CHAPTER 2

REGIONAL TECTONIC CONTEXT

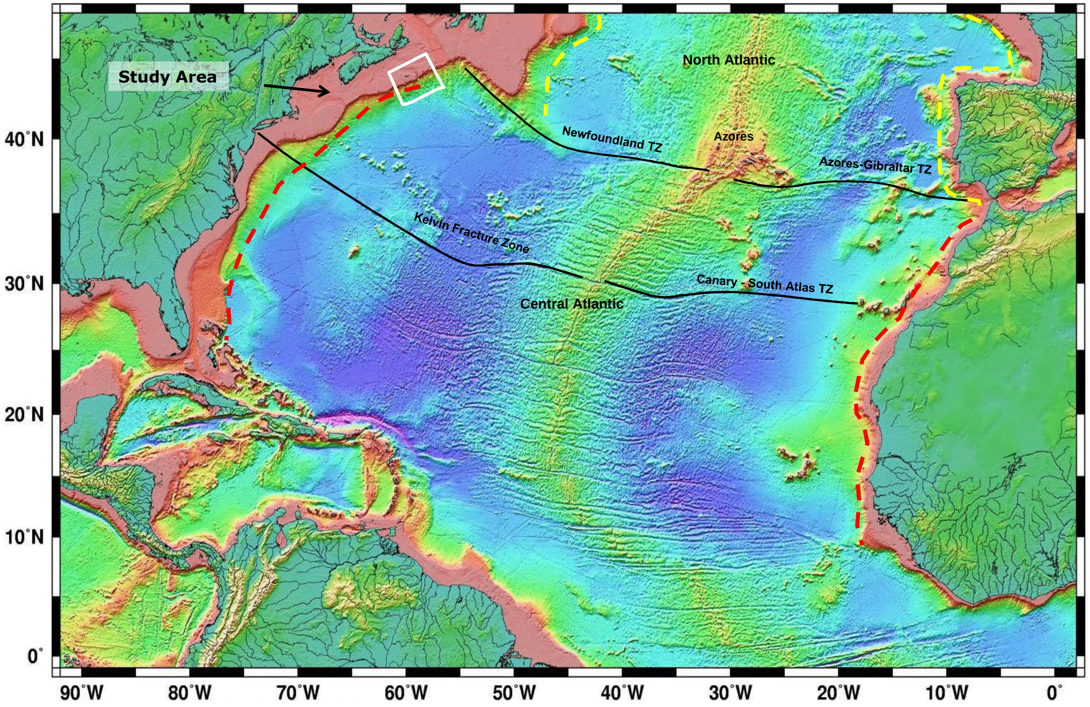


Figure 1: Bathymetric prediction of Central and North Atlantic (Sandwell and Smith, 1996) showing the location of the study area and distribution of different continental margin types (Boillot & Coulon, 1998; Louden et al. 2010). The magma rich (volcanic) margins of the Central and North Atlantic (red dashed line) are differentiated from the magma poor (non volcanic) margins along Newfoundland and Europe (yellow dashed line).

Origin and Architecture of the Nova Scotia Margin

The Nova Scotia passive continental margin resulted from the break-up of the Pangean continental block at the end of Triassic period, approximately 225-220 million years ago. As described in Beaumont (2010) "*we have only a rudimentary understanding of the way the final crustal structure, as observed today, is linked to the Triassic-Jurassic lithospheric extension and rifting between Nova Scotia and Morocco. In particular, the form of the syn-rift and early post-rift sedimentary basins can only be determined approximately from either the large-scale crustal structure or interpretation of the ION/GXT NovaSpan seismic reflection images of the deep sedimentary structures.*"

Magma Dominated or Magma Poor Margin

The Nova Scotia Margin is located between a magma dominated province (i.e., a volcanic margin) to the south along the US East Coast and a magma poor province (i.e., a non volcanic margin) to the north along the east coast of Newfoundland (Figure 1). The southwestern part of the margin (until 62° W) has all the characteristics of a magma dominated margin with clear seaward dipping reflectors (SDRs; Figure 2b). In contrast, the north-eastern part of the margin (between 62°W and 55°W), just south of the Newfoundland-Azores fault zone (shown on Figure 1) cannot be characterized by direct seismic reflection imaging.

Magma-poor margin

The magma-poor margin is characterized by a wide area of highly attenuated continental crust where the upper crust is deformed by deep listric faults that may sole out on a common detachment surface (brittle/ductile crust boundary; Figure 2a). In the distal margin the listric faults may crosscut the entire crust leading to a detachment at MOHO discontinuity. Further seaward extensional allochthonous blocks may be situated on an exhumed/serpentinized mantle, before a relatively thin oceanic crust is reached. The serpentinized mantle appears as a high velocity body with velocities ~7.2 km – 7.6 km (double arrow on Figure 2a).

Volcanic margin

A volcanic rifted margin is characterized by a thick wedge of volcanic flows manifested in multichannel seismic reflection data as seaward dipping reflectors (SDRs) and high seismic velocity lower crust also called an underplated body (*Vp* > 7.2 km/s), both located seaward of the rifted continental margin (Figure 2b). Because the rapid generation of voluminous amounts of magma requires rapid and significant melting in the mantle, White and McKenzie (1989) proposed that an anomalously hot mantle (100-200°C above normal) must be present under the rift shortly before continental breakup to enable the formation of volcanic rifted margins. Subsequently it has been proposed that either such temperature anomalies, or mantle plumes by themselves, may cause the breakup of continents.

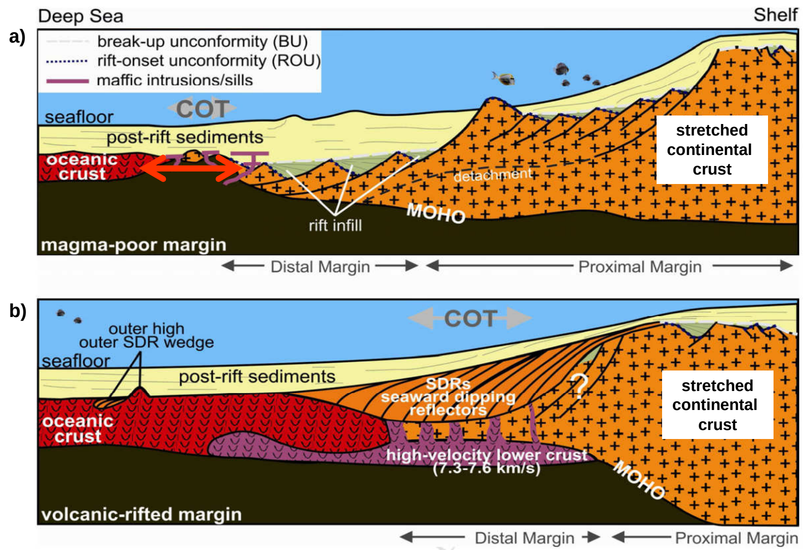


Figure 2: Schematic sketch of end-members of passive continental margins. a: the magma-poor margin sketch; b: the volcanic margin sketch. Red double arrows: exhumed mantle (from Franke, 2012).

Magma dominated or magma poor: why does it matter for the petroleum system?

The deep rifting processes and the deep architecture of the margin have a direct impact on thermal regime and subsidence history that control the deposition and maturation of source rocks. In particular, the radiogenic heat of the continental crust contributes to the maturation of the source rocks meanwhile serpentinized mantle and oceanic crust have no primary influence on the thermal regime. The Seaward Limit of the Continental Crust (SLCC) or Continent Ocean Transition (COT or OCT) zone, as well as the thinning factor of the continental crust are important parameters of the basin modeling.

A primarily non-magmatic rift to drift process (Figure 2a) would most likely imply a deep water environment, due to relatively rapid subsidence during the syn-rift and early post rift phases. This would almost certainly preclude the possibility of a shallow restricted anoxic marine environment and hence the possibility of a rich Early Jurassic source rock system.

If the rift followed an essentially volcanic process (Figure 2b) it would imply uplift and sub-aerial extrusives characterized by seaward dipping reflectors (SDRs), which would in turn imply a much longer period of shallow restricted marine environment during the late phase of the rift to early post rift. During this relative uplift phase, one could expect deposition of evaporite, carbonate and shallow marine source rocks. Such a source bed system of Early Jurassic age could then be argued to be present along the whole margin and be of a high richness because of the restricted marine environment of deposition.

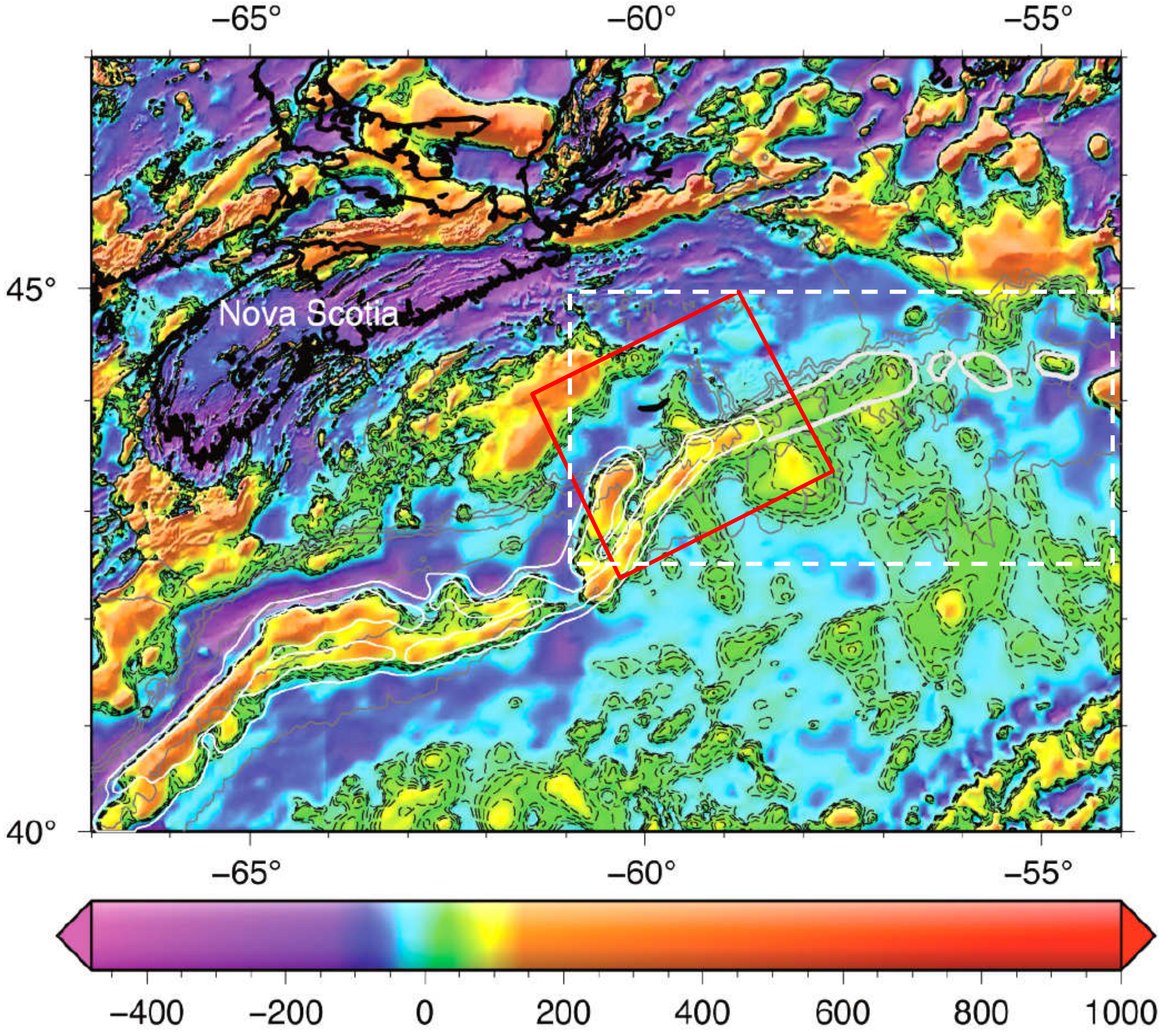
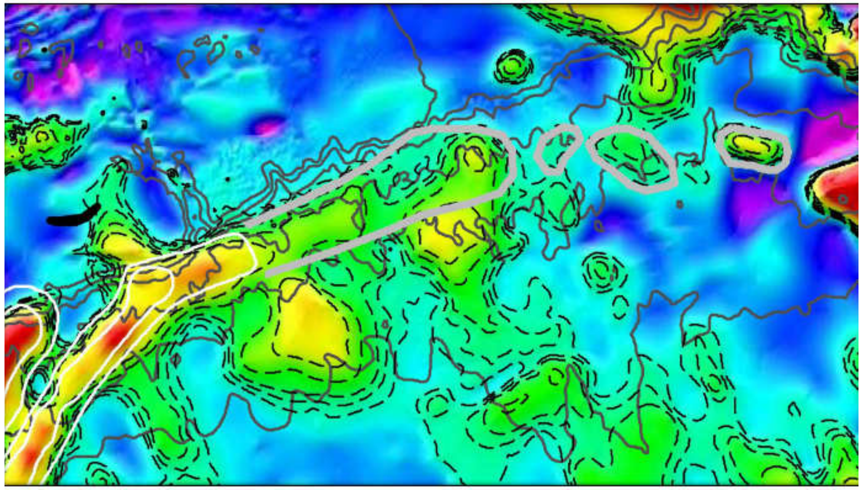


Figure 3: Updated magnetic grid of the northwestern central Atlantic Ocean (Dehler, 2010) (top) and detail of the northeastern termination of ECMA (top right). Dashed lines are contoured magnetic anomalies and continuous black lines are bathymetric contours every km. In white the portion of ECMA already identified by Sahabi et al. (2004) and in grey what Sibuet et Rouzo (2012) suggest as a reasonable northern prolongation of ECMA. Red box: study area; dashed white box: zoomed area shown at top right.



The magnetic grid of Verhoef et al. (1996) has been updated by introducing the detailed magnetic data acquired by Fugro in the northeastern part of the map (Dehler, 2010). Thin continuous grey lines are bathymetric contours (200 m, 500 m and then every km, extracted from ETOPO1 data set (Amante and Eakins, 2009).
The East Coast Magnetic Anomaly (ECMA) constitutes the prominent positive anomaly (higher than 100 nT) underlined by white contours (ECMA already identified by Sahabi et al. (2004). The heavy gray lines show the present interpretation of the northern ECMA prolongation. Dashed black lines are a selection of iso-values used to interpret the ECMA prolongation.

ECMA Northeastward Termination

East of 58.5° W, the amplitude of the ECMA is considerably reduced and associated magnetic anomalies broaden. Whereas the ECMA features a single positive anomaly that locally splits in two branches southward, the northward continuation shows a rather different signature (Figure 3). First, the amplitude is two to three times lower. Second, the prolongation mostly appears as a linear borderline between a large negative anomaly to the north (blue in the Figure3) and a gently varying positive anomaly to the south (cyan and green in the Figure3). Thus, the northeastward prolongation of the ECMA is straightforward for its landward side and, using the magnetic iso-contours up to 56.5°W, could be define as a stripe not wider than the ECMA southward. Further east, only individualized, shorter extent and discontinuous anomalies could be interpreted as a possible prolongation of the ECMA.

What Caused the ECMA?

The East Coast Magnetic Anomaly (ECMA) seems to be directly related with volcanics producing SDRs all along the US Atlantic margin (Plate 2-1). However toward the northeast, the magnetic anomaly is much more diffuse and faint even if new reprocessed data acquired by Fugro suggest a more continuous anomaly. On seismic lines seaward dipping reflectors are impossible to identify in this part of the margin.

Dehler (2010) suggested by forward modeling that the ECMA can be related to a very thin layer of volcanics located above the eastern part of the high velocity body, slightly extending to the northwest of the high velocity body. The ECMA may also be partly caused by the edge effect due to the juxtaposition of a magnetized serpentinized body and a poorly magnetized thinned continental crust.

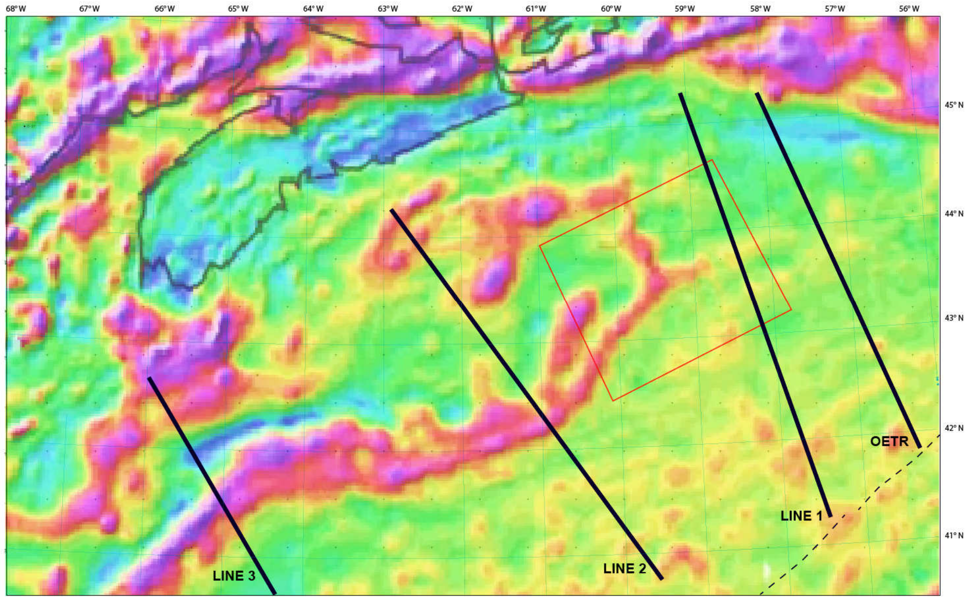


Figure 4: Magnetic map of offshore Nova Scotia (from Sibuet et al., 2011).

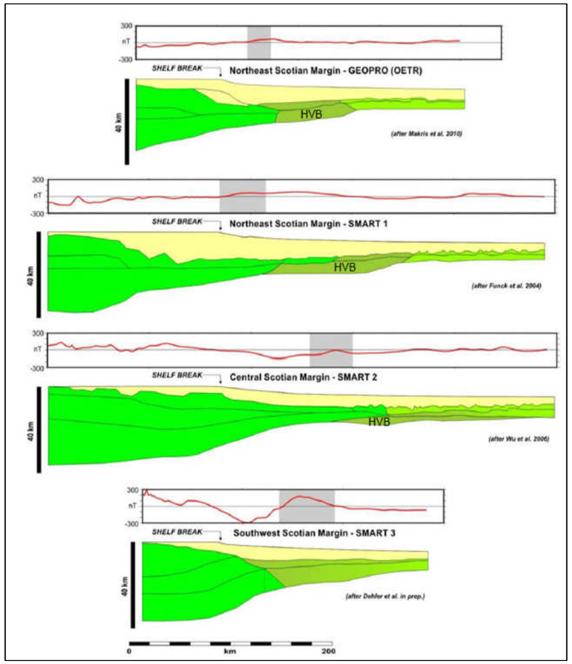


Figure 5: Seismic refraction lines shown in Figure 4 and magnetic anomaly profiles. The positive anomaly identified as the ECMA is indicated on the profiles (shaded areas). HVB= High Velocity Body.

Magnetic Data

The East Coast Magnetic Anomaly (ECMA) can easily be followed along the US Atlantic Margin and much of the Nova Scotia Margin. The anomaly varies in character along the margin, with several areas where offsets are noted, such as the vicinity of the New England Sea Mounts near 40°N (Figure 4). The changes are more notable offshore Nova Scotia, where both the trend and the character of the anomaly change. The anomaly becomes more difficult to follow to the northeast, and appears to break into two parallel components, one of which terminates near Sable Island at 60°W. The outer most branch continues until at least 58°W. It is difficult to determine the trend of the anomaly east of this location on the regional compilation of magnetic data, which was compiled and processed by the Geological Survey of Canada (GSC) from a variety of sources including aeromagnetic grids and marine surveys (Oakey and Dehler, 2004).

A grid of high resolution magnetic data, acquired by Fugro in 1999 through 2001, covers the northeastern end of the margin where GSC coverage is sparse. This grid was processed and merged into the regional GSC grid for this study to allow examination of the eastern end of the ECMA. On the merged map the ECMA appears to continue until at least 57°W (Figure 4).

Refraction Profiles and ECMA (Figure 5)

The “margin anomaly” does not coincide with the high-velocity bodies or shelf break except on the SMART Line 3.
The “margin anomaly” could be explained by thin volcanic units on lines 1 and 2. At the intersection with the OETR refraction linea ~2km thick volcanic unit is required to model the strength of the magnetic anomaly (PFA 2011).
The position of the volcanic unit is roughly coincident with the seaward edge of thinned continental crust (except for Line 3).

ECMA vs Magnetic Anomaly (Figure 6):

A set of magnetic profiles has been acquired along the US East Coast. Wu et al. (2006) compiled the magnetic data and compared the position of the peak with the location of the ECMA. A good fit is observed in the south of the margin (from US Atlantic Margin to the south of the Nova Scotia Margin, Figures 4 & 6) between the ECMA position and the magnetic anomaly's positive peak. The correlation becomes more difficult to follow further north. Indeed the peak becomes broader and the correlation with the ECMA position is uncertain (Figures 5 & 6).

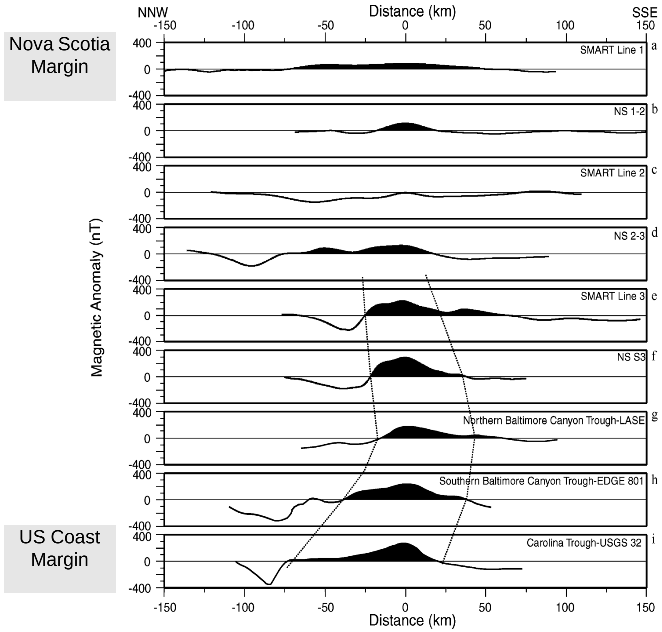


Figure 6: Magnetic profiles across the East Coast margin of North America. A good fit is observed on southern profiles (e-i) between the peak of positive anomaly (filled in black) and the ECMA position (width of ECMA defined by two dotted lines). On the northern side (a-d) this correlation seems to disappear and the anomaly becomes broader and weaker (from Wu et al., 2006).

The Nova Scotia Margin is located in between the volcanic U.S. Atlantic Margin to the southwest and the non-volcanic Newfoundland Margin to the northeast. This observation, added to the gradual disappearance northward along the Scotian Margin of both the East Coast Magnetic Anomaly (ECMA) and the Seaward Dipping Reflector sequence on the seismic reflection profiles, tends to designate the Nova Scotia Margin as the transition area from volcanic to non-volcanic rifting.

In order to image the transition from continental to oceanic crust and to assess the lateral variation in crustal structure, three wide-angle refraction seismic lines perpendicular to the coastline were acquired in 2001 along the margin. These lines, called Scotian Margin Transects refraction lines (SMART line) were processed and interpreted by the Department of Earth Sciences of the Dalhousie University (Funck et. al. 2004 and Wu et. al. 2006).

A new wide-angle refraction reflection line was acquired in 2009 by GeopPro GmbH. This line (OETR), parallel to an existing wide-angle reflection line (GXT NovaSPAN line 2000), was undertaken to clarify the crustal structure of the ocean continental transition and for use in the plate reconstruction before oceanization along the northern part of the Nova Scotia Margin.

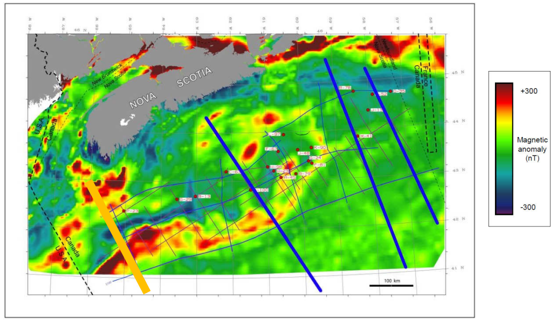


Figure 7: Location of the wide-angle seismic line (SMART3, shown in yellow) on the regional magnetic anomaly map.

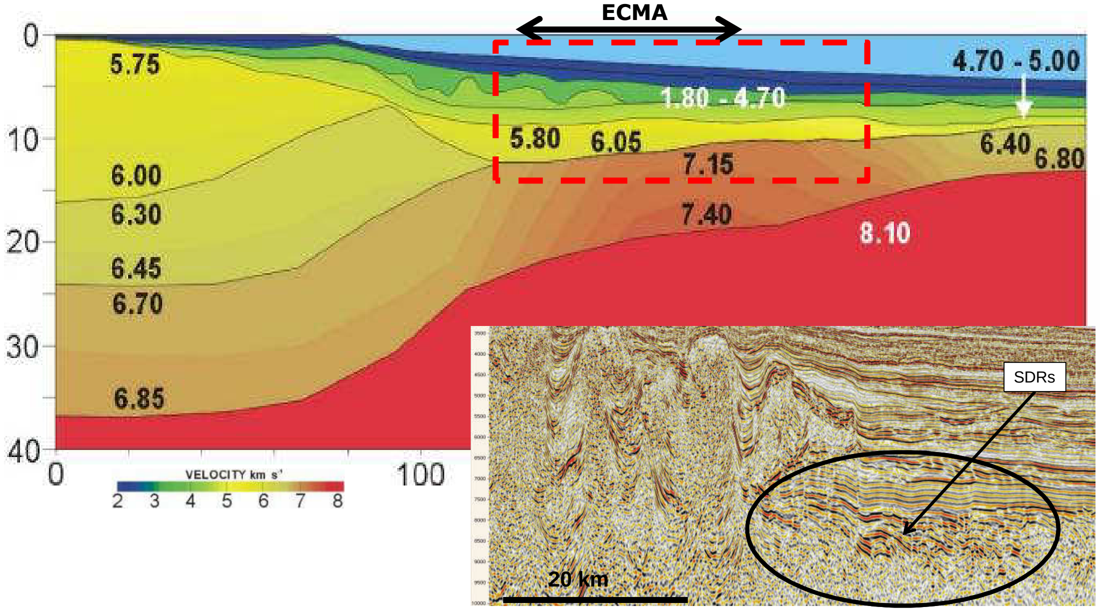


Figure 8: P-Wave velocity model: SMART3 line. Red box: seismic line zoom position (NovaSpan Line 1100PSDM).

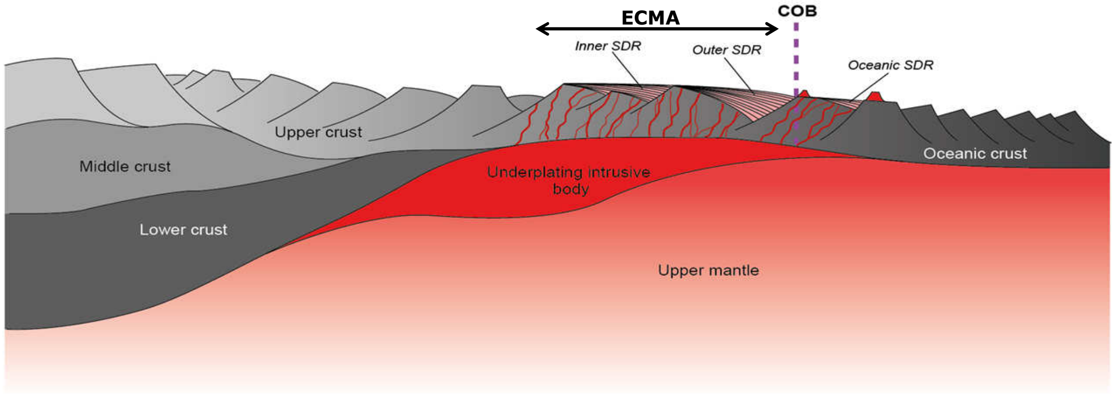


Figure 9: Sketch of a magma rich or volcanic margin. It is characterized by a narrow thinning continental crust which is separated from the oceanic crust by a thick magmatic underplating body. This underplating is expressed by thick volcanic wedges (SDR) on surface.

Velocity Modeling Description

The continental crust is divided into three layers: the upper, middle and lower crust (Figures 8 & 9).

SMART3 shows a continental crust thinning over 100 km wide zone. The middle crust dips landward and appears to have been removed, perhaps eroded, after x = 80 km. In this area, the sediments are directly overlying the lower crust.

A high velocity lower crustal (HVLC) layer with Vp > 7.2 km/s has been identified at x = 120 km, exactly at the interception of the ECMA. This HVLC is interpreted as an underplated intrusive body. This hypothesis is supported by the Seaward Dipping Reflector sequence observed on the seismic reflection line (Figure 8).

Interpretation

The comparison of various crustal transects from the US East Coast and SMART3 illustrates that the character of SMART3 is consistent with the magma-rich U.S. margins to the south. These transects are similar in terms of the total width of stretched continental crust, thickness of oceanic crust, the presence of an interpreted magmatic underplate, and seaward dipping reflectors (Figures 8 & 9).

This magma-rich part of the margin has therefore the following set of characteristics:

1. Narrow region (100km wide) of continental crust thinning;
2. Thick magmatic underplate separating thinned continental and oceanic crust;
3. Seaward dipping reflectors (SDRs) above the seaward end of the high velocity body;
4. Normally-thick oceanic crust;
5. Thin sedimentary basin;
6. Wedge of low velocity material above underplated region.

Based on these characteristics, the southwest Nova Scotia Margin can be interpreted as the northern extension of the magma-rich margin domain that characterizes the US East Coast (Figure 9).

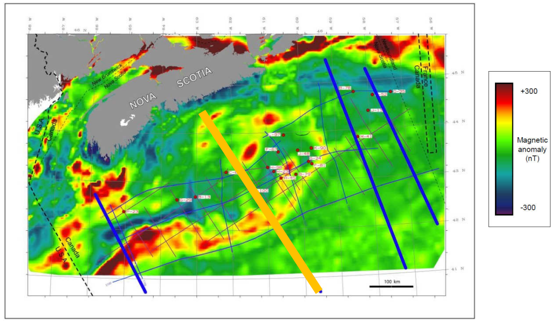


Figure 10: Location of the wide-angle seismic line (SMART2, shown in yellow) on the regional magnetic anomaly map.

Velocity Modeling Description

The velocity model shown in Figure 11 is from Wu et al. 2006.
The continental crust is divided into three layers: the upper, middle and lower crust. The upper crust thins from 4 to 2 km over a distance of 285 km with Vp ranging from 5.5 to 5.7 km/s. The unstretched middle crust is 12 km thick and thins over a distance of 245 km until it is completely absent. Thus, beyond x = 260 km, the upper crust directly overlies the lower crust within a 60 km wide zone. The zone where velocity ranges between 5.2 and 5.6 km/s from 180 to 300 km distance could be attributed to sediments. The lower crust thins from 18 to 2 km over a distance of 300 km with P-Wave velocity ranging between 6.8 and 6.9 km/s (Figure 11).
Immediately following continental breakup, a relatively thin (4 km) oceanic crust forms with a distinct continent-ocean boundary (COB). The oceanic crust is composed of two layers: the lower unit with Vp ranging between 6.95 and 7.4 km/s and the upper unit with Vp ranging between 5.5 and 6 km/s. A constant sedimentary sequence is overlying these two layers. The oceanic continental transition (OCT) zone also encompasses an anomalously High Velocity Lower Crust (HVLC) with Vp between 7.6 to 7.95 km/s which is 200 km wide with a maximum thickness of 4 km.
The maximum sedimentary thickness is 6 km with Vp ranging between 1.8 to 5 km/s and part of what is attributed to the upper crust between km 160 and 300 (Vp = 5.2 -56) is probably of sedimentary origin (Figure 11).

Interpretation

SMART2 is representative of the central margin. Its velocity model reveals an OCT zone consisting of a HVLC.
This High Velocity Body has been interpreted as partially serpentinized mantle (PSM) (e.g. Wu et. al. 2006) or as an underplated magmatic body (PFA 2011).
However, along these profiles there are four noteworthy characteristics:
1. The lateral extent of the middle crust;
2. The total width of the crustal thinning zone that defines the margin;
3. The location and extent of the HV body; and
4. The thickness of the overlying sedimentary basin.
On this part of the margin, no SDRs have been observed at the location of the ECMA on the seismic reflection (Figure 12) but they may be obscured by the prominent diapiric salt structures. Also note that SDRs are observed at the conjugate location offshore Morocco in a zone that intersects the WACMA (West Africa Continental Magnetic Anomaly).

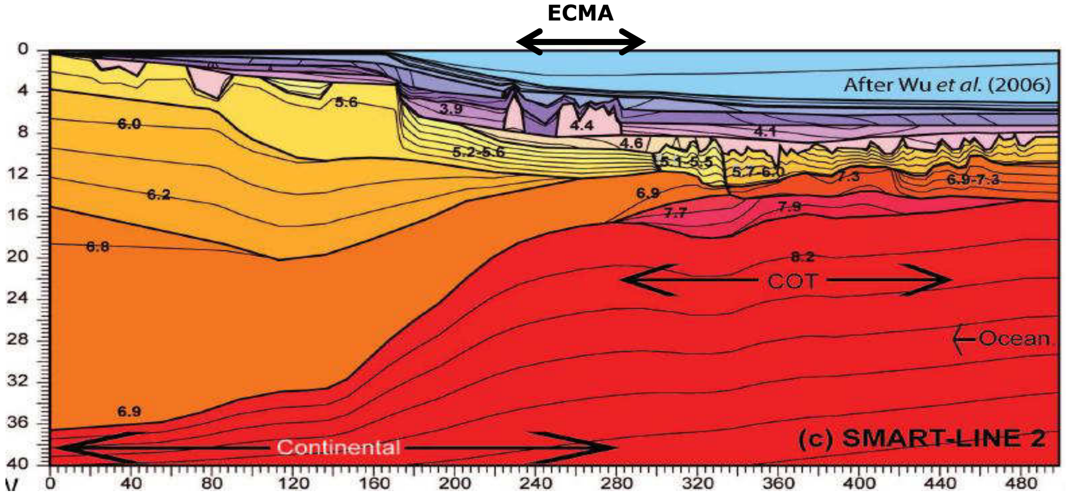


Figure 11: P-Wave velocity model: SMART2 line.

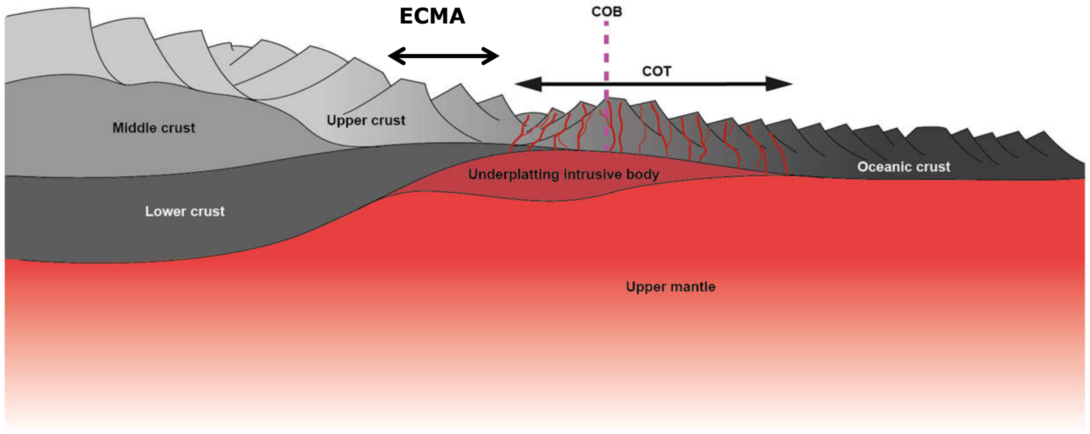


Figure 12: Sketch of the central margin. The middle crust extension is wider than on the other areas. There is still an underplating body between the thin continental crust and the oceanic domain. SDRs are not observed (salt?) and the ECMA does not fit well with underplating body position.

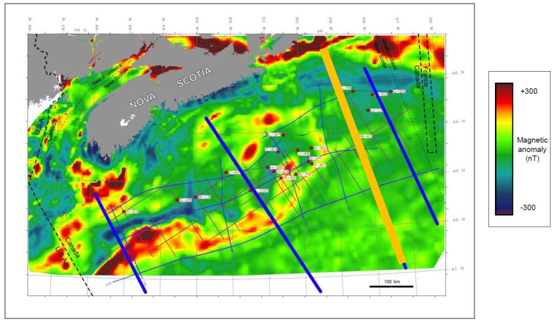


Figure 13: Location of the wide-angle seismic line (SMART1, shown in yellow) on the regional magnetic anomaly map.

Velocity modeling description of SMART1

The velocity model has been developed and interpreted by Funck et al. in 2004 (Figure 14).
The continental crust is divided into three layers: the upper, middle and lower crust. The upper crust thins from 10 to 3 km over a distance of 230 km with Vp ranging from 5.7 to 6 km/s. The unstretched middle crust is 6.5 km thick and thins over a distance of 25 km until it is completely absent. Thus, beyond x = 80 km, the upper crust directly overlies the lower crust within an approximate 160 km wide zone. The lower crust thins from 20 to 5 km over a distance of 180 km with P-Wave velocity ranging between 6.7 and 6.9 km/s.
The oceanic crust is 4 km thick and contains 2 layers with a lower unit of Vp ranging between 6.4 and 6.6 km/s and an upper unit with Vp ranging between 4.7 and 5 km/s. A sedimentary sequence is overlying these two layers (Figure 14).
The upper crust and the oceanic crust are separated by a relatively low velocity domain (5.2-5.4 km/s) which has been interpreted as an exhumed and highly serpentinized mantle 70 km wide (Louden et al. 2010). The continental oceanic transition encompasses a high velocity lower crust body 130 km wide and a maximum of 6 km thick, which separates the lower continental crust and the lower unit of the oceanic crust.
The sedimentary basin can be up to 14 km thick with Vp values ranging between 1.8 to 5 km/s.

Interpretation

SMART1 is representative of the northern margin. In contrast to SMART2, the upper continental and the oceanic crust of SMART1 are separated by a wide Low Velocity Body (Vp =5.2 km/s) overlying a High Velocity one (7.6-7.3 km/s). The Low Velocity Body has been interpreted as exhumed and highly serpentinized mantle while the High Velocity Body is considered as to be partially serpentinized mantle (Louden et al. 2010; Figures 14 & 15). The same alternate interpretation with an underplated magmatic body has also been proposed (Luheshi et al. 2012).

On this part of the margin, the ECMA has almost disappeared and no SDRs have been recognized on the seismic reflection. However, as noted above for SMART2, evidence of SDRs at the offshore Morocco conjugate location has been reported by Maillard et al. (2006) and Roeser et al. (2002) (Figure 15).

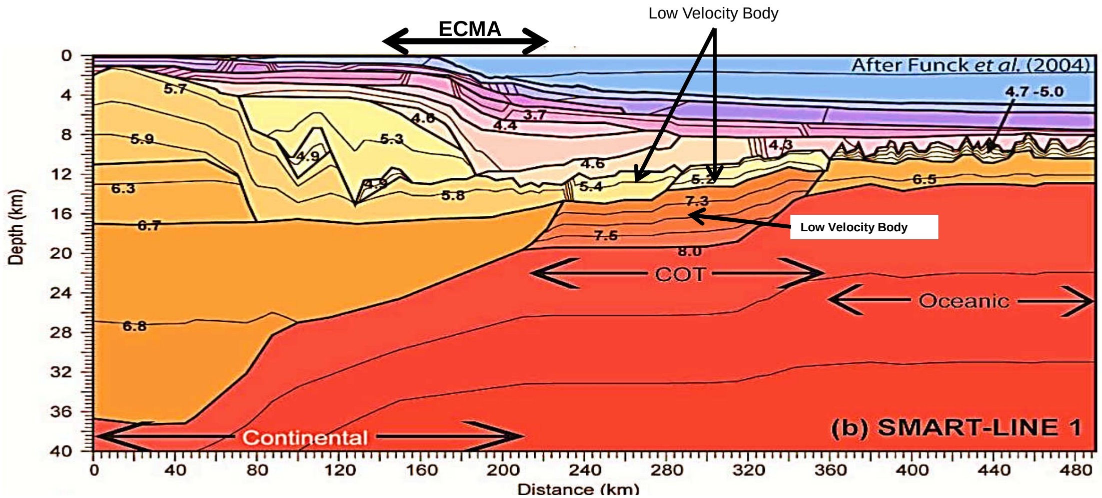


Figure 14: P-Wave velocity model: SMART1 line.

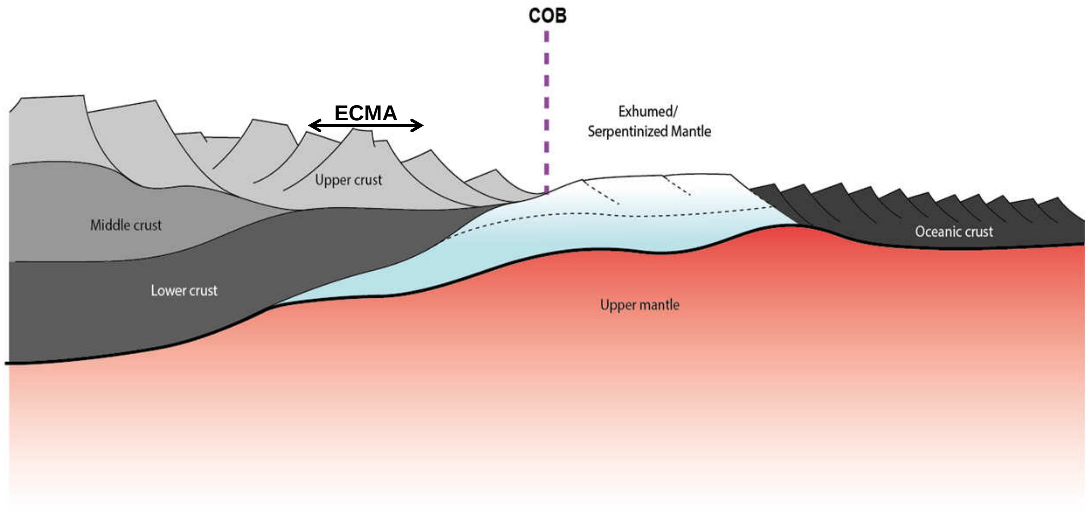


Figure 15: Sketch of the northern margin. The thin continental crust is separated from the oceanic crust by exhumed or seprentinized mantle.

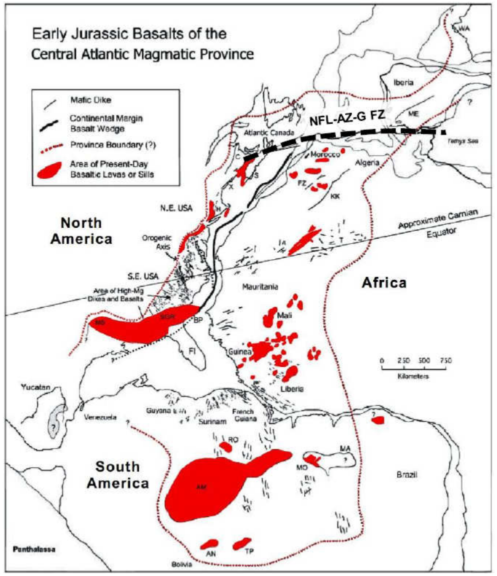


Figure 16: Mesozoic volcanism in the Central Atlantic Magmatic Province (CAMP). The base map is the pre-volcanism reconstruction modified from Klitgord & Schouten (1986). Mesozoic volcanism is mainly developed south of Newfoundland-Azores-Gibraltar Fault Zone (Louden et al. 2010).

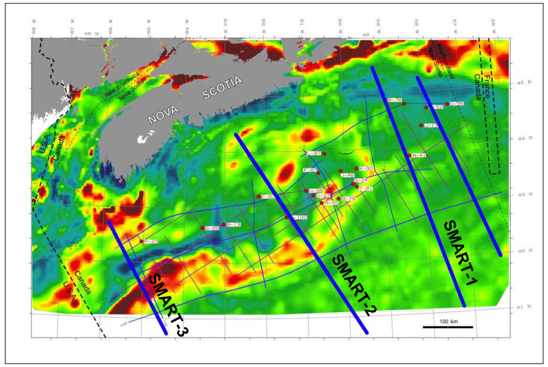


Figure 17: Location of the wide-angle seismic line on the regional magnetic anomaly map.

An overview of the different margin architectures on both the Canadian and Moroccan sides of the Atlantic has been prepared by Klingelhoeffer et al. 2016. They combined the SMART profiles from the Canadian side and the SISMAR, MIRROR and DAKHLA profiles from the Moroccan margin in order to compare the architecture along the entire length of the margins. The comparison reveals a major variation in deep crustal structure along the strike of the margin (Figure 18).

The southern part of the margin (SMART-3/DHAKLA Nord) shows characteristics typical of a volcanic margin, with narrow rifted continental crust, thick oceanic crust and thick SDR wedges between these two domains (Figure 18C).

On the northern part (SMART-1/SISMAR4), the continental thinning is wider on both sides. Moreover the SDRs are clearly imaged on the Moroccan side (on 3D seismic), whereas they are absent on the Canadian margin. A high velocity body is also highlighted on this margin (Figure 18A).

These differences indicate that the rifting and initial opening of the oceanic basin did not follow identical mechanisms from north to south.

The conjugate passive margin of Nova Scotia is characterized by a decrease of the amount of volcanism associated to the original break-up from south to north also observed at the ECMA.
The evolution of margin characteristics from south to north suggests that the Nova Scotia Margin is a transition margin with a typical volcanic (magma-rich) margin to the south (US coast) to a magma-poor margin to the North (Newfoundland–Azores-Gibraltar transform).

However the precise nature of the margin is still an open question and more data is required to clarify its structure and history.

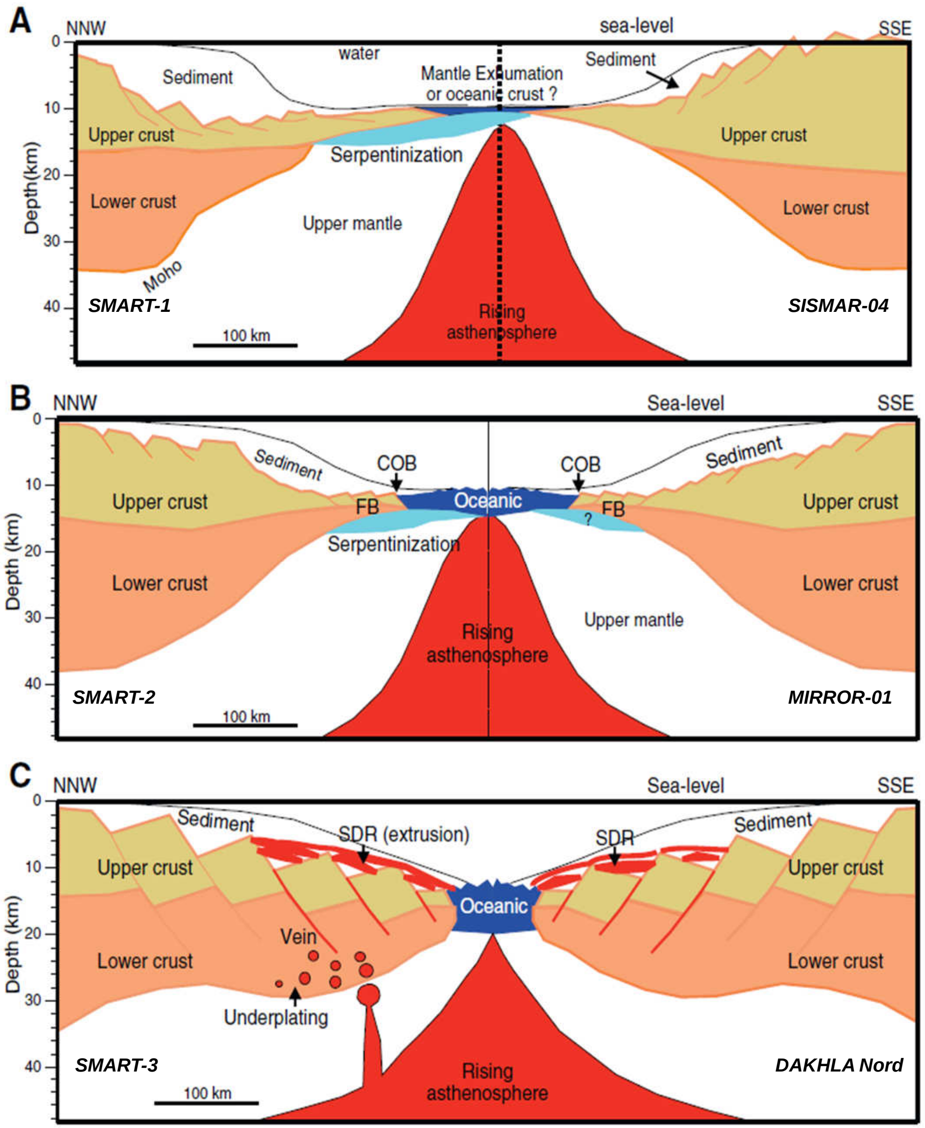


Figure 18: Variation of volcanism along the Nova Scotia–Morocco conjugate margins. A) conceptualized crustal structure across the NE segment; B) crustal structure across the central segment; C) crustal structure across the SW segment; COB: Continental Oceanic Boundary; FB: Faulted Block; SDR: Seaward Deeping Reflectors (from Klingelhoefer et al. 2016).

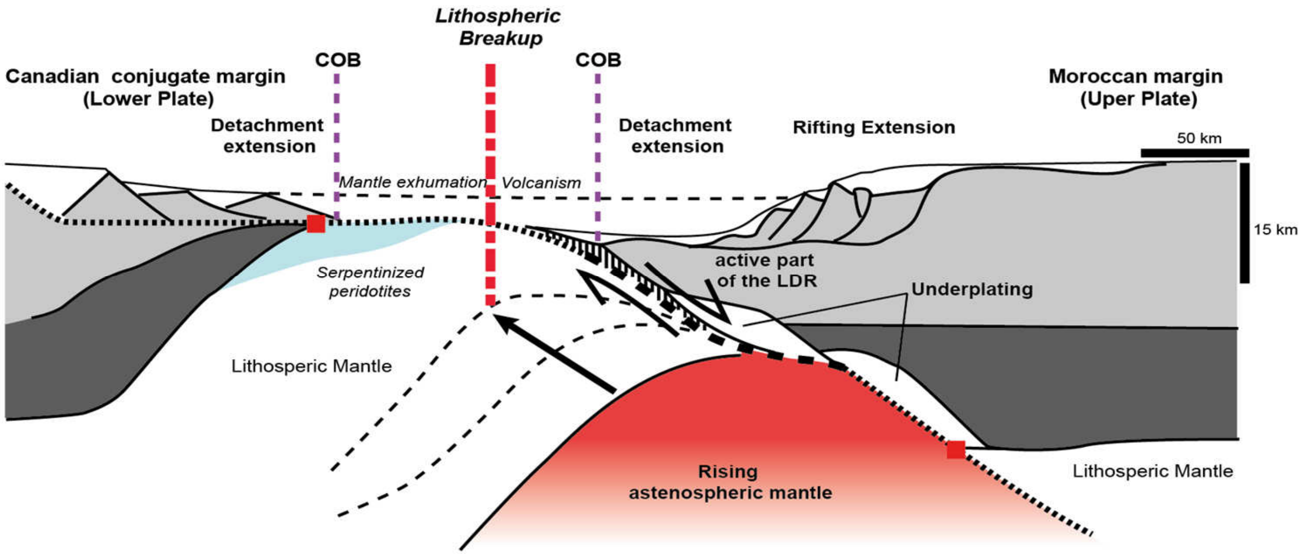


Figure 19: 2D conceptual model of the conjugate margins of Morocco and Nova Scotia at a pre-rupture stage (based on SISMAR and SMART1 profiles, Figure 14). These conjugate margins result from a simple shear rifting type with an upper plate (Morocco) and lower plate (Canada). The result is a mantle exhumation in the Canadian Margin and an underplating with extrusion in the Moroccan Margin. From Klingelhoefer et al. 2016.

Maillard et al. (2006) and Klingelhoeffer et al. (2016) proposed a composite model to explain the margin architecture of the northern portion of Nova Scotia-Morocco conjugate margins.

They proposed a synthetic model with a simple shear rifting, where the Canadian Margin is the lower plate and Moroccan the upper plate (Figure 19).

The crustal thinning results from two stages:

1: a widespread crustal extensional stage during continental rifting;

2: exhumation of lower continental crust or mantle close to the surface throws a large detachment.

This sketch shows the stage immediately before lithospheric rupture and further asthenospheric rise. This model integrates all observations presented by Maillard et al. 2006.

The lithospheric detachment fault (simple shear) explains the asymmetry of both margins. Because the Moroccan Margin is the upper plate, crustal thinning is wider here than along the Canadian Margin. At the end of continental thinning, a mantle exhumation stage will appear before true oceanic accretion occurs (Figures 19 & 20C).

In this sketch, the Landward Dipping Reflector (LDR) on the Moroccan side represents the latest active part of the detachment. Along the LDR the rising asthenospheric mantle may have produced magmas feeding underplated bodies or volcanics along it (Figure 19).

An interesting point regarding the rate of motion between North America and Africa was highlighted by Maillard et al. 2006. Considering homologous points on both sides on the detachment, as indicated by red squares (Figure 19), the overall displacement between these two plates during detachment was about 300km. This implies a very different rate of initial motion between both margins and thus that the volcanics associated to the LDR is younger than the CAMP magmatism.

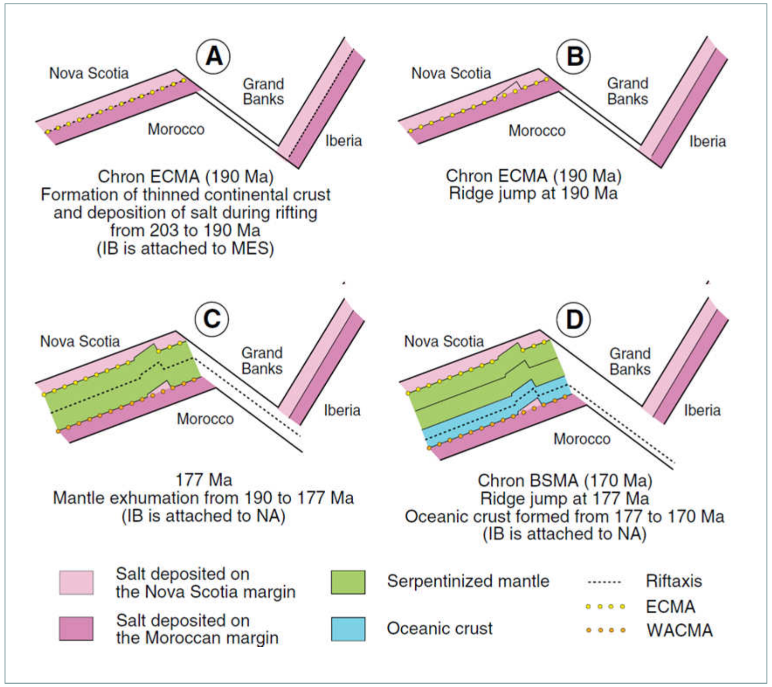


Figure 20: Sketch showing that the westward shift clipped out a portion of the Nova Scotia margin and its overlying salt features on the Moroccan side (190Ma), while a later eastward shift left the serpentinized mantle on the Canadian side (177Ma). IB: Iberia; MES: Meseta; NA: North America (From Sibuet et al. 2012).

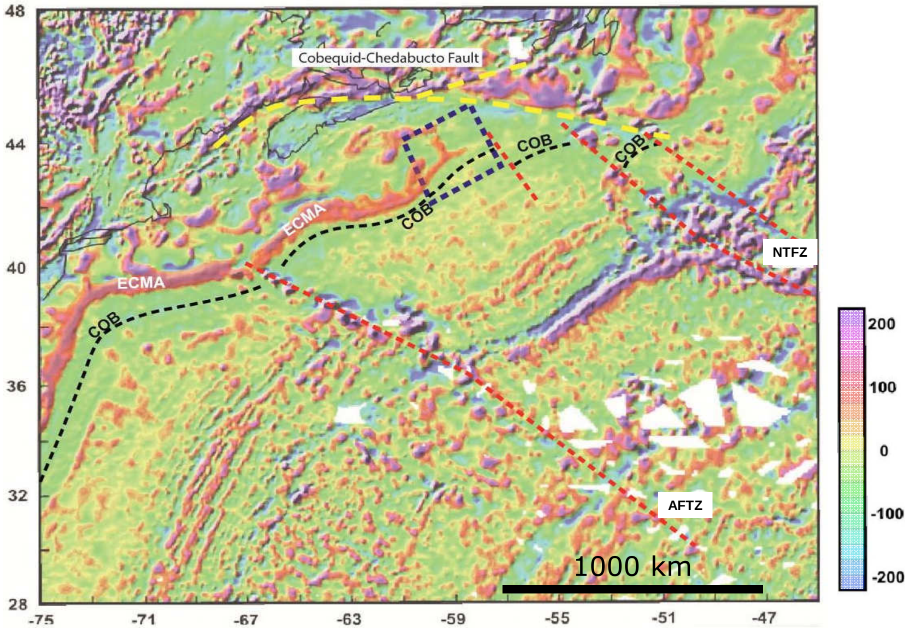


Figure 21: Magnetic anomaly map (Dehler, 2010) and main tectonic features. The oceanic domain is characterized by typical succession of normal and reverse linear magnetic anomalies. Georges Bank and the Scotian Shelf are separated from the oceanic domain by the strong positive regional East Coast Magnetic Anomaly (ECMA). Continent Ocean Boundary (COB) is deduced from previous PFA studies (Sibuet, 2011; Labails et al. 2010).

Using gravimetric and magnetic anomaly maps, several major tectonic elements of the Nova Scotia Margin area can be traced (Figure 21):

1. The East Coast Magnetic Anomaly (ECMA) and the Continent Ocean Boundary (COB). The COB is clearly delineated and easy to define in the Georges Bank in the overall margin (Figure 21). It is more difficult to define in the Laurentian sub-basin where the ECMA and COB signals are weak and along the Grand Banks where the ECMA no longer exists (Figures 21 & 22).
2. The Newfoundland Transform Zone (NFTZ) is one of the most prominent fault systems in the North Atlantic. It constitutes the main escarpment of the South Bank High and forms a well-defined limit just north of the J ridge and Newfoundland ridge anomalies. Its trace along the South Bank High is less well expressed on gravimetric and magnetic maps and was constrained by seismic data.
3. The Cobequid-Chedabucto Fault is clearly imaged by gravimetric and magnetic data. Its trace is evident both offshore and onshore, where it constitutes a major fault separating the Middle Paleozoic Terranes of the Meguma Block from the Upper Paleozoic (mainly Carboniferous) Avalon units. In the offshore east of Nova Scotia, the main fault appears to split into a number of splay faults but does not appear to connect directly to the NFTZ.
4. The Appalachian Thrust Front, illustrated primarily through the contrast between (Figure 21):
   a) magnetic "lows" of the Appalachian foreland corresponding to the Carboniferous flexural basin; and
   b) magnetic "highs" of the deformed belt.

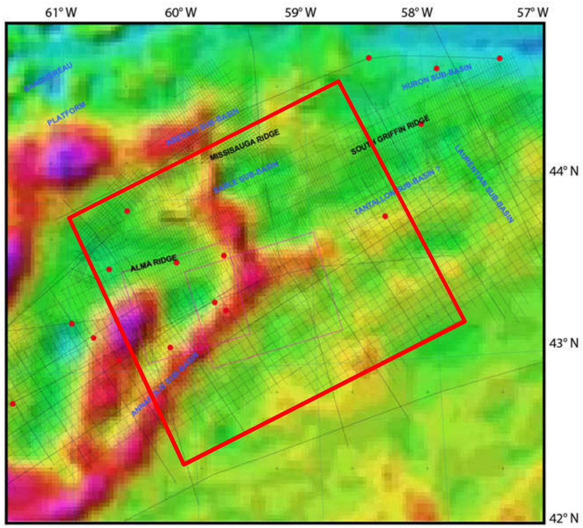


Figure 22: Enlargement of the magnetic anomaly map from Dehler, 2010. Study area is shown by the red polygon.

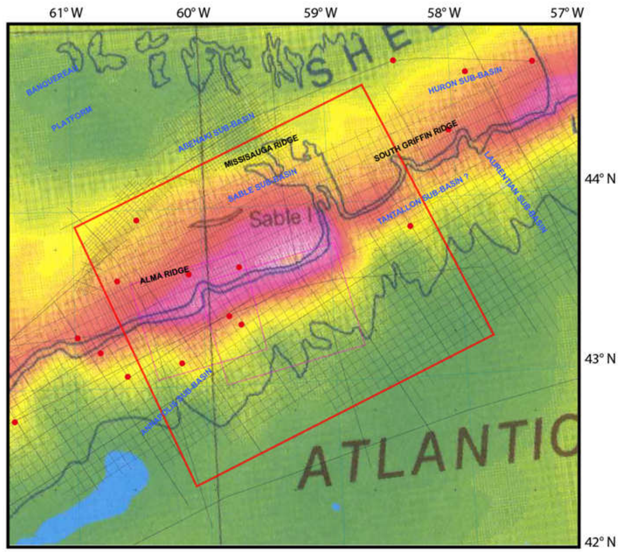


Figure 23: Bouguer Gravimetric Anomaly Map and main tectonic features. Mesozoic age sedimentary basins of Annapolis sub-basin are clearly imaged by positive values of the isostatic gravity anomaly. Study area is shown by the red polygon.

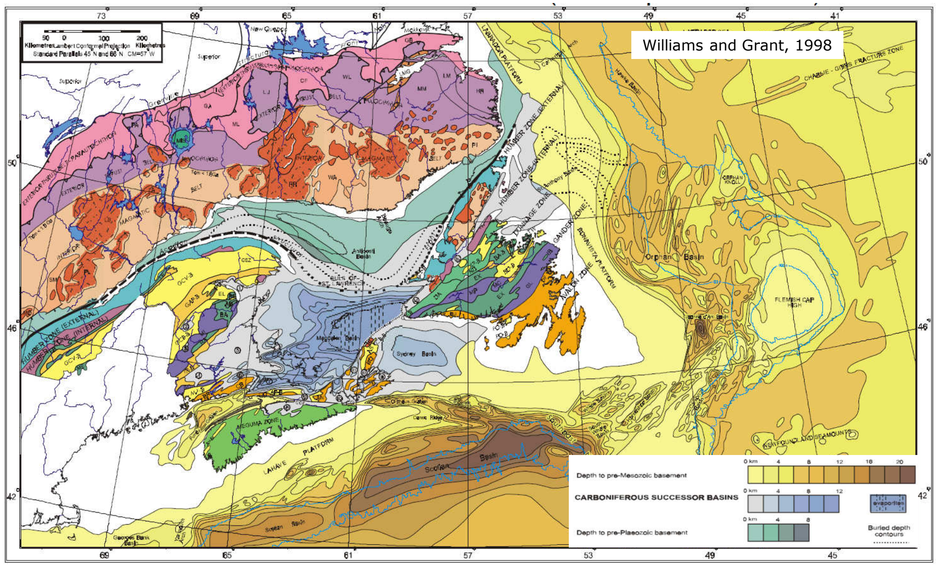


Figure 24 : Geological map of Proterozoic and Paleozoic basement. Contours indicate:
1) *Depth to pre-Mesozoic basement (brown and yellow colors);*
2) *Depth to pre-Carboniferous basement (grey-blue colors);*
3) *Depth to pre-Paleozoic basement (grey-green colors).*

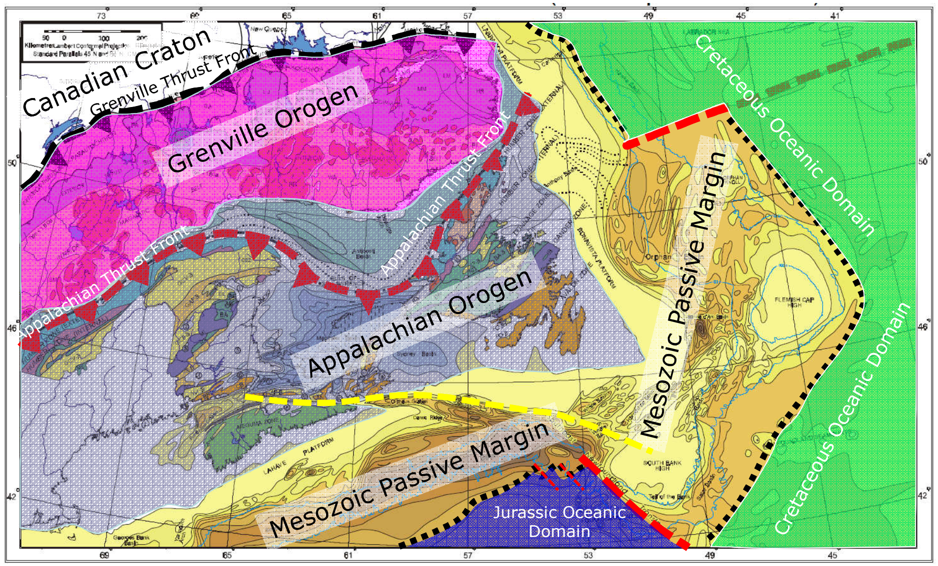
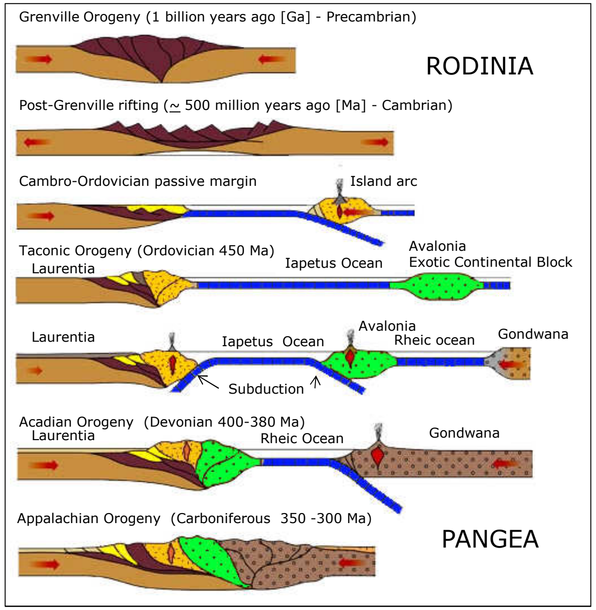


Figure 25: Nova Scotia and Newfoundland main tectonic units. 1) Canadian Archean and Proterozoic Craton (white); 2) Proterozoic Grenville Orogeny (pink); 3) Paleozoic units including Appalachian foreland and Appalachian orogeny (grey); 4) Mesozoic passive margin (yellow and brown contour map); 5) Jurassic oceanic domain (purple); and 6) Cretaceous oceanic domain (green). The Cobequid-Chedabucto Fault is shown by the yellow dashed line.



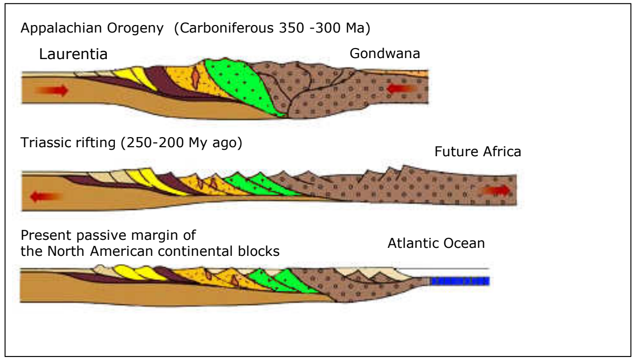


Figure 26: Schematic cross-section showing the pre-Mesozoic evolution of the Nova Scotia continental margin.
Modified from https://mountainbeltway.wordpress.com/category/west-virginia

The present Nova Scotia Margin is the result of a series of complex geological events that began in Proterozoic times (Figures 24 and 25). The present day passive margin developed after the completion of a full Wilson cycle (the opening and closing of an ocean basin) beginning with the breakup of the Rodinia supercontinent, the formation of the Pangea supercontinent and the subsequent breakup of the Pangea. The initial breakup of Rodinia and formation of Pangea resulted from the following mountain building and rifting events (Figure 26).

- **Grenville Orogeny** (1,000 Ma)

Grenvillian rocks are subdivided into a set of allochthonous terranes arranged in the form of a south-easterly dipping thrust stack emplaced over an Archean age continental margin and intruded by numerous post-orogenic plutons.

- **Post-Grenville rifting:** Iapetus Ocean (~ 500 Ma - Cambrian)

Following the Grenville Orogeny, which resulted from the collision of two continental blocks during Precambrian times (Figure 26), the newly formed continent named Rodinia began to break apart during a rifting event in Cambrian times. This break-up led to the formation of the Iapetus Ocean separating two large continental blocks, Laurentia and Gondwana.

- **Taconic Orogeny** (Ordovician 450 Ma)

During Ordovician times, island arc material was accreted to the Cambro-Ordovician passive margin resulting in a major orogeny or mountain building event.

- **Acadian Orogeny** (Devonian 400-380 Ma)

During the Devonian period, the collision of an isolated continent (the Avalonia block) induced the closure of the Iapetus Ocean and a second major orogeny.

- **Appalachian Orogeny** (Carboniferous 350-300 Ma)

During the Carboniferous, the collision of the future African continental block (Gondwana) with the North America continental plate (Laurentia) caused the closure of the Rheic Ocean, the Appalachian Orogen (the formation of the Appalachian Mountains) and the Pangea supercontinent.

- **Triassic rifting**: Atlantic Ocean (250-200 My ago)➔ Passive margin history

The next step of this evolution was rifting during the Triassic period, initiating a new Wilson Cycle.

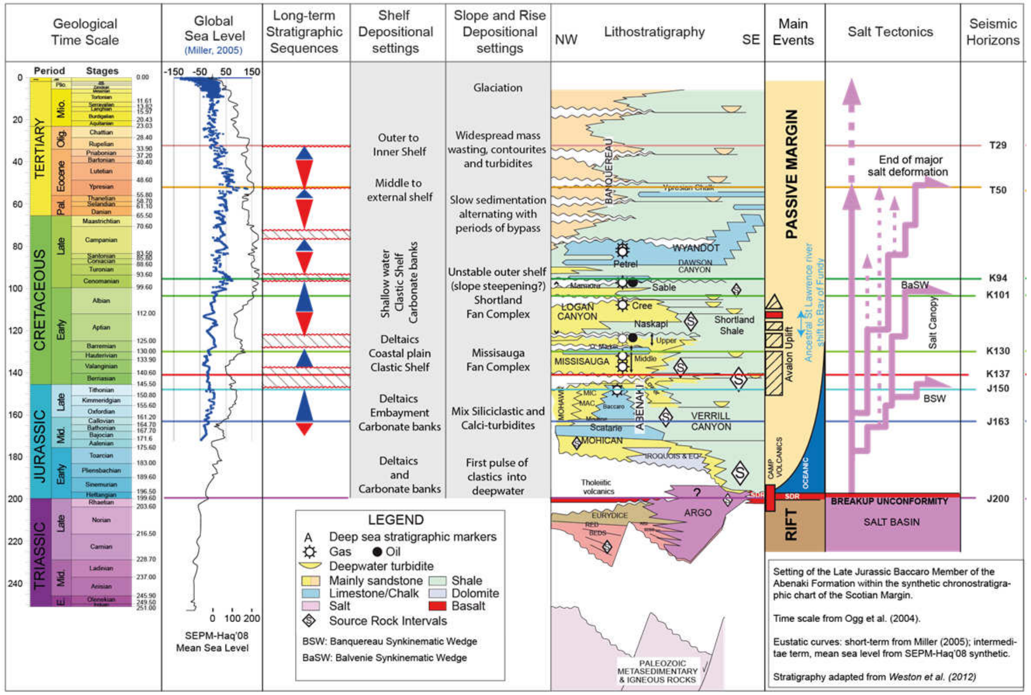


Figure 27: Stratigraphic chart of the Nova Scotia Margin from rifting to drifting stages.

Following initial rifting at the end of the Triassic / beginning of Jurassic periods, the oceanic accretion stage begins (Figure 27) and the Nova Scotia Margin begins to subside; passive margin processes dominate.

To accommodate the oceanic accretion in the Central Atlantic and the propagation of Atlantic rifting, a set of transform faults appears, in particular the Azores Fracture Zones (AFZ) and the Newfoundland Fracture Zone (NFZ). These faults will affect the entire margin (Figure 28).

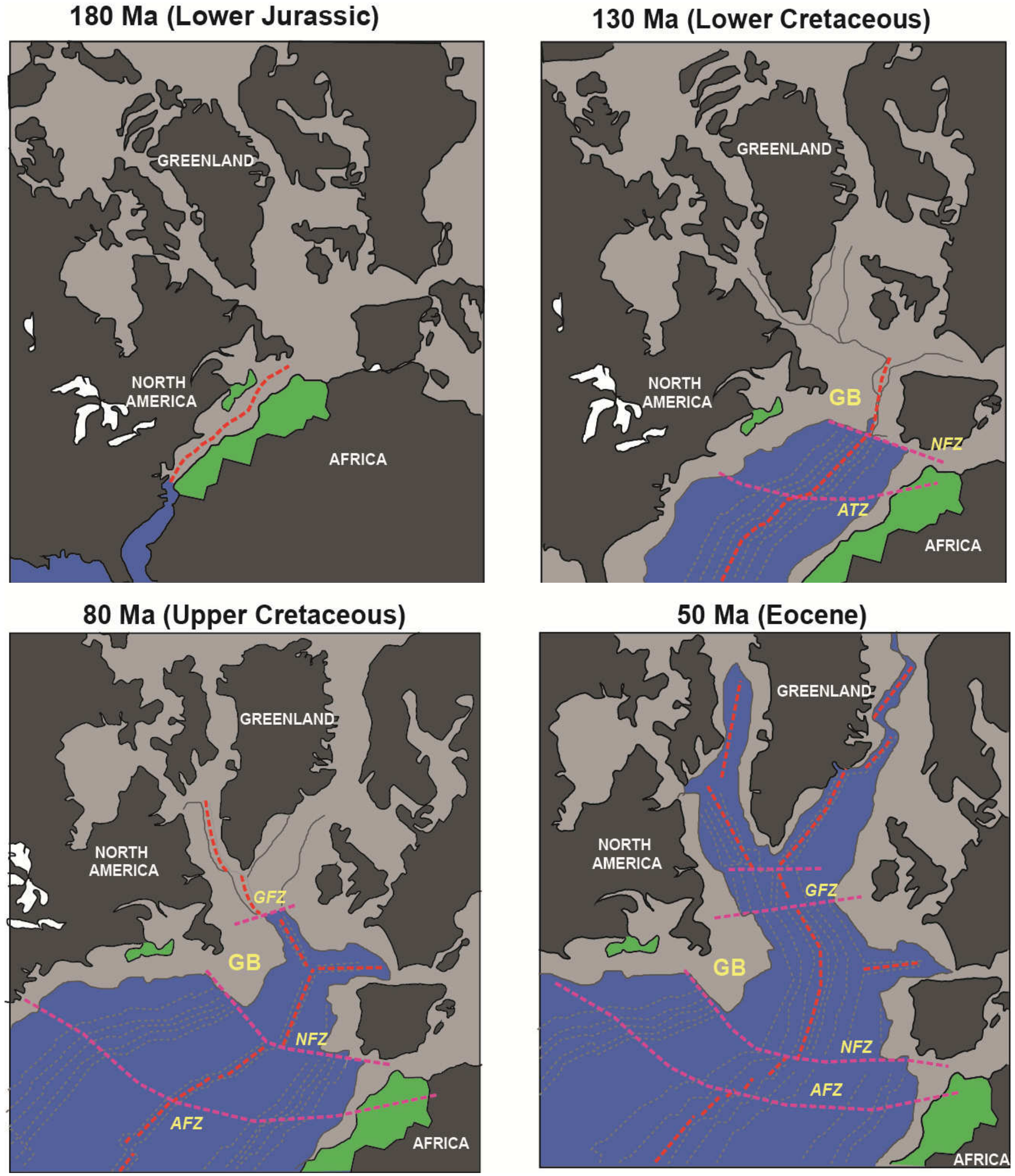
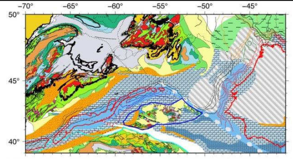
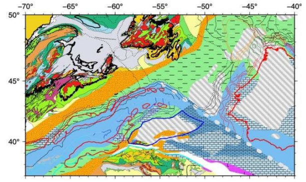


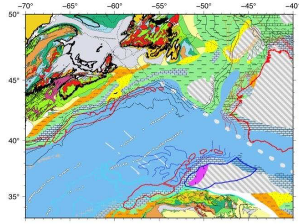
Figure 28: Plate tectonic reconstruction during the opening of Central Atlantic at 180 Ma, 130Ma, 80Ma and 50Ma. Dashed grey lines give the location of sea-floor spreading anomalies. NFZ: Newfoundland Fracture Zone; AFZ: Azores Fracture Zone; CFZ : Charlie-Gibbs Fracture Zone; GB: Grand Bank. Modified from Coffin 1992.



Sinemurian/Pliensbachian limit (190 Ma)
Mohican–Iroquois times

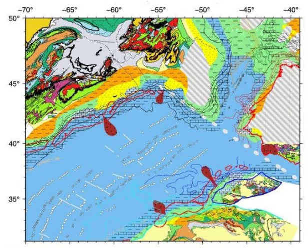


Middle Bajocian (170 Ma)
Scatarie time

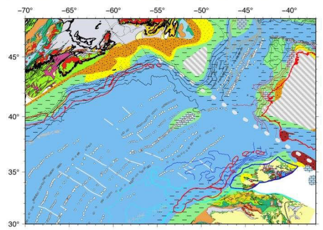


Tithonian (150 Ma)
Mic Mac time

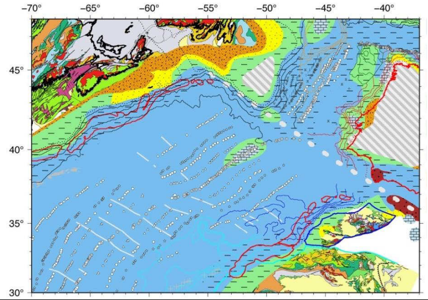
Avalon uplift 145-135 Ma



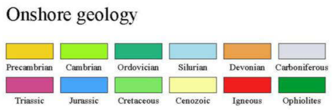
Valanginian (136 Ma)
Lower Mississauga time



Late Barremian-early Aptian (125 Ma)
O Marker time



Santonian (83.5 Ma)
Wyandot time



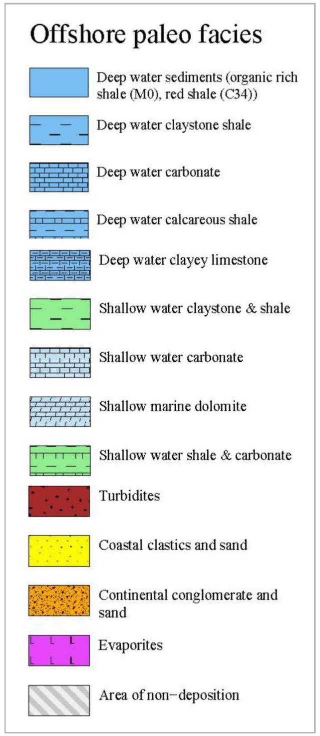
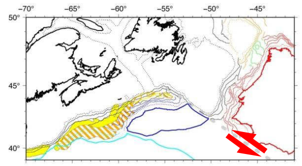
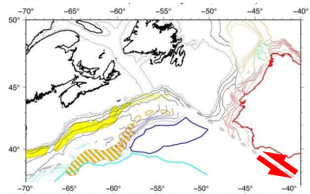


Plate Reconstruction Sibuet et al., 2011, PFA 2011

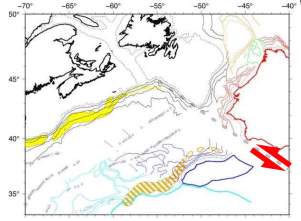
Figure 29a: Offshore paleofacies maps (Sibuet et al. 2012).



Kinematic reconstruction at chron ECMA (end of rifting)
Sinemurian/Pliensbachian boundary 190 Ma
(Upper Eurydice –Argo time)

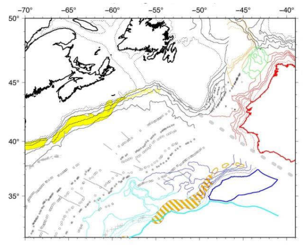


BSMA reconstruction
Middle Bajocian 170 Ma
(Mohican time)

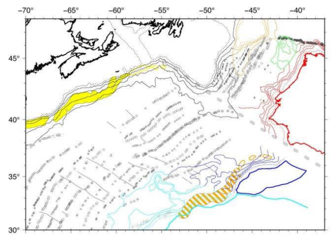


M22 reconstruction
Tithonian 150 Ma
(Mic Mac time)

Avalon Uplift 145-135 Ma



M11 reconstruction
Valanginian 136 Ma(
Lower Mississauga time)



M0 reconstruction
Late Barremian / early Aptian, 125 Ma
(O marker time)

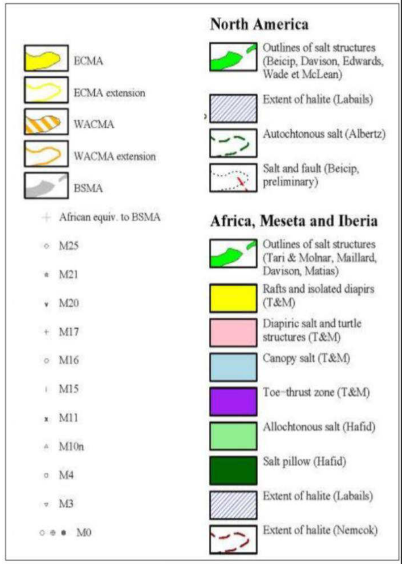
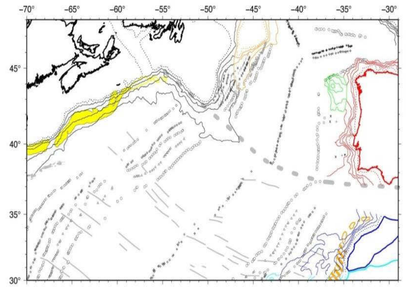


Plate Reconstruction Sibuet et al., 2011, PFA 2011

The NFZ is a left-lateral strike-slip fault between Grand Banks during Early Jurassic



C34 reconstruction
Santonian, 83.5 Ma
(Wyandot time)

Figure 29b: Isochron magnetic anomaly maps (Sibuet et al. 2012).

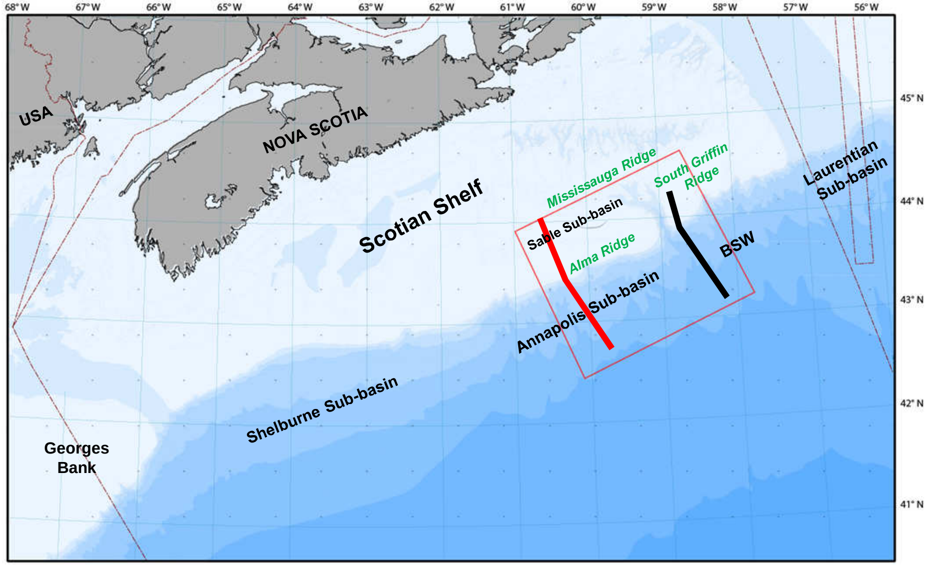


Figure 30 : Study area location. BSW (Banquereau Synkinematic Wedge)

The study area, located mainly on the slope, is on the central part of the Nova Scotia Margin ~~in~~ between the Annapolis sub-basin and the Banquereau Synkinematic Wedge (Figure 30).
This area is characterized by two distinct domains.

The **Annapolis Sub-basin** is a rifted basin, between the Shelburne Sub-basin to the SW and the Laurentian Sub-basin to the NE (Figure 30). Structurally, it is characterized by a complex salt structure with two salt bodies:

- The basal autochthonous salt on top of which Jurassic series were faulted and tilted.
- The upper allochthonous salt body, expelled from the Sable Sub-basin and creeping toward ocean driven by Cretaceous deltaic deposits. The uppermost salt forms large canopies (Canopy province, Figure 31).

The **Sable Sub-basin** is a NE-SW intra-shelf rifted basin, bounded by two ridges: the Mississauga Ridge to the north and the Alma Ridge to the south. This basin is the source of the allochthonous salt expelled in the Annapolis Sub-basin. The autochthonous salt layer in place ~~moves~~ develops into diapirs in the overall basin (Figure 32).

The **Banquereau Synkinematic Wedge** is located between the Annapolis and the Laurentian Sub-basins. Bounded by the South Griffin Ridge to the north, it is characterized by a number of piercing diapirs and a large Jurassic landslide gliding on top of an allochthonous salt tongue (Shimeld 2004; PFA 2011; Figure 32).

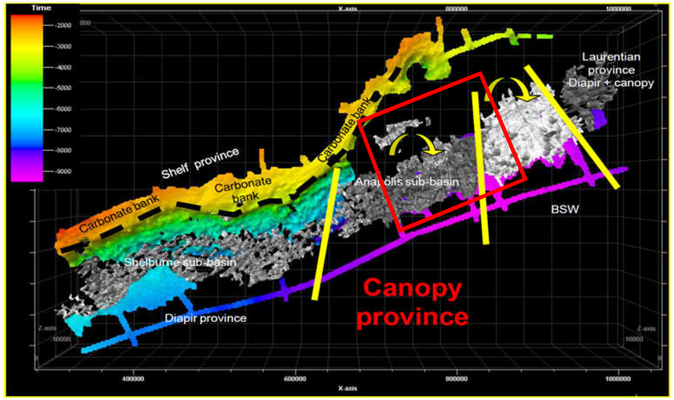


Figure 31: Scotian Margin Zonation (from PFA, 2011)

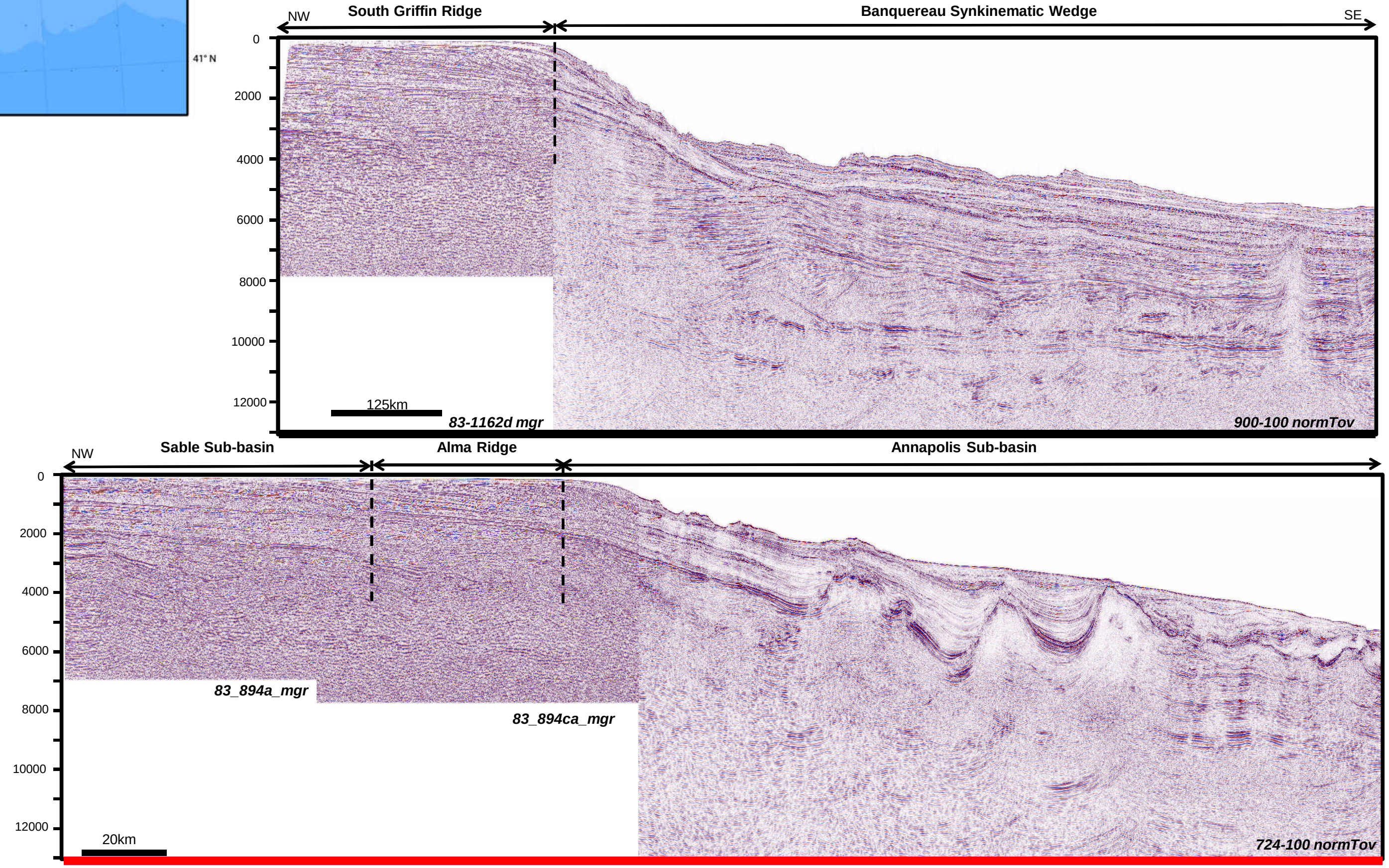
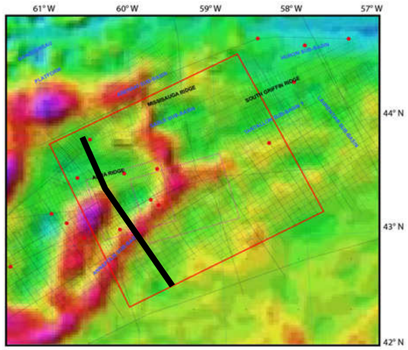


Figure 32: Morphologic and tectonic characteristics of the study area: a) in the Banquereau area b) in the Annapolis Sub-basin.



Along this seismic line different key structures are observed:

- a growth faulting set related to autochthonous and also to allochtonous salt.

- On the Jurassic, landward tilted blocks related to gliding of autochtonous salt whereas on the Cretaceous oceanward tilted blocks related to the allochtonous salt: "fishbones architecture".

- A roho system with autochtonous salt feeders and canopy with allochthonous and autochtonous salt connection.

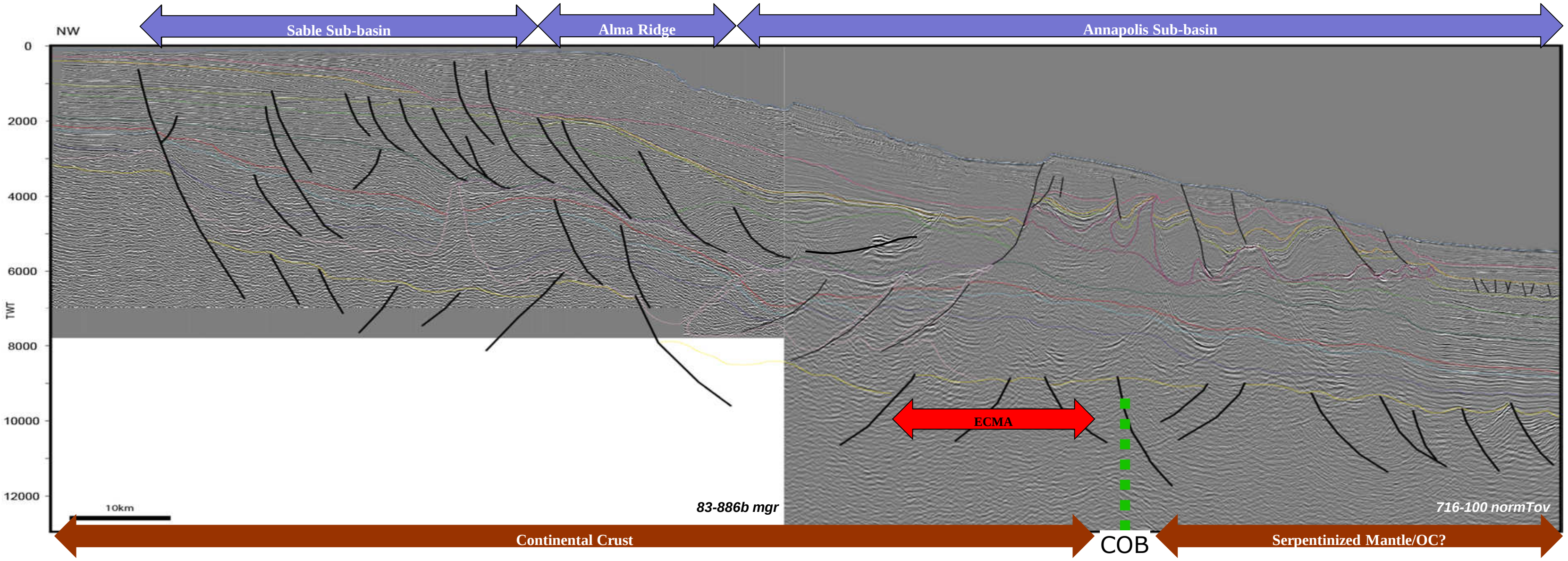
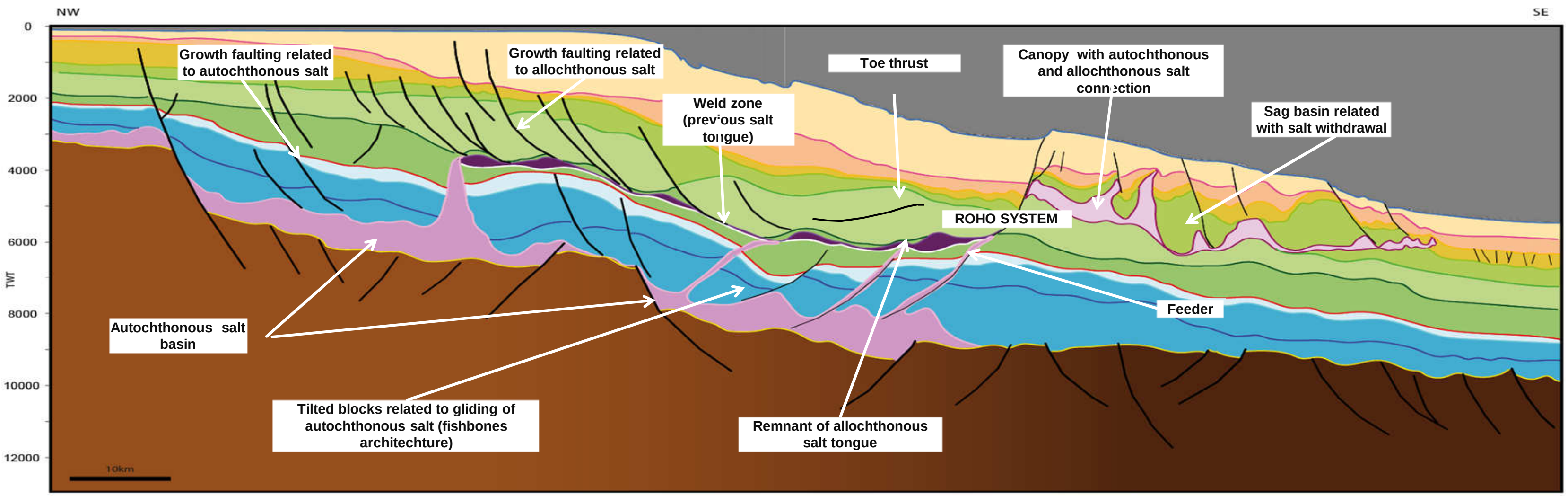


Figure 33: Seismic line across the SW of the study area in the Annapolis Sub-basin .



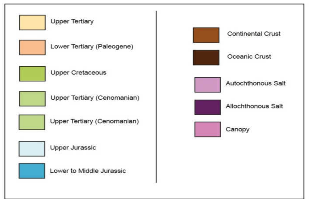
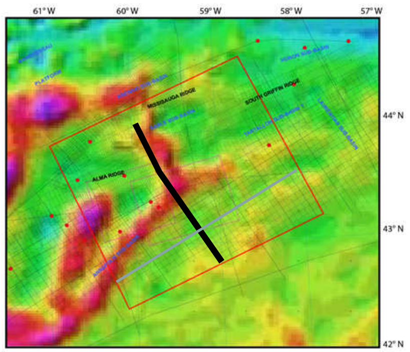


Figure 34: Interpreted seismic line across the SW of the study area in the Annapolis Sub-basin.



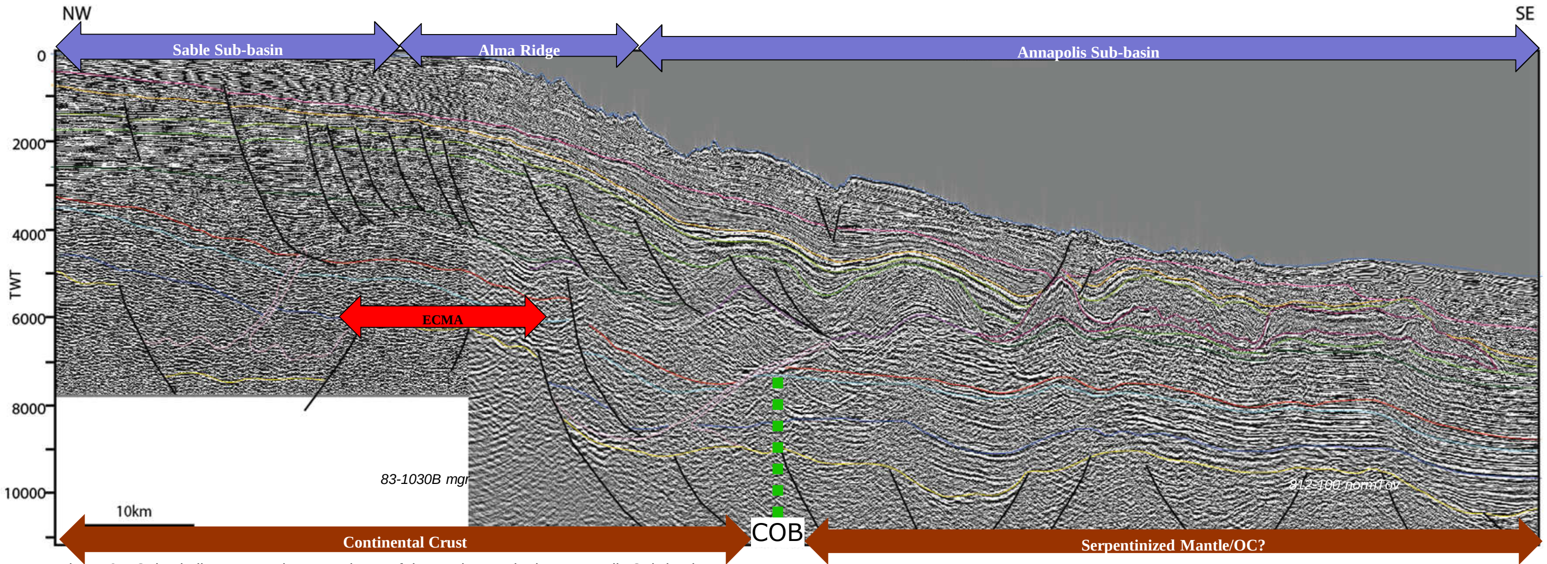
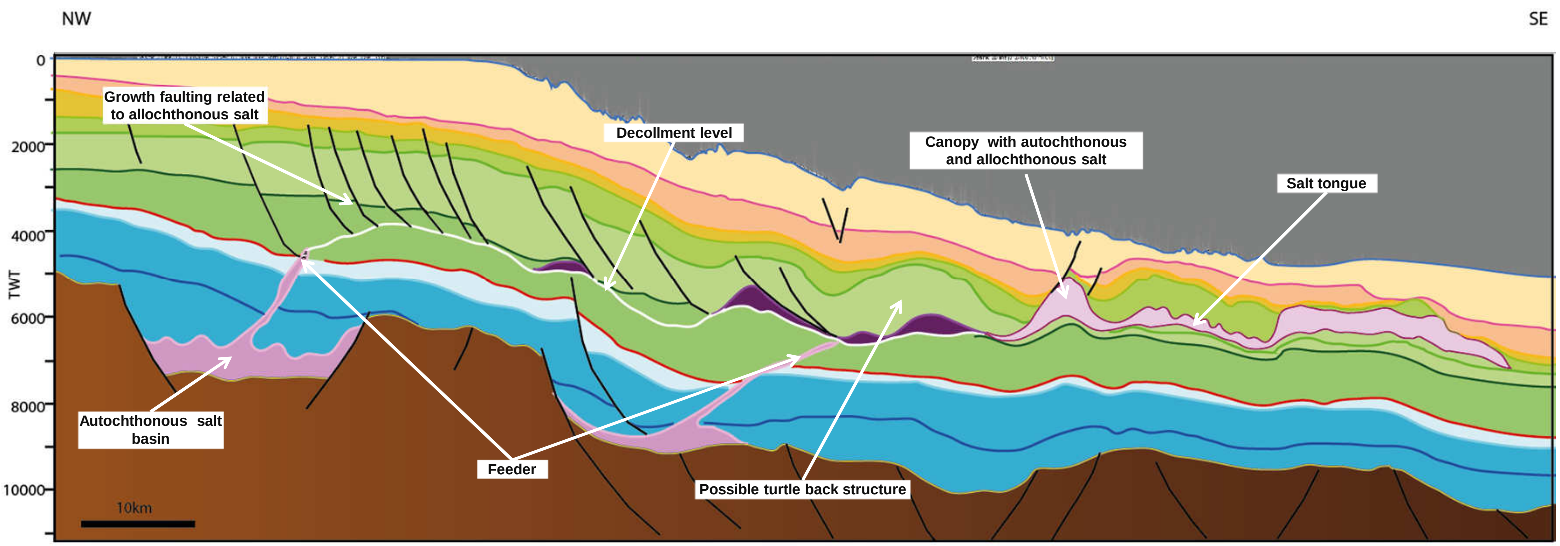


Figure 35: Seismic line across the central part of the study area in the Annapolis Sub-basin.

Along this seismic line different key structures are observed:

- A growth faulting set related to allochtonous salt in the Cretaceous interval rooted on a decollement level.
- A synkinematic wedge on the slope in the Cretaceous interval with extension in the upper part and compression in the down part (turtle back structures).
- A roho system with autochtonous salt feeders and canopy with allochthonous and autochtonous salt connexion



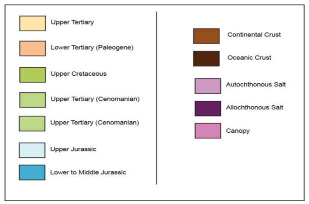
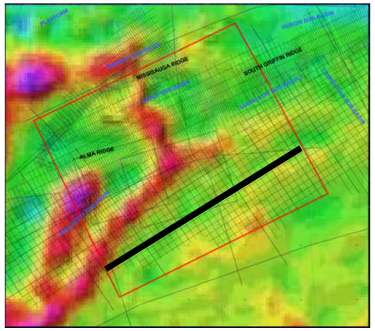


Figure 36: Interpreted seismic line across the central part of the study area in the Annapolis Sub-basin.



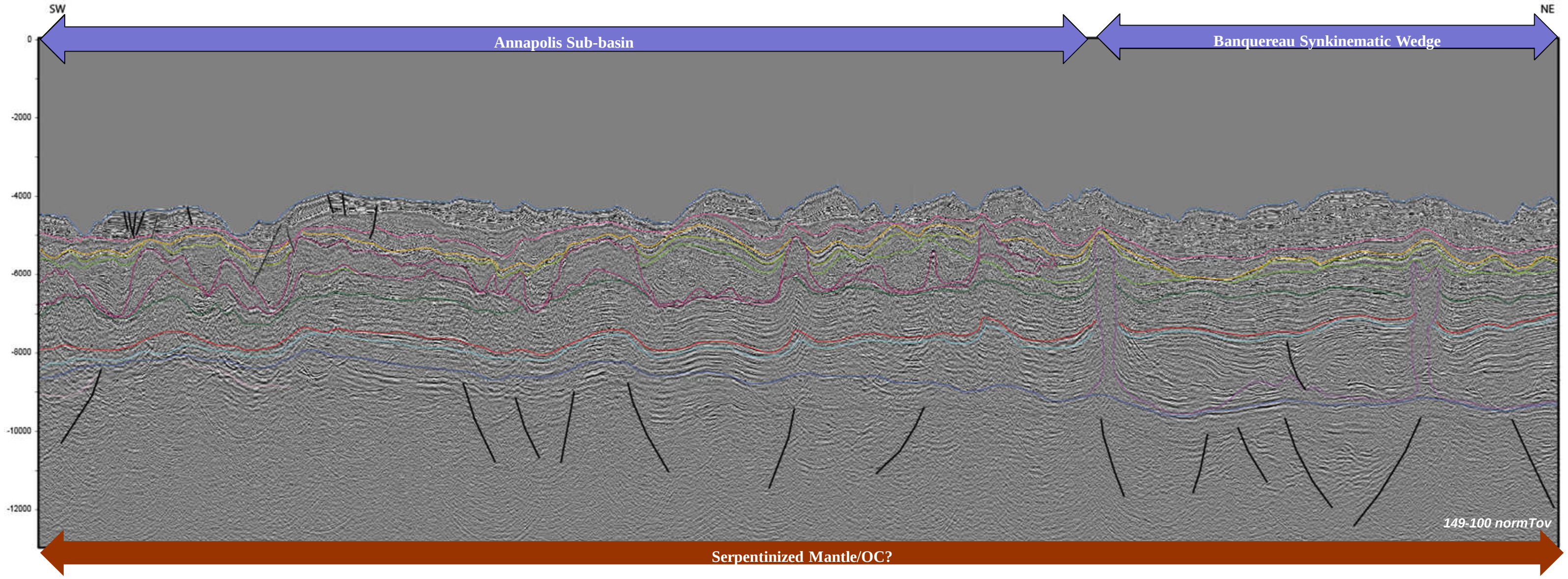


Figure 37: Longitudinal seismic line across the study area from Annapolis Sub-basin to BSW.

Along this seismic line different key structures are observed:

- two salt layers: the autochtonous salt located on the top of the continental crust and the canopy within the allochtonous salt on the Cretaceous interval.

- The NE part is characterized by the Banquereau Synkinematic Wedge (BSW) which is bounded by diapirs on each side.

- A structural high oriented NNE-SSW corresponds to the boundary of the BSW on the western side (also highlighted by the diapir position).

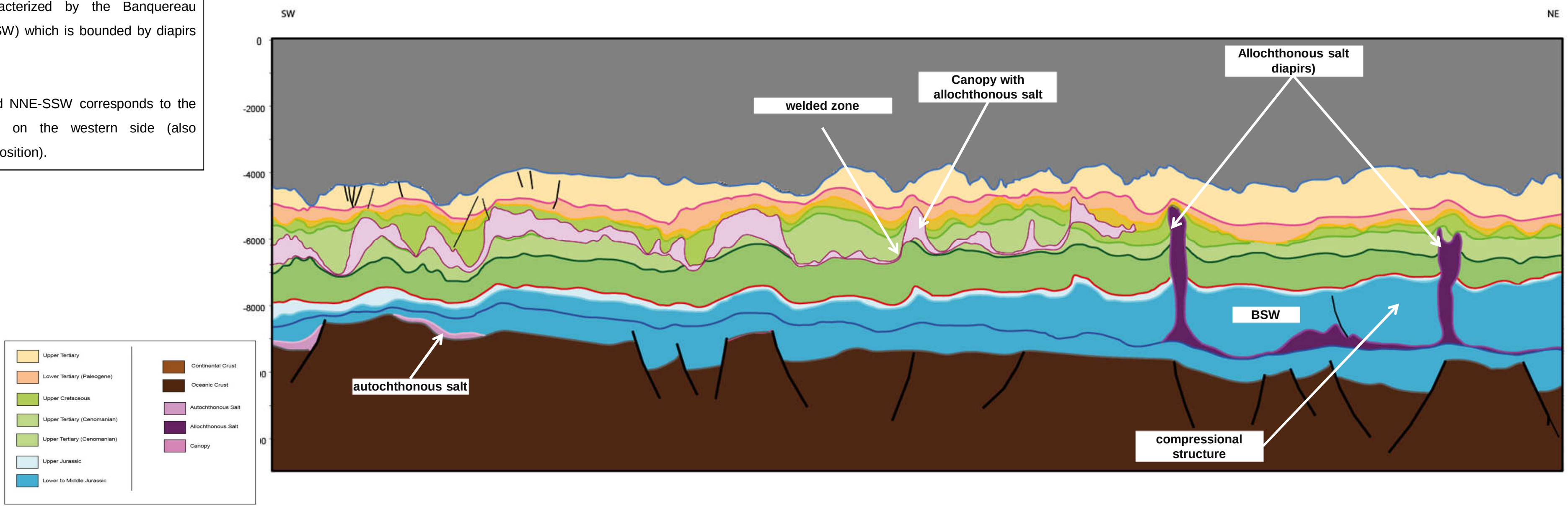
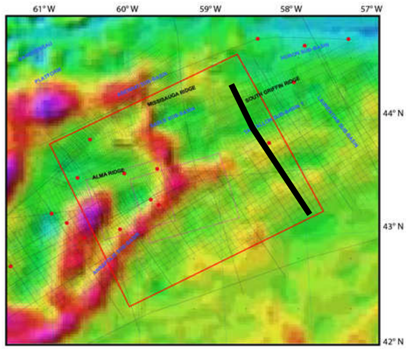


Figure 38: Interpreted longitudinal seismic line across the study area from Annapolis Sub-basin to BSW.



Along this seismic line different key structures are observed:

- A growth faulting set related to autochtonous and allochtonous salts in the Jurassic interval rooted on a decollement level.
- A synkinematic wedge in the Jurassic interval: the Banquereau Synkinematic Wedge (BSW), with allochthonous extensional block on the bottom and compressional structures.

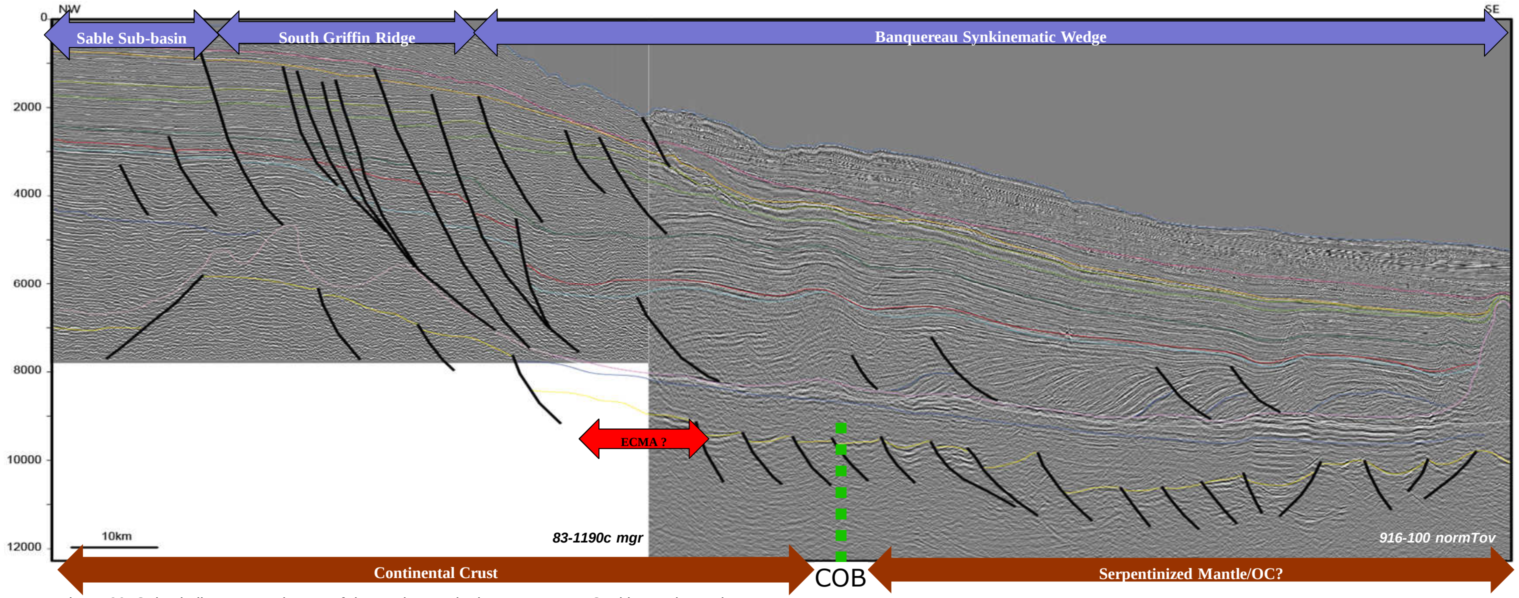
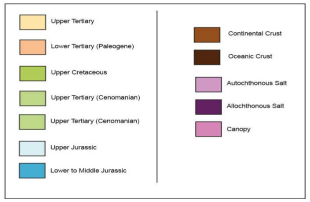


Figure 39: Seismic line across the NE of the study area in the Banquereau Synkinematic Wedge.



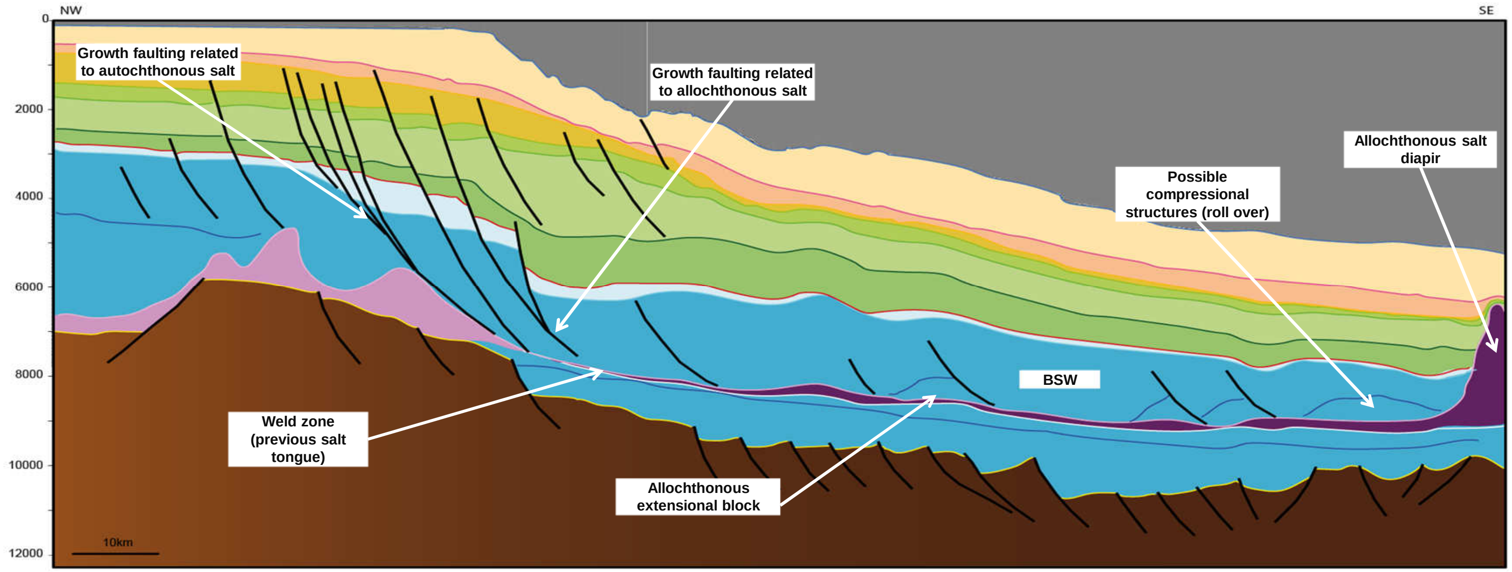


Figure 40: Interpreted seismic line across the NE of the study area in the Banquereau Synkinematic Wedge.

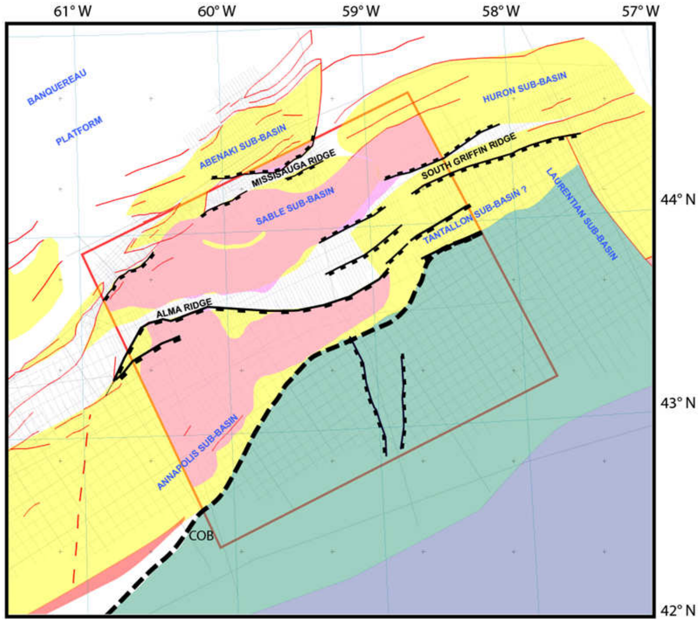
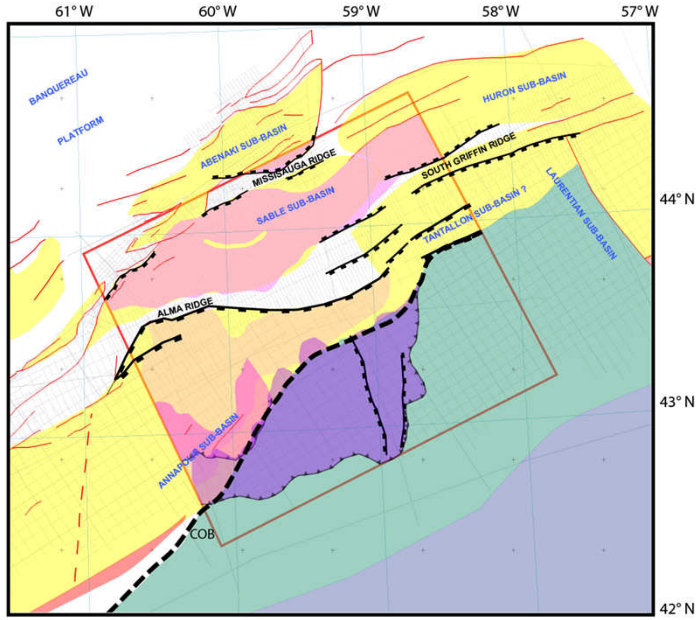


Figure 41: Structural sketch map of the study area showing the main crustal elements and autochthonous salt basins.



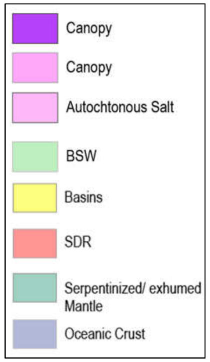


Figure 42: Structural sketch map of the area showing the main allochthonous salt occurrences.

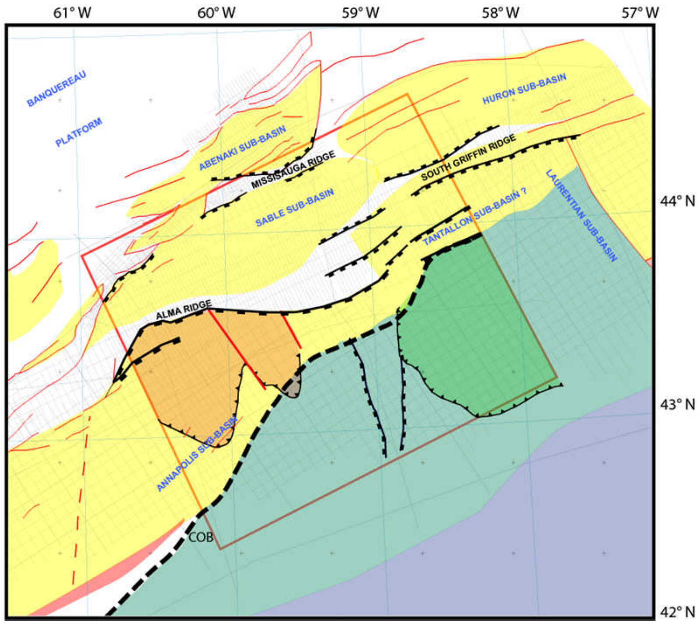


Figure 43: Structural sketch map of the area showing the main synkinematic sedimentary wedges (Balvenie wedge in orange and Banquereau wedge in green).

➢ Structurally, three main domains are observed (Figure 41):

- **The continental crust**, where the NE-SW rifted basins (Sable Sub-basin, Annapolis Sub-basin, etc.) are located and which are separated by ridges (Alma, South Griffin, etc.).
- The **serpentinized/exhumed mantle**, bounded to the NW by the Continent-Ocean Boundary (COB). Structurally, this domain is affected by small normal oceanward faults (trending NE-SW). In the central part of the study area, between the Annapolis Sub-basin and BSW, two normal faults have been identified with a different trend: NNW-SSE (Figure 38). They seem to be the SW boundary of the BSW (Banquereau Synkinematic Wedge).
- The **oceanic crust** in the deepest part. It is difficult to determine its actual landward extension due to the Jurassic Quiet Magnetic Zone (Figure 41). The oldest known age corresponds to the magnetic anomaly M25, Tithonian (148.5 Ma).

➢ Two salt domains (Figure 42):

- **Autochthonous salt** located in the rifted basins of Sable and Annapolis. In the Annapolis Sub-basin, it is bounded to the south by the ECMA (Figure 33, plate 2.14). Its gliding on top of a paleosalt tongue created tilted blocks in the underlaying Jurassic series.
- **Allochthonous salt** expelled from the Sable Sub-basin area and squeezed oceanward. This uppermost salt forms large canopies locally connected by several feeders to the autochthonous salt: **Sable Canopy** (true canopies, Figures 33 & 34 plate 2.14). In the Sable Sub-basin area the uppermost salt has been almost completely expelled from its original province and is only connected by remnant weld feeders so called: **Balvenie Roho System** (Deptuck et al. 2009; Deptuck, 2010a, 2010b; Kendell and Deptuck, 2010). On the shelf and shelf break area, the structure is controlled by listric faults rooted either in the basal autochthonous salt or in the uppermost salt body.

➢ Two distinct wedges (Figure 43):

- **Banquereau Synkinematic Wedge**, in the north-eastern side and bounded by a structural high (NNW-SSE faults in serpentinized domain). This wedge is actived between the J163 and J150.
- **Balvenie Synkinematic Wedge**, in the south-western side. It is bounded by a detachment level above the K137 at the bottom, by the K94 at the top and by the Sable Canopy oceanward.

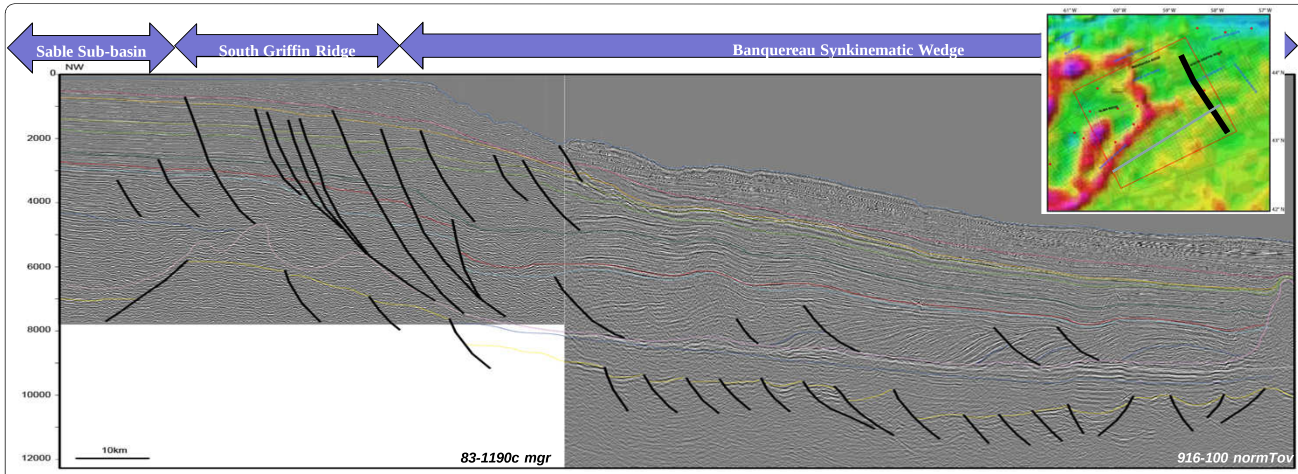


Figure 44: Seismic line across the Banquereau Synkinematic Wedge (BSW).

This province was described in detail by Ings and Shimeld (2004), PFA (2011) and Deptuck et al. (2014). It corresponds to a "mega landslide" gliding event during the Upper Jurassic.
This gliding developed on top of a salt tongue expelled from the Sable Sub-basin due to sedimentary loading from the paleo Laurentian River (Figure 44).

Based on the different studies and analyses of this wedge, a series of sketches has been proposed to explain its formation.
At the end of rifting, a salt basin forms on both the Canadian and Moroccan Margins (Figure 45a).

From the Lower to Middle Jurassic, salt begins to creep from the Sable Sub-basin into the Banquereau area. The salt progrades oceanward on the paleoseafloor (Figure 45b). A set of listric faults are initiated in the Sable Sub-basin and on the South Griffin Ridge.

Salt continued to be expelled from the Sable Sub-basin during the Late Jurassic and propagate oceanward on the seafloor. Due to active deltaic sedimentation, an overload is created. Thus, a system of listric growth faults (rooted to the salt detachment tongue) is created and the sediments of the Lower to Middle Jurassic begin to move downward (Figure 45c). Some allochthonous blocks (rafts) of Lower to Middle Jurassic age are trapped in the distal part of the wedge (Figure 45c).

This wedge ceased during the Tithonian (150Ma), possibly due to a decreased sedimentation rates during the Valanginian. The growing fault system on the shelf break is still active since the salt continues to creep (diapirism) (Figure 45d & e).

During the Tertiary (Eocene), the activity of growth fault systems on the shelf break progressively decreases and ceases (Figure 45f).

Neogene	Hauterivian	Salt (Late Trias-Early Jur.?)
Paleogene	Tithon.-Berrias-Valang.	
Late Cretaceous	Late Jurassic	
Barrem.-Apt-Albian	Early-Middle Jurassic	

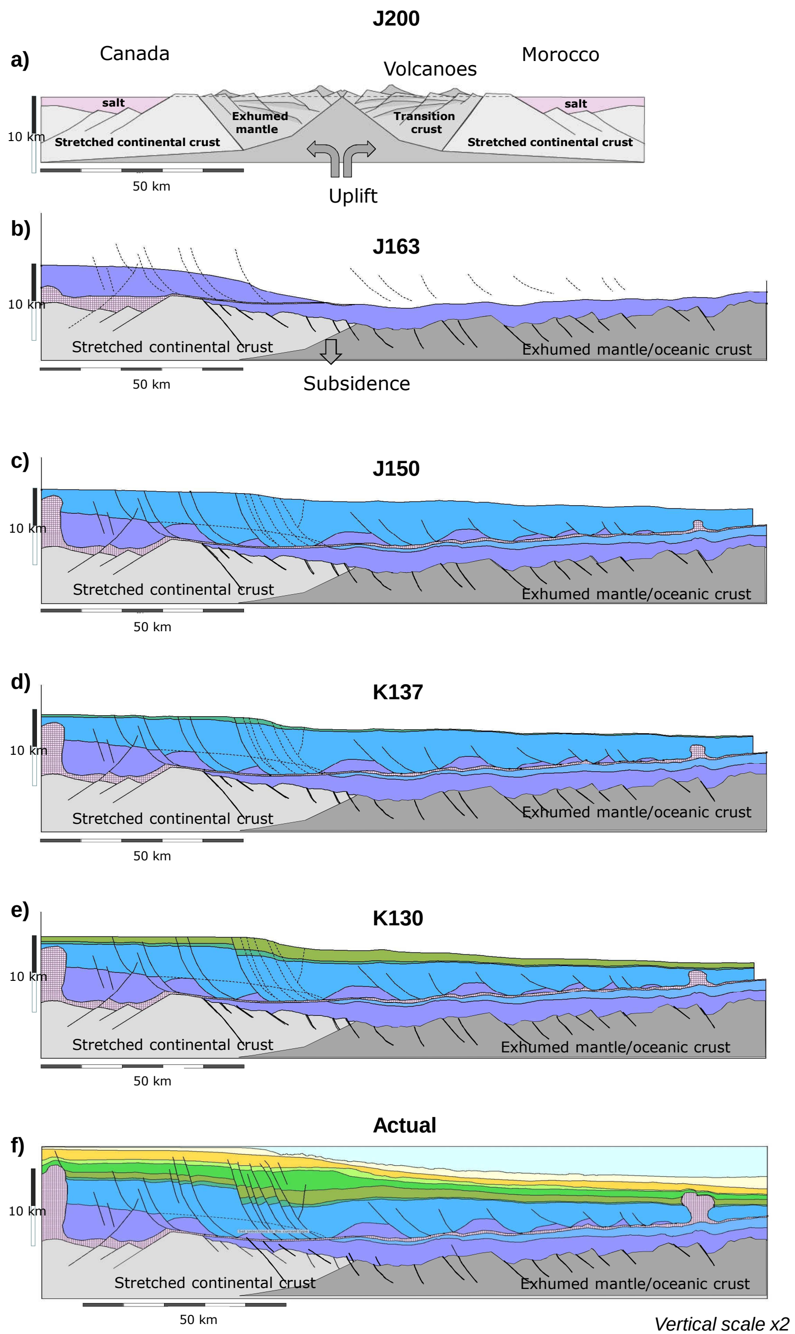


Figure 45: Balanced cross-sections showing the evolution of the Banquereau Synkinematic Wedge.

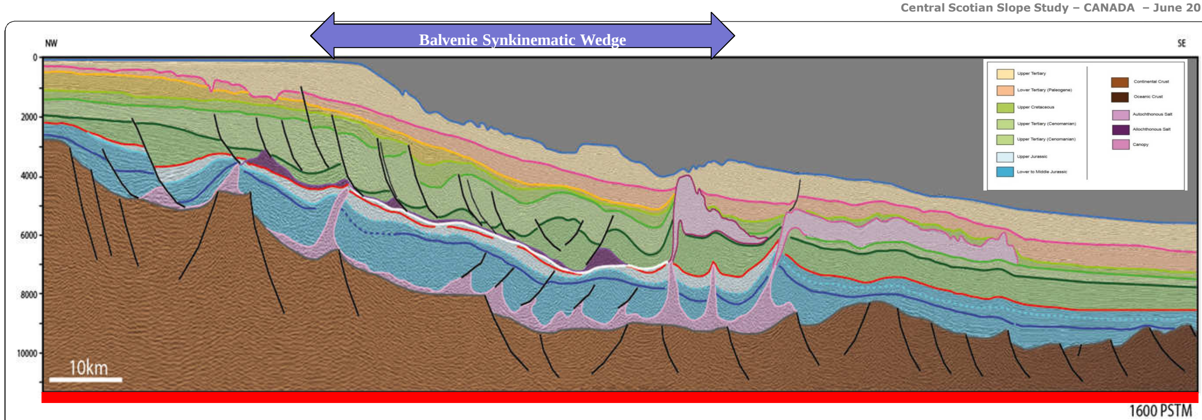


Figure 46: Dip line across the Balvenie Synkinematic Wedge.

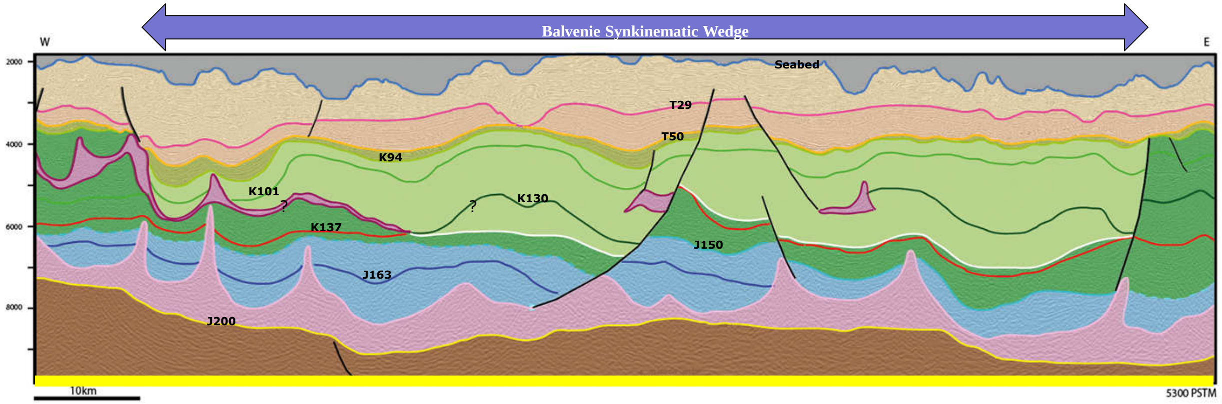


Figure 47: Longitudinal line across the Balvenie Synkinematic Wedge.

The Balvenie Synkinematic Wedge (BaSW) is a part of a wider system described by Kendell et al. 2012 and Pe-Pipper et al. 2015: the **Balvenie Roho System** (Figure 34, Plate 2.14).

This wedge is located on the western side of the study area and formed during the Lower Cretaceous. As with the BSW, it resulted from salt creeping from the Sable Sub-basin in a time of high sedimentation rates.

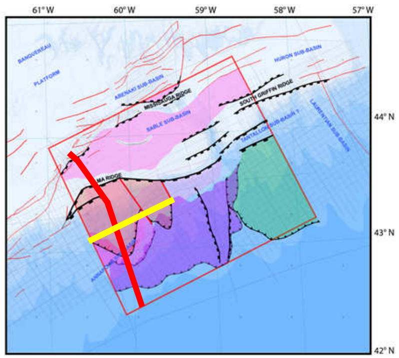
Thus during the Valanginian (after 137 Ma), salt is expelled from the Sable Sub-basin over the Alma Ridge onto the paleo seafloor of the Annapolis Sub-basin. A set of listric growth fault is initiated (Figure 46).

In contrast to the BSW, the detachment level is also powered by feeders of autochthonous salt from the Annapolis Sub-basin. All this salt (allochthonous and autochthonous) is pushed away and creates a canopy system in front of the wedge: the Sable Canopy (Figures 46 & 48).

Structurally, the BaSW is divided into two domains:

- Extensional in the upper part with listric faults rooted on the detachment level along the shelf break (Figure 46).
- Compressional in the lower part with toe thrust or roll over structures (Figure 46).

Moreover, the wedge is affected by strike-slip faults which create compartments (Figures 47 & 48). These faults might be inherited from basement created during rifting and drifting stages.



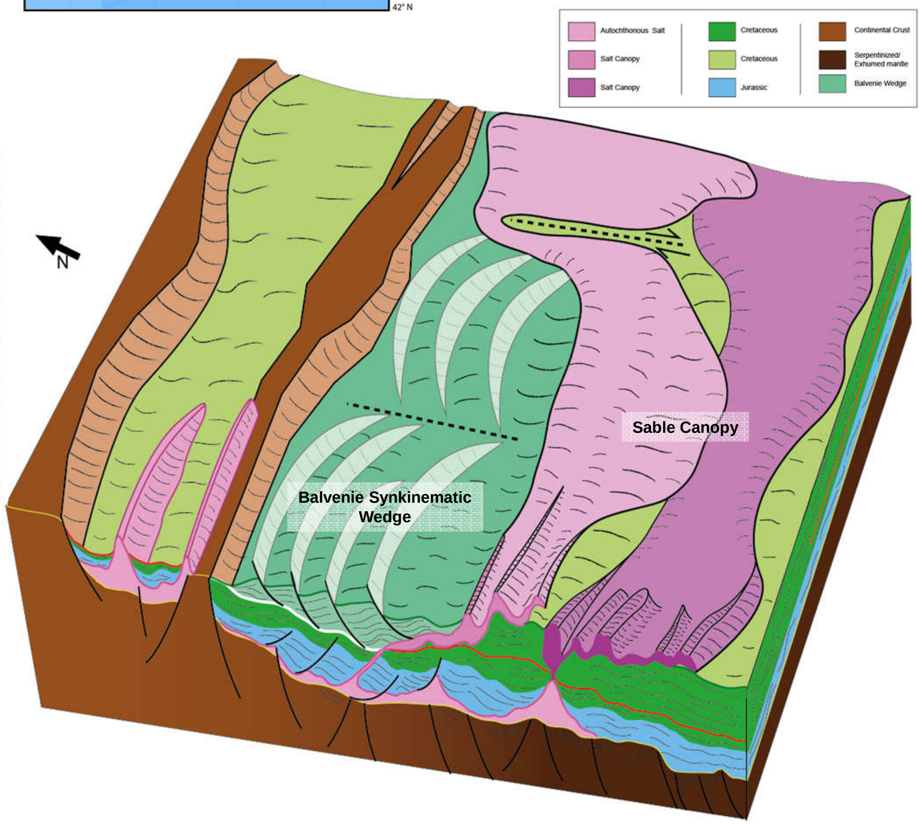


Figure 48: 3D block diagram showing geometry and position of the BaSW with the Sable Canopy.

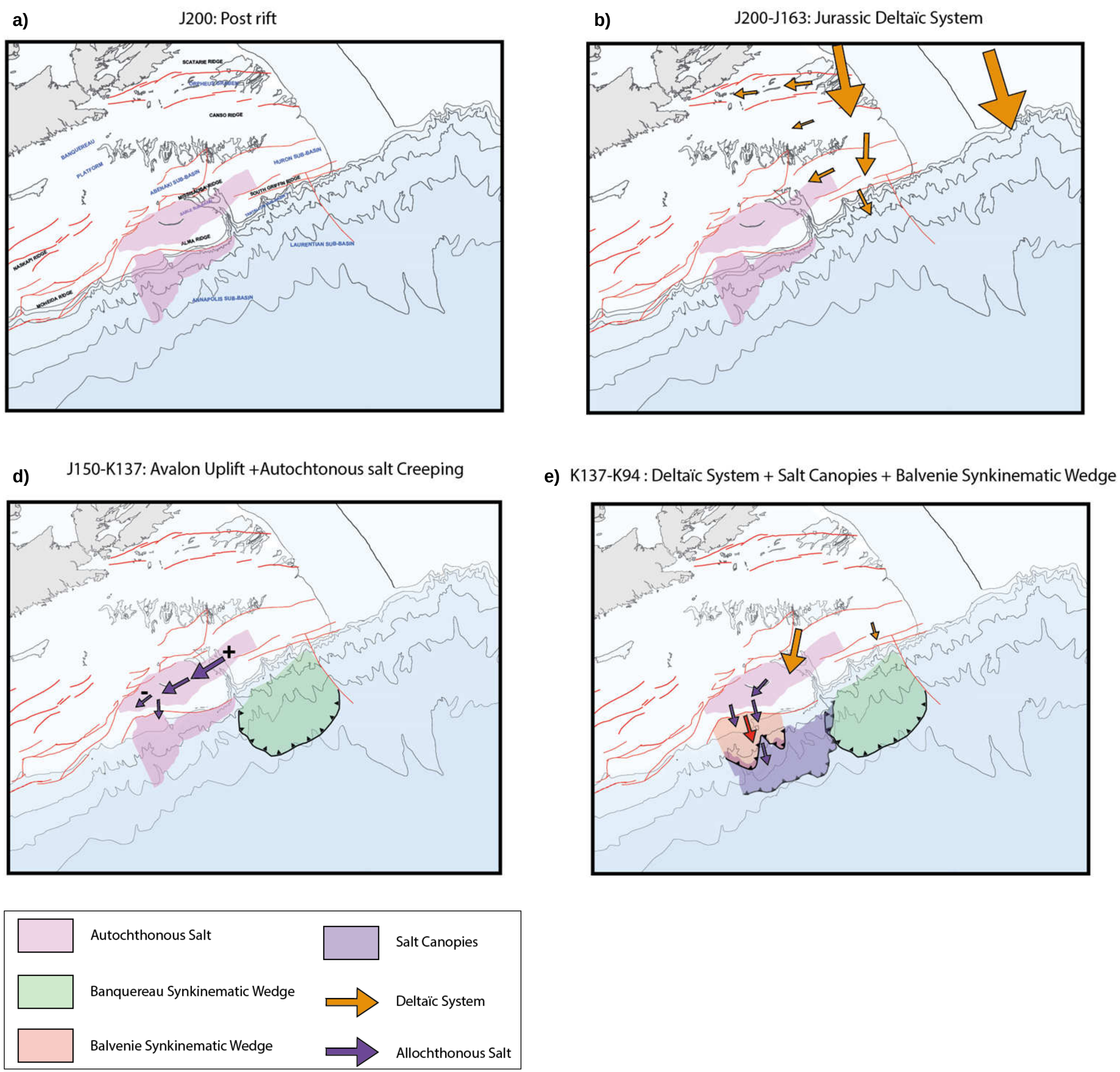


Figure 49: Tectonic evolution of the study area from the Early Jurassic to Late Cretaceous

At the Early Jurassic (200Ma) the rifting ceased and the Nova Scotia Margin began to subside. The Argo salt is deposited in the rifted basins of Sable and Annapolis (Figure 49a).

At the same time, a deltaic system is established with primary sedimentary input from the NE. The deposits are mainly localized in the Sable Sub-basin and the Banquereau area (Figure 49b). As described in the plate 2.19, the salt begins to be expelled from the Sable Sub-basin toward oceanic domain, over the South Griffin Ridge.

During the Upper Jurassic (from Oxfordian to Tithonian) the deltaic system continues to provide sediments as salt creep continues. The sedimentary deposit becomes unstable and the Banquereau Synkinematic Wedge forms (Figure 49,c).

From the Late Jurassic to the Early Cretaceous and during the Avalon Uplift (Valanginian), the autochthonous salt in the Sable Sub-basin continues to glide following a NE-SW gradient (Figure 49d). This event is characterized by a strong diapirism and growth fault activity.

From the Lower Cretaceous (Valanginian) to early Late Cretaceous (Cenomanian), the salt starts to overpass the Alma Ridge and spread into the Annapolis Sub-basin (Figure 49e). Thus, a listric fault system begins to develop. The autochthonous salt also begins to creep and merge with allochthonous salt through feeders.
In the late Early Cretaceous (Albian) the Banquereau Synkinematic Wedge forms as the Sable Canopy develops (Figure 49e).

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