. S. Holbrook, B. E. Tucholke, J. R. Hopper, K. E. Louden, H. C. Larsen, H. J. A. Van Avendonk, S. Deemer, and J. Hall (2004), Data report: Marine geophysical data on the Newfoundland nonvolcanic rifted margin around SCREECH Transect 2, in Proc. Ocean Drill. Program Initial Report, 210.

Shillington, D. J., W. S. Holbrook, H. J. A. Van Avendonk, B. E. Tucholke, J. Hopper, K. Louden, H. C. Larsen, G. T. Nunes, and ODP Leg 210 Scientific Party (2006) Evidence for asymetric nonvolcanic rifting and slow incipient seafloor spreading from seismic reflection data on the Newfoundland margin, *J. Geophys. Res.*, *111*, B09402, doi:10.1029/2005JB003981.

Sibuet, J. C., W. B. F. Ryan, et al. (1979), Initial Reports of the Deep Sea Drilling Project, *47*, Part 2, U.S. Government Printing Office, Washington, DC.

Sibuet, J. C., S. Srivastva, M. Enachescu, and G. D. Karner (2007), Lower Cretaceous motion of Flemish Cap with respect to North America: Implications on the formation of Orphan Basin and SE Flemish Cap/Galicia Bank conjugate margins, in *Imaging, Mapping and Modelling Continental Lithosphere Extension and Breakup*, vol. 282, edited by G. D. Karner, G. Manatschal, and L. M. Pinheiro, Geol. Soc. of London, Special Publication.

Srivastava, S. P., J. C. Sibuet, S. Cande, W. R. Roest, and I. D. Reid (2000), Magnetic evidence for slow seafloor spreading during the formation of the Newfoundland and Iberian margins, *Earth Planet. Sci. Lett.*, *182*, 61–76.

Stipp, M., H. Stünitz, R. Heilbronner, and S. M. Schmid (2002), Dynamic recrystallization of quartz: Correlation between natural and experimental 1174 conditions, *Geol. Soc. London, Spec. Publ.*, *200*(1), 171–190.

Sutra, E., and G. Manatschal (2012), How does the continental crust thin in a hyper-extended rifted margin? Insights from the Iberia margin, *Geology*, *40*, 139–142, doi:10.130/G32786.1.

Tankard, A. J., H. J. Welsink, and W. A. M. Jenkins (1989), Structural styles and stratigraphy of the Jeanne d'Arc basin, Grand Banks of Newfoundland, in *Extensional Tectonics and Stratigraphy of the North Atlantic Margins*, vol. 46, edited by A. J. Tankard and H. R. Balkwill, pp. 265–282, AAPG Memoir.

Tucholke, B. E., and J. C. Sibuet (2007), Leg 210 synthesis: tectonic, magmatic, and sedimentary evolution of the Newfoundland-Iberia rift, in *Proceedings of the Ocean Drilling Program, Scientific Results*, vol. 210, edited by B. E. Tucholke, J.-C. Sibuet, and A. Klaus.

Tucholke, B. E., J.-C. Sibuet, A, Klaus, et al. (2004), *Proceedings of the ODP Initial Reports*, *210*, Ocean Drilling Program, College Station, Tex.

Van Avendonk, H. J. A., W. S. Holbrook, G. T. Nunes, D. J. Shillington, B. E. Tucholke, K. E. Louden, H. C. Larsen, and J. R. Hopper (2006), Seismic velocity structure of the rifted margin of the eastern Grand Banks of Newfoundland, Canada, *J. Geophys. Res.*, *111*, B11404, doi:10.1029/2005JB004156.

Van Avendonk, H., L. L. Lavier, D. J. Shillington, and G. Manatschal (2009), Extension of continental crust at the margin of the eastern Grand Banks, Newfoundland, *Tectonophysics*, *468*, 131–148, doi:10.1016/j.tecto.2008.05.030.

Whitmarsh R. B., F. Avedik, and M. R. Saunders (1986), The seismic structure of thinned continental crust in the northern Bay of Biscay, *Geophys. J. R. Astron. Soc.*, *86*(2), 589–602.

Whitmarsh R. B., P. R. Miles, and M. Pinheiro (1990), The seismic velocity structure of some NE Atlantic continental rise sediments; A lithification index?, *Geophys. J. Int.*, *101*, 367–378.

Whitmarsh, R. B., M.-O. Beslier, P. J. Wallace, et al. (1998), *Return to Iberia: Proceedings of the Ocean Drilling Program, Initial Reports*, *173*, Ocean Drilling Program, College Station, Tex.

Whitmarsh, R. B., S. M. Dean, T. A. Minshull, and M. Tompkins (2000), Tectonic implications of exposure of lower continental crust beneath the Iberia Abyssal Plain, Northeast Atlantic Ocean: Geophysical evidence, *Tectonics*, *19*(5), 919–942.

Whitmarsh, R. B., G. Manatschal, and T. A. Minshull (2001), Evolution of magma-poor continental margins from rifting to seafloor spreading, *Nature*, *413*, 150–154.

Williams, S. E., J. M. Whittaker, and R. D. Müller (2011), Full-fit, palinspatic reconstruction of the conjugate Australian-

Antarctic margins, *Tectonics*, *30*, TC6012, doi:10.1029/2011TC002912.

Wilson, R. C. L., G. Manatschal, and S. Wise (2001), Rifting along non-volcanic passive margins: Stratigraphic and seismic evidence from the Mesozoic successions of the Alps and western Iberia, in *Non-Volcanic Rifting of Continental Margins: A Comparison of Evidence From Land and Sea*, vol. 187, edited by R. C. L. Wilson et al., pp. 429–452, Geol. Soc. London, Special Publication.

Zalán, P. V., M. C. G. Severino, C. A. Rigoti, L. P. Magnavita1, J. A. B. de Oliveira, and A. R. Vianna (2011), An entirely new 3D-view of the crustal and mantle structure of a South Atlantic Passive Margin–Santos, Campos and Espírito Santo Basins, Brazil, *Search Discov. Article* 30177.

Zelt, C. A., K. Sain, J. V. Naumenko, and D. S. Sawyer (2003), Assessment of crustal velocity models using seismic refraction and reflection tomography, *Geophys. J. Int.*, *153*, 609–626.