# The role of pre-magmatic rifting in shaping a volcanic continental margin: An example from the Eastern North American Margin

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## Abstract

Both magmatic and tectonic processes contribute to the formation of volcanic continental margins. Such margins are thought to undergo short-lived extension across a narrow zone of lithospheric thinning (~100 km). New observations from the Eastern North American Margin (ENAM) contradicts this hypothesis. With ~64,000 km of 2D seismic data tied to 40 wells combined with published refraction, deep reflection, receiver function and onshore drilling efforts, we quantified along-strike variations in the distribution of rift structures, magmatism, crustal thickness, and early post-rift sedimentation on the shelf of Baltimore Canyon trough (BCT), Long Island Platform and Georges Bank Basin (GBB) of ENAM. Results indicate that BCT is narrow (80-120 km) with a sharp basement hinge and few rift basins. The Seaward Dipping Reflectors (SDR) there are ~50 km seaward of the hinge line. In contrast, GBB is wide (~200 km), has many syn-rift structures, and SDR there are about 200 km away from the hinge line. Early post-rift depocenters at the GBB coincide with thinner crust suggesting "uniform" thinning of the entire lithosphere. Models for the formation of volcanic margins do not explain the wide structure of the GBB. The different characteristics between BCT and GBB point to different modes of rifting. The BCT underwent little, or highly localized, thinning prior to the volcanic phase. Thinning of the GBB segment was broader. These variations result from either diachronous rifting, heterogenous rheology or a lateral asthenosphere temperature gradient.

# The role of pre-magmatic rifting in shaping a volcanic continental margin: An example from the Eastern North American Margin

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**Key Points:**

- Rift structure, crustal thickness and distribution of breakup volcanism of the Eastern North American volcanic margin are presented
- Georges Bank Basin experienced substantial pre-magmatic thinning whereas Baltimore Canyon Trough thinning was magma-assisted
- Inherited distribution of crustal rheology determined the nature and intensity of pre-magmatic strain

## Abstract

Both magmatic and tectonic processes contribute to the formation of volcanic continental margins. Such margins are thought to undergo extension across a narrow zone of lithospheric thinning (~100 km). New observations based on existing and reprocessed data from the Eastern North American Margin contradict this hypothesis. With ~64,000 km of 2D seismic data tied to 40 wells combined with published refraction, deep reflection, receiver function and onshore drilling efforts, we quantified along-strike variations in the distribution of rift structures, magmatism, crustal thickness, and early post-rift sedimentation under the shelf of Baltimore Canyon trough (BCT), Long Island Platform and Georges Bank Basin (GBB). Results indicate that BCT is narrow (80-120 km) with a sharp basement hinge and few rift basins. The Seaward Dipping Reflectors (SDR) there extend ~50 km seaward of the hinge line. In contrast, the GBB is wide (~200 km), has many syn-rift structures, and the SDR there extend ~ 200 km seaward of the hinge line. Early post-rift depocenters at the GBB coincide with thinner crust suggesting “uniform” thinning of the entire lithosphere. Models for the formation of volcanic margins do not explain the wide structure of the GBB. We argue that crustal thinning of the BCT was closely associated with late-syn rift magmatism whereas the broad thinning of the GBB segment predated magmatism. Correlation of these variations to crustal terranes of different compositions suggests that the inherited rheology determined the pre-magmatic response of the lithosphere to extension.


## 1 Introduction

Deep-rooted tectonic and magmatic processes accompany the extension and breakup of continents, leading to the formation of passive continental margins. The resultant rifted margins are broadly divided into volcanic and magma-poor margins (Fig. 1; e.g. Doré, & Lundin, 2015; Franke, 2013; Menzies et al., 2002; Mutter et al., 1988). The structures and petrological properties of these two archetype margins are described as dichotomic. Whereas, magma-poor margins usually consist of a wide zone of crustal necking, hyperextension and exhumation of lower crust and mantle rocks (Fig. 1B; e.g. Franke, 2013; Peron-Pinvidic et al., 2013; Reston, 2009), volcanic margins are often described as having narrow zones of crustal thinning (<100 km) adjacent to thick intrusive and extrusive magmatic additions (Fig. 1A; e.g. Franke, 2013; Lizarralde and Holbrook, 1997; Stica et al., 2014).

The processes that thin the continental crust and mantle lithosphere giving rise in magma-poor margins were extensively modelled in recent years (e.g. Brune et al., 2014, 2017; Huismans & Beaumont, 2011, 2014; Lavier & Manatschal, 2006; Peron-Pinvidic et al., 2013; Reston, 2009; Sutra et al., 2013). The formation of volcanic margins on the other hand, remains unsettled. Volcanic margins may result from heating of the upper mantle by either a plume head (White & McKenzie, 1989; White et al., 1987) or non-plume related processes (Kelemen & Holbrook, 1995; McHone, 2000) such as continental insulation (Brandl et al., 2013; Anderson, 1982) or small-scale convection induced by sharp lithosphere necking (Mutter et al., 1988; King & Anderson, 1998). However, it is not clear whether the initial lithosphere thinning mechanisms leading to the formation of volcanic margins are distinct (e.g. Mutter et al., 1988; White & McKenzie., 1989) or are mostly similar to the mechanical rifting processes that form magma-poor margins (Guan et al., 2019; Eldholm et al., 2000). It is widely accepted that the inherited structure and composition of the pre-rift lithosphere controls the deformation and thinning patterns at rifts and passive margins (e.g. Manatschal et al., 2015; Brune et al., 2017; Misra &

Mukherjee 2015). However, less is known about the role that inheritance plays during the formation of volcanic margins, as weakening by heating and intrusions might overwhelm the inherited rheological signal.

We use an extensive set of seismic reflection and auxiliary data along the volcanic Eastern North American Margin (ENAM; Fig. 2) to constrain syn-rift crustal and lithosphere thinning patterns at a margin-wide scale. We show that: a) the width of the zone of crustal thinning varies along the margin. b) extensive (>200 km wide) crustal and lithosphere thinning predated volcanic breakup in the Georges Bank Basin (GBB) segment, contradicting some existing models for the formation of volcanic margins; c) rifting of the ENAM can be divided into pre-magmatic and magmatic rifting stages d) the distribution, width, and nature of pre-magmatic thinning is controlled by the pre-rift rheology and e) magmatic rifting is accompanied by major strain localization and intense crustal thinning.

### 1.1 Crustal structure

The most pronounced characteristic of volcanic margins is the magmatic addition related to their latest stage of formation. These include a thick (<20 km) wedge of subaerially emplaced volcanic rocks, which were imaged on seismic reflection data as oceanward/seaward dipping reflectors (SDR) (Fig. 1B; Hinz, 1981; Mutter et al., 1982; Planke et al., 2000) and an intruded and/or underplated lower crust (e.g. Abdelmalak et al., 2017; Eldholm et al., 1995; Holbrook et al., 1992; Menzies et al., 2002; White et al., 1987). SDR emplacement occurs on top of seaward tilting blocks composed of intruded continental or oceanic crust (Stica et al., 2014; Geoffroy et al., 2005). Alternatively, they tilt as a response to flexural subsidence of gabbroic dikes that form their base (Mutter et al., 1982; Paton et al., 2017; Tian & Buck, 2019). The SDR transform seaward into an abnormally thick oceanic crust that gradually thins to typical oceanic thicknesses away from the continent (Menzies et al., 2002). In most volcanic margins, the transition from an unthinned continental crust to an igneous/oceanic crust occurs over relatively short distances (50-100 km, indicated by the “Necking domain” in Fig. 1A; Ebinger & Casey, 2001; Franke, 2013; Paton et al., 2017; White & McKenzie, 1989; White et al., 1987). Nevertheless, volcanic margins might exhibit wider geometries where older rifting episodes predated volcanic breakup (Guan et al., 2019). Another phenomenon often associated with volcanic margins is the emplacement of large igneous provinces shortly before or during rifting (Menzies et al., 2002; White & McKenzie, 1989; Ziegler & Cloetingh, 2004).

Magma-poor margins seldom include the magmatic components described above. However, they are associated with other unique characteristics such as hyperextended crust (<10 km thick and composed of brittle hydrated crust), detachment faults and exhumed mantle rocks (Fig. 1B; Lavier & Manatschal, 2006; Manatschal, 2004; Sibuet et al., 1987). The along-dip extent of the thinned continental crust is usually wider than that found in volcanic margins and may reach up to 350 km (e.g. profile SMART 2 in Nova Scotia which appears at Wu et al., 2006).

### 1.2 Modes of rifting

The sequence of events leading to the formation of volcanic and magma-poor margins is also different. In a broad sense, the formation of magma-poor margins involves the breakup of the continental crust before the breakup of the mantle lithosphere (e.g. Reston, 2009), whereas rifting of volcanic margins is thought to involve the breakup of the mantle lithosphere before or concomitantly with the total breaking of the crust (Franke, 2013). Magma-poor margins often experience polyphase rifting and relatively low strain rates during their formation (<15 mm/year half extension rate, Lundin et al., 2014 and references therein). This slow and protracted rifting promotes a broad zone of crustal thinning (Reston, 2009 and references therein). The formation of volcanic margins, on the other hand, is associated with high strain rates (25-30 mm/year half extension, Schreckenberger et al., 2002; Hopper et al., 2003), increasing weakening of the lithosphere and strain localization toward the rift axis (Buck, 2004, 2006).

A widely accepted model for the formation of the igneous material that characterizes volcanic margins, considers rifting over a mantle hotter than normal by at least 150ºC (White & McKenzie, 1989). The increased mantle temperature is attributed to the presence of a mantle plume under a continental rift (White & McKenzie, 1989; White et al., 1987) or to upper mantle convection (e.g. Anderson et al., 1992; Kelemen & Holbrook, 1995). This model treats the co-occurrence of rifting and mantle heating as incidental, yet it requires both. Once the lithosphere has been thinned by a factor of ~5 it breaks, allowing melt to migrate to the surface. Part of the melt might not reach the surface and accumulate at the base of the crust (White & McKenzie, 1989; White et al., 1987).

Other models suggest convective partial melting under rifts as an explanation for melt production during the formation of volcanic margins (Mutter et al., 1988). These models do not necessarily require increased temperatures to produce melts. Rather, they require rapid and localized lithospheric thinning that promotes a sharp relief at the lithosphere-asthenosphere boundary under the rift (Mutter et al., 1988; Van Wijk et al., 2001). The asthenospheric material that rises into the region of thinned lithosphere is hotter than its surroundings. Lateral temperature and density differences drive small-scale convection under the rift, bringing more hot asthenosphere from below and increasing the generation of melts. (Simon et al., 2009; Van Wijk et al., 2001).

Although the convective partial melting models outline an inverse cause-and-effect scenario to the one depicted by rifting over hotter than normal mantle models, both types of models predict margins with narrow zones of crustal and lithospheric thinning (Fig. 1A). The sharp lithosphere-asthenosphere boundary, a requisite for convective partial melting models, implies that the thinning must be limited to a narrow zone (~100 km; Mutter et al., 1988). According to White and McKenzie (1989), the presence of hot asthenosphere under a rift weakens the lithosphere and promotes strain localization toward the rift axis. If breakup is achieved, strain localization leads to the formation of a narrow margin. Later works further proposed that large quantities of magma generated during rifting over a heated mantle would intrude and heat the lithosphere, reducing the tensile stress required to split it (Buck, 2004, 2006). This "magma-assisted rifting" mechanism was used to explain observations of minor crustal thinning coincident with large amounts of breakup magmatism at the east Africa rift system (Buck, 2006; Kendall et al., 2005). Recently, Geoffroy et al. (2015) proposed that two conjugate syn-volcanic crustal-scale detachment faults accommodate most of the crustal thinning at volcanic margins. The subsiding hanging walls of these faults accommodate extrusive flows (SDR), forming a relatively sharp hinge between the untinned and igneous crust (Stica et al., 2014).

Despite the considerable amount of research on the evolution of volcanic margins, the nature of crustal deformation, the processes that involve the pre-magmatic extension and the implication these have for the post-rift evolution of such margins, remain unclear. To investigate these unresolved issues, the current study examines the ENAM. The ENAM is chosen due to its relatively continuous and well-constrained rifting phase, and the availability of recently released seismic and borehole data (Triezenberg et al., 2016). These data, coupled with the availability of modern interpretation and visualization software allow the documentation of along-margin variations in greater detail than was previously possible. We examine the syn- and post-rift evolution of the Baltimore Canyon Trough (BCT) and Georges Bank Basin (GBB) (Fig.2) and specifically, the extent and geometry of their crustal thinning and distribution of SDR.

## 2 The Eastern North American Volcanic Margin

The geology of the ENAM records two full Wilson cycles. The last cycle included the closure of the Iapetus and Rheic Oceans (e.g. van Staal et al., 2009) and the formation of the supercontinent Pangea between 420 Ma and 270 Ma (Thomas, 2006, and references therein). Late Triassic to Early Jurassic rifting of Pangea (e.g. Olsen, 1997; Withjack et al., 2012) was accompanied by the formation of a series of asymmetric rift basins (i.e. half-grabens, Fig. 2). The North American remnant of this rift system is bounded by the Appalachian Mountains to the NW and the continent-ocean boundary to the SE (roughly at the present-day continental slope, Fig.2; e.g. Leleu et al., 2016; Withjack et al., 2012). The basins accumulated a well-documented Triassic-early Jurassic syn-rift sequence (e.g. Leleu & Hartley, 2010; Olsen, 1997; Schlische, 1992). The syn-rift sequence records the emplacement of an intense magmatic event that occurred at ~200 Ma known as the Central Atlantic Magmatic Province (CAMP; e.g. Hames et al., 2000; Marzoli et al., 1999, 2011, 2018; Nomade et al., 2007; Olsen, 1999; Olsen et al., 2003; Whiteside et al., 2007). Rift-basin subsidence in central North America ended soon after the CAMP magmatism (Withjack et al., 2012). Cessation of rifting was attributed to lithospheric breakup associated with the opening of the Atlantic Ocean. Estimates for the age of breakup range between 175 Ma (Klitgord & Schouten, 1986), to 190 Ma (Labails et al., 2010; Sahabi et al., 2004; Sibuet et al., 2012) to 200 Ma (Schettino & Turco, 2009). It was proposed that breakup was diachronous, starting at ~200 Ma in southern North America, advancing to central North America at 195-175 Ma (Withjack et al., 1998, 2012). Shuck et al. (2019) suggest that accretion of proto-oceanic crust occurred over an unbroken lithosphere starting at ~200 Ma. They claim that full lithospheric breakup was achieved at 175 Ma when normal seafloor spreading began. By the end of the rifting phase, post-rift thermal subsidence dominated the vertical motions on the continental margin (e.g. Sawyer, 1985; Steckler & Watts, 1978; Swift et al., 1987).

The discovery of magmatic material, that was accreted during the latest stages of rifting and earliest seafloor spreading, led to the recognition of the volcanic nature of the ENAM (Austin et al., 1990; Holbrook & Kelemen, 1993; Holbrook et al., 1992; Holbrook et al., 1994; Keen & Potter, 1995; Kelemen & Holbrook, 1995; LASE, 1986; Lizarralde & Holbrook, 1997; Talwani et al., 1995; Tréhu et al., 1989). Holbrook and Kelemen (1993) correlated intrusive and extrusive bodies, recognized on several wide-angle seismic profiles along the margin, to a margin-parallel positive magnetic anomaly known as the East Coast Magnetic Anomaly (ECMA, Fig.2). Hence, magmatism was regional, spanning over ~2000 km from the Blake Plateau Basin to offshore southern Nova Scotia. This East Coast Margin Igneous Province (ECMIP) is comprised of an SDR wedge inferred to be extrusive basalt above its intrusive counterpart in the form of a high-velocity lower crust ($V_p$=~7.5 km/s). Wide-angle seismic data reveal that the continental crust thins rapidly seaward toward a point of convergence between the high-velocity lower crust and SDR. Seaward of this point, the crust is entirely igneous (LASE, 1986; Tréhu et al., 1989). At the BCT, the maximum thickness of the igneous crust is 13-24 km (Talwani et al., 1995).

Models for the emplacement of ECMIP favor minor pre-breakup lithospheric thinning over an abnormally hot asthenosphere. A mantle plume was suggested as the source of excess heat (White & McKenzie, 1989). The plume was probably situated at the southern part of the rift system, near Florida (e.g. Wilson, 1997; Ruiz-Martínez et al. 2012). Other proposed heating mechanisms include continental insulation (e.g. Hole, 2015), edge-driven convection (McHone, 2000) and slab delamination processes (Whalen et al., 2015). Kelemen and Holbrook (1995) suggested that the magma originated in partial melting of hotter-than-normal mantle (>1500ºC) under high pressure (>4 GPa). They proposed a scenario in which the lithosphere acted as a thick lid due to a minor amount of thinning until the final stages of rifting. Reprocessing of the dataset used by Kelemen and Holbrook (1995) led Talwani and Abreu (2000) to suggest that a 30 km-thick continental crust juxtaposes an igneous crust of comparable thickness at the BCT. They inferred that crustal thinning was minimal and required high mantle temperatures. Farther south, under the Carolina Trough (Fig. 2), a similar crustal structure was observed and may also imply minor thinning prior to breakup (Tréhu et al., 1989). Since ECMIP rocks have not been sampled offshore, the exact age of the ECMIP and its relation to the CAMP are unresolved issues. Age estimates for the ECMIP are 172-179 Ma (Benson, 2003), 175 Ma (Klitgord & Schouten, 1986) and 190 Ma (Labails et al., 2010; Sibuet et al., 2012). Recently, Davis et al. (2018) suggested that ECMIP is the offshore continuation of CAMP and that its emplacement took between 6 to 31 Myr, starting at ~201 Ma and ending between 195 to 170 Ma.

Although the ENAM is volcanic from the Blake Plateau Basin in the south to the Scotian Basin in the north, previous studies have noticed that it is segmented. The segmentation is reflected in the location of the hinge zone, geometry of the rift basins, characteristics of the post-rift unconformity, post-rift sedimentation, elastic thickness of the lithosphere and details of gravity and magnetic anomalies along the strike of the margin (Klitgord et al., 1988; Behn & Lin, 2000; Wyer & Watts, 2006). When suggesting an explanation for the along-strike heterogeneity of the ENAM, some of the cited studies emphasize allogenic factors such as sediment supply (Poag & Sevon, 1989) whereas others suggested autogenic controls such as rift-related variations in lithospheric strength (Wyer and &, 2006). Works predating the recognition of the margin as volcanic explained the along-strike variations using rifting models that are more suitable for magma-poor settings (e.g. upper plate vs. lower plate, Klitgord et al., 1988). The current study aims at explaining these variations in the context of a volcanic margin.

## 3 Data and Methods

We used a comprehensive set of seismic reflection data acquired on the continental shelf and slope from the U.S.-Canada border to Cape Hatteras (Fig. 3; Table S1 Supporting information). The 64,000 km of 2D seismic profiles were acquired as 4147 lines using a variety of acquisition parameters during 23 cruises for industry and research from the 1970s to the 1990s (e.g. Benson & Doyle, 1988; Klitgord et al., 1988; Poag, 1991; Poag & Sevon, 1989; Schlee & Fritsch, 1982). The industry data are archived at the USGS National Archive of Marine Seismic Surveys (Triezenberg et al., 2016). Ca. 4000 km of the seismic data were reprocessed as part of an offshore $CO_2$ sequestration evaluation project (Cumming et al., 2017; Fortin et al., 2018).

Forty offshore wells were incorporated (Fig. 3). Well data includes paleontological reports, check-shot records and geophysical well logs such as sonic and density logs (Table S3). The data were scanned and digitized as part of the offshore $CO_2$ sequestration project (Cumming et al., 2017).

A compilation of published results of wide-angle seismic, deep reflection seismic, and receiver function data helped constrain crustal thicknesses (Fig. 3, Table S2 supporting information). As part of this compilation, depth domain data were converted into two-way travel time (TWT) based on refraction results (Fig. 3, Table S2 supporting information). The domain conversion was done from depth to TWT and not vice versa for three reasons. First, most of the data used are in the TWT domain. Second, depth domain data are restricted to areas of thin or no sediment cover. This makes their domain conversion function more straightforward compared with most of the TWT data which are found in areas with thicker (>3 km) sediment cover. Third, the TWT domain allows the interpretation of crustal boundaries and large thickness changes using few assumptions and without having to rely on the choice of conversion velocities. For onshore depth data, an average of 6.3 km/s conversion velocity was used for the continental crust (Lizarralde & Holbrook, 1997; Pratt et al., 1988). A depth to Moho grid by Li et al. (2018) was used for constraining Moho onshore the northern BCT. The grid is the outcome of interpolation of multiple receiver function stations. For offshore data at the northern BCT, lithological boundaries (Figure 5 in LASE, 1986) were digitized following the interpretation of Talwani et al. (1995). Since no refraction data crosses the GBB and LIP, constraints on the crustal structure in these areas rely on reflection data alone.

Magnetic anomaly data were used to constrain the ECMA and infer on its relation to the margin structure and especially the SDR. The EMAG2v3 (version 3) global magnetic anomaly grid used here incorporates satellite, ship, and airborne magnetic measurements and features a 2-arc-minute resolution (Meyer et al., 2017).

Depth to the base of the post-rift (BPR) beneath the coastal plain was constrained using a Digital Elevation Map by Pope et al. (2016). The map illustrates the structure of the base of the US North Atlantic coastal plain aquifer from New York in the north to the southern part of North Carolina in the south (Fig. 3). The coastal plain aquifer is composed of the post-rift sequence. Hence, the base of the aquifer separates pre-rift basement rocks and syn-rift strata below from the overlying post-rift sequence. The mapping of the base of the aquifer (post-rift) by Pope et al. (2016) relies on a regional amalgamation of results of previous studies, which defined the aquifer based on well-log data. The Pope et al. (2016) Digital Elevation Map was only used onshore and was smoothed using a 1 km by 1 km window. The map was converted to TWT using an average velocity of 2.5 km/s based on the average velocity observed for the equivalent depth interval at the wells located on the outer shelf (e.g. COST B-2, Smith et al., 1976).

### 3.1 Seismic Interpretation

Four horizons/horizon packages have been mapped to identify and understand the rifting, basement, and crustal geometries: top of basement, seaward dipping reflectors (SDR), the Moho, and BPR. An additional six post-rift horizons have been mapped and will be reported elsewhere.

#### 3.1.1 Top Basement

Since only one well, the COST G-1 well, penetrated pre-rift basement rocks in the study area, the main input for mapping the top basement is seismic reflection data. On seismic sections, the sediment-basement interface usually appears as a high amplitude reflector that separates continuous sedimentary reflectors above from discontinuous, chaotic reflectors below (Figs. 4, 5 and 6). In several locations (e.g. the Long Island Platform and some rift basins at the GBB) along the margin, the upper part of the basement appears to be reflective as well. This phenomenon may be attributed to pre-rift sediments or metasediments or to ‘ghost’ artifacts, and it sometimes obscures picking the top of basement. Where those upper crust reflectors appear, the interpretation follows a high amplitude reflector that is onlapped by post-rift reflectors (Fig. 4). Inside rift basins, where dipping, divergent reflectors mark syn-rift strata (e.g. Klitgord et al., 1988), the top of basement is regarded as the base of the divergent wedge (red line, Figs. 4B, 5B). At the deepest parts of GBB and BCT the interpretation of top basement is ambiguous. To reduce the uncertainty in picking top basement at these areas, the results of published refraction surveys were used to guide the interpretation of reflection data (Figs. 3, 4 and 6). The absence of deep refraction data at the GBB makes the interpretation of its deepest part (>5 s TWT) less certain.

#### 3.1.2 Seaward Dipping Reflectors (SDR)

Multichannel seismic reflection, together with published refraction data, were also used to map the extent of SDR along the continental shelf, slope and rise. The SDR were mapped based on their reflection geometry following the definition of Mutter et al. (1982). In addition, published wide-angle seismic data were used to constrain the interpretation and to increase data coverage. The TWT values of the top of the SDR in northern BCT were re-picked on published Expanded Spread Profile velocities (LASE, 1986). The top of the SDR was assigned to an increase in P-wave velocity from ~5.7 km/s to ~6.1 km/s. The corresponding TWT values were then placed on the USGS profile 25 at each Expanded Spread Profile location and compared to the seismic reflection data. Previous interpretations of the three EDGE profiles (Sheridan et al., 1993) were digitized for mapping the top of the SDR at the southern BCT. The top SDR horizon, as recognized on both reflection and refraction data, was then traced regionally using seismic reflection profiles.

#### 3.1.3 Moho

The base of the seismic crust (Moho) was mapped according to both deep seismic reflection and published refraction data. Moho reflection were interpreted as deep (9-12 s), mostly continuous, low-frequency reflectors at the base of a reflective interval that can be distinguished from an underlying transparent zone (pink line, Fig. 5). These reflectors appear only on data collected by the USGS. The interpretation of these reflectors to be the Moho agrees with previous interpretations of the same data at the Long Island Platform (Hutchinson et al., 1985; 1986), the Gulf of Maine (Hutchinson et al., 1988; Hutchinson et al., 1987) and other seismic data in the ENAM (Keen et al., 1991; LASE, 1986; Lizarralde & Holbrook, 1997; Sheridan et al., 1993). Previous interpretations of the Moho underneath the continental shelf were extended by using two seismic attributes with seismic interpretation: structural smoothing to increase reflector continuity and time-varying gain.

### 3.1.4 Base post-rift

The base post-rift (BPR) horizon is a combination of three stratigraphic tops: the top of SDR, the top of syn-rift strata, and the top of basement. Where rift basins are present, the BPR is interpreted as an erosional surface that separates the divergent syn-rift strata from onlapping and sagging post-rift strata (Figs. 4 and 5). Where SDR are apparent, the BPR is placed at the top of the seaward dipping package (Figs. 5 and 6). In places where neither SDR nor syn-rift strata appear, the BPR coincides with top basement. The time span of the hiatus across the BPR unconformity should generally increase landward. Though diachronous, the BPR unconformity should correspond to the time interval during which rifting had ceased and post-rift subsidence commenced seaward of the hinge line. Early estimates for rift cessation point to early Hettangian age (201 Ma; Walker et al., 2018) while the latest estimates for initiation of seafloor spreading are of early Aalenian (174 Ma; Walker et al., 2018; for further discussion see Withjack et al., 2012).

### 3.1.5 Post-rift horizons

Interpretation of post-rift horizons follows standard seismic interpretation procedures of sedimentary units (e.g. Mitchum et al., 1977; Vail et al., 1977). Available wells were tied to sequence bounding surfaces to constrain the ages of the interpreted horizons (For a detailed description of seismic-well tie procedures and paleontological data see Table S3). In total, six post-rift horizons were mapped along the margin (Fig. 4, Table 1). Paleontological reports are in general agreement regarding the ages of Cretaceous and younger strata. Age determination for the Cretaceous sequences follows Jordan et al. (2019), Miller et al. (2018) and Schmelz et al. (2019). There is, however, no consensus regarding the pre-Cretaceous chronostratigraphy (For further discussion see Cousminer & Steinkraus, 1988; Poag, 1991; Poag & Valentine, 1988). The Jurassic chronostratigraphy presented here follows Poag and colleagues' interpretations (Poag, 1991; Poppe et al., 1992a; b). No rocks older than Kimmeridgian were penetrated in the BCT. Thus, the age assignment of the deeper MJ horizon at the BCT follows Poag (1985), which estimated it to be Top Callovian.

## 4 Interpretation

### 4.1 Top Basement and basement faults

The following paragraphs describe the structure of the top basement surface and the rift basins found in the research area. Some of the rift basins were previously described (e.g. Hutchinson & Klitgord, 1988; Hutchinson et al., 1985; Klitgord et al., 1982). However, the tight grid (<7 km line spacing at the GBB) used here uncovers details that were previously concealed. It provides accurate estimates of the extent, orientation and lateral terminations of previously recognized rift basins and the detection of new basins not identified in earlier surveys.

#### 4.1.1 Georges Bank Basin

The top basement at the GBB has the highest density of faults of all the margin segments examined in this study (Fig. 7). The faults accommodate normal displacement and form a complex array of rift basins that generally deepen toward the shelf edge. Two main fault orientations appear: NNE-SSW (AB, FB, IYB, OYB in Fig. 7) and ENE-WSW (PB, F2 in Fig. 7). Smaller, secondary faults inside the Atlantis Basin are sub-parallel to the ENE trend. Both the existence ENE-WSW direction and secondary faults are presented here for the first time.

The basement faults at the GBB dip both landward and seaward forming horsts, grabens and half grabens. The Atlantis Basin is composed of three main NNE striking normal faults (Figs. 4 and 7). The two faults that bound the basin dip toward each other, forming a full graben with two fault-bounded highs/horsts. On a cross-section, the faults appear listric with a maximum displacement of ~2 s (Fig. 4). They can be traced to travel times of 5-6.5 s. The southern ending of the Atlantis Basin is unclear on the seismic data: the three main faults either terminate abruptly toward the present-day shelf edge or continue under the continental slope where data are ambiguous. A newly identified basin is named here Poag Basin after USGS scientist emeritus C. Wylie Poag, who made seminal contributions to the study of the Atlantic margin stratigraphy. The Poag Basin bounds the northern extent of the Atlantis Basin (Fig. 7). It is a 130 km long half graben with a SW dipping listric border fault that is seismically visible to travel times of 5.5 s. North of the Poag Basin, the Franklin Basin is the shallowest basin under the GBB (Fig. 7). On its western side it is bound by three en-echelon normal listric faults that dip ESE and penetrate to a maximum travel time of 5.5 s. The maximum vertical displacement on the main faults is ~1.5 s. Antithetic and synthetic faults of smaller displacement are mappable to the east of the main faults.

The deepest part of the GBB, the Georges Bank Trough, is located east of the Poag and Atlantis Basins. Two normal faults bound the Georges Bank Trough to the north and west (F1 and F4 in Fig. 7) whereas the Yarmouth Arch bound it to the east. Although seismic penetration does not provide clear determination of its maximum travel-time, the data provide information about its fault orientations, surface dips, and general geometry. It consists of two fault-bounded steps (the bounding faults are marked F1 and F2 in Fig. 7). Both steps plunge to the SE toward N-S faults that bound the Trough to the SW (F3 and F4 in Figs. 4 and 7).

The area east of the Franklin Basin and north of the Georges Bank Trough diverts from the general seaward deepening trend of the margin. There, two rift basins, the Inner and Outer Yarmouth Basins are separated by a prominent basement horst - the Yarmouth Arch. The Inner Yarmouth Basin is a half-graben 50 km wide by 90 km long that extents to travel times greater than 4 s (Figs. 5 and 7). The basin and faults that bound it to the east strike NNE-SSW and gradually terminate towards the LeHave Platform (Fig. 7). A convergent transfer zone, where two opposing normal faults dip toward each other, separates the Inner Yarmouth Basin from the Georges Bank Trough. The dip of the eastern border faults of the Inner Yarmouth Basin is WNW making the Yarmouth Arch the footwall of this fault system. The fault system forms 2-4 tilted blocks between the Yarmouth Arch and the Inner Yarmouth Basin (Figs. 5 and 7). Cumulative vertical displacement of the Inner Yarmouth Basin fault system reaches ~3 s. Assuming no erosion of the footwall and seismic velocity of 5 km/s for the syn-rift section, that is equivalent to more than 7 km. The cumulative heave of this fault system reaches ~18 km. On a section view, these faults appear listric (Fig. 5). In their shallowest part, their inclination is 40 º to 30º. The inclination decreases as they penetrate ~3.5 s into the crust.

The Inner Yarmouth Basin and its bordering fault system comprise the upper crustal manifestation of a possible crustal-scale shear zone. Fig. 5 illustrates a zone of reflective lower crust <2 s above the Moho. Above this zone, at the northwestern part of the section, is a series of reflectors that mildly (<13º) dip landward. These reflectors are traceable over ~80 km, shallowing to the southeast. In the upper continental crust, these reflectors coincide with the fault system that forms the Inner Yarmouth Basin. Following the interpretation of similar observations at other rifts and continental margins (e.g. Clerc et al., 2015; Clerc et al., 2018; Fazlikhani et al., 2017; Phillips et al., 2016; Reston et al., 1996), these inner crustal reflectors may indicate detachment faulting, crustal shearing, and ductile deformation of the crust.

The Yarmouth Arch is a ~120 km long, 30 km wide, NNE-SSW trending elongated horst found east of the Inner Yarmouth Basin. Steep, east-dipping faults bound the Arch to the east and separate it from the Outer Yarmouth Basin. An E-W fault, oblique to the Yarmouth Arch, marks its southern termination and separates it from the Georges Bank Trough. The structure of the south-eastern corner of the Arch is not well constrained by the available data. However, the trend of neighboring areas to the south and east suggests that an elevated branch of the Arch may extend SE, toward the shelf edge. The Outer Yarmouth Basin is composed of two subbasins separated by an east-dipping fault. Overall, the entire ~200 km wide GBB, from the western Franklin Basin to the shelf edge, represents a zone of deformed and faulted basement.

#### 4.1.2 Long Island Platform

The top basement in the Long Island Platform is the shallowest of the three margin segments (Figs. 4 and 7). It descends from near sea-surface elevation at the shoreline to about 5 s under the continental slope along a convex trajectory (Fig. 7). The seismic data reveal three known rift structures: Nantucket Basin, Long Island Basin and New York Bight Basin (Fig. 7). Nantucket Basin is located in the eastern part of Long Island Platform, NW of Atlantis Basin. It is interpreted here as an arcuate half-graben with a down to the SE boundary fault. Reaching a maximum of ~3 s TWT, it is the deepest rift basin at the Long Island Platform. At the center of Long Island Platform is the Long Island Basin. Its border fault dips toward the ESE, down throwing its hanging wall to more than 2 s. The New York Bight Basin in the western Long Island Platform is composed of five identified faults. Due to the sparsity of data in this area, its faults' orientations are not well constrained, and the interpreted dips shown in Fig. 7 are apparent dips. Nevertheless, the easternmost fault of the Basin was identified on two profiles as having a westward dip. Thus, the other faults of the New York Bight Basin were assigned with a similar westward dip.

#### 4.1.3 Baltimore Canyon Trough

Offshore New Jersey, the top basement reaches more than 8 s TWT (Figs. 6 and 7). Reflection data do not allow identification of a single top basement reflector or a seismic facies boundary in these deep basin areas (Fig. 6). Hence, interpretation relies mostly on published refraction control points (LASE, 1986) that are tied to reflection profiles. In areas shallower than ~6 s, the top basement is identifiable on reflection data as well. In map view, the BCT has an asymmetric arcuate shape. To the north, the top basement plunges steeply southward from 1.5 s under the western Long Island Platform, to 8 s over less than a 100 km. Farther SW, offshore New Jersey, the top basement dips southeastward with the same amount of deepening occurring over ~150 km. SW of New Jersey and offshore Delaware Bay, the top basement deepens to about 6 s on an ESE trajectory. At the southern BCT the top basement dips mostly to the east. There, a sharp hinge separates a shallow (<3 s), gentle top basement surface under the inner shelf from the deeper part under the outer shelf (Figs. 7 and 8).

Few faults involving basement were identified at the BCT. The sparsity of faults in the deepest part, over 6 s, may be attributed to poor seismic resolution. A near-vertical, down-to-the-north, fault (Named here the Delaware Bay Fault, Figs. 4 and 7) separates the deep northern BCT from the shallower southern BCT. The fault has an E-W strike and a maximum vertical displacement of ~0.5 sec. A similar fault might be present at the opposing northern flank of the northern BCT (Fig. 4), although data sparsity does not allow it to be clearly identified and mapped.

Only one rift basin can be identified at the BCT in the offshore seismic grid, the Norfolk Basin, which is located under the inner continental shelf of the southern BCT (Fig. 7). Its border fault dips to the east and has a maximum displacement of ~1.5 s. A series of synthetic faults are located east of the border fault. East of the Norfolk Basin, two structural ridges plunge eastward under the outer shelf. It is not clear from the seismic data whether these structures are bounded by faults. About 70 km to the south of the Norfolk Basin, lies a ~20 km wide basement depression. Its imaging does not reveal clear faults that might bound it. South of that depression, the top basement is shallower (<3 s), dipping moderately eastward toward the shelf edge. Three elongated rift basins along the northern BCT hinge line that were previously described by Klitgord et al. (1988) and Benson and Doyle (1988) based on seismic reflection data were not identified using the denser dataset presented here.

### 4.2 Base Post-Rift (BPR)

The general structure of the BPR surface is that of a smooth surface along the top basement, along the top of the rift basins and along the top of the SDR where these overlay the top basement (Figs. 4 and 9). In the GBB area, the BPR descends towards the southeast from less than 0.5 s at the eastern Long Island Platform. Further east, seaward of the Gulf of Maine, the BPR first descends above syn-rift strata of the Inner Yarmouth Basin, forming a trough that plunges to the southwest. East of the Inner Yarmouth Basin, the BPR rises along the top basement of Yarmouth Arch, forming a 170-km long by 70-km wide elongated ridge that also plunges to the southwest (Figs. 5 and 7). The BPR then descends to the southeast above the syn-rift strata within the Outer Yarmouth Basin (Figs. 5 and 9). The trough above Inner Yarmouth Basin connects to a deeper and wider south-trending trough coincident with the Georges Bank Trough (as seen in the top basement map, Fig. 7). With travel times of 4.5 s, this is also the deepest part of the BPR under the GBB shelf. The descent from the ~0.5 s deep Gulf of Maine to the deepest trough occurs gradually over ~150 km.

The BPR surface at the Long Island Platform coincides with the top basement where rift basins are absent (Figs. 4 and 10). The BPR has a southward plunging convex structure along most of the Long Island Platform (Figs. 9 and 10). A steep E-W slope separates the Long Island Platform from the northern BCT.

The asymmetry of the BCT, as observed in the top basement surface, also characterizes the BPR. Similarly to the top basement, the BPR morphology shifts from convex (shallower parts) to concave in the deeper areas (Fig. 9). The dip in the deepest part of the BPR (> ~5 s) is gentler than the dip of top basement in the same locality. The gentler BPR dip is attributed to the filling of the space trapped between the top basement and BPR by SDR and possibly syn-rift strata. (Figs. 4, 6 and 8). The BPR at the outer northern BCT reaches more than 6.5 s (Figs. 6 and 9). To the south, the BPR dips mostly eastward. The faults, troughs and highs apparent in the southern BCT top basement have no expression on the BPR.

At the onshore Salisbury Embayment, the BPR is concave, deepening toward the BCT (Fig. 9). It outcrops at the landward edge of the coastal plain from New York City to the southern extent of the study area and reaches a maximum depth of ~2 s TWT beneath the coastline. In the northern part of the embayment, the BPR forms a concentric structure, plunging towards the central BCT.

### 4.3 Seaward Dipping Reflectors (SDR)

SDR appear on seismic data along the entire studied margin. Although their spatial extent and down-dip position change along the margin strike, several geomorphic characteristics remain similar. In all the sections that show both SDR and their underlying top basement surface, the SDR packages have a wedge-shaped geometry that thickens seaward and pinches out landward (Figs. 6 and 8). The SDR themselves toplap with respect to the BPR. At the GBB and Long Island Platform, the SDR landward termination is 10-30 km seaward of the present-day shelf edge, taken here as the 200 m isobath (Figs. 5, 10, 11 and 12). At the BCT, however, the SDR pinch outs are located more landward, underneath the continental shelf. The landward distance between the pinchout and the 200 m isobath decreases gradually from ~100 km at the northernmost BCT to ~30 km at the southern BCT. The seaward termination of the packages is less distinctive than their landward termination.

### 4.4 Moho depth

Moho reflectors in the USGS seismic lines were identified on dip profiles at the GBB, Long Island Platform and the southern BCT (Fig. 13A). At the GBB, four profiles revealed Moho reflectors at 8-10.5 s (Figs. 5 and 13A). Four dip profiles and one strike profile show a relatively continuous series of reflectors at depths of 9-11 s under the Long Island Platform. Moho reflectors are sparsely imaged on the USGS lines covering the BCT. They appear over short distances (tens of kilometers) as discontinuous reflectors on one strike profile and 6 dip profiles, mostly at the southern BCT.

Interpolation of interpreted Moho reflectors combined with published Moho picks yielded a regional structural map (Fig. 13B). Travel times to the Moho mostly range between 9 to 12 s. At the GBB the interpolated map shows a ~100-km-wide by 400 km long ridge in the Moho surface. This elevated Moho extends in a southerly direction from the inner Gulf of Maine to outer GBB and is located mostly in the region between the Franklin Basin and the Inner Yarmouth Basin (Fig. 7). The ridge is higher than its surroundings by 1-1.5 s. Under the Long Island Platform, the Moho exhibits general southward dips. Under the offshore portion of the northern BCT the Moho is deeper (~11 s) than under New Jersey coastal plain (~10 s). At the southern BCT, however, there is no clear distinction between the depth to the Moho offshore and onshore.

The heterogeneous distribution of seismic velocities above the Moho may cause the appearance of artificial structures on the TWT structural map. In that sense, the presence of thick, low-velocity sedimentary basins will increase the underlying Moho travel times. Some of the bias is resolved by looking at the crustal thickness map (See description of the BPR to Moho interval and supporting information).

## 4.5 Isochron maps

### 4.5.1 Base Post-Rift to Moho interval

The isochron between the BPR surface and the Moho was calculated regionally (Fig. 14A). We chose this interval and not the more orthodox top basement to Moho interval for two main reasons. First, the interpretation of the BPR surface is more straightforward than that of the top basement. Therefore, its spatial extent and degree of accuracy are higher, especially where thick syn-rift or SDR successions occur. Second, the use of the BPR as an upper datum for the calculation filters out short-wavelength (<50 km) thickness variations associated with rift basins. These basins manifest crustal deformation restricted to the upper crust that does not necessarily have mantle compensation. The BPR surface smooths these basin structures, thus emphasizing regional crustal thickness variations. The presented thickness could be treated as an upper limit for crustal thickness as the thickness trapped between the BPR and top basement is added to its calculation. On the deeper troughs (outermost BCT and the GBB trough), the difference between the crustal thickness and BPR to Moho thickness may reach >2 s. This difference nulls where rift-basins are absent.

The travel time interval of the BPR to Moho varies along and across the margin. It ranges between extreme values of <4 s at the outer northern BCT to ~12 s landward of southern BCT (Fig. 14A). The thickness in ~70% of the region is between 8 and 11 s. GBB is bisected by an NNE-SSW-oriented travel-time minimum which coincides with Inner Yarmouth Basin and Georges Bank Trough. There, thick syn-rift infill (up to 3 s) with velocities slower than the surrounding basement rocks (<5 km/sec for the syn-rift vs. ~6.3 km/sec for the continental crust) is expected to increase the travel time interval. This, in turn, causes artificial inflation of the BPR-to-Moho interval. Thus, the thickness minimum under the Georges Bank Trough and Inner Yarmouth Basin is probably even more dramatic than is observed in the time domain. Farther south, toward the GBB shelf edge, the thickness of the interval decreases to less than 5 s.

Unlike the GBB, the Long Island Platform is almost devoid of syn-rift basins with velocities slower than crustal velocities (Fig. 7). Travel-time crustal thickness at the Long Island Platform, is relatively constant, between 8.5 and 9.5 s (Fig. 14A). Similar values extend south west under the New Jersey coastal plain. At the BCT, the BPR to Moho interval has an asymmetric thickness minimum close to the shelf edge offshore New Jersey. The transition from >9 s thickness at the Long Island Platform and New Jersey coastal plain to the thinnest part at the BCT (<4 s) occurs over less than 110 km. Under the outer southern BCT shelf the interval thickness is 6-7 s; 2-3 s thicker than under the LASE profile ~250 km to the north. The thickness gradient is steepest under the western flank of the southern BCT, where the interval thins by 4 s over ~50 km.

The gradient map of the BPR to Moho travel-time thickness shows a “hinge line” where rapid seaward thinning of the crust (in TWT) begins (red line in Fig. 14B). The hinge line roughly bounds the BCT and GBB on the west and the Long Island Platform on the east and south. At the BCT, the steepest local gradient is found immediately east of the hinge line.

#### 4.5.2 Early Post-rift

The thickness of post-rift Jurassic sediments, described below, indicates the distribution of the depocenters that developed in the early stages of the drift phase, 30-45 Myr after the continental breakup. Post-rift Jurassic sediments are concentrated in two depocenters under the continental shelf, filling the GBB and the BCT (Fig. 15). The GBB depocenter is an NNE-SSW trough with a maximum travel-time thickness of ~1.8 s at its southern half. It decreases gradually northward to ~ 1 s at the northern edge of the map. Sediment thickness is much thinner (<800 milliseconds) east of the GBB depocenter. At the Long Island Platform post-rift Jurassic sediments are found only at the outer shelf (Figs. 4 and 15). The BCT Jurassic depocenter is asymmetric, thicker in the north (>3.5 s) than in the south. North of there, the Jurassic thins rapidly toward the Long Island Platform (Figs. 4 and 15) and pinches-out after ~100 km. The western edge of the BCT depocenter is not constrained by the offshore seismic data at the northern BCT.

### 4.6 Thermal subsidence and lithospheric structure of the Georges Bank Basin

Since the formation of a volcanic margin is to a large extent a thermal process, the rift-stage structure of the thermal lithosphere should be examined. To estimate the lithospheric thinning patterns at the time of rifting, we evaluate the thermal relaxation of GBB as expressed by the thickness of the early post-rift sequence. The connection between early post-rift thicknesses and lithospheric thinning is valid assuming that the thinning occurred shortly before breakup and ended with the onset of seafloor spreading (McKenzie, 1978). This assumption is supported by direct age dating of the syn-rift sequence in drill holes at the GBB (e.g. Poag, 1991) and by seismic stratigraphic analysis that shows the rift basins and basement rocks all being truncated by the post-rift unconformity (i.e. BPR in Figs. 4 and 5; Klitgord et al., 1988). The inference of a spatial connection between lithospheric thinning and early post-rift depocenter also assumes very low flexural rigidity of the lithosphere. Such low rigidities characterize regions of upwelled asthenosphere (Watts et al., 1982) and young volcanic margins specifically (Tian & Buck, 2019).

The post-rift Jurassic deposits represent the first 30-45 Myr of deposition on the ENAM after breakup. During this initial post-rift phase where the lithosphere had been thinned, thermal gradients are expected to be steep and thermal subsidence high (McKenzie, 1978). Thermal subsidence indeed peaked during the early post-rift of ENAM, forming most of the Jurassic accommodation space (Poag & Sevon, 1989; Steckler & Watts, 1978). Hence, the post-rift Jurassic thickness (Fig. 15) can be treated as a proxy for identifying thermal subsidence patterns and thus areas of lithospheric thinning. Fig. 16 shows that the thicknesses of the BPR to Moho across the GBB is inversely proportional to the thickness distribution of the early post-rift Jurassic unit. For example, areas where the BPR to Moho interval is thinnest (5.8 s, 17.4 km, assuming an average velocity of 6 km/s) are overlain by the greatest thickness of post-rift Jurassic sediments (1.85 s, 4.1 km, assuming an average velocity of 4.5 km/s based on well data (Taylor & Anderson, 1982)). Areas with thicker BPR to Moho (8 s, ~24 km) are overlain by thinner Jurassic strata (0.8 s, ~1.8 km). The spatial relations between crustal thinning and early post-rift thermal relaxation are evident on a map view (Fig. 17). The crustal hinge line outlines the western and northern bounds of the GBB Jurassic depocenter and, seemingly the zone of lithospheric necking. This suggested spatial coincidence of crustal and lithospheric boundaries, together with the thickness relations shown in Fig. 16 allude that thinning of the crust and mantle lithosphere under GBB spatially overlapped. It is possible that not only the crust deformed and thinned over a ~200 km wide zone, but so did the lithosphere.

## 5 Discussion

### 5.1 Breakup volcanism, the East Coast Magnetic Anomaly and the width of the extended continental crust

The final stages of the formation of the ENAM were accompanied by voluminous magmatic eruptions and the emplacement of the ECMIP. The results presented here show that the landward extent of the volcanism, as marked by the pinch-out location of the SDR wedge, spatially correlates with the western limit of the ECMA (Figs. 5, 6, 8, 10, 11 and 12). This observation supports previous correlations that were based on a few isolated 2D seismic lines (e.g. Austin et al., 1990; Holbrook & Kelemen, 1993). However, the relationship between the landward extent of the SDR wedge and the corresponding magnetic anomaly varies along the margin. Whereas at the GBB and the Long Island Platform the SDR pinch-out correlates with the landward edge of a narrow (~80 km) high amplitude anomaly that is regarded as the axis of the ECMA (Behn & Lin, 2000; Benson & Doyle, 1988; Klitgord et al., 1988), at the BCT, and in particular at its northern part, the SDR terminate where a low amplitude extension of the anomaly feathers out (Figs. 6 and 12). It is noteworthy that this extension also appears in a reduced to pole version of the magnetic anomaly map, as presented by Behn and Lin (2000).

To evaluate the extent of crustal thinning west of the breakup line, it is crucial to define both the landward and seaward bounds of the area of thinned continental crust. Rift structures are widely spread (up to 400 km) between the eastern Appalachians and the continental slope (Fig. 2; Withjack et al., 2012). Yet, onshore rift basins usually overlay continental crust of normal or thicker-than-normal thickness (>35 km, Li et al., 2018). Stretching in these areas appears to be restricted to the upper crust and does not involve local mantle compensation (Harry & Sawyer, 1992; Sawyer & Harry, 1991; Li et al., 2018). Most of the thinning occurs farther seaward, along a margin-parallel belt (Fig. 14). Whilst the data presented here provides a good estimate of the landward boundary of this thinning belt (i.e. the hinge line, Fig. 14), its seaward edge, where the crust turns entirely igneous, is more elusive (for further discussion regarding the challenges in determining the edge of the continental crust see Eagles et al., 2015). The high amplitude pick of the ECMA was previously regarded as the approximate position of the seaward edge of the continental crust (i.e. ocean-continent transition; e.g. Austin et al., 1990; Greene et al., 2017; Klitgord et al., 1988; Withjack et al., 2012). In addition, interpretations of refraction profiles along the ENAM suggest that the crust located seaward of the ECMA axis is entirely igneous or oceanic (Figs. 6, 8 and 12; Austin et al., 1990; Holbrook et al., 1994; Talwani et al., 1995; Talwani & Abreu, 2000; Shuck et al., 2018). Considering the paucity of available refraction data, the ECMA is assumed here to mark the seaward edge of the continental crust. Therefore, the crust in the area bounded by the hinge line and the axis of the ECMA is considered thinned continental crust, probably intruded and partially overlaid by breakup volcanism. The width of this area, when measured perpendicular to the ECMA, reaches ~220 km at the GBB and ~110 km at the northern BCT (Fig. 12). It is narrowest at the Long Island Platform and southernmost BCT where it extends for ~60 km.

### 5.2 Along margin variability: key differences between the segments

Our data reveal an along-margin variability in crustal structure, deformation style, volcanic addition and post-rift sedimentation of the ENAM. The variability is especially noteworthy between the GBB and the BCT- two parallel segments, oriented perpendicular to the rifting-related extensional regime (Withjack et al., 2012). Variations are manifested in several ways: (a) whereas a narrow band of thinned continental crust lies seaward of a steep hinge zone at the BCT (<110 km), a gentle hinge zone borders a wide (up to 220 km) thinned zone at GBB; (b) few rift basins are observed at the BCT whereas a complex system of well-developed rift basins and detachment faulting constitutes the base of GBB; (c) volcanism in the form of SDR at the BCT, reaches landward <50 km east of the hinge line whereas the landward boundary of the SDR at GBB is located much farther seaward under the continental rise, separated from the hinge line by up to 200 km; (d) the early post-rift sediment fill of the BCT consistently thickens seaward, whereas at the GBB the thickness increases toward the middle shelf and decreases again towards the Yarmouth Arch (basement high) under the outer shelf.

A broad zone of thinned crust landward of the ECMA is also observed in the volcanic Scotian margin of Canada, immediately north of the GBB (Fig. 2; Deptuck & Kendell, 2017; Savva et al., 2016). Water depth at the Scotian Margin reaches ~2.5 km (Savva et al., 2016), Jurassic sediment thickness is ~3 km (Deptuck & Kendell, 2017) and crustal thickness is 20 km (Dehler, 2012). It, therefore, appears that a broad zone of crustal and likely lithospheric thinning landward of the magmatic outpouring extends along a substantial (650 km) portion of the Atlantic margin, which includes both the GBB and the volcanic SW-most Scotian margin.

The Long Island Platform, located between the GBB and BCT, has a relatively thick crust (8-10 s or ~31-25 km), few extensional structures and minor early post-rift subsidence (0-3 s or 0-5 km top basement depth; Figs. 4 and 7). Its hinge line, top basement dip, and ECMA trend are oblique to those found at BCT and the GBB. At the eastern Long Island Platform, the BPR-to -Moho interval maintains its thickness from the inner shelf to the shelf edge (Fig. 14A) and forms a steep BPR-to -Moho hinge, about 50-km-away from the ECMA and the SDR. The obliquity of the Long Island Platform, relative to its neighboring segments and the minor thinning of its crust were previously interpreted as the result of transform or wrench motion during rifting (Hutchinson & Klitgord, 1988; Klitgord & Behrendt, 1979; Klitgord et al., 1988; Thomas, 2006). Some have linked the obliquity of the Long Island Platform and its suggested transform motion to the intersection of the margin at this segment by oceanic fracture zones (Klitgord et al., 1988; Le Pichon & Fox, 1971). Yet, recent studies have rejected the genetic connection between oceanic fracture zones and syn-rift strike-slip faults (e.g. Taylor et al., 2009). While tensile strain in an oceanic lithosphere tends to localize in an orthogonal or parallel direction (Dauteuil & Brun, 1996), strain in a continental lithosphere may be accommodated by oblique rifting (e.g. Gulf of California (Bennett & Oskin, 2014) and Gulf of Aden (Autin et al., 2013)). Thus, inference regarding the transform nature of the Long Island Platform cannot be based solely on its spatial relation to oceanic fracture zones. The intrinsic characteristics of the Long Island Platform do not match these expected from a transform margin. It lacks fundamental structures of transform margins such as a marginal ridge, continent-ward tilted horizons and a marginal plateau (Mercier de Lépinay et al., 2016). On the other hand, the presence of a sharp hinge, minor crustal thinning, and post-rift subsidence fits an obliquely rifted margin (Davison, 1997). From a kinematic perspective, the Long Island Platform might have served as an accommodation/transfer zone (e.g. Morley et al., 1990; Schlische & Withjack, 2009) between two orthogonal rift segments.

### 5.3 Examination of models for the creation of volcanic margins

Models of magmatic rifting and volcanic margin formation predict a narrow zone (<100 km) of crustal and lithospheric thinning and steep relief at the base of the lithosphere. The narrow geometry is considered to be either the result of weakening and localizing processes that stem from the steep geothermal gradient at volcanic rifts (Buck, 2004; 2006; Geoffroy, 2005; Geoffroy et al., 2015; White & McKenzie, 1989) or the initial conditions required for melt generation (Mutter et al., 1988; Simon et al., 2009; Van Wijk et al., 2001). The proposed models are supported by globally distributed observations of narrow volcanic margins (e.g. Franke, 2013; Franke et al., 2007; Hopper et al., 2003; Hopper et al., 1992; Paton et al., 2017; Schnabel et al., 2008; Tréhu et al., 1989) including the crustal structure of the BCT (Figs. 6, 8 and 14; Holbrook et al., 1994; LASE, 1986; Lizarralde & Holbrook, 1997).

Although the GBB is volcanic, it does not fit the observations and models of a narrow thinning zone that is usually ascribed to volcanic margins. The observations presented here indicate a ~220 km wide zone of crustal thinning at the GBB (Figs. 5, 7, 14). The thinning is manifested by well-developed brittle extensional structures possibly coupled with ductile deformation of the middle crust (or below). The crust is considerably thinner than typical continental crust (35-40 km; Christensen & Mooney, 1995) and reaches a minimum thickness of 4-6 s or 12-19 km, assuming an average crustal Vp of 6.3 km/s (Fig. 5; Fig. S1 in supporting information). The wide extent of thinned crust, together with the presence of middle crust detachment faulting and developed surface extensional structures, are usually ascribed to magma-poor margins. At such margins, the zone in which such features occur is referred to as the 'necking domain' (Peron-Pinvidic et al., 2013; Reston, 2009; Sutra et al., 2013). The necking domain represents a thinning phase during which strain localization and deformation of the middle and possibly lower crust occurs, promoting drastic crustal thinning. In the sequence of events that leads to the formation of magma-poor margins, thinning follows a phase of tectonic stretching that is locally uncompensated by mantle uplift (i.e. 'stretching phase') and predates hyperextension of the crust and exhumation of mantle rocks (i.e. 'hyperextension/exhumation phase'; Peron-Pinvidic et al., 2013). The juxtaposition of a wide necking domain and SDR makes the structure of the GBB (and likely also the southwest Scotian margin) a hybrid between an underdeveloped magma-poor margin and a volcanic margin.

The broad (> 200 km) syn-rift thinning under the GBB challenges the understanding of the thermomechanical conditions suggested for the formation of volcanic margins. The initial conditions required for a volcanic breakup, as proposed by Mutter et al. (1988), include a sharp near-vertical asthenosphere-lithosphere boundary that would induce convective partial melting. This condition was most probably not met at the GBB where the relief of the base of the thermal lithosphere was moderate and thinning of the lithosphere probably took place over 200 km across the margin. Buck (2004, 2006) proposed that a considerable amount of lithosphere extension over a hotter-than-normal asthenosphere would be accommodated by dike intrusions. Moreover, high heat flux around the intrusions would weaken the lithosphere and promote strain localization toward the rift axis. This mechanism would result in a minor and localized thinning. Although this model might successfully explain the narrow structure of the BCT, it fails to explain the broad necking zone under GBB. Kelemen and Holbrook (1995) also proposed that lithospheric necking was minor prior to the formation of the volcanic BCT and originated in melts formed under high pressure (up to 4 GPa) and temperatures, which they attributed to the presence of a thick lithospheric lid above the melt. At the GBB, however, pre-magmatic necking reduced the thickness of such a lid. Geoffroy et al. (2015) emphasized the role of continentward-dipping detachment faults play during crustal necking at volcanic margins. The abundance of oceanward dipping faults at the GBB (Fig. 7), the ~200 km offset between the crustal necking and the ECMIP (Fig. 17), the lack of evidence supporting continentward dipping faults associated with the SDR along the entire ENAM (Figs. 8, 10 and 11; Lizarralde & Holbrook, 1997) do not support the model proposed by Geoffroy et al. (2015).

A possible reconciliation between lithospheric thinning and the melting under high pressure might include a time-varying geotherm. In this scenario, initial rifting would take place over a "cold" mantle (potential temperature is <1300ºC, Reston, 2009) forming a wide, magma-poor structure. If mantle temperature were to rise later, this magma-poor structure would be superimposed by a narrower volcanic structure. If this is the case for the rifting of the GBB, then the increase in mantle temperature is not expected to result from the geometry of the rift as in the edge-driven convection models (Mutter et al., 1988; King & Anderson, 1998). Similarly, elevated mantle temperature could not be related to a heated pre-rift mantle such as in the continental insulation models (e.g. Anderson, 1982; Brandl et al., 2013; Hole, 2015) since the initial rifting took place over a cold mantle. Rather, it should stem from processes not related to the rift itself, such as a mantle plume (White & McKenzie., 1988). If, as some suggested, the plume was situated at the southern part of the rift (Wilson, 1997; Ruiz-Martínez et al. 2012), the amount of magmatic additions to the margin should decrease northward. Yet, the intensity of the ECMA does not decay northward (Fig. 2). Since the amplitude of the ECMA correlates with the added magmatic volume (Holbrook & Kelemen; 1993; Talwani et al., 1995), there is also no sign of northward decrease in the volume of the breakup magmatism. The independence of the reduced-to-pole ECMA and the SDR burial depth supports the connection between the intensity of the ECMA and the volume of the volcanic rocks (Figures 7b and 7c in Behn & Lin, 2000). Moreover, some geochemical (Whalen et al., 2015; Shellnutt et al. ,2018; Elkins et al., 2020) and geophysical (Shuck et al., 2019) evidence cast doubt on a mantle plume origin of CAMP and ECMIP melts. Other mechanisms such as volatile enrichment of the mantle (Elkins-Tanton, 2007) and slab break-off (Whalen et al., 2015; Elkins et al., 2020) may also explain the sudden initiation of magmatism. Unfortunately, these cannot be confirmed or disproved using the data presented here.

The eastern North Atlantic volcanic margin in northern Europe was formed by successive rifting events dating from the Late Devonian to the early Cenozoic volcanic breakup (Doré et al., 1999; Roberts et al., 1999). This led some authors to suggest that wide rifting, like rifting that predates to the formation of magma-poor margins, also predates the formation of volcanic margins (Eldholm et al., 1995, 2000). However, the protracted nature of rifting of the eastern north Atlantic implies that although the crust under that margin was thin, the lithosphere was not necessarily thin at the onset of rift magmatism. Cooling of upwelled mantle between rifting phases should have resulted in the re-thickening of the lithosphere. Unlike the European North Atlantic, the Central Atlantic, and the ENAM in particular, had experienced a relatively short and continuous rifting that was immediately followed by seafloor spreading (Withjack et al., 2012). Recently, Guan et al. (2019) proposed that volcanic margins that experienced non-magmatic rifting shortly before their volcanic breakup exhibit narrow necking zones, whereas longer time spans between failed rifting and volcanic breakup result with wide volcanic margins.

This is in contradiction with the observed variations between the GBB and BCT, which experienced similar rifting histories before their volcanic breakup.

The African side of the Atlantic South Austral margin is a possible example of a volcanic margin that was tectonically thinned soon before its magmatic phase. Like the GBB, the southernmost part of the margin exhibits a wide area of thinned continental crust, high-strain extensional structures and detachment faulting along with SDR that correlate with a prominent magnetic anomaly (Blaich et al., 2011). The geometry of the adjacent segment to the north exhibits the typical narrow and steep margin. Examining the Brazilian margin, Stica et al. (2014) interpreted a 280 km wide zone of necked and intruded crust between the hinge line and the first oceanic crust of the Pelotas Basin. Yet, unlike the GBB, most of this zone underlies a thick SDR wedge, which the authors interpret as "continental igneous crust". A modern analogue for the rifting of the GBB may be found at the Manda Hararo active rift in central Afar. There, Stab et al. (2016) observed a wide zone (~200 km) of crustal necking, mid-crustal detachment faulting along with abundant volcanism.

Although it is clear from our results that the style of thinning varied along the ENAM, the causes for these variations remain unsettled. Trying to explain the difference in crustal structure and post-rift subsidence, Klitgord et al., (1988) and Wernicke and Tilke, (1989) proposed a simple shear model (Wernicke, 1985; Lister et al., 1991) with alternating polarities between the segments. Modeling efforts have shown, however, that simple-shear rifting does not allow enough melt production for the formation of volcanic margins (Buck et al., 1988; Latin & White, 1990; Simon et al., 2009). More recent numerical modeling addressed the width of the lithosphere necking zone at rifts and passive margins (e.g. Svartman Dias et al., 2015; Tetreault & Buiter, 2018). According to these models, two main factors appear to determine the architecture of a rift system: the extensional strain rate and the rheology of the lithosphere. Estimates of syn-rift divergence rates at ENAM range between 2-6 mm/year for the Carolina Trough (Kneller et al., 2012; Ruiz-Martínez et al. 2012, respectively) to 8 mm/year for the BCT (Schettino & Turco, 2009). The margin-wide distribution of slow to ultra-slow divergence of similar orientation cannot account for the lateral variation in margin architecture. Thus, we suggest rheological rather than kinematic contrasts were dominant in shaping the margin's width.

### 5.4 The origin of along-margin variability at the ENAM

Previous interpretations and numerical modeling of the rifting and breakup of the Central Atlantic margin mostly assumed initial conditions of homogenous rheology of the continental lithosphere subjected to tensile stresses and perhaps underlying heat and melt source (Klitgord et al., 1988; Wernicke & Tilke, 1989; Dunbar & Sawyer; 1989). Furthermore, most margin-scale rifting models lack the crustal and likely lithospheric lateral heterogeneity as manifested in the crustal fabric of eastern North America and the time-varying geotherm imposed by the emplacement of CAMP and ECMIP. The lithosphere in which rifting and breakup occurred was the outcome of ~160 Myr of west-dipping subduction, collision and right-lateral translation (Hatcher, 2010; Van Staal et al., 2009; Hibbard et al., 2007,2010). The convergence phase ended with the collision of Gondwana along the Rheic/Allegahanian suture at ~280 Ma, leaving a heterogenous pre-rift lithosphere (Figs. 2b and 18). In addition to the spatial rheology variations, the introduction of heat by the emplacement of CAMP and ECMIP added a time-varying component to the rheological structure of the lithosphere (Kelemen & Holbrook, 1995; Marzoli et al., 1999). To try and address these complexities, we first examine the along-strike variability of ENAM's crustal building blocks and their response to the pre-magmatic rifting and later examine the effect of magmatism on the rift architecture.

5.4.1 Rheological controls on the pre-magmatic rifting

Examining the pre-rift crustal fabric reveals major compositional differences along the strike of the ENAM. The outboard portion of the Appalachian crust is composed of peri-Gondwanan Terranes that were accreted to Laurentia before the Alleghenian orogeny. Meguma terrane at the northern part of the margin (Figs. 2, 18b and 18c), is the easternmost and latest accreted terrane to Laurentia (Hibbard et al., 2007; Hatcher et al., 2010). Exposed in Nova Scotia, the Meguma terrane overthrusts the Avalon terrane to the NW (Figs. 18b and 18c). The Avalon terrane overthrusts the Gander terrane from New England to Newfoundland, but probably abuts the older Appalachian belts (the Goochland or Piedmont domains) landward of the BCT (Figs. 2 and 18a; Hatcher et al., 2010; Hibbard et al., 2006 Sheridan et al., 1993). Basement rocks under the GBB and the Scotian Shelf are interpreted to belong or be closely related to the Meguma terrane (Hutchinson et al., 1988; Pe Piper & Jansa, 1999; Kuiper et al., 2017). To the south, the Avalon Terrane was suggested to underlie the BCT constituting the most outboard Paleozoic terrane of this segment (Sheridan et al., 1993; Hatcher et al., 2010).

The Meguma and Avalon terranes have different compositions. The Meguma terrane is composed of 10-12 km of metasedimentary sequence (White et al., 2010) that overlies crystalline rocks of Gondwanan passive margin affinity. Both metamorphic and crystalline rocks are intruded by mostly felsic plutons of Devonian age (van Staal et al., 2009). The Avalon terrane is composed of several arc-related volcano-sedimentary belts. The oldest exposed Avalonian rocks in Newfoundland represent oceanic crust and are composed of plutonic and volcanic rocks of gabbroic composition (O'Brien et al., 1996). These rocks are overlain and intruded by Neoproterozoic sediments and arc-related magmatic rocks of bi-model composition (O'Brien et al., 1996; van Staal et al., 2009). Although a full lithological description of the two terranes is lacking, the thick metasedimentary sequence and presumably felsic basement of the Meguma terrane should result in a weaker rheology compared to the rheology expected from the intermediate-mafic Avalonian composition.

The compositional differences between the terranes were manifested during the pre-magmatic Mesozoic extension. In areas where the two terranes juxtapose, extension-related crustal thinning remained confined to the Meguma terrane. Inboard of the Meguma-Avalon suture, the Avalon terrane is observed to be mostly unbroken and unthinned (Figs. 18b and 18c). For example, Pe Piper and Jansa (1999) showed that crustal necking offshore Nova Scotia was limited to the Meguma basement. Similar relations exist farther south between the unthinned Avalon crust of the Gulf of Maine and the thinned Meguma crust under the GBB (Hutchinson et al., 1988; Keen et al., 1991). Our suggested hinge line in the GBB coincides with the Hutchinson et al. (1988) and Keen et al. (1991) boundary between the Avalon and Meguma terranes (Fig. 14) and implies that the Meguma terrane had a weaker, more easily deformed crust in which extensional strain concentrated. More generally, where the Eastern North American margin included the Meguma terrane, the distribution of rift basins is restricted to the Meguma belt (Figs. 2, 18b and 18c). Where the Meguma terrane is absent and the Avalon terrane constitutes the outboard terrane, rift basins developed farther inland on top of older Appalachian domains (Figs. 2 and 18a; Hatcher et al., 2010). If our hypothesis is correct, the weaker Meguma terrane accommodated the extensional stresses, whereas the stronger Avalon terrane resisted the extensional deformation and transferred the stress to adjacent areas. Furthermore, post-CAMP-intrusions faulting at the rift-basins onshore the BCT (Withjack et al., 2012) implies that strain localization, and thus necking (Buck et al., 1999) of the crust under the BCT did not occur earlier than 200 Ma. We argue that the necking of the BCT was made possible only when rifting was magma-assisted, later than ~195 Ma (see later discussion).

The weaker inherited rheology of the GBB allowed rifting to progress from stretching to necking without the need for magmatic softening. The weak Meguma rheology facilitated deep detachment faulting, shearing, and ductile behavior of the middle to lower crust (Fig. 5) along with intense brittle deformation of the upper crust (Fig. 7). The fault-bounded rift basins in the GBB are coincident with the zone of crustal thinning. The age of these basins is considered pre-SDR (Carnian-Norian age: 237-208.5 Ma; Poag ,1991). Thus, the 200 km wide crustal and possibly lithospheric necking zone observed at the GBB resulted from pre-magmatic rifting. The presence of the weak rheology of the Meguma terrane probably enabled wide necking (Svartman Dias et al., 2015). Thus, we propose that a composition-controlled strain distribution determined the along-margin variations in the pre-magmatic necking stage as observed on our data.

#### 5.4.2 Magma-assisted rifting at ENAM

The Eastern North American Rift System entered its magmatic phase with the emplacement of CAMP at ~200 Ma, 40-30 Myr after rifting began. Fault-controlled subsidence onshore the BCT segment mostly ceased a few Myr after the emplacement of CAMP (Withjack et al., 2012). The abandonment of faults landward of the ECMIP in conjunction with the initiation of volcanism is also observed at the GBB. There, the SDR emplacement follows the Post-Rift Unconformity (Klitgord et al., 1988). Le Roy and Pique (2001) describe oceanward migration of strain simultaneously with volcanism at the African conjugate of ENAM. Early passive margin models would attribute the cessation of faulting to the onset of seafloor spreading (Falvey, 1974; McKenzie, 1978), suggesting the emplacement of CAMP and the SDR are related to the initiation of seafloor spreading. However, Shuck et al. (2019) and Kelemen and Holbrook (1995) showed that the generation of the magmas that formed ECMIP and the subsequent proto-oceanic crust took place under a lithospheric lid 15-70 km thick. In other words, the tectonic transition associated with the emplacement of the ECMIP does not signify the breakup of the lithosphere or the rift-drift transition but rather a change in nature of strain accommodation that was from this point dominated by the intense magmatism instead of faulting.

Models predict that dike intrusion would reduce the tectonic force required for mechanical stretching and promote strain localization, thus narrowing a rift system (Buck et al., 1999; Buck, 2004, 2006). The reduction in lithospheric strength is attributed to heating caused by the magmatic intrusions. The applicability of the suggested relationship between the magmatism of the rift and strain localization at ENAM could be examined by comparing its volcanic and magma-poor segments. The ENAM volcanic to non-volcanic transition occurs north of the GBB, offshore southern Nova Scotia (Fig. 2; Keen & Potter, 1995; Dehler, 2012; Deptuck, 2020). The rift basins north of the transition and landward of the ECMA (Fundy, Mohican, and Orpheus basins) continued accumulating sediments 5-25 Myr after the emplacement of CAMP (Withjack et al., 2012). That is, strain localized and faulting ceased only in segments where CAMP magmatism was followed by magmatic rifting associated with the emplacement of extrusive basalts (SDR). Similar magmatic localization occurred at the Afar region in east Africa where localized volcanism replaced faulting along widely distributed border faults (Wolfenden et al., 2005; Keir et al., 2006).

The crustal structure of the BCT fits observations at the currently active magma-assisted East African Rift. The necking zone of the BCT is narrow (80-110 km) and is overlaid by SDR. The hinge line roughly parallels the landward edge of the SDR alluding to a genetic relation between volcanism and crustal thinning (Figs. 6, 8 and 17). Similarly, at the northern part of the East African Rift, zones of localized crustal thinning overlap areas of voluminous basaltic flows interpreted as early-stage SDR (Bastow & Kier, 2011). To explain the tight connection between volcanism and crustal thinning, Bastow and Kier (2011) proposed that initially, repetitive, localized magmatic intrusions reduced lithospheric strength without reducing crustal thickness. Once sufficiently weakened, the lithosphere thinned mechanically along a narrow band. The narrow thinning resulted in decompression melting and extrusion of voluminous basaltic flows above the area of intruded and thinned continental crust. The BCT crustal structure and its relation to the distribution of SDR lead us to suggest that a similar sequence of events occurred during the ENAM magmatic phase.

With the transition to the magmatic phase later than 200 Ma, the dominant factor in determining the rheology, and thus the locus of straining, was no longer the composition of the crust but the strength reduction by magmatic intrusions. At this stage, the rift basins west of the ECMA were abandoned and strain migrated toward areas weakened by diking and heating (Figs. 18a.3 and 18b.3). Therefore, the structures inboard of the ECMA represent the pre-magmatic deformation, whereas the structures overlapping ECMA resulted from superposition of pre-magmatic and magmatic rifting. Offshore central and northern Nova Scotia, where the rift never turned magmatic (Keen & Potter, 1995), crustal thinning continued after 200 Ma as indicated by the presence of hyperextended crust offshore (Fig. 18c.4; Funck et al., 2004; Wu et al., 2006). An alternative explanation for continued rifting in central and northern Nova Scotia up to ~175 Ma was that the breakup was diachronous being earlier in the south than in the north (Withjack et al., 2012). Recent work by Shuck et al. (2019) suggests however, that extension without seafloor spreading also persisted until around that time (175 Ma) offshore Cape Hatteras, just south of our study area. Therefore, the breakup does not appear to have been diachronous.

#### 5.4.3 Rheology and across-ocean asymmetry

The previous paragraphs discussed the along-strike heterogeneity of the ENAM. Recent studies of the west African margin show that the structure also varies between the conjugate pairs across the Atlantic Ocean (e.g. Labails et al., 2009; Biari et al., 2017: Klingelhoefer et al., 2016). The African conjugate of the BCT has a narrower necking zone, more moderately thinned crust and fewer or no SDR compared to the BCT (Labails et al., 2009; Biari et al., 2017). Data regarding the crustal structure of the conjugate of the GBB is lacking. The Moroccan conjugate of northern Nova Scotia is also narrower and thinner than its American pair (Biari et al., 2017). Similar to the ENAM, the African conjugate underwent oceanward strain localization associated with late Triassic-early Jurassic volcanism (Le Roy & Pique, 2001). We speculate that, also like the ENAM, the African inherited pre-rift rheology determined the nature of the pre-magmatic rifting. We propose that the structural asymmetry might reflect the asymmetry in rheological properties between the conjugate pairs. Following a prolonged history of westward subduction and collision, the Permian North American side of the rift was made of a series of peri-Gondwanan accreted terranes overlying a wedge of Laurentian (Grenville) crust that thinned toward Gondwana (Fig. 18; Hibbard et al. ,2006; Hatcher et al., 2010; Cook & Vasudevan, 2006; Sheridan et al., 1993; Sheridan et al., 1999, Hughes & Luetgert, 1991; Marillier et al., 1989). The Rheic/Alleghenian suture separated the peri-Gondwanan terranes from the over-thrusted African Craton (McBride & Nelson, 1988; Villeneuve, 2005). McBride and Nelson (1988) suggested that breakup and the emplacement of the ECMIP followed the Rheic/Alleghenian suture and the suture served as a zone of weakness during the Mesozoic rifting (Figs. 18a.3 and 18b.3). The coincidence of the ECMIP with the suture would have left the Appalachians and their accreted terranes on the Laurentian (North American) side of the ocean and the African Craton on the Gondwanan side. If pre-magmatic extensional deformation concentrated on the peri-Gondwanan terranes (see previous discussion) and other Appalachian weakness zones, then the African side of the rift should have remained mostly unthinned. A full model describing the interaction between the dying convergent Paleozoic boundary and the birth of the Mesozoic ocean is beyond the scope of this paper. However, we note that such model will have to consider the inherited asymmetry and the uneven distribution of the crustal and lithospheric rheology.

## 6 Conclusions

A full crustal model of the ENAM shelf from Cape Hatteras to the U.S-Canada border was constructed and incorporated with seismic interpretation and mapping of upper crustal structures, breakup volcanism and early post-rift sedimentation patterns to examine the nature of the pre-magmatic thinning of the crust and mantle lithosphere in a volcanic margin setting. The results are based on seismic interpretation of more than 64,000 km of seismic reflection profiles tied to 40 wells and of published data. Dense data and newer processing and visualization techniques provided significantly more detailed crustal and fault structures of the ENAM shelf than was previously available. We found that the structure of the southern and northern BCT is typical of a volcanic continental margin with a narrow (~50 km) transition zone between a normal thickness continental crust and the breakup volcanism. The crustal structure of the GBB shows a broad zone (≤200 km) of crustal thinning landward of the SDR inferred to be coupled with a broad zone of lithospheric thinning. To explain these differences, we divide the rifting into pre-magmatic (prior to the emplacement of ECMIP) and magma-assisted rifting. While the GBB underwent intense pre-magmatic thinning, the BCT experienced no or minor thinning prior to the emplacement of ECMIP. We suggest that the nature and vigor of pre-magmatic rifting were determined by the spatial distribution of the pre-rift crustal rheology. Weaker rheology of the Meguma terrane underlying the GBB allowed intense faulting and crustal thinning, whereas the stronger rheology of the Avalon terrane underlying the BCT inhibited crustal thinning and transferred the tensile stresses westward to the older Appalachian domains. Magma-assisted rifting started with the emplacement of ECMIP (later than 200 Ma). It included localized magmatic heating and intrusion. Heating overwhelmed the compositional constraints on the rheology and facilitated oceanward strain localization. Localized straining resulted in a narrow necking zone overlaid by SDR. We speculate that the cross-ocean asymmetry in deformation and magmatism between the passive margins of Africa and North America may have also been governed by the heterogeneous distribution of the rheology.

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**Figure 1**. Schematic comparison between A) a volcanic continental margin (modified after Doré, & Lundin, 2015; Franke, 2013; Eldholm et al., 1995) and B) a magma-poor continental margin (modified after Doré, & Lundin, 2015; Franke, 2013; Peron-Pinvidic et al., 2013; Sutra et al., 2013). Abbreviations: HVLC = High Velocity Lower Crust; SDR = Seaward Dipping Reflectors; ZECM = Zone of Exhumed Continental Mantle.

**Figure 2**. **a)** Major geological features of eastern North America. Light grey contours are 1 km spaced bathymetry contours. East Coast Magnetic Anomaly (ECMA) data is after Meyer et al. (2017). Locations of early Mesozoic rift basins are marked with red shading after Klitgord et al. (1988) and Withjack et al. (2002) and references therein. Oceanic fracture zones and onshore faults (dark gray lines) are after Klitgord et al. (1988) and Hibbard et al. (2006) respectively. The transition from a volcanic to a non-volcanic margin south of Nova Scotia is marked after Deptuck and Kendell (2017). Locations of major cities are indicated as stars. The segments of the Eastern North American Margin are BCT = Baltimore Canyon Trough, LIP = Long Island Platform, GBB = Georges Bank Basin, SB = Scotian Basin. Main rift basins: C = Culpeper; CV = Connecticut Valley; F = Fundy; G = Gettysburg; O = Orpheus; T = Taylorsville. Other abbreviations: CH = Cape Hatteras; CC = cape cod; DB = Delaware Bay; GOM = Gulf of Maine; NESM = New England Seamount Chain; NJ = New Jersey; NS = Nova Scotia. **b)** Distribution of crustal building blocks and terranes (after Hibbard et al. (2006) and (2007), Hatcher et al. (2010) and Sheridan et al. (1993)). Br = Brunswick; Ca = Carolina; DD = Dunage Domain; G = Goochland; LR = Laurentian Realm; PD = Piedmont Domain; Sw = Suwannee.

**Figure 3**. Distribution of data used superimposed on bathymetry. Black and blue lines mark the locations of the present-day shoreline and 200 m isobath, respectively. Red diamonds are locations of LASE (1986) Expanded Spread Profile data. Onshore depth to base of coastal plain aquifer is from Pope et al. (2016). Bathymetry data are from Andrews et al. (2013). BOS = Boston; NY = New York; WA = Washington.

**Figure 4**. **A)** Composite multichannel seismic reflection section of pre-stack time migrated USGS profile 12 and industry data, along the strike of the ENAM. **B)** Interpretation of A. Inset shows stratigraphy color code (see table S1 for the ages of the horizons). Red circles mark locations of Moho reflectors as they appear on crossing dip-oriented reflection profiles. Red rhombuses are locations of the Moho, Top Basement and Base Post-Rift horizons based on crossing seismic refraction profiles, which are indicated by vertical dashed lines. Projections of two wells, located less than 2 km NW of the profile, are shown in the Georges Bank Basin. **C)** Map showing the profile location. AB = Atlantis Basin; DBF = Delaware Bay Fault; GBT = Georges Bank Trough; YA = Yarmouth Arch.

**Figure 5**. **A)** USGS multichannel seismic reflection profile 18 across the northern GBB continental shelf, slope and rise. **B)** Interpreted section. Inset shows stratigraphy color code (see table S1 for the ages of the horizons). **C)** Map showing the profile location. Magnetic anomaly profile is shown across the top of the section A. ECMA = East Coast Magnetic Anomaly; IYB = Inner Yarmouth Basin; OYB = Outer Yarmouth Basin; SDR = Seaward Dipping Reflectors; YA = Yarmouth Arch.

**Figure 6**. **A)** Dip-oriented section across the northern Baltimore Canyon Trough composed of reprocessed, pre-stack time migrated USGS multi-channel reflection profile 25 offshore and base

of coastal plain aquifer Digital Elevation Map and results of receiver function analysis used to mark the BPR and Moho onshore. **B)** Interpreted section. Red rhombuses are locations of the Moho, high-velocity lower crust, Top Basement, Seaward Dipping Reflectors package and top carbonate bank based on re-interpretation of wide-angle seismic results (LASE, 1986). Positions of the Base Post-Rift, the Moho west of the hinge line and the seaward limit of continental crust are after Pope et al. (2016), Li et al. (2018) and Talwani et al. (1995), respectively. See figure 3 for description of the stratigraphy. Dashed rectangle marks location of C. **C)** Uninterpreted, vertically exaggerated magnification of the part in A that show SDR. **D)** Map showing the section location. ECMA = East Coast Magnetic Anomaly; HVLC = High Velocity Lower Crust; SDR =Seaward Dipping Reflectors; SLCC = Seaward Limit of Continental Crust.

**Figure 7**. Structural map of Top Basement (in Two Way Travel Time) based on interpretations of seismic reflection and published results of seismic refraction data. Black patches mark fault heaves. Cross-hatched pattern at the GBT represents an area where interpretation is less certain. AB = Atlantis Basin; BOS = Boston; DB = Delaware Bay; DBF = Delaware Bay Fault; FB = Franklin Basin; GBB = Georges Bank Basin; GBT = Georges Bank Trough; IYB = Inner Yarmouth Basin; LIB = Long Island Basin; LIP = Long Island Platform; NaB = Nantucket Basin; NBCT = Northern Baltimore Trough; NoB = Norfolk Basin; NY = New York; NYB = New York Bight Basin; OYB = Outer Yarmouth Basin; PB = Poag Basin; SBCT = Southern Baltimore Canyon Trough; WA = Washington; YA = Yarmouth Arch.

**Figure 8**. **A)** Dip-oriented section across the southern Baltimore Canyon Trough coastal plain to continental rise based on MA-032 time-migrated multi-channel reflection profile. Magnetic anomaly profile is shown across the top of the section. Position of the Base Post-Rift west of the coastline is after Pope et al. (2016). Position of the Moho, top basement under the SDR, seaward limit of continental crust and the presence of high-velocity lower crust are interpolated based on adjacent (~13 km) refraction data (Lizarralde & Holbrook, 1997; Sheridan et al., 1993; Talwani et al., 1995). Dashed rectangles mark locations of B and C. **B)** and **C)** Uninterpreted and interpretation of magnifications of the parts in A that show SDR. Note the overlap between the positive East Coast Magnetic Anomaly and the distribution of seaward dipping reflectors. **D)** Map showing the profile location. See figure 3 for description of the stratigraphy. ECMA = East Coast Magnetic Anomaly; HVLC = High-Velocity Lower Crust; SDR = Seaward Dipping Reflectors.

**Figure 9**. Two-way travel time structural map of the Base Post-Rift. Abbreviations of names of structures underlying the BPR: AB = Atlantis Basin; GBT = Georges Bank Trough; IYB = Inner Yarmouth Basin; NoB = Norfolk Basin; OYB = Outer Yarmouth Basin; YA =Yarmouth Arch. Other abbreviations: BOS = Boston; GBB = Georges Bank Basin; GOM = Gulf of Maine; NBCT = Northern Baltimore Canyon Trough; NY = New York; SBCT = Southern Baltimore Canyon Trough; SE = Salisbury Embayment; WA = Washington.

**Figure 10**. **A)** Interpreted dip-oriented time-migrated multichannel seismic profile 288-AN-16744 across the Long Island Platform outer continental shelf, slope and rise. Magnetic anomaly profile is shown across the top of the section. See figure 3 for description of the stratigraphy. The East Coast Magnetic Anomaly and the Seaward Dipping Reflectors spatially overlap. Dashed rectangle marks location of B. **B)** Magnifications of uninterpreted and interpretation of the part in A that show diverging Seaward Dipping Reflectors. **C)** Map showing the profile location.

**Figure 11**. **A)** Interpreted dip-oriented seismic reflection section (part of USGS profile 4) across the northern Georges Bank Basin continental slope and rise. Magnetic anomaly profile is shown across the top of the section. Dashed rectangle marks location of B. Note the overlap between the East Coast Magnetic Anomaly and the distribution of Seaward Dipping Reflectors. **B)** Magnified seismic expression of the Seaward Dipping Reflectors and its interpretation. **C)** Map showing the profile location.

**Figure 12**. Magnetic anomaly map (adopted from Meyer et al., 2017). Locations of the landward pinch-outs of SDR identified on seismic reflection sections are shown as red circles. Yellow triangles mark the pinch-out location of the Base Post-Rift horizon on seismic sections that do not clearly show an SDR geometry (Strike profiles or profiles of insufficient imaging quality). Outlined green squares indicate locations of the seaward limit of the continental crust as observed on seismic refraction data (after Talwani et al., 1995). BOS = Boston; NY = New York; WA = Washington.

**Figure 13**. **A)** Distribution of data and published results used for constraining base of the crust (the Moho) depths (in two-way travel time). Moho picks (this study) are marked in red-yellow-green-blue-purple spectra. Dashed red polygon marks the area used for interpolation of the Moho depths. Legend show data sources. **B)** Time-domain structural map of the Moho interpolated from A. GBB = Georges Bank Basin; LIP = Long Island Platform; NBCT =Northern Baltimore Canyon Trough; SBCT =Southern Baltimore Canyon Trough.

**Figure 14**. **A)** Thickness (in two-way travel time) of the interval between Base Post-Rift and the Moho and B) Gradient of Base Post-Rift-to-Moho thickness expressed as dip angle in pseudo-degrees, assuming the 1 millisecond equals 1 meter. Contours on both maps represent Base Post-Rift-to-Moho thickness. Red line is the hinge line as defined by the location of increasing Base Post-Rift-to-Moho thickness. Histogram below color scales represent the relative abundance of values. BOS = Boston; BCT = Baltimore Canyon Trough; CH = Cape Hatteras; DB = Delaware Bay; GB = Georges Bank; GOM = Gulf of Maine; NJ = New Jersey; NY = New York, WA = Washington.

**Figure 15**. Two-way travel-time thickness of the post-rift Jurassic sequence. Histogram below color scales represent the relative abundance of a specific values. BOS = Boston; GBB = Georges Bank Basin; NBCT = Northern Baltimore Canyon Trough; NY = New York, WA-Washington.

**Figure 16**. Post-rift Jurassic thickness (in two-way travel time) against the thickness of the Base Post-Rift to Moho interval at Georges Bank Basin.

**Figure 17**. Magnetic anomaly map (Meyer et al., 2017) overlaid by key results, including location of the hinge line, locations of the SDR landward pinch-out (red circles) and Jurassic thickness contours (colored according to a thickness spectrum). Green squares mark locations of the seaward limit of the continental crust as observed on seismic refraction data (after Talwani et al., 1995). BOS = Boston, GBB = Georges Bank Basin; LIP = Long Island Platform; NBCT = Northern Baltimore Canyon Trough; NY = New York; SBCT = Southern Baltimore Canyon Trough; WA = Washington.

**Figure 18**. Schematic model for the formation of ENAM along the BCT (a), GBB (b) and Central and Northern Nova Scotia (c) segments (not to scale). Where Meguma terrane is present, it focused the pre-magmatic extensional strain. Strain had localized oceanward when rifting at the BCT and the GBB turned magmatic. General pre-rift crustal configuration of ENAM follows Hibbard et al. (2006) and Hatcher et al. (2010). Specific additions include the BCT crustal composition (Sheridan et al., 1993), the extension of Laurentia under the peri-Gondwanan terranes (Cook & Vasudevan, 2006; Pratt et al., 1988; Marzen et al., 2019), the nature of the Gondwanan crust (Villeneuve, 2005; Le Roy & Pique, 2001), the structural relations between Avalon and Meguma terranes (Hutchinson et al., 1988; Keen et al., 1991; Pe Piper & Jansa, 1999), the proto-oceanic stage structure of the BCT (Lizerralde & Holbrook, 1997; LASE, 1986; Labails et al., 2009; Shuck et al., 2019; Biari et al., 2017), GBB (Dehler, 2012) and Central and Northern Nova Scotia (Maillard et al., 2006; Klingelhoefer et al., 2016; Wu et al., 2006) segments, the role of the Alleghenian suture as a magma conduit during the emplacement of ECMIP (McBribe & Nelson, 1988) and the possible existence of a Rheic slab under Laurentia (Whalen et al., 2015; Van Staal et al., 2009)

**Table 1.** *Seismic Horizons and Their Corresponding Ages*

| Horizon | Geological Period | Age (Ma) [a] |
|---|---|---|
| T1 | Top Oligocene | 23 |
| UK | Top Cretaceous | 66 |
| MK | Middle Cenomanian | ~97 |
| LK | Top Barremian | 126 |
| UJ | Top Tithonian | 145 |
| MJ | Top Callovian (?) | 164? |
| BPR | Hettangian (?) -early Aalenian (?) | 201-174 |
| Top Basement | Paleozoic | >252 |
| Moho | NA | |

*Note.* [a]Walker et al. (2018)

**Figure 1.**

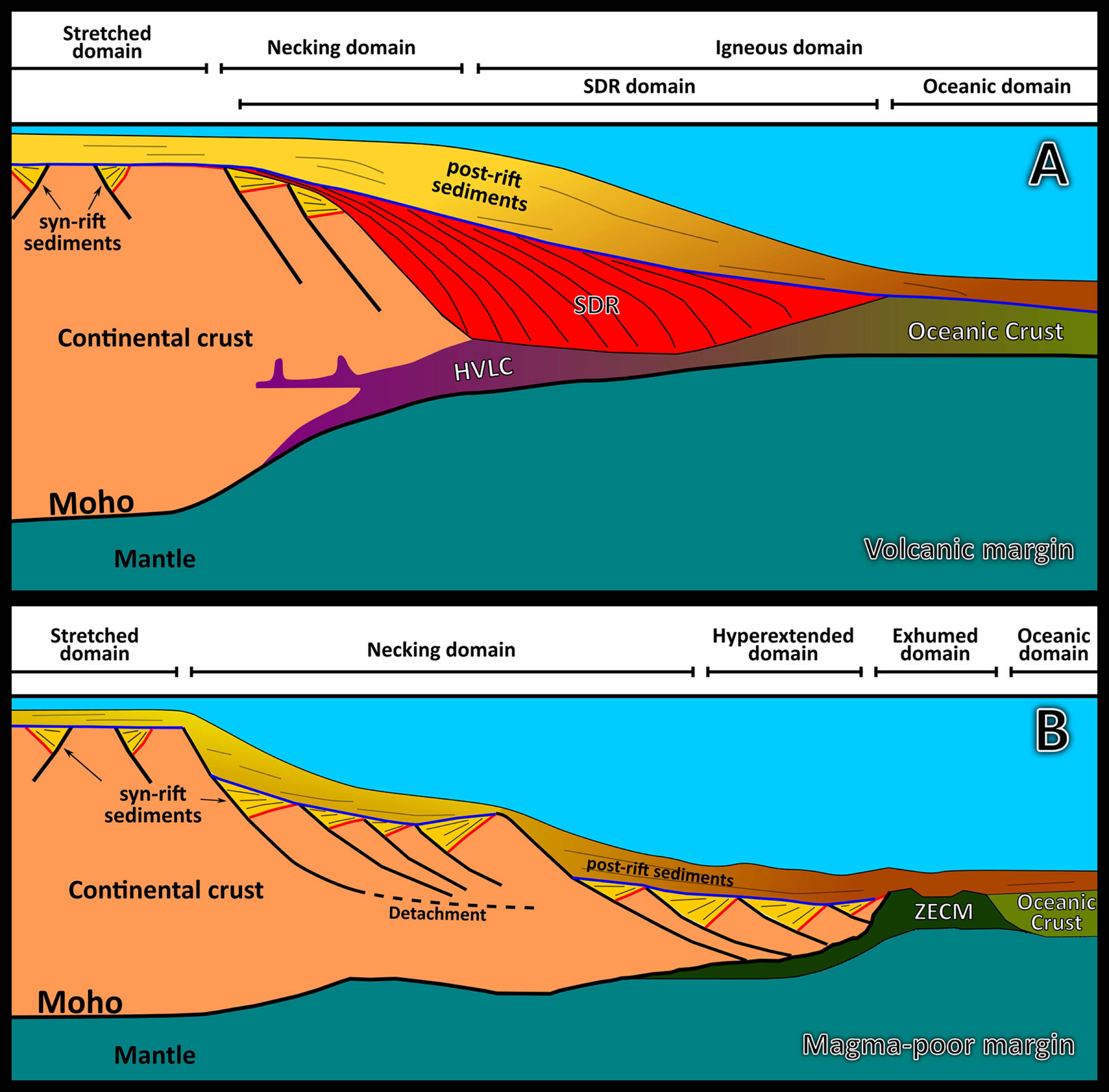

Stretched domain
Necking domain
Igneous domain
SDR domain
Oceanic domain
A
post-rift sediments
syn-rift sediments
SDR
Oceanic Crust
Continental crust
HVLC
Moho
Mantle
Volcanic margin
Stretched domain
Necking domain
Hyperextended domain
Exhumed domain
Oceanic domain
B
syn-rift sediments
Continental crust
Detachment
post-rift sediments
ZECM
Oceanic Crust
Moho
Mantle
Magma-poor margin

**Figure 2.**

N
80°W
65°W
55°W
75°W
40°N
35°N
DD
Ganderia
Meguma
Rift Basins
Avalonia
LR
ECMA
G
PD
(b)
Sw
Ca
Br
70°W
65°W
60°W
45°N
O
F
NS
SB
GOM
BOS
GC
GBB
Volcanic
non-Volcanic
Appalachian Mountains
CV
NY
N
LIP
G
NJ
NESM
ECMA
WA
C
SE
DB
BCT
(a)
T
CH
Normal fault
Thrust/strike-slip fault
Edge of Coastal Plain
200m isobath
Axis of ECMA
Fracture Zone
Rift Basin
Magnetic Anomaly [nT]
300
200
100
0
80°W
75°W
70°W
65°W
40°N
35°N

**Figure 3.**

40°N
45°N
80°W
35°N
75°W
70°W
WA
NY
BOS
N
Fig. 4
Fig. 6
Fig. 8
Fig. 10
Fig. 5
Fig. 11
Depth [km]
1
2
3
4
5
6
0
200 km
Seismic reflection line
Published seismic reflection line
Reprocessed PSTM line
Published reciever function data
Borehole
Published refraction data
Edge of coastal plain post-rift deposits
Axis of ECMA
Figure location
Depth to base of coastal plain aquifer data

**Figure 4.**


SW
Baltimore Canyon Trough
Long Island Platform
Georges Bank Basin
NE
EDGE MA-801
EDGE MA-803
LASE
133-1
975-1
TWT [sec]
C
A
0
100
200 km
B
Seafloor
T1
UK
MK
LK
UJ
MJ
BPR
TB
Moho from reflection data
TB/BPR/Moho from refraction data
DBF
AB
F4
F3
F2
GBT
YA

**Figure 5.**


NW
SE
Magnetic Anomaly [nT]
200
100
0
-100
ECMA
TWT [sec]
A
Intra crustal reflections
IYB
YA
OYB
?
SDR
Moho
0
50km
B
C
Seafloor
UK
MK
UJ
MJ
BPR
TB
Moho

**Figure 6.**

NW
SE
Magnetic Anomaly [nT]
200
100
0
-100
ECMA
TWT [sec]
0
1
2
3
4
5
6
7
8
9
10
11
12
6
7
C
0
10 km
A
Fall line
Coastline
Hinge line
SLCC
D
Continental Crust
Carbonate bank
SDR
HVLC
Moho
0
50 km
B

**Figure 7.**

WA
NY
BOS
SE
DB
NJ
NYB
LIB
NaB
NoB
Gulf of Maine
Avalonia
Meguma
LeHave Platform
DBF
LIP
NBCT
SBCT
CH
GBB
N
TWT [sec]
0
1
2
3
4
5
6
7
8
Coastline
Edge of Coastal plain
-200 isobath
Axis of the ECMA
Landward extent of Meguma
0
200km
40°N
35°N
75°W
70°W
PB
FB
AB
IYB
YA
OYB
F1
F2
F3
F4
GBT
Fig. 5

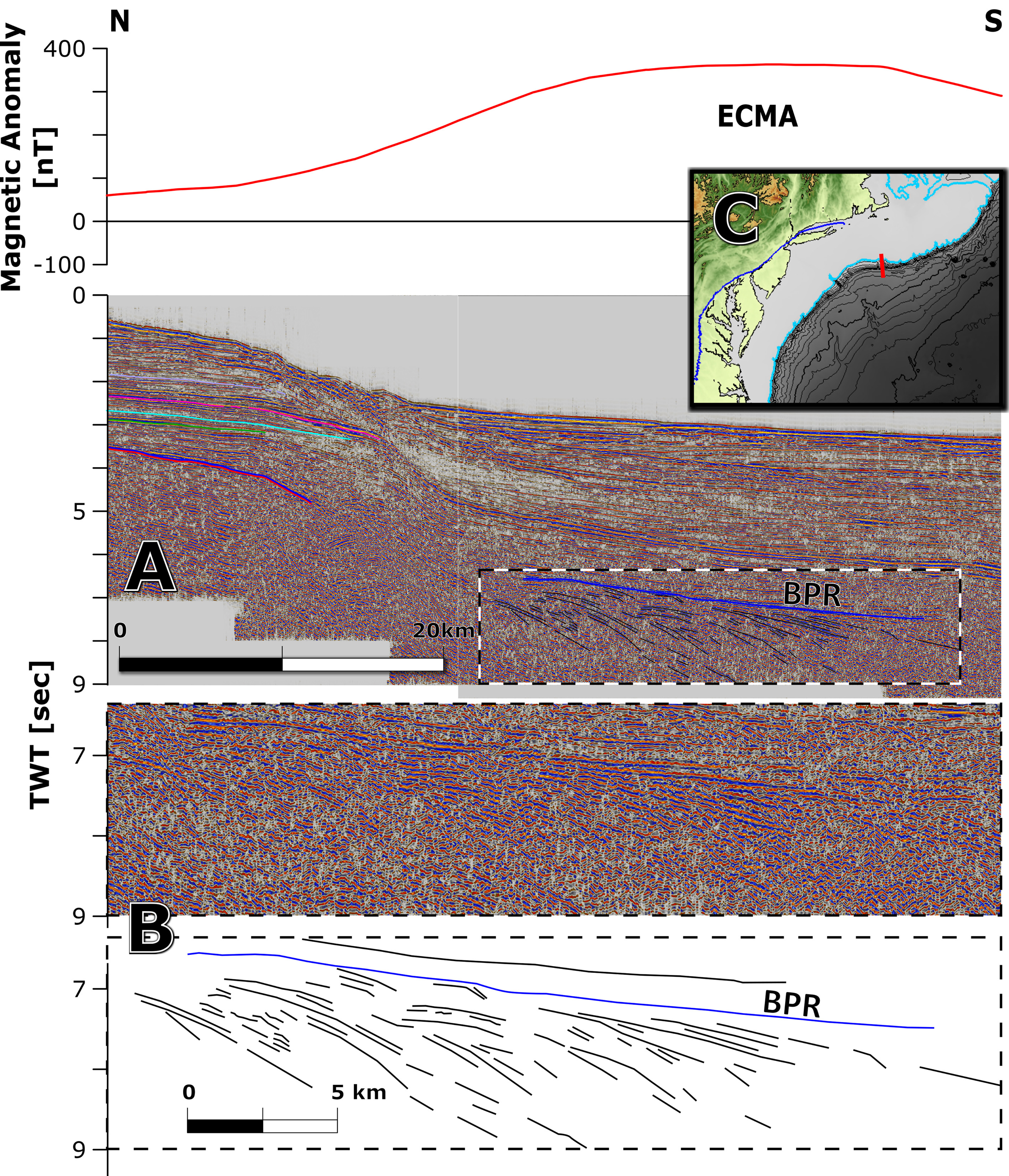
**Figure 8.**

W
E
Magnetic Anomaly [nT]
300
200
100
0
-100
Coastline
Hinge line
ECMA
SLCC
TWT [sec]
0
1
2
3
4
5
6
7
8
9
10
11
12
Continental Crust
SDR
B
?
HVLC
C
0
100 km
Moho
A
D
3
4
5
6
7
B
BPR
Top Basement
7
8
9
C
BPR

**Figure 9.**

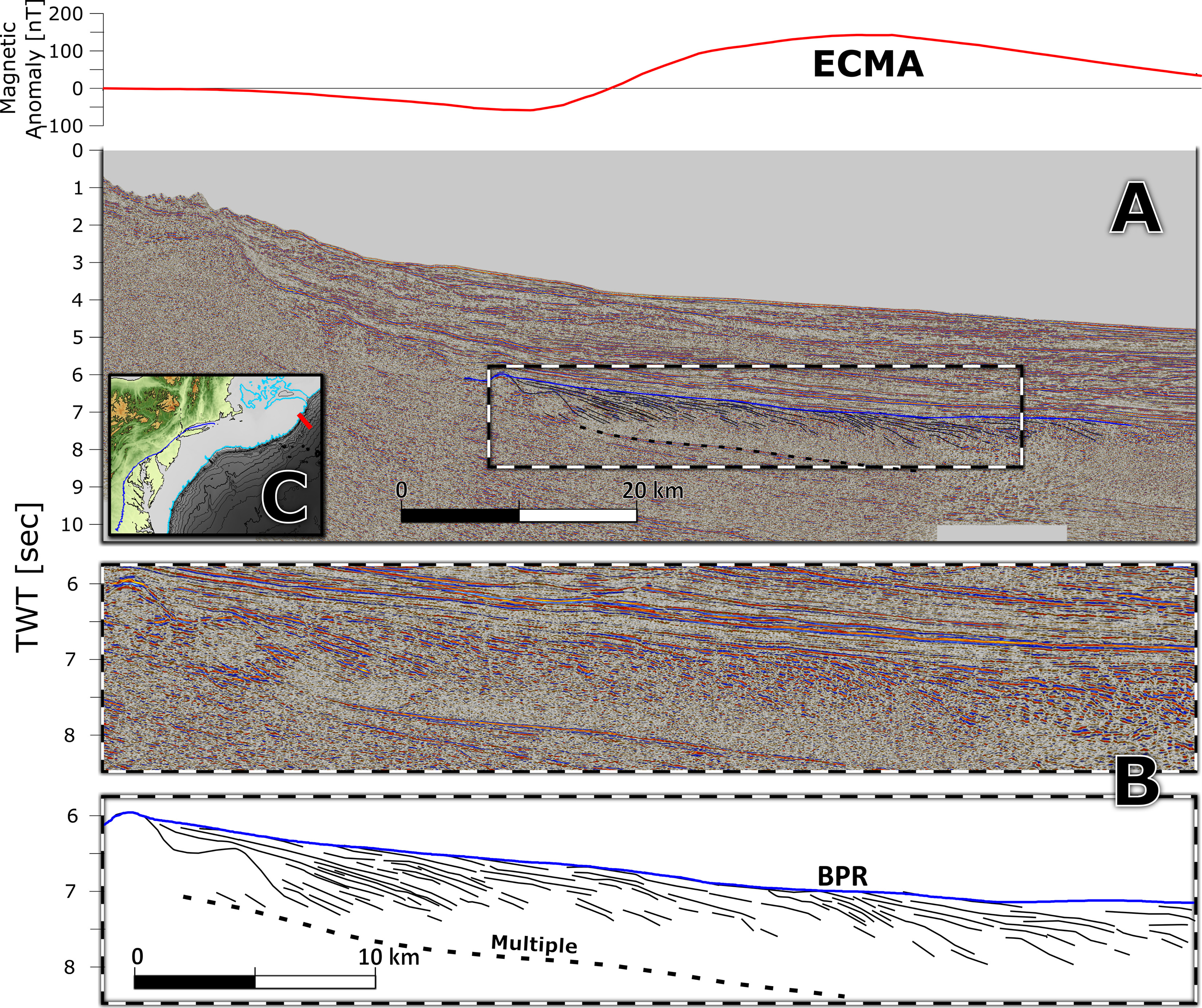
WA
40°N
45°N
NY
BOS
SE
GOM
35°N
NoB
LIP
1000
2000
3000
4000
5000
6000
NBCT
SBCT
AB
IYB
YA
OYB
GBT
GBB
TWT [sec]
6 5 4 3 2 1
Coastline
Edge of Coastal plain
-200 isobath
Axis of the ECMA
0
100 km
75°W
70°W

**Figure 10.**



N
S
Magnetic Anomaly [nT]
400
0
-100
ECMA
C
0
5
9
A
0
20km
BPR
TWT [sec]
7
9
B
7
BPR
0
5 km
9

**Figure 11.**



Magnetic Anomaly [nT]
200
100
0
100
ECMA
A
B
C
0
20 km
10 km
TWT [sec]
BPR
Multiple

**Figure 12.**

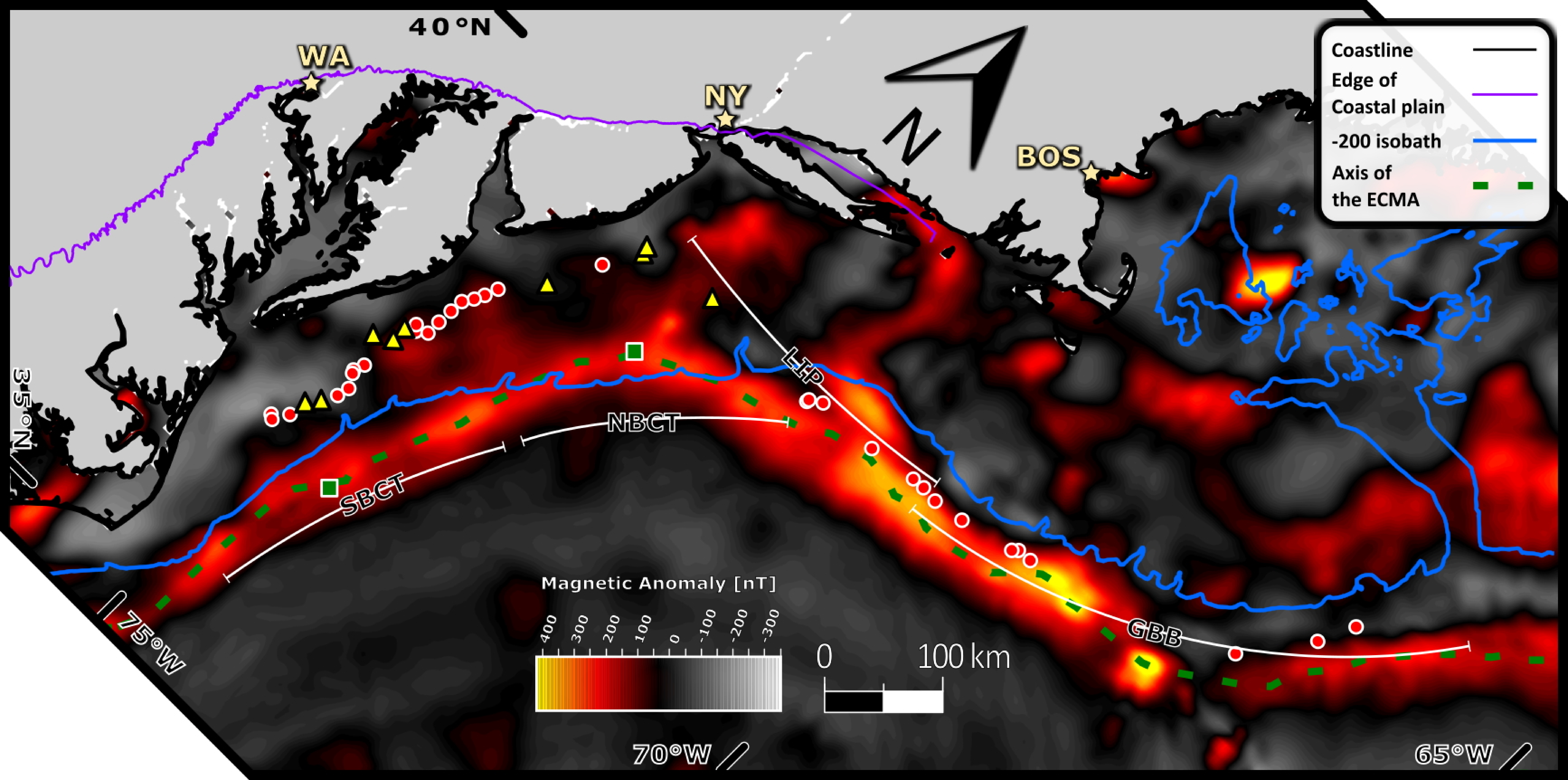

40°N
WA
NY
BOS
N
Coastline
Edge of Coastal plain
-200 isobath
Axis of the ECMA
35°N
LIP
NBCT
SBCT
GBB
75°W
70°W
65°W
Magnetic Anomaly [nT]
400 300 200 100 0 -100 -200 -300
0
100 km

**Figure 13.**

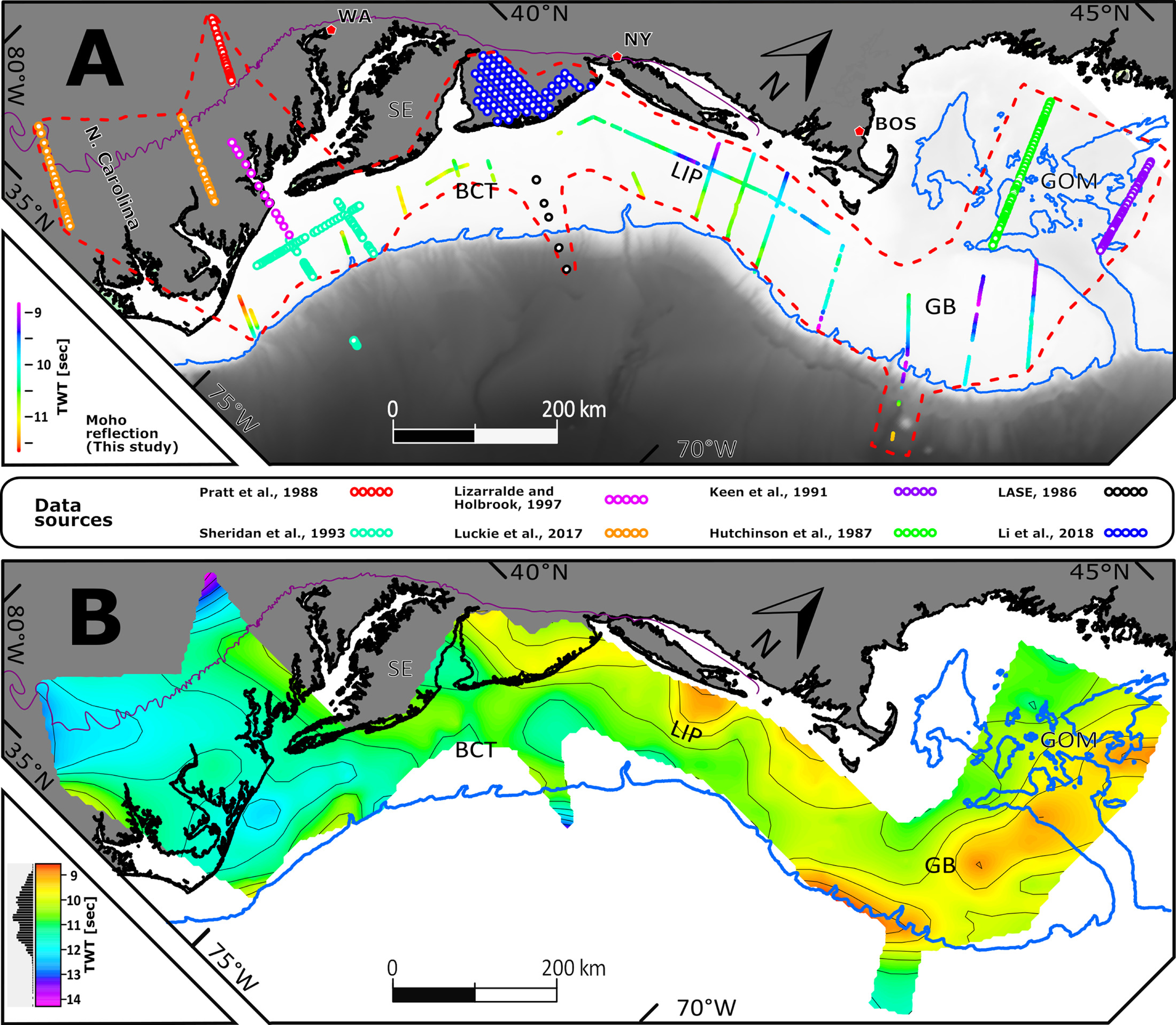

A
WA
NY
BOS
40°N
45°N
80°W
35°N
75°W
70°W
SE
N. Carolina
BCT
LIP
GOM
GB
N
0
200 km
TWT [sec]
-9
10
11
Moho reflection (This study)
Data sources
Pratt et al., 1988
Lizarralde and Holbrook, 1997
Keen et al., 1991
LASE, 1986
Sheridan et al., 1993
Luckie et al., 2017
Hutchinson et al., 1987
Li et al., 2018
B
9
10
11
12
13
14

**Figure 14.**

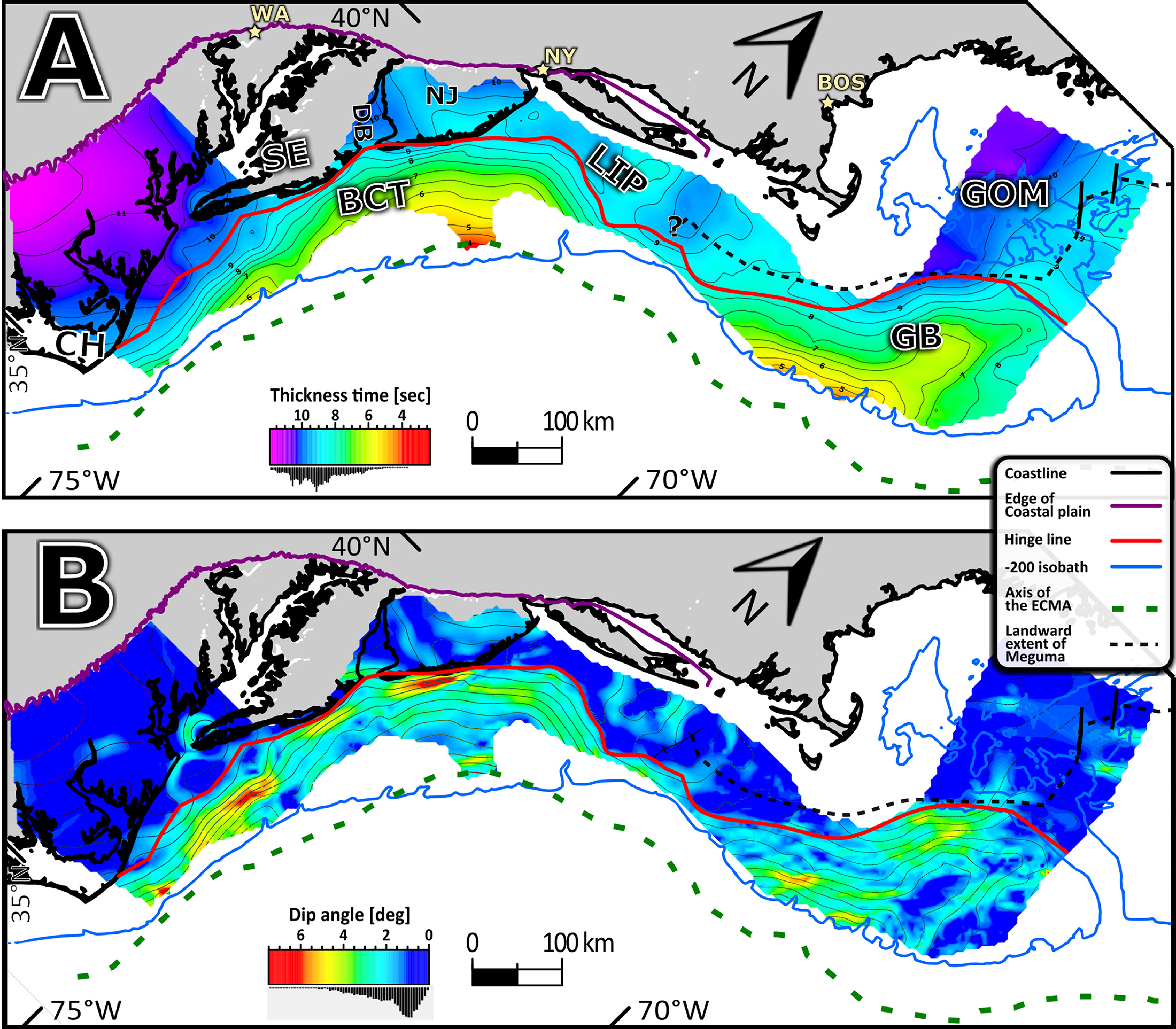
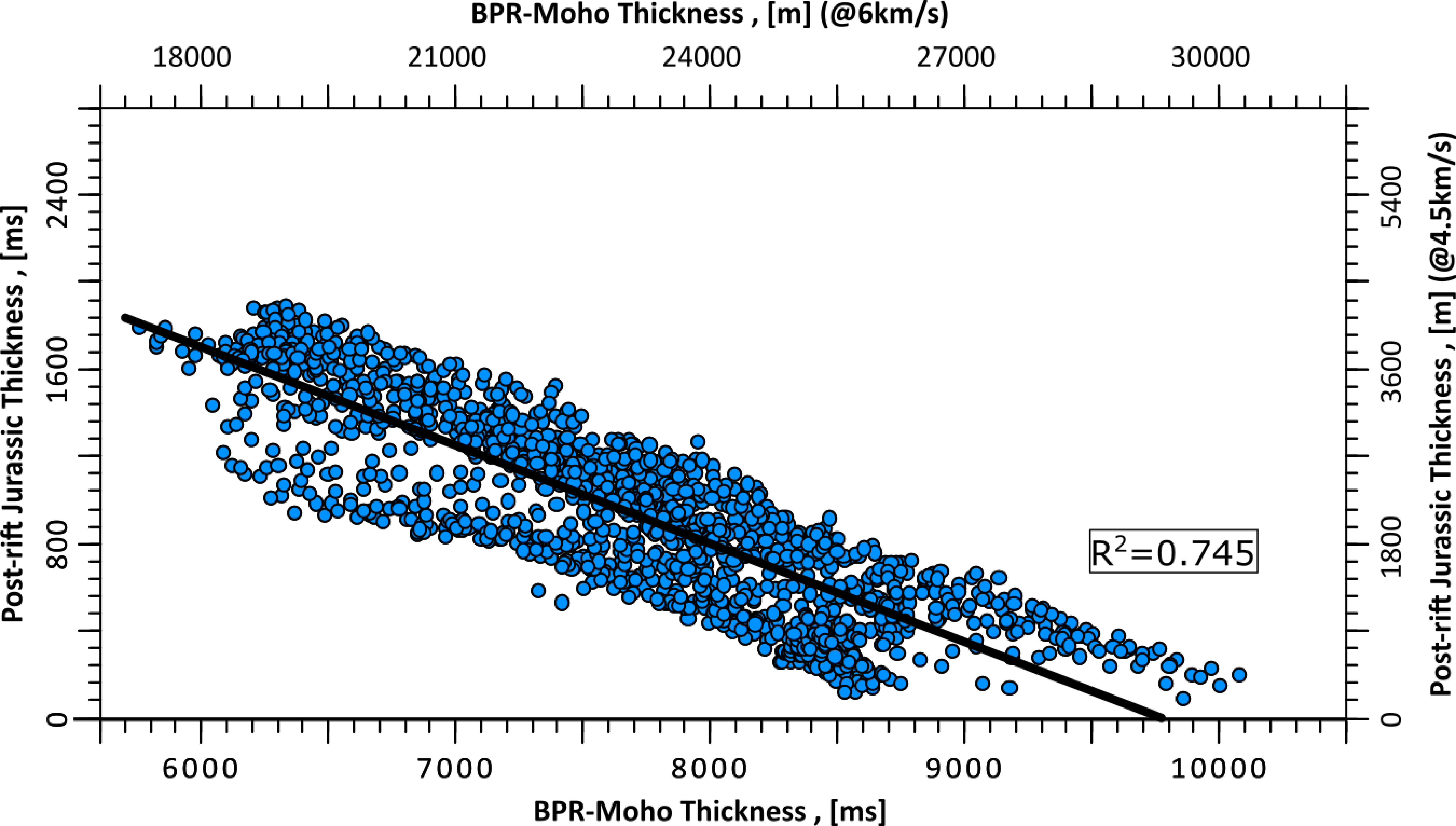

A
WA
40°N
NY
BOS
N
NJ
DB
SE
LIP
BCT
GOM
?
CH
GB
35°N
Thickness time [sec]
10
8
6
4
0
100 km
75°W
70°W
B
40°N
N
35°N
Dip angle [deg]
6
4
2
0
0
100 km
75°W
70°W
Coastline
Edge of Coastal plain
Hinge line
-200 isobath
Axis of the ECMA
Landward extent of Meguma

**Figure 15.**

40°N
WA
NY
BOS
N
Coastline
Edge of
Coastal plain
Hinge line
-200 isobath
Axis of
the ECMA
LIP
NBCT
SBCT
GBB
35°N
Thickness time [ms]
3000
2000
1000
0
0
100 km
75°W
70°W

**Figure 16.**



BPR-Moho Thickness , [m] (@6km/s)
18000
21000
24000
27000
30000
Post-rift Jurassic Thickness , [ms]
2400
1600
800
0
Post-rift Jurassic Thickness , [m] (@4.5km/s)
5400
3600
1800
0
$R^2$=0.745
6000
7000
8000
9000
10000
BPR-Moho Thickness , [ms]

**Figure 17.**

40°N
WA
NY
BOS
N
Coastline
-200 isobath
Edge of Coastal plain
Axis of the ECMA
Hinge line
Landward extent of Meguma
?
LIP
NBCT
SBCT
GBB
0
1000
2000
3000
35°N
75°W
70°W
65°W
Jurassic Travel-time Thickness [ms]
3500 3000 2500 2000 1500 1000 500 0
Magnetic Anomaly [nT]
400 300 200 100 0 -100 -200 -300
0 100 km

**Figure 18.**

Baltimore Canyon Trough
Georges Bank Basin
Central and Northern Nova-Scotia
(Magma-poor)
Late Alleghenian Orogeny >280 Ma
volcanism ~230-~195 Ma
Emplacement of ECMIP ~195 Ma
Proto-Oceanic stage ~180 Ma
Collision
Laurentia
Gondwana
Piedmont domain
Goochland
Avalon
Dunnage domain
Gander
Meguma
Rheic Slab?
Rehic Slab?
a.1
b.1
c.1
Normal Faulting
Normal Faulting + Crustal Thinning
a.2
b.2
c.2
Magma-Assisted Rifting
Hyper-extension
a.3
b.3
c.3
Accretion of Proto-Oceanic Crust
Accretion of Proto-Oceanic Crust ?
Hyper-extension + Exhumation
a.4
b.4
c.4
Continental crust
Continental mantle lithosphere
Asthenosphere
Peri-Gondwanan accreted terranes
Proto-Oceanic crust
High Velocity Lower Crust
Alleghenian Suture
Subaerially erupted basalt flows
Rift basin
Intrusions and extrusions
Zone of extension

*Journal of Geophysical Research: Solid Earth*

Supporting Information for

**The role of pre-magmatic rifting in shaping a volcanic continental margin: An example from the Eastern North American Margin**

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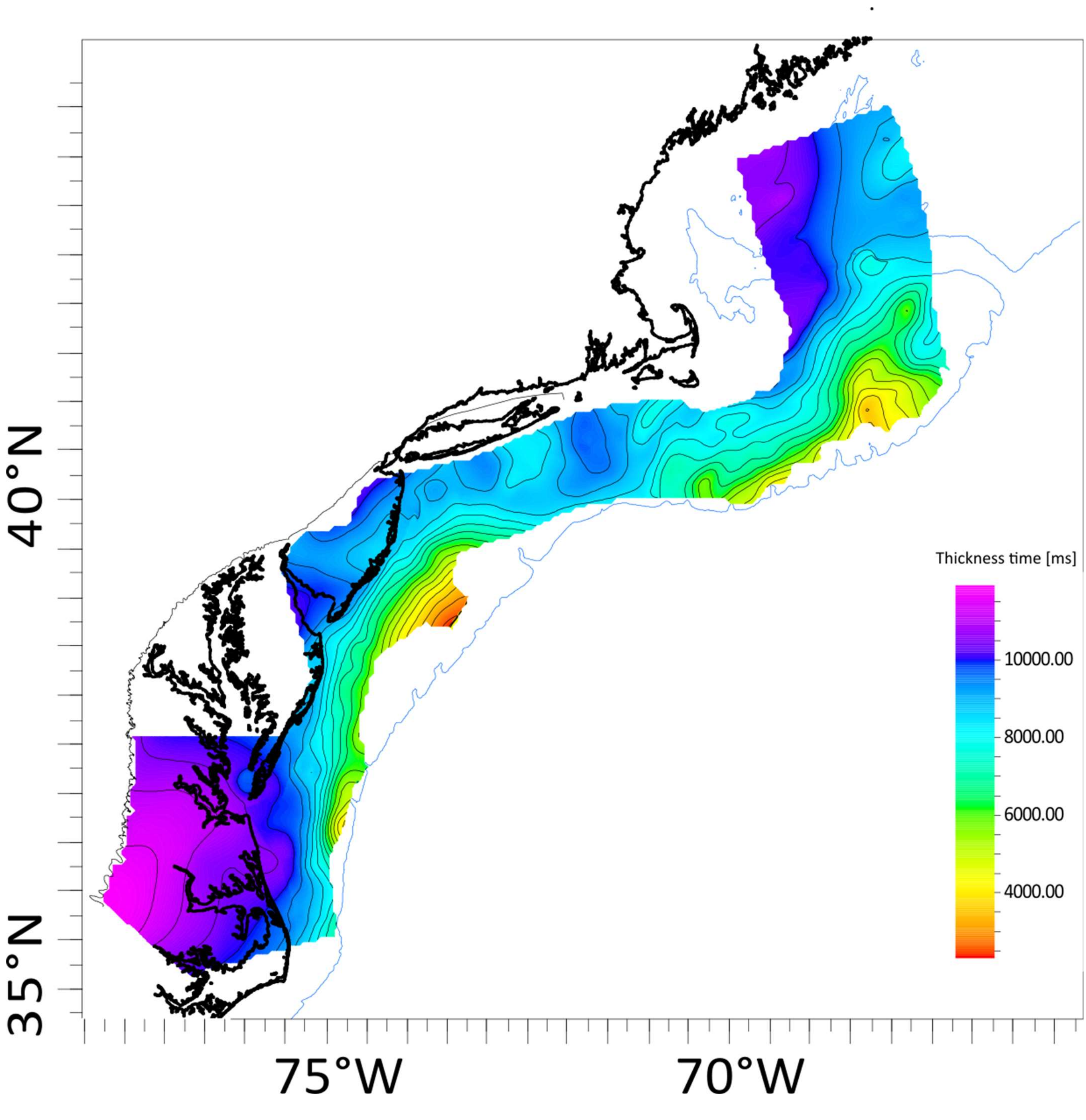


**Figure S1.** Crustal thickness along ENAM (In TWT) as represented by the interval bounded by the Top Basement and Moho. Black and blue lines mark the locations of the present-day shoreline and 200 m isobath, respectively.

| Survey | # lines | Length [km] | Domain | Acquisition year | Source | Receivers | Record length [ms] | Final processing step | Reference |
|---|---|---|---|---|---|---|---|---|---|
| 1** | 12 | 570 | TWT | 1977 | 980 cubic inches | NA | 6900 | Migration | [*Triezenberg et al.*, 2016] |
| 388-a | 29 | 1441 | TWT | 1981 | NA | NA | 6900-9900 | Migration | [*Triezenberg et al.*, 2016] |
| 80PMA- | 15 | 859 | TWT | 1980 | 14 X 2,682 cubic inches | 48 | 6600-9920 | Migration | [*Triezenberg et al.*, 2016] |
| 81- | 129 | 3823 | TWT | 1981 | 25 X 2220 cubic inches Airgun | 96 | 5600-7925 | Migration | [*Triezenberg et al.*, 2016] |
| 88 GBB | 16 | 414 | TWT | 1988 | NA | NA | 8000 | Migration | [*Triezenberg et al.*, 2016] |
| a-E01-75-Mig-123-251 | 141 | 8853 | TWT | 1975 | 10 X 1,700 cubic inches Airgun | NA | 6900-9000 | Migration | [*Triezenberg et al.*, 2016] |
| d- | 172 | 12011 | TWT | 1975 | 18 X 1,700 cubic inches | NA | 6824-8224 | Migration | [*Triezenberg et al.*, 2016] |
| Dan Lizarralde LsP | 1 | 142 | TWT | 2009 | 45 in.^3/105 in.^3, generator–injector (GI) air gun | 48 | 4000 | Migration | [*Siegel* et al., 2012] |
| de- | 92 | 5348 | TWT | 1975 | 18 X 1,700 cubic inches | NA | 7000-8500 | Migration | [*Triezenberg et al.*, 2016] |
| ma- | 1 | 23 | TWT | 1977 | 1,080 cubic inches Airgun | NA | 8000 | Migration | [*Triezenberg et al.*, 2016] |
| mmg-15 | 1 | 25 | TWT | 1976 | 5,400 cubic inches Airgun | 96 | 5900 | Migration | [*Triezenberg et al.*, 2016] |
| npr | 29 | 1166 | TWT | 1978 | 7 X 1,341 cubic inches | 48 | 6000 | Migration | [*Triezenberg et al.*, 2016] |
| PR-82 | 84 | 1485 | TWT | 1982 | 14 X 3,050 cubic inches | 96 | 7900-8000 | Migration | [*Triezenberg et al.*, 2016] |
| PRI | 5 | 254 | TWT | 1979 | 1,940 cubic inches | 96 | 6744-7900 | Migration | [*Triezenberg et al.*, 2016] |
| Reprocessed USGS | 21 | 4187 | TWT | 1973-1978 | 4 to 23 airguns with a total volume of 1200 to 2160 cubic inches | 24-48 | 7000-15000 | Pre-stack time migration | [Fortin et al., 2018] |
| Southern BCT wide grid | 80 | 5522 | TWT | 1976 | 18 X 1,700 cubic inches | NA | 6900-10000 | Migration | [*Triezenberg et al.*, 2016] |
| sx- | 20 | 1102 | TWT | 1988 | Airgun | NA | 6970-8700 | Migration | [*Triezenberg et al.*, 2016] |
| TX | 1 | 43 | TWT | 1976 | 5 X 660 cubic inches Airgun | NA | 6800 | Migration | [*Triezenberg et al.*, 2016] |

| | | | | | | | | | |
|---|---|---|---|---|---|---|---|---|---|
| USGS CDP | 50 | 8657 | TWT | 1973-1978 | 4 to 23 airguns with a total volume of 1200 to 2160 cubic inches | 24-48 | 3500-15000 | | [*Triezenberg et al.*, 2016] |
| Southern BCT tight grid | 43 | 1270 | TWT | 1982 | 5,600 cubic inches | NA | 6800 | Migration | [*Triezenberg et al.*, 2016] |
| XPR-78 | 27 | 1060 | TWT | 1978 | 7 X 1,341 cubic inches | 48 | 6000-7000 | Migration | [*Triezenberg et al.*, 2016] |
| JGM | 75 | 2774 | TWT | 1984 | 18 X 3,000 cubic inches | 120 | 5000-7000 | Migration | [*Triezenberg et al.*, 2016] |
| GB-75 | 44 | 1261 | TWT | 1975 | 1,200 cubic inches | 48 | 6000 | Migration | [*Triezenberg et al.*, 2016] |
| na GBB | 65 | 1599 | TWT | 1983 | 4,000 cubic inches | NA | 6800 | Migration | [*Triezenberg et al.*, 2016] |

**Table S1.** Seismic Reflection Surveys Used for Interpretation. Data published by Triezenberg et al. [2016] are available at the USGS National Archive of Marine Seismic Surveys: https://walrus.wr.usgs.gov/namss/search/

| Survey | Line | Region | Type | Vertical dimension | Horizons | Conversion velocity | Reference |
| --- | --- | --- | --- | --- | --- | --- | --- |
| -- | 88-2 | Gulf of Maine | Deep reflection | TWT | Base post-rift, Moho | NA | [*Keen et al.*, 1991] |
| -- | USGS 1A | Gulf of Maine | Deep reflection | TWT | Base post-rift, Moho | NA | [*Hutchinson et al.*, 1988; *Hutchinson et al.*, 1987] |
| LASE | 6 | N. Baltimore Canyon Trough | Seismic refraction/wide-angle reflection | TWT | Base of extended continental crust, Moho, Base post-rift (reinterpreted) | NA | [*LASE*, 1986] |
| -- | I-64 | Virginia Piedmont | Deep reflection | TWT | Top Basement, Moho | NA | [*Pratt et al.*, 1988] |
| EDGE | MA-801 (offshore), MA-802, MA-803 | S. Baltimore Canyon Trough | seismic refraction/wide-angle reflection | TWT | Base post-rift, Base SDRs, Moho | NA | [*Sheridan et al.*, 1993] |
| EDGE | MA-801 (onshore) | S. Baltimore Canyon Trough | Seismic refraction/wide-angle reflection | Depth | Moho | 6.3 [km/s] | [*Lizarralde and Holbrook*, 1997] |
| -- | -- | New England (Only the coastal plains of New Jersey and New York were used in the current study) | Teleseismic receiver functions | Depth | Moho | 6.3 [km/s] | [*Li et al.*, 2018] |

**Table S2.** Published Deep Seismic Results Incorporated in the Analysis.

| Well | Region | Total depth [m] | Checkshots | Vp log | Density log | Seismic-well tie procedure | Paleontological report reference | Remarks |
|---|---|---|---|---|---|---|---|---|
| **COST G-1** | GBB | 4898.4 | NA | V | V | * | [Poag, 1991] | *Time-Depth-Relationships are taken and digitized from Taylor and Anderson [1982]. A synthetic seismogram was constructed to evaluate the tie to the seismic data |
| **COST G-2** | GBB | 6667.2 | NA | V | V | * | [Poag, 1991] | |
| **Exxon 133-1** | GBB | 4303.2 | V | V | V | ISWT | [*Edson et al.*, 2000a] | |
| **Conoco 145-1** | GBB | 4419.6 | V | V | V | ISWT | [Poppe et al., 1992] | |
| **Tenneco 187-1** | GBB | 5525.1 | V | V | V | SC | [*Edson et al.*, 2000d] | |
| **Mobil 273-1** | GBB | 4748.8 | V | V | V | ISWT | [*Edson et al.*, 2000b] | |
| **Mobil 312-1** | GBB | 6096 | V | V | V | ISWT | [Poppe and Poag, 1993] | |
| **Shell 357-1** | GBB | 5921.3 | V | V | V | ISWT | [*Edson et al.*, 2000c] | Shallow (<3670m) checkshots data is taken from Mobil 312-1 |
| **Shell 410-1** | GBB | 4745.1 | V | V | V | ISWT | [Poppe and Poag, 1993] | |
| **Exxon 975-1** | GBB | 4451.6 | V | V | V | ISWT | [Poppe and Poag, 1993] | |
| **COST B-2** | BCT | 4838.8 | NA | V | V | * | [*Poag*, 1985] | *Time-Depth-Relationships are taken and digitized from *Scholle* [1977]. A synthetic seismogram was constructed to evaluate the tie to the seismic data |
| **COST B-3** | BCT | 4807.2 | NA | V | V | * | [*Poag*, 1985] | *Time-Depth-Relationships are taken and digitized from [*Scholle*, 1980]. A synthetic seismogram was constructed to evaluate the tie to the seismic data |
| **Mobil 17-2** | BCT | 4115.0 | | | | | [*Edelman et al.,* 1979] | |
| **Murphy 106-1** | BCT | 5610.0 | V | V | V | SC | [*Adinolfi*, 1986] | |
| **Shell 272-1** | BCT | 4115.0 | NA | V | V | * | [*Poag*, 1985] | *Time-Depth-Relationships are taken and digitized from the neighboring Shell 273-1 well |
| **Shell 273-1** | BCT | 4826.0 | V | V | V | ISWT | [*Steinkraus*, 1979] | |
| **Shell 372-1** | BCT | 3515.4 | V | V | V | ISWT | [*Edson*, 1987a] | |
| **Tenneco 495-1** | BCT | 5547.0 | V | V | V | ISWT | [*International_Biostratigraphers_Incorporated*, 1979b] | |
| **Exxon 500-1** | BCT | 3316.0 | V | V | V | ISWT | [*Crane*, 1979c] | |
| **Mobil 544-1A** | BCT | 4806.7 | V | V | V | ISWT | [*Gauger*, 1979] | |
| **Shell 586-1** | BCT | 4828.0 | V | V | V | ISWT | [*Edson*, 1986] | |

| | | | | | | | | |
|---|---|---|---|---|---|---|---|---|
| **Shell 587-1** | BCT | 4420.0 | NA | V | V | SC | [*Edson*, 1987b] | |
| **Conoco 590-1** | BCT | 3607.3 | V | V | V | ISWT | [*International_Biostratigraphers_Incorporated*, 1978a] | |
| **Texaco 598-1** | BCT | 4884.0 | V | V | V | ISWT | [*Kobelski*, 1987] | |
| **Exxon 599-1** | BCT | 5199.3 | NA | V | V | * | [*Cousminer et al.*, 1986] | *Time-Depth-Relationships are taken and digitized from the neighboring Texaco 598-1 well |
| **Shell 632-1** | BCT | 4241.6 | V | V | V | CS | [*Picou*, 1978] | |
| **Texaco 642-1** | BCT | 5377.0 | NA | V | V | * | [*Amato and Bielak*, 1990] | *Time-Depth-Relationships are taken and digitized from the neighboring Tenneco 642-2 well |
| **Tenneco 642-2** | BCT | 5554.0 | V | V | V | ISWT | [*Bielak*, 1986] | |
| **Tenneco 642-3** | BCT | 4785.2 | V | V | V | ISWT | NA | This well had no available paleontological report. It was used only for calibration of seismic-well tie |
| **Exxon 684-2** | BCT | 5096.9 | V | V | V | ISWT | [*Crane*, 1979b] | |
| **Exxon 684-1** | BCT | 5243.0 | V | V | V | ISWT | [*Crane*, 1979a] | |
| **Gulf 718-1** | BCT | 3882.5 | NA | V | V | * | [*Poppe et al.*, 1990] | *Time-Depth-Relationships are taken and digitized from the neighboring Shell 632-1 well |
| **Exxon 728-1** | BCT | 4609.2 | NA | V | V | * | [*Stough*, 1981] | *Time-Depth-Relationships are taken and digitized from the neighboring Exxon 684-2 well |
| **Exxon 816-1** | BCT | 5386.3 | V | V | V | CS | [*Crane*, 1981] | |
| **Homco 855-1** | BCT | 5305.0 | NA | V | V | * | [*International_Biostratigraphers_Incorporated*, 1979a] | *Time-Depth-Relationships are taken and digitized from the neighboring Gulf 857-1 well |
| **Gulf 857-1** | BCT | 5320.0 | V | V | V | ISWT | [*Bifano*, 1978] | |
| **Exxon 902-1** | BCT | 4802.0 | V | V | V | ISWT | [*Crane*, 1979d] | |
| **Shell 93-1** | BCT | 5407.0 | V | V | V | ISWT | [*Amato*, 1987] | |
| **Homco 676-1** | BCT | 3781.0 | NA | V | V | * | [*International_Biostratigraphers_Incorporated*, 1978b] | *Time-Depth-Relationships are taken and digitized from the neighboring Shell 632-1 well |

**Table S3.** Wells Used for Stratigraphic Division.