Smooth velocity model from traveltime tomography of an offshore dense wide-angle profile in Nova Scotia: preliminary results

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Abstract

Understanding the mechanisms that generate rifted continental margins is important in several aspects: (1) scientific - how continents break apart; (2) geopolitical - delimitation of the territorial sea through the United Nations Convention on the Law of the Sea (UNCLOS, United Nations Convention on the Law of the Sea) and; (3) economic, de-risking of oil and gas exploration. The goal of the work proposed here is to prestack migrate in a joint fashion the OETR-2009 OBS and NovaSpan 2000 MCS data. To the best of our knowledge, this would be the first attempt to carry out this type of data analysis and we anticipate significant improvements in the imaging with this approach, in particular of the crystalline basement, crust and the Moho. In order to construct an accurate and detailed but smooth velocity model needed for joint prestack migration of the OBS and MCS data, we will first run a detailed tomographic inversion of the OETR profile. The proposed work will show how velocity modeling of detailed wide-angle refraction observations (such as the OETR profile) can significantly improve the definition of the lower (deep) sediment and basement geometries. We also hope that our results will have a major impact on the sighting of future 3D reflection surveys and exploration wells. The preliminary result of smooth velocity model shows: the depth of the Moho discontinuity is estimated to be ~33 km in the continental crust over a distance of ~60 km, rapidly rising to ~24 km at the distance of ~100 km and, gradually, rising to ~13 km at 280 km within the oceanic crustal domain, with the same depth extending until the end of the model. Between 10-80 km distance, a LVL–HVL–LVL (LVL=Low Velocity Layer, HVL=High Velocity Layer) sequence is observed. In the OCT, between 200-260 km distance, there is a section above the Moho which has a strong gradient and high velocity in compared to normal crust velocities.

Introduction

Deep-penetration controlled source seismic data are essential for constraining the structure of a transition between the continental and oceanic crust, as are appropriate analysis of these data and subsequent result interpretation. Supporting information used for constructing a structurally reliable seismic model of an OCT (ocean-continent transition) are gravity, magnetic and well data. Recent practice of collecting coincident wide-angle ocean-bottom seismometer (OBS) and long hydrophone streamer multi-channel seismic (MCS) profiles, allows for imaging to greater depth, beyond the sediment section and through the crust all the way into the uppermost mantle.

When continents rift apart and new ocean basins form, a transition region is created between the old, thick continental crust and young, thin oceanic crust [Louden and Lau, 2001]. Based on the amount of volcanism that occurs during rifting, two primary classes of this transitional region at continental margins can develop, volcanic and non-volcanic [Louden and Lau, 2001; Keen and Potter, 1995]. Standard crustal scale MCS reflection and OBS refraction projects were carried out during the past couple of decades to better understand the ocean-continent transition (OCT) found offshore eastern Canada (Figure 1) [e.g., Funck et al., 2004; Wu et al., 2006; Louden, 2002]. The formed reflection images and velocity models [Louden and Chain, 1999; Louden and Lau, 2001; Keen and Potter, 1995; Funck et al., 2004; Wu et al., 2006; Louden, 2002, Van Avendonk et al., 2006; Gerlings et al., 2012] show significant basement structural variations between the profiles that are, together with potential field data, interpreted to indicate a volcanic-type margin offshore southern Nova Scotia, a non-volcanic margin offshore central Nova Scotia, and an extremely amagmatic margin further north. These zones of transitional crystalline crust appear to coincide with northward transitions within overlying sedimentary structures, from salt-free to salt bodies in the south, and from autochthonous salt diapirs to allochthonous salt tongues in the north [Hansen et al., 2004].

The existing knowledge about the crystalline basement structure offshore Nova Scotia (Figure 1) has, so far, been gained using sparsely spaced (100s of km) regional seismic profiles with sparsely spaced (~20 km) OBS instruments. New breakthroughs in our understanding of crustal structures offshore Nova Scotia will require new 2D OBS data with closer profile spacing (10s of km) and/or smaller instrument spacing (1-5 km) [Lau et al., 2015; Watremez et al., 2015]. Densely spaced, high-resolution OBS profile data will also require additional, more sophisticated data analysis such as are waveform tomography and prestack depth migration. Both the new data and the advanced data analysis are needed to resolve deep structures with sufficient resolution to answer first order questions such as is: What is the detailed nature of the OCT, which can exhibit complex structures, including highly extended continental crust, exposed mantle and thin oceanic crust that are difficult to distinguish with standard data and data analysis methods. Detailed imaging of these structures is critically important for an improved understanding of such widely-held concepts as the continent-ocean boundary and the breakup unconformity.

The style of rifting (volcanic or non-volcanic) also has a direct implication on the rift infill and paleo-water depth at the onset of drifting [PFA Atlas, 2011, PFA – Play Fairway Analysis]. In magma rich conditions the rift zone is uplifted with continental and shallow marine conditions prevailing, thus allowing for the deposition of confined marine source rock. In magma poor conditions the stretching of the lithosphere induces mantle exhumation and rapid subsidence resulting in 1-2 km deep marine conditions that preclude the deposition of shallow marine source rock. The presence of such a source rock can have a direct impact on the petroleum system. The general consensus among academic researchers is that the transition from magmatic to amagmatic rifting occurs offshore south-central Nova Scotia. This view is consistent with the absence of any direct evidence for a magma rich margin, such as a continuous belt of seaward dipping reflectors (SDR) or a continuous and homogeneous East Coast Magnetic Anomaly (ECMA). However, reinterpretations by consulting companies for the PFA Atlas [2011] have found several features that they believe strongly suggest that the magma rich Camp Province could extend north up to the Newfoundland-Azores transform fault. In this interpretation, the entire Nova Scotia margin could be considered as a magma rich passive margin. This suggestion indicates that there is an urgent need for detailed data analysis of the existing OBS profiles in order to derisk the petroleum exploration offshore Nova Scotia. The research presented here, using data from OETR (Offshore Energy Technical Research) refraction profile (Figure 1), directly addresses this need and will, hopefully, resolve the ongoing debate.

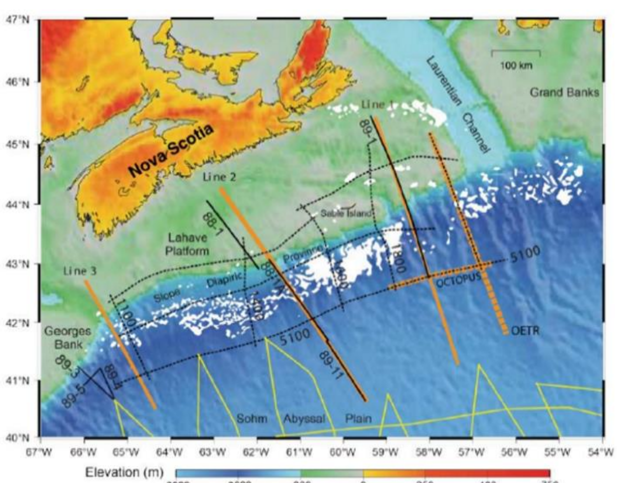


Figure 1 – Location of regional seismic lines offshore Nova Scotia. Seismic refracion profiles are marked by orange lines, dashed for OETR (NW-SE) and OCTOPUS and solid for SMART Lines 1-3 and. Regional MCS reflection lines are shown as solid black lines (Lithoprobe and Frontier Geoscience Program), dashed black lines (NovaSpan Project of ION-GXT) and yellow lines (UNCLOS Project). Jurassic salt over the Slope Diapiric Province is indicated by white patches.

Method

The purpose of tomographic inversion is detailed exploration of the velocity pattern of a medium. Such exploration is based on first arrival traveltimes obtained from a set of source-to-receiver pairs. Any geometry of sources and receivers may be considered. In our case, this was a 400 km long profile that permitted recording of refracted or diving waves and reflected waves from great depths. It is essential that the rays penetrate all parts of the area that we wish to map seismically and that the rays form a complete net, crossing all parts of the model we wish to penetrate.

Tomographic inversion requires as input picked arrivals and an initial velocity model. The following consists of two main steps: 1) solving the direct problem; 2) solving the inverse problem. While fundamentally different, both Rayinvr and Tomo2D software using these two steps. The aim of the first step is the computation of arrival traveltimes/amplitudes and corresponding raypaths. Traveltime residuals (i.e. differences between observed traveltimes and computed ones) are input information for the second step. A traditional way for solving the inverse problem is to divide the investigated region into cells and to find perturbations of the initial model, provided the perturbation for each cell is constant. Adding these perturbations to the initial model, one obtains the refined model which reduces the traveltime residuals.

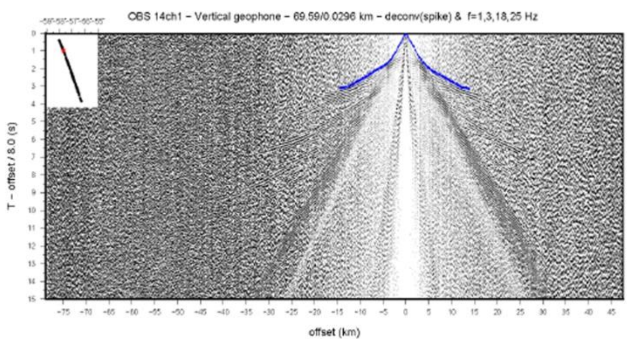
The main difference between Rayinvr and Tomo2D are the input data (the picks) and the initial/resulting models of each software. Rayinvr allows the construction of a structured (layered) initial model, both in the definition of the layers and in the definition of the grid of velocities. It is possible to define floating reflectors within the layers and also allow jumps in the velocity gradients (velocity discontinuities), for example a layer can have a velocities between 3-5 km/s and the underlying layer starts with a velocity of 6 km/s. In addition to first arrivals, Rayinvr relies on secondary arrivals (reflections) from various interfaces. Unlike Rayinvr, Tomo2D requires a smooth model without velocity discontinuities (layer interfaces). Only one reflection interface is allowed but even this is a floating reflector (there is no velocity discontinuity across this boundary). Only first arrivals (direct waves and refractions) can be used with an exception of reflection from a signle interface. While Tomo2D seems to be limited relative to the Rayinvr sofware, obtaining smooth minimum structure velocity models using only first arrivals, as is the case with Tomo2D, is critical because (a) this result is less interpretative than the one obtained by Rayinvr and therefore more objective and (b) only this type of a result can be used for prestack depth migration of reflection signals.

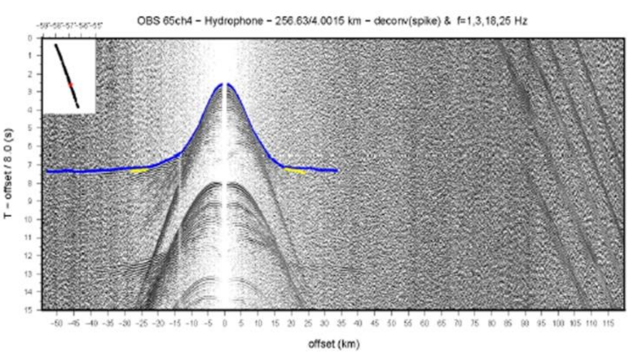
DATA AND DATA ANALYSIS

The data were collected as part of the PFA by GeoPro GmbH under contract to OETR along the pre-existing MCS reflection profile ION/GXT NovaSPAN-2000 (Figure 1). The data are stroed in SEGY format and each OBS has four components: Ch1-vertical geophone; Ch2 and Ch3-horizontal geophone; Ch4-hydrophone. The OETR refraction profile (Figure 1) is 400 km long and has 100 OBS, mostly spaced only 2-5 km between stations as opposed to the typical spacing for standard surveys of 15-25 km. A few datasets have these characteristics, thus the results are expected to show significant improvements in image resolution and details captured, particularly for the crystalline basement, crust and Moho discontinuity.

The first task was to analise the date quality and headers that contains all position information. This was in part needed in order to evaluate if there is a need to relocate the OBS, a procedure executed in Matlab that uses water depth, water velocity, deployment position, recovery position, and the direct arrivals through water layer picked on OBS data. This analysis confirmed that the original relocation exercise was carried out well except for only six OBS found at the seaward end of the OBS line, which needed improved positons at the seafloor. Data for 22 could not be recovered. However, even with this unusually large loss of data, this seismic refraction profile with 78 four-component OBS recordings is considered to have a very dense coverage.

Figure 2 shows the quality of the data in the OBS records, together with the first arrival picks (refractions with some direct waves at short offsets) and picks of reflection Moho. OBS 14 is along the shelf, OBS 39 within the ocean-contitnet trasition zone, and OBS 65 further seaward. In the seismic section of OBS 14, rapid attenuation of the first arrivals, probably, due to a low velocity layer. This structure results in a velocity inversion. Seismic recordings for OBS 39 show clear first arrivals and several reflections (offsets -20, -35, -45-, -55 km) indicating a complex structure of the crust. In the seismic section of OBS 65, firsrt arrivals are observed with a good signal-to-noise ratio but the reflections from the Moho are not clear.





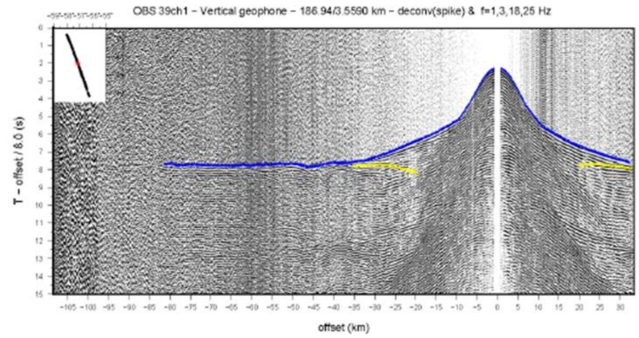


Figure 2 – Seismic sections of OBS 14, 39 and 65. The blue picks are the first arrivals, yellow picks are Moho reflection. In the upper left corner the relative position of the OBS in the seismic refraction profile is shown.

Reflection phases other than the PmP (Moho reflection) can be identified, but the joint refraction and reflection inversion using Tomo2D can utilize only one reflection phase. The PmP arrivals where chosen because they are most important for undersatnding the regional structure in the study area. The total number of seismic arrival picks used in the tomographic inversion was 55,899: 45,272 first arrivals (Pg – crustal refractions, Pn – mantle refractions) and 10,627 PmP arrivals. To stabilize the inversion, it was carried out in stages with picks at larger offsets added gradually.

STARTING MODELS

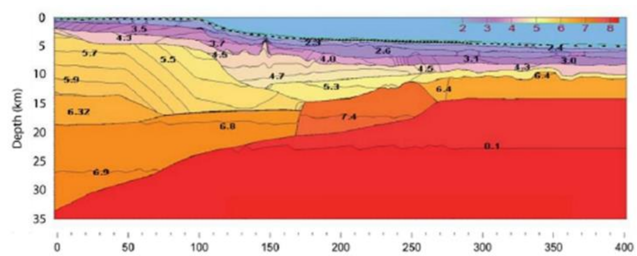


Figure 3 - GeoPro layer velocity model. Numbers in figure are P-wave velocity in km/s. The contour interval is 0.1 km/s. Triangle are OBS.(Luheshi et al., 2012).

The OETR-2009 data, used in this study, were originally used to construct a preliminary layered velocity model called GeoPro by Luheshi et al. [2012] (Figure 3). This model was a reference for the construction of the initial model to be used in Tomo2D. Two initial models were created for Tomo2D input, with the intention that the initial models are as simple as possible and with smooth velocity gradients. The first model (Figure 4), has seafloor velocity of 1.6 km/s and Moho velocity of 7.5 km/s (black dashed line is the Moho). Velocity between the seafloor and the Moho increases linearly with depth.

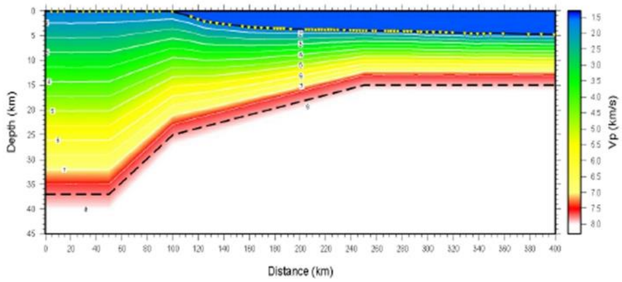


Figure 4 – Initial model, linear velocity gradient increase

TOMO2D INVERSION PROCEDURE

Tomo2D was used to carry out joint tomography inversion of refracted first-arrival traveltimes and later reflection arrivals arising from a single interface (in our case, the Moho) in a sheared mesh model, hanging from the seafloor. Cells are parallelograms with the top and bottom sides parallel to the seafloor and the two other sides being vertical [Korenaga et al., 2000]. Cells in our model are 200-m wide, and their height is 200 m, 45 km beneath the seafloor, with the cell heights linearly increasing with depth. The Moho is modeled as a floating reflector with depth nodes every 200 m. Traveltime tomography using Tomo2D produces a smooth velocity model using minimum a priori information. This means that there are no velocity discontinuities anywhere in the model, including the Moho.

The forward problem in Tomo2D consists of finding the shortest raypath from the shot to the receiver for each arrival time following a hybrid approach that combines graph and ray-bending methods (Moser et al., 1992; Korenaga et al., 2000). We use a tenth-order forward star (Zhang and Toksöz, 1998) for the graph method and a minimum segment length of 1 km with 10 interpolation points per segment for the bending method (Papazachos and Nolet, 1997). Tolerances are 5×10^{-4} s and 5×10^{-5} s for the conjugate gradient and Brent minimization, respectively.

Tomography inversion (the inverse problem in Tomo2D) results in a reduction in the residuals between picked and calculated traveltimes through model updates by perturbation of velocities and depth of the interface nodes, using a least-squares regularized inversion. Parameters for the inversion were chosen, similar to the forward problem, after a full parametric study. The correlation lengths control the smoothness of the model perturbations (i.e., the inversion stability). For the velocity nodes, we use horizontal correlation lengths that linearly increase from 2 to 20 km from the seafloor to the base of the model and vertical correlation lengths that linearly increase from 1 to 10 km from the seafloor to the base of the model. The correlation length for the depth of the interface nodes is set to 6 km. Weighting parameters also control the smoothing and the damping constrains. The depth kernel weighting parameter is equal to 0.5, to favor stronger velocity perturbations relative to the interface depth perturbations. Indeed, we favor strong velocity perturbations over Moho depth perturbations because we have more picks for first arrivals than for wide-angle reflections, and their uncertainties are much lower. Finally, we set the least-squares tolerance to 1 ms.

For each iteration, the computed synthetic traveltimes are compared to the picked traveltimes, and the inversion reduces the misfit between the two sets of traveltimes by perturbing the input velocity model. The forward and inverse problems run until the misfit is reduced to a $\chi2$ (Chi square) of approximately 1. A $\chi2$ of 1 is achieved when the synthetic arrival times are in agreement with the picked arrival times, within their uncertainty range (Bevington, 2003). When the acceptable final model is reached, model resolution and model uncertainties are evaluated using checkerboard tests, restoration resolution tests, and Monte Carlo analysis [Korenaga & Sager, 2012; Korenaga et al., 2000].

Results

Figure 6 (last page) shows the layered Rayinvr velocity model. The depth of the Moho discontinuity at the most landward section of the profile (left side in Figure 6) is estimated to be 33-35 km in the continental crust over a distance of about 50 km. The continental crust rapidly thins to ~23 km at profile distances from 50 to 100 km. From there the Moho discontinuity gradually rises to ~15 km at the distance of 280 km remaning mostly flat further into the oceanic crust domain up to the end of the profile at distance of 400 km. At the continental shelf, between 50-90 km distance and at the depth of ~5 km, a LVL–HVL–LVL (LVL=Low Velocity Layer, HVL=High Velocity Layer) sequence is observed. In the OCT, between 190-270 km, there is a section above the Moho which has a strong gradient and high velocity in compared to normal crust velocities.

More than a hundred models were generated while searching for optimal inversion parameters for the Tomo2D tomography. Figure 7 (last page) shows the lastest model with a very smooth grid of velocities, emphasizing the area where there is ray coverage. It can be observed that even with less detail, compared to the Rayinvr model, the main features are repeated. The depth of the Moho discontinuity is estimated to be ~33 km in the continental crust over a distance of ~60 km, rapidly rising to ~24 km at the distance of ~100 km and, gradually, rising to ~13 km at 280 km within the oceanic crustal domain, with the same depth extending until the end of the model. Between 10-80 km distance, at the depth of ~5 km, the same LVL-HVL-LVL feature is present below the continental shelf. In the OCT, between 200-260 km distance, a high velocity structure that could be correlated to Rayinvr model shown in Figure 6 is present.

Conclusions

The careful and detailed layered modeling using Rayinvr software and carried out in this work (Figure 6) shows that the original layered modeling carried out by GeoPro (Figure 3) likely has utilized only a fraction of the collected data as it outlined only the most regional structures along the OETR profile, similar to what was possible to delineate by analysis of the nearby SMART 1 OBS profile [Funck et al., 2004] (Figure 1) with OBS spacing 5 to 10 times greater than for the OETR profile. The new modeling shows that the large increase in velocity on the shelf is not caused by a shallow continental crust basement, as proposed by Luheshi et al. [2012] based on the GeoPro model, but is rather due to carbonate layers, which were drillied in the area crossed by the profile [wells: Hesper P-52 and Sachem D-76]. These carbonate layers are found from 50-90 km distance along the profile at ~5 km depth. Moreover, there is a velocity inversion below the carbonates caused by lower velocity clastic sediments. Further seaward, in the OCT region, the new model also shows that the controvesial interpretation of magmatic underplating, between 190-270 km, proposed by Luheshi et al. [2012] and based on the constant 7.4 km/s velocity body (Figure 4, between 170-270 km distance) is rather a complex area with high velocity gradients which are strongly suggestive of mantle exhumation and serpentinization. Finally, at the very seaward end of the profile and again contrary to the results and interpretation of the GeoPro velocity model, the oceanic crust appears thiner than normal and is characterized by a velocity gradient, suggesting that the transition from rifting to drifting likely took time during which anomalous oceanic crust was produced.

The less informative but more objective minimum structure velocity model, here produced using Tomo2D software (Figure 7) and used to verify the results of Rayinvr modeling, confirms the existance of the shallow (~5 km depth, 10-80 km distance) carbonate basement on the shelf and an unusual oceanic crust at the seaward end of the profile. Joint inversions results for the Moho reflection interface and the deep velocity at the OCT show a complex relationship between the iso-velocity lines and the reflection Moho (between 200-260 km distance) but this area of the model still requires additional work before a conclusive intepretation is possible to carry out. Completion of this smooth traveltime tomography work will open the door for the last and perhaps most exciting stage of the proposed work, the joint prestack depth migration of OBS and MCS data.

Acknowledgments

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References

Bevington, Philip R., and D. Keith Robinson. "Data reduction and error analysis." McGraw-Hill (2003).

Funck, Thomas, et al. "Crustal structure of the northern Nova Scotia rifted continental margin (eastern Canada)." Journal of Geophysical Research: Solid Earth 109.B9 (2004).

Gerlings, Joanna, Keith E. Louden, and H. Ruth Jackson. "Crustal structure of the Flemish Cap Continental Margin (eastern Canada): an analysis of a seismic refraction profile." Geophysical Journal International 185.1 (2011): 30-48.

Hansen, Dorthe Møller, et al. "Development of a major polygonal fault system in Upper Cretaceous chalk and Cenozoic mudrocks of the Sable Subbasin, Canadian Atlantic margin." Marine and Petroleum Geology 21.9 (2004): 1205-1219.

Keen, C. E., and D. P. Potter. "The transition from a volcanic to a nonvolcanic rifted margin off eastern Canada." Tectonics 14.2 (1995): 359-371.

Korenaga, J., and W. W. Sager. "Seismic tomography of Shatsky Rise by adaptive importance sampling." Journal of Geophysical Research: Solid Earth (1978–2012) 117.B8 (2012).

Korenaga, J., et al. "Crustal structure of the southeast Greenland margin from joint refraction and reflection seismic tomography." Journal of Geophysical Research: Solid Earth 105.B9 (2000): 21591-21614.

Korenaga, Jun, et al. "Gravity anomalies and crustal structure at the southeast Greenland margin." Journal of Geophysical Research: Solid Earth106.B5 (2001): 8853-8870.

Lau, KW Helen, et al. "Structure of thinned continental crust across the Orphan Basin from a dense wide-angle seismic profile and gravity data." Geophysical Journal International 202.3 (2015): 1969-1992.

Louden, K, Lau, H, Wu, Y & Nedimovíc, M, 2011. Refraction Crustal Models and Plate Reconstruction of the Nova Scotia and Morocco Margins. Annex 14, Play Fairway Analysis Atlas – Offshore Nova Scotia. http://energy.novascotia.ca/oil-and-gas/offshore/play-fairway-analysis/analysis.

Louden, K. E., and D. Chian. "The deep structure of non-volcanic rifted continental margins." PHILOSOPHICAL TRANSACTIONS-ROYAL SOCIETY OF LONDON SERIES A MATHEMATICAL PHYSICAL AND ENGINEERING SCIENCES (1999): 767-799.

Louden, Keith, and Helen Lau. "Insights from scientific drilling on rifted continental margins." Geoscience Canada 28.4 (2001).

Louden, Keith. "Tectonic evolution of the east coast of Canada." CSEG Recorder 27.2 (2002): 37-48.

Luheshi, Matt, et al. "Atlantic Canada Offshore R&D: Nova Scotia Play Fairway Analysis." Offshore Technology Conference. Offshore Technology Conference, 2011.

Luheshi, Matt, et al. "The impact of conjugate margins analysis on play fairway evaluation–an analysis of the hydrocarbon potential of Nova Scotia." first break 30.1 (2012): 61-72.

Moser, T. J., G. Nolet, and R. Snieder. "Ray bending revisited." Bulletin of the Seismological Society of America 82.1 (1992): 259-288.

Papazachos, C., and G. Nolet. "P and S deep velocity structure of the Hellenic area obtained by robust nonlinear inversion of travel times." Journal of Geophysical Research: Solid Earth 102.B4 (1997): 8349-8367.

Watremez, Louise, et al. "Traveltime tomography of a dense wide-angle profile across Orphan Basin." Geophysics 80.3 (2015): B69-B82.

Wu, Yue, et al. "Crustal structure of the central Nova Scotia margin off Eastern Canada." Geophysical Journal International 166.2 (2006): 878-906.

Zelt, C. A., and R. M. Ellis. "Practical and efficient ray tracing in two-dimensional media for rapid traveltime and amplitude forward modeling." Canadian Journal of Exploration Geophysics 24.1 (1988): 16-31.

Zhang, Jie, Uri S. ten Brink, and M. Nafi Toksöz. "Nonlinear refraction and reflection travel time tomography." Journal of Geophysical Research: Solid Earth 103.B12 (1998): 29743-29757.

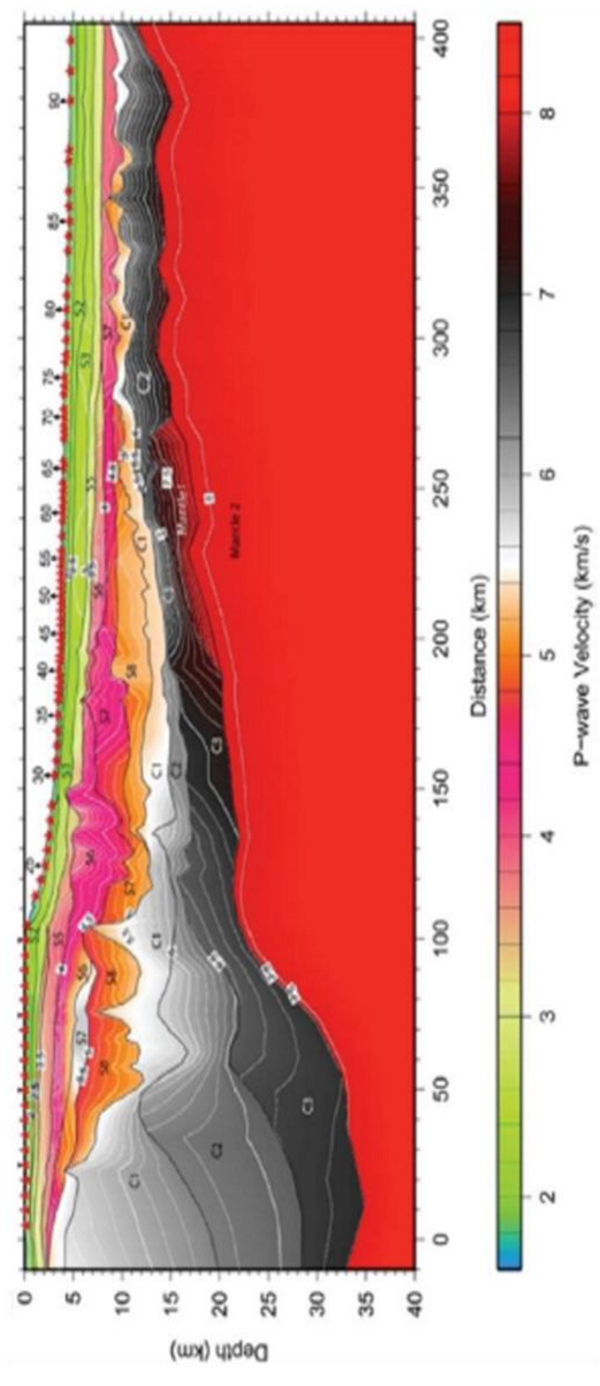

Figure 6 – Final P-wave velocity model for profile OETR-2009. Numbers on the red stars are Ocean Bottom Seismometers (OBS) numbers. White lines are velocity contours at 0.1 km/s interval. S1-S8: layers interpreted as sedimentary layers. C1-C3: interpreted crustal layers.

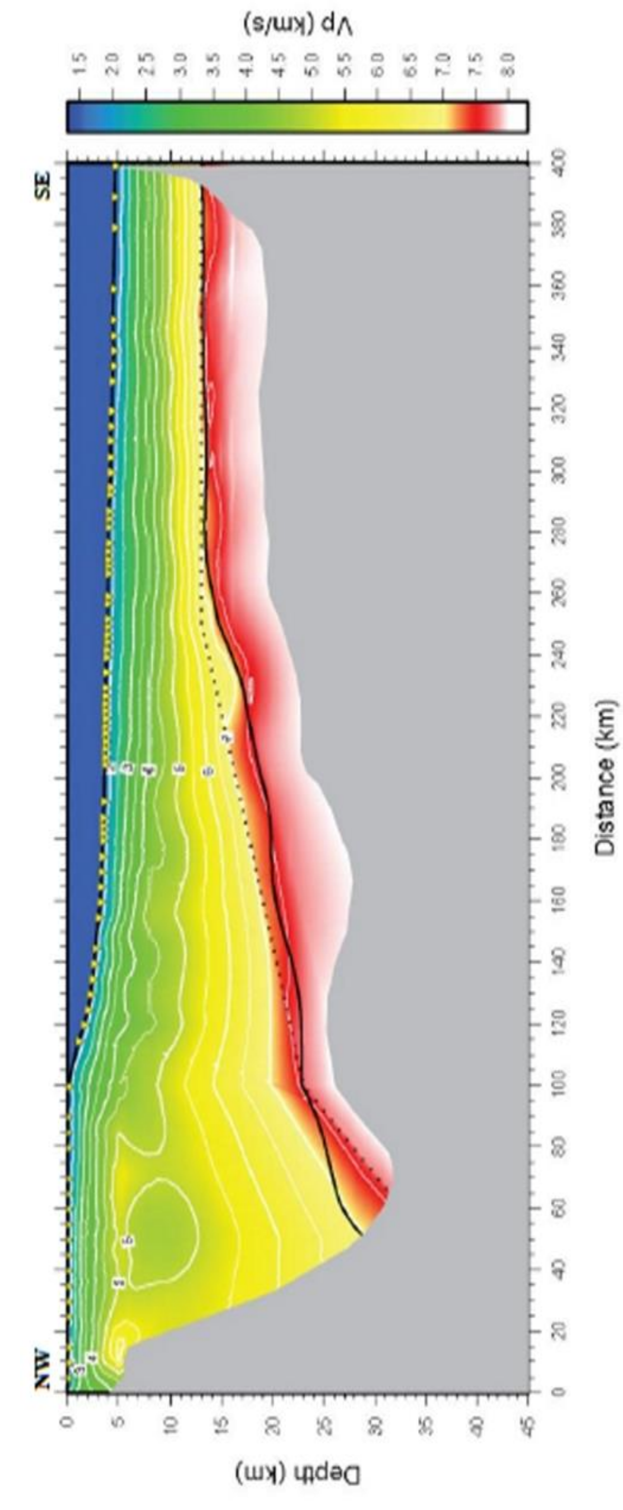

Figure 8 – Most recent preliminary result obtained using Tomo2D.