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Crustal and Upper Mantle Investigations of the Caribbean – South American Plate Boundary

by

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ABSTRACT

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The evolution of the Caribbean – South America plate boundary has been a matter of vigorous debate for decades and many questions remain unresolved. In this work, and in the framework of the BOLIVAR project, we shed light on some aspects of the present state and the tectonic history of the margin by using different types of geophysical data sets and techniques. An analysis of controlled-source traveltime data collected along a boundary-normal profile at ~65°W was used to build a 2D P-wave velocity model. The model shows that the Caribbean Large Igenous Province is present offshore eastern Venezuela and confirms the uniformity of the velocity structure along the Leeward Antilles volcanic belt. In contrast with neighboring profiles, at this longitude we see no change in velocity structure or crustal thickness across the San Sebastián - El Pilar fault system. A 2D gravity modeling methodology that uses seismically derived initial density models was developed as part of this research. The application of this new method to four of the BOLIVAR boundary-normal profiles suggests that the uppermost mantle is denser under the South American continental crust and the island arc terranes than under the Caribbean oceanic crust. Crustal rocks of the island arc and extended island arc terranes of the Leeward Antilles have a relatively low density, given their P-wave velocity. This
may be caused by low iron content, relative to average magmatic arc rocks. Finally, an
analysis of teleseismic traveltimes with frequency-dependent kernels produced a 3D P-
wave velocity perturbation model. The model shows the structure of the mantle
lithosphere under the study area and clearly images the subduction of the Atlantic slab
and associated partial removal of the lower lithosphere under northern South America.
We also image the subduction of a section of the Caribbean plate under South America
with an east-southeast direction. Both the Atlantic and Caribbean subducting slabs
penetrate the mantle transition zone, affecting the topography of the 410-km and 660-km
discontinuities.
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Introduction

The margin between the Caribbean and South American plates has a history of oblique collision that has resulted in a variety of geologic features observed on both sides of the margin. These include accretionary wedges, fold and thrust belts, and extensional as well as foreland basins [e.g. Ostos et al., 2005]. Additionally, the Caribbean plate underwent extensive volcanism in the Cretaceous that resulted in the formation of the Caribbean Large Igneous Province and a thickening of large areas of the Caribbean crust [e.g. Hoernle et al., 2004]. Current relative plate motions make the boundary predominantly strike-slip, bounded on the east by the Atlantic subduction and in the west by partial subduction of the Caribbean plate under South America (Figure 1).

For the past six years, the BOLIVAR (Broadband Offshore-Onshore Lithospheric Investigation of the Venezuela and Antilles Arc Region [Levander et al., 2006]) project in the United States and its Venezuelan counterpart GEODINOS (Geodinámica Reciente del Límite Norte de la Placa Suramericana) have been studying this margin in a multidisciplinary way. The goals of the projects are to understand the current state and the past evolution of the margin and to evaluate island arc accretion as a mechanism for continental crust growth. The collection, modeling and interpretation of geophysical data on a lithospheric and sublithospheric level constitute one of the main pillars of the projects. This dissertation is composed of slightly modified versions of three manuscripts in different stages of the publication process that describe the geophysical characterization of the plate boundary at different levels and utilizing different types of data and techniques.
Figure 1: Tectonic setting of the study area. Red arrows show GPS vectors from a compilation by Calais and P. Mann [2009], plate boundaries from the UTIG plates database. SA – San Andrés island; LA – Leeward Antilles; SS-EP – San Sebastián-El Pilar fault system

The first chapter describes the analysis of the controlled-source, wide-angle seismic data along BOLIVAR boundary-normal profile 65W and interprets the results in the context of velocity models derived for the other BOLIVAR profiles. At the northern end of profile 65W we find that the Venezuela Basin is underlain by Caribbean Large Igneous Province crust, unlike the normal oceanic crust found on profile 67W ~ 250 km to the west. The profile crosses the trend of the Leeward Antilles islands at the La Blanquilla High which has a velocity structure that is consistent with that found in the other BOLIVAR profiles for the Leeward Antilles, Aves Ridge and Lesser Antilles. In contrast with BOLIVAR profiles 67W and 64W, where the San Sebastian - El Pilar strike-slip fault system was found to separate two distinct types of crust; along profile
65W we see no significant differences in crustal velocity or thickness across the fault zone. This suggests that the crustal blocks that meet at the margin have similar characteristics and a fragment of Caribbean island arc crust may have been accreted to South America at this location. Our findings show that the evolution of the margin is not a simple progression of events occurring in the same manner and diachronously from west to east, but that the geologic processes resulting from the oblique collision were likely significantly affected by the paleogeography and pre-existing structures of the South American side of the margin.

In chapter two I present a new methodology for 2D gravity modeling using seismically derived initial density models and apply the algorithm to four of the BOLIVAR boundary-normal profiles. Crustal density adjustments to the initial models in the crystalline crust found to fit the gravity data are of a different nature on either side of the strike slip margin for profiles 67W and 64W (negative for the island arc crust and weakly positive for continental South American crust). On profile 65W negative density adjustments were also found for the island arc crust, but there is no change across the strike slip margin and the step to positive values that we associate with continental South American crust occurs further inland. This analysis of the gravity data supports the conclusion drawn from the seismic velocity models regarding the nature of the crust to the north and south of the strike-slip fault zone: Along profiles 67W and 64W the two crustal blocks that meet at the fault zone have different characteristics, while along profile 65W they are of a similar nature. The negative density adjustments in the island arc crust suggest a lower iron content with respect to typical island arc rocks. Island arc crust must loose some of its iron in order to eventually become continental crust [e.g. Lee
et al., 2007], and our analysis suggests that this has indeed occurred at the Leeward Antilles. However, more work from a geophysical and geochemical perspective is required to further explore this hypothesis.

In the third and last chapter, I use teleseismic data recorded by the BOLIVAR broadband seismometer array to build a 3D tomographic P-wave velocity model of the upper mantle under the margin, revealing lithospheric features and imaging the present configuration of Atlantic and Caribbean subduction. Shallow subduction of a section of the Caribbean plate under northern Colombia is followed by a stage of steep subduction that is imaged in our model. Limitations due to station coverage do not allow us to image the down going Caribbean slab at the location of the Bucaramanga seismic nest. The work presented in this dissertation suggests that efforts should be made to extend the mantle image westward in order to investigate the relationship between the slab and this seismically active zone.

Our image of the Atlantic slab suggests that while subduction terminates at the strike slip boundary for the upper lithosphere, part of the lower South American lithosphere subducts along with the Atlantic slab. This partial removal of the lithosphere would potentially cause isostatic rebound which may have played a role in the uplift of the coastal mountain ranges in northern Venezuela. Future work will examine lithosphere-asthenosphere boundary depths in northern Venezuela and this information should be analyzed in conjunction with uplift and erosion rate data for the coastal ranges to test this hypothesis.

The Atlantic as well as the Caribbean slab we have imaged reach depths greater than 700 km, affecting mantle transition zone topography. In the near future we can
expand on this study by updating the transition zone topography results using our mantle velocity model and building a tomographic S-wave velocity model for the study area. Combining all these results we can try to place constraints on the temperature and water content of the two slabs.

The research presented here adds to the growing body of geophysical information about the Caribbean – South America plate boundary, adds new constraints, and helps refine our understanding of this much debated margin, while pointing towards new directions for future research.
Chapter 1

The Caribbean - South American plate boundary at 65°W: Results from wide-angle seismic data

M. J. Bezada, M. B. Magnani, C. A. Zelt, M. Schmitz, A. Levander

Abstract

We present the results of the analysis of new wide-angle seismic data across the Caribbean – South American plate boundary in eastern Venezuela at about 65°W. The ~500 km long profile crosses the boundary in one of the few regions dominated by extensional structures, as most of the southeastern Caribbean margin is characterized by the presence of fold and thrust belts. A combination of first-arrival traveltime inversion and simultaneous inversion of PmP and Pn arrivals was used to develop a P-wave velocity model of the crust and the uppermost mantle. At the main strike-slip fault system, we image the Cariaco Trough; a major pull-apart basin along the plate boundary. The crust under the Southern Caribbean Deformed Belt exhibits a thickness of ~15 km, suggesting that the Caribbean Large Igneous Province extends to this part of the Caribbean plate. The velocity structures of basement highs and offshore sedimentary basins imaged by the profile are comparable to those of features found in other parts of the margin, suggesting similarities in their tectonic history. We do not image an abrupt change in Moho depth or velocity structure across the main strike-slip system, as has been observed elsewhere along the margin. It is possible that a terrane of Caribbean island arc origin was accreted to South America at this site, and was subsequently bisected by the strike-slip fault system. The crust under the continental portion of the
profile is thinner than observed elsewhere along the margin, possibly as a result of thinning during Jurassic rifting.

1.1. Introduction

The Caribbean-South American (CAR-SA) plate boundary in the southeast Caribbean is a ~300 km wide diffuse zone of deformation that accommodates the ~20 mm/yr eastwards displacement of the Caribbean plate relative to South America. Along the central and eastern Venezuelan coast most of this motion is accommodated along the right-lateral strike-slip San Sebastián-El Pilar (SS-EP) fault system. The main tectonic elements of this area include an accretionary wedge, active and extinct island arcs, a large dextral strike-slip system, and coastal thrust belts with associated foreland basins (Figure 1.1).

Many factors contribute to making this an area of broad geological interest including 1) the potential accretion of the Leeward Antilles to the South American continent as a mechanism for creating new continental crust, 2) the relationship between margin processes and the formation and accumulation of large oil and gas reserves, and 3) the high population density in northern Venezuela that is at risk from geohazards, particularly earthquakes and landslides, resulting from margin dynamics. Due to all these factors, the boundary has been the subject of geological and geophysical studies for decades [e.g. Hess and Ewing, 1940; Hess, 1966; Silver et al., 1975; Diebold et al., 1981; Burke, 1988; van der Hilst and Mann, 1994]. Despite the abundance of studies and extensive seismic exploration for petroleum in this region, data that image the lithospheric structure at the plate boundary is limited [e.g. Schmitz et al., 2005].
The BOLIVAR (Broadband Ocean-Land Investigation of Venezuela and the Antilles arc Region) project was designed to image the structure of the lithosphere along the entire southeastern Caribbean margin [Levander et al., 2006]. The active-source component of BOLIVAR provided wide-angle and reflection seismic data along a series of 5 boundary-normal profiles (Figure 1.1) intended to image the margin at different stages of evolution after the early Cenozoic CAR-SA collision. This paper will focus on
the profile at ~65°W longitude, hererafter referred to as 65W. We present a P-wave velocity model along the profile, derived from a hybrid inversion technique that combines first-arrival tomography and layer-based inversion of deeper seismic phases.

The 65W profile approximately bisects the plate boundary zone at the strike-slip fault system, extending from the South Caribbean Deformed Belt (SCDB) in the north, to the Oriental Basin and Espino Graben to the south, crossing the La Blanquilla High, the Los Roques Canyon, the Margarita Basin, and the Cariaco pull-apart basin in the Barcelona Bay (Figure 1.1). Understanding of the margin in this area has evolved with the increasing availability of data through time. In the 1990’s, several studies were published that focused on producing balanced cross-sections constrained by the abundant exploration-scale reflection seismic profiles that have imaged the region’s sedimentary basins. These balanced cross-sections were extended to a crustal scale and structures at depth were constrained by potential field data [e.g. Passalacqua et al., 1995]. Schmitz et al. [2005] presents the first wide-angle seismic survey to study a north-south profile across the Oriental Basin (OB). The study proposed a depth to basement of ~13 km and placed some constraints on the depth of the Moho. The profile was extended offshore, where no wide-angle seismic information was available at the time, using gravity data.

The profile presented here differs from its companion BOLIVAR profiles [Christeson et al., 2008; Clark et al., 2008; Guedez, 2007; Magnani et al., 2009] in that it meets the strike-slip system not at the coastal mountains of the Cordillera de la Costa, but rather at a bay that encompasses a pull-apart basin offshore, and a broad lowland onshore. Some other distinctive characteristics of the profile include crossing the Los Roques Canyon, and the eastern end of the SCDB offshore and the Espino Graben
onshore. Additionally, the profile was less well instrumented offshore than the other BOLIVAR profiles with one sixth as many ocean bottom seismometers (OBSs) deployed. The smaller number of OBSs results in decreased coverage, particularly towards the northern end of the profile.

1.2. Tectonic Setting

A review of the tectonic history of the study area begins at the Mid-Jurassic, long before this region hosted any interactions between the CAR-SA plates, indeed before a Caribbean plate, per se, existed. Opening of the Atlantic Ocean in the Mesozoic resulted in the formation of many grabens located today around the Atlantic’s margins [e.g. Burke, 1976]. In northeastern South America, a series of such grabens with a SW-NE orientation include the Takutu Graben in Guyana [Crawford et al., 1985] and the Espino Graben in Venezuela [Moticska, 1985]. Continued rifting between North and South America in the Late Jurassic - Early Cretaceous resulted in the opening of the proto-Caribbean seaway [Pindell, 1985] while, to the west, the eastward subduction of the Farallon plate beneath the Americas created a volcanic arc known as the Great Arc of the Caribbean [Burke, 1988; Pindell and Dewey, 1982]. In the Early Cretaceous (~110 Ma) a modification in plate configurations resulted in a change in subduction polarity beneath the Great Arc, which caused a fragment of the Farallon plate to start overriding the proto-Caribbean; this fragment became the Caribbean plate. During the same time, the Caribbean Large Igneous Province (CLIP) was formed by processes that remain unclear [Hoernle et al., 2004], covering much of what is now the Caribbean plate with 5 to 15 km of additional basaltic crust. As it advanced over the proto-Caribbean, the Caribbean plate
migrated northeastwards relative to SA until it collided with the Bahamas platform in the Eocene. The collision caused clockwise rotation of the plate and modified its trajectory to establish a generally eastward motion relative to stable SA [Pindell and Kennan, 2001, 2007] with a small component of convergence that led to highly oblique collision between CAR and SA. The collision caused the diachronous (eastward younging) emplacement of the fold and thrust belts onshore Venezuela (known as the Caribbean Mountain System to the west of the study area and as the Serranía del Interior to the east) (Figure 1.1), as well as associated foreland basins onshore [Ysaccis, 1997; E. J. Hung, 1997; Ostos et al., 2005]. Emplacement of the onshore fold and thrust belts was progressively followed by backthrusting north of the Leeward Antilles which resulted in the formation of the SCDB [Mann et al., 2007]. Fission track data show that shortening in the eastern Serranía del Interior ceased at ~12 Ma [Locke and Garver, 2005] possibly reflecting the end of collision. After the end of oblique convergence between CAR and SA in the Late Miocene the CAR plate acquired its present motion and the plate boundary evolved to the nearly purely strike-slip margin observed today [Pindell et al., 2005; Ysaccis, 1997]. Motion along the El Pilar fault system seems to have initiated at this time coinciding with the start of deposition in the Cariaco-Tuy Basin [Jaimes, 2003]. Recently, Higgs [2009] suggested that onset of motion along the El Pilar fault occurred in the Pliocene and could be as young as 2.5 Ma. Today, the plate boundary is a 300 km wide zone of deformation. It is most clearly defined in central and eastern Venezuela where the SS-EP fault systems account for 80% of the 20 mm/yr of right-lateral displacement [Perez et al., 2001; Weber et al., 2001].
Profile 65W begins to the north in the Venezuela Basin floored by undeformed oceanic crust (Figure 1.1) and continues south crossing the SCDB near its eastern end (only a small expression of the SCDB is observed in a companion profile ~120 km to the east, suggesting minor underthrusting of CAR under SA [Clark et al., 2008]). Accordingly, the accretionary prism at this longitude is thinner than in the west, where the onset of deformation is older. South of the SCDB the profile crosses a bathymetric high, the La Blanquilla High, part of a tectonic belt revealed by gravity and bathymetry data containing the Leeward Antilles arc and the Aves Ridge (Figure 1.1b, [Mann, 1999]). Further south, the profile crosses the Los Roques Canyon, a NW-SE trending basin that exhibits a bathymetric and gravity signature (Figure 1.1). Little has been published about the tectonic origin of this basin. The Los Roques fault runs along the southwestern flank of this basin. It has been hypothesized that, prior to the onset of pure strike-slip conditions; this fault was connected to the Urica fault onshore. This continuous system would have served as a transfer zone between the underthrusting of CAR under the SCDB and the compressional structures of the Serranía del Interior [Higgs, 2009]. On its path between the Margarita and La Tortuga islands, profile 65W crosses the Margarita Basin (Figure 1.1), an extensional structure that opened in the Paleogene as a result of rifting between the two islands. Sediment deposition in this basin remained modest until the Pliocene, when the transtensive regime responsible for the opening of the Cariaco Trough renewed fault-controlled subsidence [Jaimes, 2003]. Profile 65W encounters the Cariaco Trough further south, where the seafloor drops to 1400m below sea level, making this the deepest basin in the southeast Caribbean (Figure 1.1). The Cariaco Trough has an area of 20 km$^2$ and is composed of two rhomboidal features, the West and
East Cariaco Trough. The profile crosses the smaller East Cariaco Trough. The Cariaco trough has been interpreted as a pull-apart basin occurring due to transfer of strike-slip motion from the San Sebastián to the El Pilar fault zones [Schubert, 1982; Jaimes, 2003]. Opening of the Cariaco Trough initiated in the Late Miocene and fully developed in the Pliocene through a transtensive regime that has since weakened [Jaimes, 2003].

Jaimes [2003] hypothesized that the releasing bend formed as a consequence of misalignment of the SS-EP fault system by motion along the Urica fault; a lateral ramp connecting the Guárico and Pirital thrust fronts.

As the profile enters land, it crosses a thick sedimentary sequence with a multi-stage subsidence history. The bottom of the sequence is comprised of Paleozoic sediments that predate the rifting associated with the opening of the Atlantic. The pre-rift sequence is followed by Jurassic syn-rift and post-rift sediments that filled the Espino Graben and that are sealed by a Cretaceous and Paleogene passive margin sequence that rests on Paleozoic pre-rift sediments in areas not affected by the Jurassic rifting episode. The Neogene foreland basin sediments of the ESE-WNW trending Oriental Basin were deposited over the passive margin wedge and cap the sedimentary sequence [Salazar, 2006]. A depth to basement of 6-8 km at the axis of the Espino graben has been estimated from potential field methods [Feo-Codecido et al., 1984]; however, more recent depth estimates based on reflection seismic data place the basement of the graben (bottom of Jurassic sediments) at 10-11 km [Salazar, 2006].
1.3. Wide-angle Seismic Data Acquisition

The seismic wide-angle data were acquired using sources and receivers located both onshore and offshore along the profile. Offshore, 7 OBSs were deployed by the R/V Seward Johnson II. The OBSs were LC2000 instruments from the Institute of Geophysics and Planetary Physics at Scripps Institute of Oceanography, with one vertical component and one pressure component. They were deployed at intervals of ~ 27 km in water depths ranging from 250 to 2000 m and used a sampling rate of 8 ms. Locations for these offshore receivers were obtained by inverting the travel times of the water-wave arrivals. On land, 514 recording stations consisting of one (48 hr recording capacity) or two (24 hr recording capacity each) RefTek 125-01 ("Texans") single-channel digitizer/recorders with 4.5 Hz geophones as sensors were deployed at intervals of ~ 500 m. Land stations were located using handheld GPS units with errors ranging from 5 to tens of meters depending on the site and atmospheric conditions.

All receivers recorded shots from the airguns of the R/V Maurice Ewing. The R/V Maurice Ewing's source consisted of a tuned 20-element airgun array with a combined capacity of 114 l (6400 in³). The profile was shot twice, once at 50 m shot spacing to achieve a high fold for multi-channel seismic reflection recordings and once at 150 m shot spacing to reduce noise from the wave train of previous shots on OBS records for the wide-angle analysis. In addition to the offshore sources, two land shots (600 kg of pentoilite) located 60 km and 115 km from the coast, provided some reversed coverage of the land portion of the profile (Figure 1.1).
1.4. Wide-angle Seismic Data

The quality of the OBS data used for wide-angle analysis is generally good. Data were examined in both the 50 m and 150 m shot interval records as well as both the hydrophone and vertical component records. Signal processing applied consisted of an Ormsby bandpass filter with corner frequencies typically at 2 and 15 Hz (varying slightly on a case-by-case basis). Picks were made in whichever of the records showed the clearest arrivals. First arrivals are clearly visible in OBS records typically to offsets of 50-70 km. Occasionally, arrivals are identifiable at larger offsets reaching, in one case, 110 km. Moho reflections (PmP) were identified and picked in records from five of the seven OBSs at offsets ranging from 35 to 95 km (Figure 1.2a).

The quality of the Texan recordings of the offshore airgun shots was variable, but generally poor. Only a handful of the recording stations (a fraction of those closest to the coast, up to 35 km inland) yielded record sections with clearly visible arrivals. First arrivals were picked for 12 of these at offsets as large as 115 km, but more commonly up to 50-80 km. PmP arrivals were also picked for 10 of these stations (Figures 1.2b, 1.2c). Unlike airgun shots, the land shots were recorded very well by the Texan stations (Figure 1.2d). First arrivals were clearly visible to offsets as large as 160 km. The PmP phase was identified and picked in the Texan record sections of both land shots. An additional reflected phase was also picked in the Texan records of the northern land shot (Figure 1.2d), we interpret this phase as a reflection from the base of the sedimentary sequence (i.e. the base of the Espino Graben). OBS recordings of the land shots were poor and no picks were made. This is most likely due to the large offsets (> 90 km) involved as well as the large spacing between OBSs which makes it difficult to correlate arrivals between
instruments. At a later stage of picking, arrivals corresponding to waves traveling through the uppermost mantle (Pn) were identified with the aid of arrival times predicted from preliminary models. Larger uncertainties were assigned to these picks due to the low signal-to-noise ratio (SNR).

![Figure 1.2](image)

**Figure 1.2**: Record sections of wide-angle data for (a) OBS station 07 located at model distance 164 km, (b) Texan station 004 located at model distance 186 km, (c) Texan station 039 located at model distance 221 km and (c) the northern landshot, located at model distance 257 km. Data are plotted with a reduction velocity of 8 km/s and band-pass filtered from 3 to 15 Hz. First arrivals shown in red, PmP arrivals are shown in green and an intracrustal phase identified in the northern landshot is shown in blue. Offsets are plotted positive to the south.

Uncertainties were assigned to each pick based on the SNR using a 500 ms time window centered on the pick [Zelt and Forsyth, 1994]. Uncertainty is proportional to SNR within pre-defined bounds. The lower and upper bounds were 50 ms and 250 ms for
the first arrivals used in the tomographic inversion and 150 ms and 350 ms for the PmP, Pn and intracrustal arrivals used in the layer-based inversion.

1.5. Wide-angle Traveltime Analysis

The wide-angle data were analyzed using a hybrid, layer-stripping approach. In the first stage, first arrival traveltime tomography was used to constrain the velocity structure of the upper ~20 km of the model. This was followed by the inversion of PmP and Pn arrivals in a layer-based algorithm to constrain the location of the Moho and the velocity of the lower crust and uppermost mantle. Model assessment was performed as a final step, to provide an estimate of the reliability of the results. This modeling strategy was applied to all BOLIVAR wide-angle profiles and a detailed description is available in Magnani et al. [2009]. For this reason, it will only be broadly described here and we will focus on the aspects that are different for this profile.

1.5.1. First-arrival inversion

To obtain the tomographic model we use the regularized, iterative, first arrival traveltime inversion algorithm of Zelt and Barton [1998]. For the forward traveltime calculations, a finite difference method to solve the eikonal equation based on that of Vidale [1990] as modified by [Hole and Zelt, 1995] is used. The model is parameterized in terms of nodes for the forward calculation of traveltimes and as cells for the inversion of slowness values. The inversion algorithm solves for the slowness values that minimize an objective function that includes norms for model roughness and data misfit [Zelt and Barton, 1998]. If the inversion is successful, the resulting model will fit the observed
traveltime data appropriately ($\chi^2$ of 1, RMS data misfit equivalent to data uncertainty) while having minimum roughness in the vertical and horizontal directions.

The model that we present here was parameterized with a node spacing and cell size of 0.5 km in both the X and Z directions for the forward and inverse step, respectively. With a total model length of 560 km and a depth of 60 km this results in 137,883 nodes and 136,640 cells. The 0 km horizontal reference point on the model was established at 12°N in keeping with the convention for all BOLIVAR profiles. The reference depth of 0 km corresponds to the mean sea level with positive values indicating subsurface depths.

Choosing an appropriate starting model proved to be essential to the success of the inversion. The first starting model was a simple, pseudo-1-D model consisting of the following velocity values: 3 km/s at the surface/seafloor, 6 km/s at a depth of 8 km and 8 km/s at every depth equal to or greater than 40 km (Figure 1.3). When using this starting model, the inversion failed to converge to an acceptable $\chi^2$ value. However, the inverted models showed low velocity structures that matched the known location of sedimentary basins. Therefore, a new starting model was constructed by incorporating the laterally varying distribution of sedimentary (2.0 -5.5 km/s) velocities (Figure 1.3). The traveltimes calculated through this new starting velocity model had an RMS misfit of 227 ms or $\chi^2$ of 4.24. Using this starting model, a final model was reached after 9 iterations of the inversion algorithm (Figure 1.3). The final model achieved a $\chi^2$ of 1.14, corresponding to an RMS travelttime misfit of 90 ms. Ray coverage from the first arrival inversion is only adequate for the uppermost ~17 km of the model with only a few rays
Figure 1.3: Pseudo-1D starting model (top left) and the corresponding best model (bottom left) which achieved a normalized misfit $\chi^2$ of 4.24. The model shallow velocity structure in this model was used as a template to produce a new starting model (top right) which resulted in the final first arrival model (bottom right) with a $\chi^2$ of 1.14. Regions of the final first arrival model (bottom right) not covered by rays are blanked out in the Figure.
reaching greater depths. The distribution of rays suggests that constraint on the model from first arrivals is strictly limited to the upper crust (Figure 1.4).

![Figure 1.4: Ray coverage from first arrival inversion. Every fifth ray is shown. Arrow indicates location of the El Pilar fault. Dots indicate the location of the Moho resulting from ray-tracing inversion (gray) and tomographic inversion (black).](image)

1.5.2. Layer-based inversion

The velocity model obtained through first arrival traveltime inversion was re-parameterized to be used as starting model for a new, layer-based inversion (Figure 1.5). This second inversion method is based on ray tracing and allows for inverting the vertical position of boundary nodes and the velocities of the nodes that define each layer [Zelt and Smith, 1992]. The shallowest 20 km of the model (shallowest 10 km at the northern end) were represented by 10 layers with laterally varying velocities to replicate the final first-arrival inversion model, including a first layer that represents the seawater. The 11th layer represents the mid to lower crust, down to the Moho, and the 12th layer represents the uppermost mantle. The first ten layer boundaries were fixed throughout the inversion
process and only the depth of the nodes representing the Moho and the velocity values directly above and below the Moho were inverted.

The data used for the inversion consists of PmP traveltime picks from the OBS and Texan records of airgun shots, as well as the record sections from the two land shots. A total of 7920 PmP picks were included in the inversion. Additionally, Pn arrivals were included for a total of 11,670 traveltime picks. Uncertainties in these arrivals are higher than those assigned to the Pg arrivals used in the tomographic inversion. Larger uncertainties were given to these arrivals because the larger offsets and smaller SNR make it more difficult to pick the arrival time and, in the case of PmP, since these events are not first arrivals and are mixed in the coda of earlier phases.

The Moho in the starting model is mostly flat (Figure 1.5) and originally parameterized by 24 nodes located at 25 km intervals. Where lateral variability across the model was required to fit the data, the spacing of the nodes was reduced to 12.5 km resulting in a total of 26 nodes. The inversion process was guided by alternatively fixing or inverting the positions and velocities of different nodes until a satisfactory model was reached. The final model resulting from this inversion will be referred to as the preferred model henceforth (Figure 1.6). The final misfit of this model has a $\chi^2$ value of 1.228 and an RMS of 231 ms. More specifically, the PmP arrivals are better fit with a $\chi^2$ of 1.089 and RMS of 213 ms, while the Pn arrivals were fit with an RMS of 295 ms and a $\chi^2$ of 1.821. Given the PmP and Pn ray coverage, the Moho in this model is constrained mostly under the center of the profile and only a weak constraint on the velocity of the uppermost mantle is achieved, also near the center of the profile (Figure 1.7). To model the reflected phase picked in the northern land shot record, a floating reflector was
introduced. Since this phase is only constrained by one shot, there are no multiple observations that need to be reconciled making it easy to fit the data very accurately. Thus, the RMS misfit for this phase is 87 ms, much smaller than the assigned uncertainties, resulting in $\chi^2$ of 0.15.

1.5.3. Model Assessment

Before any interpretations based on the velocity model are made, it is important to have a clear notion of the robustness and reliability of the results. We do this by following two procedures: 1) Checkerboard tests provide insight into the resolution of the first arrival traveltime tomography, and 2) A semi-independent inversion procedure estimates the depth of the Moho interface. This is the same assessment carried out by Magnani et al. [2009] where a more detailed description can be found.

Figure 1.5: Starting velocity model for the layer-based inversion. Velocity structure above the dotted white line is taken from the first arrival inversion model and held fixed. Starting velocities directly above and below the Moho are indicated in white.
Figure 1.6: Final Velocity model for profile 65W shown with a 5X vertical exaggeration (top) and no vertical exaggeration (bottom). Velocities in the lower crust and upper mantle labeled in white in km/s. Isovelocity contour interval is 0.5 km/s; with thicker contours labeled. Green dots indicate the location of OBSs, white dots indicate the location of Texan stations with picked records, red dots indicate the location of land shots. Yellow dots in the bottom of the model correspond to bounce points of modeled PmP reflections referred to in the text as ray-trace Moho, the tomographic Moho is indicated in black. Orange dots at a depth of ~10 km between model distances 260 and 280 km represent bounce points of the intracrustal phase interpreted as a reflection from the base of the Espino Graben. VB: Venezuela Basin, SCDB: Southern Caribbean Deformed Belt, LBH: La Blanquilla High, LRC: Los Roques Canyon, MB: Margarita Basin, CT: Cariaco Trough, EG/OB Espino Graben / Oriental Basin. Black arrow shows the location of the San Sebastian Fault, white arrow shows the location of the El Pilar Fault and yellow arrow shows the location of the coastline.
1.5.3.1. First-arrival Checkerboard Tests

We applied a 2-D adaptation of the 3-D checkerboard tests described in Zelt [1998] to the first arrival inversion process as a mean of assessing model resolution. A grid of alternating positive and negative velocity anomalies was superimposed on the starting velocity model, synthetic traveltimes through this perturbed model were calculated using the actual experiment geometry, and Gaussian noise with a standard deviation equal to the uncertainties of the real data was added. This was followed by an attempt to recover the anomaly pattern with our inversion algorithm, keeping the same values for the free parameters used in the inversion of the real data. The success of the inversion was measured by calculating the semblance of the recovered and known perturbed models. Any point in the model with a semblance value greater than 0.7 was considered to be resolved, meaning that resolution in that point was better than the anomaly size. To obtain results independent of the polarity and registration of the
anomaly pattern we switched the polarity of the anomalies and spatially shifted the patterns for each grid size and averaged the results of the semblance calculations for each case. The procedure was then repeated for different grid sizes (25x2.5 km, 50x5 km, 75x7.5 km, 100x10 km and 150x15 km) and a resolution estimate plot was generated by interpolating the resolved areas at the grid sizes tested (Figure 1.8).

**Figure 1.8:** Results of checkerboard resolution tests at different grid sizes for the first arrival inversion and estimated lateral resolution plot (bottom right panel). Areas colored in black in the lateral resolution plot have an estimated lateral resolution better than 25 x 2.5 km, while regions colored in white have an estimated lateral resolution poorer than 150x50 km. Moho as determined by the layer-based inversion is indicated by the bold black line for reference, white dots mark the location of OBS stations, Texan stations with offshore arrivals and land shots. See text for a description of the checkerboard grid sizes and testing procedure.
The checkerboard test results indicate that the shallowest 15 km of the first arrival model are well constrained, whereas constraint at depths greater than 20 km is poor. The poor constraint provided by first arrivals at mid and lower crustal depths is not of great concern however, since this area of the model is constrained by the PmP and Pn arrivals used in the layer-based inversion. The results of the checkerboard tests give us confidence that the velocity structure in the shallowest 20 km of the model, which is held fixed in the second (layer-based) inversion, is reasonably well constrained by the first-arrival tomography.

1.5.3.2. Moho Reflection Tomography

To test the results of the layer-based inversion we employ a separate regularized inversion to find the depth to Moho using PmP traveltimes exclusively. The Moho was modeled as a floating reflector while the entire velocity model was fixed. The velocity model is a slightly modified version of the final first arrival model, in which we allowed a maximum velocity of 7.0 km/s by resetting all higher velocities in the model to that value. Taking the velocity model as a priori information, the inversion algorithm then seeks to find the flattest (minimum first-spatial-derivative) Moho interface that minimizes the PmP traveltime residuals to a value consistent with their assigned uncertainties. The method works iteratively and is analogous to the one described earlier for the first arrival tomography. This approach is different from the layer-based inversion in that it uses a finite difference method for calculating the traveltimes [Hole and Zelt, 1995], it is regularized, and does not include Pn traveltimes. The method is not entirely independent however, since the PmP traveltimes and upper crustal structure are identical to those used
in the layer-based ray tracing approach. After six iterations, the final interface was reached, achieving a normalized misfit $\chi^2$ of 1.01 and an RMS misfit of 205 ms.

The result of the tomographic PmP inversion agrees well with the ray-trace inversion Moho, as the two interfaces are within ~2km of each other throughout the profile (Figure 1.6). Differences are to be expected since the lower crustal velocities determined by the layer-based inversion are different from the homogeneous 7.0 km/s value assumed in this method. The largest absolute differences are in the south of the profile where lower crustal velocities as found by the layer-based inversion are slowest (6.7 km/s).

1.5.4. Gravity Modeling

Gravity modeling served as an additional test on the feasibility of the P-wave velocity model. An observed gravity curve (Figure 1.9) was obtained from a 3 minute (~5.5 km) grid constructed from the Simón Bolívar University gravity database with free-air anomaly values offshore and Bouguer anomaly values on land [C. Izarra et al., 2005]. The curve represents an average of the observed anomaly over a 30 km wide swath centered on the profile. A density model was generated by mapping the P-wave velocities ($V_p$) of the preferred model into density values ($\rho$) using a fifth-order polynomial fit to experimental data [Zelt, 1989]. For the gravity calculations the model is parameterized as a series of trapezoidal blocks of constant density following the method of Talwani et al. [1959].

The calculated gravity for the density model matches the broad characteristics of the observed anomalies, with an RMS misfit of 23.098 mgal (Figure 1.9). The position
of the Cariaco Trough corresponds to a narrow gravity low where the anomaly reaches ~87 mgal. The SCDB, as well as the Los Roques Canyon, correspond to broader local lows in the observed gravity anomaly. The La Blanquilla High, with its high seismic velocities, and correspondingly high densities, is aligned with a local high in the observed gravity anomaly (Figure 1.9). On land, the geometry of the sedimentary basins and the Moho topography in the model result in a 200 km wide, low gravity anomaly. Although the calculated lateral extent of this anomaly agrees well with the observations, its absolute amplitude is too large by ~75 mgal.

Figure 1.9: Results of gravity modeling. (a) Comparison of the observed gravity anomaly (black crosses) with gravity anomalies calculated from a density model derived from the preferred P-wave velocity model without any adjustments (b) and three models adjusted to fit the observed data. In each of the three adjusted models, density adjustments where restricted to one of three geologic domains: sedimentary basins (c), igneous-metamorphic crust (d) or upper mantle (e). VB - Venezuela Basin; SCDB - Southern Caribbean Deformed Belt; LBH - La Blanquilla High; LRC - Los Roques Canyon; MB - Margarita Basin; CT - Cariaco Trough; EG/OB - Espino Graben / Oriental Basin. Non-linear (top) color bar used for (b) and red-white-blue (bottom) color bar used for (c), (d) and (e).
We attribute the misfit between the observed and calculated gravity anomalies chiefly to the non-uniqueness of the Vp-\(\rho\) relationship. Given the scatter of the empirical data on which the relationship is based, a range of density values is permissible for any given velocity value [P. J. Barton, 1986; Ludwig et al., 1970]. To investigate how the data misfit could be reduced by adjusting the densities of the blocks that make up the model, we used a forward modeling approach. We explored three different scenarios considering density adjustments restricted to three geologic domains: sedimentary basins (original densities between 1100 and 2550 kg/m\(^3\)), igneous-metamorphic crust (original densities between 2550 and 3000 kg/m\(^3\)), and upper mantle (original densities larger than 3000 kg/m\(^3\)).

The modeling approach is based on defining regularly spaced density adjustment (\(\Delta\rho\)) nodes for each domain. We obtain the \(\Delta\rho\) for each model block within the domain by linearly interpolating between the nearest corresponding nodes. Through trial-and-error forward modeling we found sets of \(\Delta\rho\) for each of the three domains that substantially reduced the data misfit. The spacing of the \(\Delta\rho\) nodes increased from 10 km for the sedimentary basins to 20 km for the igneous-metamorphic crust and 40 km for the upper mantle, following the assumption that lateral variability of densities is highest in the shallow subsurface and decreases with depth. The three resulting models show an improvement in the RMS data misfit, from the original 23.10 mgal to 14.27 mgal when adjustments were restricted to the upper mantle, 10.21 mgal when they were restricted to the igneous-metamorphic crust and 10.51 mgal when they were restricted to the sedimentary basins. Because of the shallower depth and smaller size of the blocks that make up the sedimentary basins, adjustments in this domain can better fit small scale
structure but require larger amplitude adjustments to reproduce the observed gravity anomalies. The crystalline crustal adjustments achieve the maximum misfit reduction (a fraction of 1 mgal better than the sedimentary adjustments); while the mantle adjustments achieve a smaller, but still significant reduction of RMS misfit. The difficulty in fitting the data adjusting only the densities of mantle blocks is due to their depth and size that make them least capable of fitting small-scale anomalies. In all cases, the density adjustments range between -130 kg/m$^3$ and 220 kg/m$^3$ and in the case of the crust and mantle they range between -130 and 150 kg/m$^3$.

We acknowledge two major drawbacks to this forward modeling approach: 1) It does not amount to a thorough exploration of the $\Delta\rho$ model space, and 2) it is intrinsically limited due to the restrictions on the distribution of density adjustments imposed. It is possible, even likely, that the real densities of the sedimentary basins, the igneous-metamorphic crust, and the upper mantle, all vary from the values we infer from their P-wave velocities; instead of these variations being restricted to only one of these domains as we have assumed for the sake of simplicity. In spite of these limitations, the forward modeling we have conducted shows that relatively minor adjustments in the densities derived from the P-wave velocity model, that are well within the scatter of experimental Vp-$\rho$ determinations, can achieve a good fit to the gravity data.

1.6. Results

In this section we present a description of the preferred P-wave velocity model (Figure 1.6). Like all of the BOLIVAR velocity models [Christeson et al., 2008; Clark et al., 2008; Guedez, 2007; Magnani et al., 2009], the 65W model has a high degree of
lateral heterogeneity and shows considerable Moho topography. Offshore, we image a series of sedimentary basins separated by basement highs. The SCDB, the Los Roques Canyon, Margarita Basin and the Cariaco Trough sedimentary basins are characterized by Vp values ranging between 1.9 and 5.4 km/s. The coincident MCS data reveal that these basins are bounded by faults that put the sedimentary sequences in contact with igneous-metamorphic rocks (Figure 1.10), resulting in a laterally alternating pattern of high and low velocities in our model. Similar patterns have been observed in the neighboring BOLIVAR profiles [Clark et al., 2008; Magnani et al., 2009]. Onshore, there is good correspondence between the location of the Espino Graben/Oriental Basin and Vp values between 3.0-5.5 km/s. The deeper crustal velocities vary laterally and are consistent within tectonic domains. In the following paragraphs we will discuss the characteristics of each of the structures imaged starting in the north with the SCDB and continuing southwards.

At the north of the profile, at model distances -50 to 0 km we image the SCDB near its eastern end. Velocities at the seafloor are 1.9 km/s and increase to 5.4 km/s at a maximum depth of ~11 km. Lower crustal velocities appear to be high compared to the rest of the model reaching 6.9 km/s at the base of the crust; however this result may not be significant due to the relatively poor constraint at the northern edge of the model. The Moho under the SCDB is modeled at a depth of 25 km and average crystalline crustal velocities are 6.3 km/s. Due to lack of ray coverage, the Venezuela Basin, north of the SCDB, is not imaged. Immediately south of the SCDB we image the La Blanquilla High which shows very high crustal velocities even at shallow depths (6.35 km/s at 4.5 km depth) as a result of the very thin or absent sedimentary cover blanketing the igneous
Figure 1.10: Time migrated MCS line along offshore portion of profile 65W (top) with line drawing interpretation of main structures (middle) and line drawing interpretation overlain on the offshore portion of the preferred velocity model converted to time (bottom). There is good correlation of the shallow velocity structure with the sedimentary basins and basement highs observed in the MCS data. Green circles indicate the location of OBS stations.
basement. The average crystalline crustal velocity is 6.4 km/s, the highest along the profile. The Los Roques Canyon between 40 and 75 km in model coordinates is imaged as a 30 km wide basin that deepens steeply towards the south, where it reaches a sediment thickness of 8 km. The average velocity of the sedimentary column is 3.8 km/s, the slowest along the profile. The Moho follows a generally gentle, southward deepening trend throughout the offshore portion of the profile, going from ~26 km depth at model distance 25 km to ~28 km at the coastline (model distance 200 km). The Margarita Basin (model distances 90 to 150 km) also exhibits velocities of 1.9 km/s at the sea floor and shows a sedimentary thickness of ~ 6.5 km. This basin is flanked to the north and south by blocks of high velocity, slightly above 6.0 km/s, that reach shallow depths.

Between 150 and 185 km, the Cariaco Trough is clearly imaged as a narrow, deep, V-shaped basin. At model distance 163 km the basin reaches its deepest point where the contact between the sedimentary sequence (7.6 km thick) and the igneous basement is at 9 km depth and the sedimentary column in 7.6 km thick. The average velocity of the sedimentary column both on the Cariaco Trough and the Margarita Basin is 4.0 km/s and crystalline crustal velocities under the two basins average 6.10 and 6.17 km/s respectively. South of the Cariaco Trough, crystalline crustal velocities greater than 5.5 km/s reach shallow depths near the coastline. As the profile enters mainland, Moho deepens southwards from 27.5 km at the coastline to a maximum depth of 39.8 km at model distance 300 km, about 100 km inland. From its deepest point, the Moho shoals to the south to a depth of ~34 km at the southern end of PmP ray coverage (model distance 375 km). The upper crustal velocity structure on land shows a broad basin that encompasses the Espino Graben as well as the younger overlying sediments of the
Oriental Basin. The boundary between the two is not conspicuous in the velocity structure. The average velocity of the sedimentary rocks on land (4.3 km/s) is greater than the average velocity of the sedimentary cover of any of the offshore basins; reflecting older, more consolidated sedimentary units onshore, and young, less consolidated sediments offshore. The reflection identified as the base of the Espino Graben (Figure 1.2d) is modeled at 10.6 km depth between the 5 km/s and 5.5 km/s contours (Figure 1.6). The average velocity of the SA plate crystalline crust is 6.3 km/s

1.7. Discussion

1.7.1. Southern Caribbean Deformed Belt and Venezuela Basin

The profile images the SCDB near its eastern end, where the sedimentary prism is youngest and least deformed. As noted above, only a small expression of this structure is observed further east [Clark et al., 2008a]. Crustal thickness is high, as there are ~15 km of crystalline crustal velocities beneath the SCDB sediments (Figure 1.6). The unusually large thickness of the oceanic crust could immediately be attributed to the presence of the Caribbean Large Igenous Province [Coffin and Eldholm, 1994]. However; previous seismic reflections studies have concluded that the southeast edge of the Caribbean (including our study area) is floored by normal oceanic crust [Diebold et al., 1981], and a companion BOLIVAR profile at 67°W longitude found oceanic crust of normal thickness ~240 km west of our study area [Magnani et al., 2009]. Although we must acknowledge that data coverage is relatively poor in the northern edge of our profile, our observations are inconsistent with oceanic crust of normal thickness. Additionally, the bathymetry of the Venezuela Basin shows that while the depth of the sea floor is ~5 km where the crust
has been found to be of normal thickness, it is shallower (4.0-4.7 km) along our profile (Figure 1.1). The shallower seafloor suggests the presence of a thicker, more buoyant, crust, in accordance with our observations. Relatively thick oceanic crust (10-11 km) is also found ~125 km east of our study area, There, the excessive thickness of the crust has been interpreted to result from the proximity to the Aves Ridge instead of the presence of the CLIP [Clark et al., 2008a]. The wide-angle seismic observations presented here suggest that the CLIP is present in the eastern edge of the southeast Caribbean.

1.7.2. Island Arc Region

Between the South Caribbean Deformed Belt and the Cariaco Trough, the Moho depth is relatively stable, dipping gently towards the south. In this region, the profile crosses basement highs as well as deep basins. We will discuss this portion of the velocity model in the context of observations across the other BOLIVAR wide-angle seismic profiles.

Immediately south of the SCDB, our profile crosses the extinct volcanic arc of the Leeward and Venezuelan Antilles at the bathymetrically high La Blanquilla High. Based on the lateral continuity of gravity anomalies, Mann [1999] proposed that the Leeward Antilles arc, the La Blanquilla High and the Aves Ridge are part of one continuous volcanic island arc, an hypothesis strongly supported by the similarities in the average 1-D velocity curves of these features as determined by wide-angle seismic data along BOLIVAR profiles 64W, 67W, 70W and TRIN [Clark et al., 2008a; Magnani et al., 2009; Guedez, 2007; Christeson et al., 2008]. The average 1-D velocity profile at the La Blanquilla High along profile 65W is remarkably similar to the velocity structure of the
belt sampled at different locations (Figure 1.11a), lending additional support to the hypothesis of a continuous tectonic belt that includes all the aforementioned structures.

South of the La Blanquilla High, the profile crosses a NW-SE trending trough known as the Los Roques Canyon (Figure 1.1). Little information is available about this basin from the scientific literature. The geometry of the basin, as inferred from the distribution of sedimentary (<5.5 km/s) velocities in the model, appears to be distinctly asymmetric; with the sedimentary cover deepening and thickening to the southwest. The geometry of the sediments filling the narrow trough suggests the presence of a NE-dipping steep fault (Los Roques Fault), bounding the NE edge of the high velocity basement beneath the Margarita Basin. The location of the canyon between two bathymetric highs and its NW-SE orientation suggest a comparison with the West Curacao Basin, which has been imaged by BOLIVAR profile 70W [Guédez, 2007]. The West Curacao Basin formed in the Early-Middle Miocene along with a series of basins that dissected the Leeward Antilles ridge offshore western Venezuela, whose subsidence was controlled by NW-SE trending faults [Gorney et al., 2007]. A comparison of the average 1-D velocity structure of the two basins (Figure 1.11b) reveals great similarity, specially if one accounts for the difference in bathymetry that makes the West Curacao Basin ~2 km shallower than the Los Roques Canyon. One important difference is the depth to Moho, which is significantly greater beneath the West Curacao Basin. The similar sedimentary velocities suggest that the two basins may have been formed by similar processes, and the Los Roques Canyon may be a product of rifting between fragments of the Leeward and Venezuelan Antilles Arc. Extension at the Los Roques Canyon probably occurred later than at the West Curacao Basin, given the generally
Figure 1.11: One-dimensional velocity profiles. (a) Comparison of 1-D velocity profiles averaged across the Leeward Antilles Arc, the La Blanquilla High as imaged by BOLIVAR profiles 64W and 65W, the Lesser Antilles Arc and the Aves Ridge. (b) Comparison of 1-D velocity profiles averaged across the Los Roques Canyon and the West Curaçao Basin as well as the Margarita Basin and Bonaire Basin. (c) Comparison of 1-D velocity averaged over 50 km sections north and south of the main strike-slip boundary along BOLIVAR profiles 64W, 65W and 67W.
eastward younging progression of tectonics in the southeast Caribbean. The difference in crustal thickness between the two basins may explain the difference in bathymetry and thus, the shallower position of the West Curaçao Basin with respect to the Los Roques Canyon. A different interpretation of the nature of the Los Roques Canyon is given by Higgs [2009]. This author proposes that the Los Roques Fault, running along the western flank of the canyon, is the offshore continuation of the Urica Fault. This fault system would have formed part of the CAR-SA plate boundary before the Pliocene by linking the SCDB and the thrust and fold belts of the Serrania del Interior. After the Pliocene, he suggests conditions switched from oblique convergence to pure strike slip; displacing the Los Roques Canyon towards the east and causing the current misalignment with the Urica Fault. This interpretation fails to explain the relationship between the fault and the basin and whether the fault has controlled the subsidence of the basin. The latter seems unlikely, if the fault was indeed a lateral ramp connecting two thrust fronts.

Further south, our profile crosses the Margarita Basin, where deposition began in the Paleogene and resumed in greater volumes in the Neogene after a period of tectonic inversion in the Middle to Late Miocene [Jaimes, 2003]. In the case of the Margarita Basin we look for a reference in the Bonaire basin, located at similar latitudes 250-400 km to the west. The velocities of the sediment column in the Margarita Basin are significantly higher (~30-40%) than in the Bonaire Basin, but the velocities in the igneous-metamorphic crust (Vp > 5.5 km/s) as well as Moho depth, are very similar (Figure 1.11b). This observation indicates that the two basins may be underlain by a similar kind of crust but are probably not genetically linked.
In general, we observe that the crust in this portion of the profile is comparable to the corresponding areas on other BOLIVAR profiles and has been interpreted as a rifted island arc crust [Magnani et al., 2009]. The sedimentary basins overlying the crust appear to be the result of a complex tectonic history involving periods of extension in different directions, as well as episodes of compression, and clockwise rotation [Beardsley and Avé Lallemant, 2007]. In the case of the Los Roques Canyon we find it analogous to other geometrically similar basins along the margin.

1.7.3. South American Continent

As noted above, the Moho deepens southwards from ~28 km at the coastline, reaches a maximum depth of 39.8 km ~100 km inland under the Espino Graben/Oriental Basin, and then shoals again towards the south following a trend that is similar to what is observed ~300 km to the west in BOLIVAR profile 67W [Magnani et al., 2009]. This characterization of the Moho topography differs from the geometry modeled by Schmitz et al. [2005] and Schmitz et al. [2008] along a coincident profile, where the Moho depth monotonically decreases towards the north. Average crustal velocities in the continental portion of the profile are higher than those occurring north of the coastline with the exception of those found in the La Blanquilla High.

The extension that formed the Espino Graben likely weakened and thinned the continental crust that we image along our profile. Crustal thinning has been observed along similar grabens also formed during the opening of the Atlantic (eg. Benue Trough [Fairhead and Okereke, 1990]) and 3D gravity inversion along the Espino Graben itself also suggested the presence of a thinned crust [Durán, 2007]. If we interpret the top of
the crystalline basement to coincide with the 5.5 km/s contour, then crustal thickness in the continental section of this profile is ~25 km, much lower than the ~35 km observed along profile 67W [Magnani et al., 2009]. The implications of this difference in crustal thickness shall be discussed below. Our estimate of the thickness of the sediments onshore is constrained by the velocity structure as well as the modeled depths of the intracrustal reflections observed in the record of the northern land shot. Where the sedimentary column is thickest, the contact with the basement is at ~11 km depth; this is consistent with the 5.5 km/s velocity boundary in the Schmitz et al. [2005] model, as well as the most recent, reflection-seismic derived, estimates for the base of the Jurassic sequence in the Espino Graben of 11 km [Salazar, 2006].

1.7.4. Cariaco Trough and the Strike-Slip Fault System

Lastly, we will address what is perhaps the most provocative result of this study, the nature of the Moho and the crust across the EP strike-slip fault system that constitutes the current plate boundary between the Caribbean and South American plates. BOLIVAR research led to the development of a shear-tear hypothesis for the rupture of subducting Atlantic (oceanic South American) lithosphere from continental South American lithosphere. This hypothesis is based primarily on observations along profile 64W, ~100 km east of our study area, and predicts the existence of the following features [Clark et al., 2008b):

1) A step in the depth of the Moho (deeper in the south) at the strike-slip boundary that juxtaposes Caribbean and South American crust. This step should become more gradual westwards as the crust “relaxes”.
2) A downward flexure of the northern edge of South American lithosphere which should be most pronounced at the site of the tear and decrease progressively westwards. Observations along profile 67W are all consistent with these predictions [Magnani et al., 2009]. Therefore, the expected result along profile 65W would be an intermediate state between what is observed at 64W and 67W; i.e., a significant step in the Moho at the site of the strike-slip boundary and a relatively deep Moho south of the boundary. The results presented in this work differ greatly from those expectations, as our seismic observations are clearly incompatible with the predicted crustal features: A step in the Moho is not observed at the strike-slip boundary and the maximum Moho depth along the profile of 39.8 km is shallower than the 45 km depth observed along profile 67W.

The base of the crust under the major strike-slip system (coincident with the Cariaco Trough) is densely covered by PmP reflections (Figure 1.7) recorded in one OBS station (Figure 1.2a) as well as several Texan stations on land (Figures 1.2b, 1.2c), suggesting that our experiment geometry is sensitive to variations in Moho depth at the strike-slip boundary. If there was indeed a step in Moho depth, the observed travel times for PmP reflections would reflect it. First arrival ray coverage is densest slightly to the north of the strike-slip boundary (Figure 1.6), which allows us to have a high level of confidence in the upper crustal velocities in that portion of the model as well.

The velocity models, and locally averaged 1D velocity profiles on the 64W and 67W models clearly show a step in Moho depth, as well as slightly lower crustal velocities at any given depth on the southern side of the strike-slip boundary, in what is interpreted as South American continental crust (Figure 1.11c). In the model presented in
this study, not only is there no difference in Moho depth across the boundary but the velocity profiles on both sides of the boundary are practically indistinguishable from each other and from the northern 1D profile at 67W (Figure 1.11c). This suggests that the crustal blocks that meet at the El Pilar fault zone at this longitude are similar. We note that a step in crustal thickness occurs about 50 km to the south of the surface expression of the strike-slip fault system.

One possibility is that the Cariaco-Tuy Basin is underlain by a crustal block of Caribbean origin similar to what is interpreted in the 67W profile to be extended island arc crust. This idea in itself is not new. Wells drilled in the Cariaco-Tuy basin revealed that its basement is composed of Cretaceous metavolcanics interpreted to have developed in an island arc setting \([\text{Talukdar and Bolivar, 1982}]\). \textit{Passalacqua et al.} [1995] incorporate a high density “crustal indenter” into their crustal model based on gravity and magnetic data that reaches south of the El Pilar fault. In a recent overview of the margin \textit{Ostos et al.} [2005] classify the Tuy-Cariaco basin as part of the “Venezuelan Platform” Cenozoic allochthon along with the Triste Gulf, north of the strike-slip system along profile 67W.

What our observations suggest, is that the basement of the Cariaco Basin could indeed be a whole crust block that has been separated from the main body of Caribbean Island Arc crust and effectively accreted to South America (Figure 1.12) given the current plate motions as measured by GPS that show that the San Sebastián-El Pilar fault system is the main boundary between the two plates \([\text{Pérez et al., 2001}]\). The mechanism by which this accretion would have occurred is difficult to deduce given the 2D nature of this study and the fact that our profile crosses only the eastern end of this hypothetical
Figure 1.12: Schematic interpretation of the South American – Caribbean plate boundary along profile 65W based on the velocity model of Figure 1.6. Sedimentary basins are shown in white, unconstrained Moho depths are shown as thick dashed lines. Striped area south of the strike slip fault system is interpreted as either an accreted terrane of Caribbean island arc origin or South American passive margin crust. VB – Venezuela Basin; SCDB – South Caribbean Deformed Belt; LBH – La Blanquilla High; LRC – Los Roques Canyon; MB – Margarita Basin; SS-EP – San Sebastián–El Pilar.
crustal terrane. Here we merely present this as a hypothesis that will have to be further developed and tested by future work.

A second possible explanation for the similarity of the velocity and crustal structure across the strike-slip system is that the crust along the northern edge of the South American plate is heterogeneous and exhibits different characteristics in this area than elsewhere along the CAR-SA boundary. The SA continental crust at this location may simply resemble Caribbean Island Arc crust. In any case, it is important to consider that the mismatch of our observations and the predictions of the shear-tear hypothesis do not negate said hypothesis, but rather are most likely the result of the impact of the heterogeneity of the northern South American margin on the dynamic of the shear tear process. The Espino Graben is a first order crustal heterogeneity and, therefore, we are intuitively inclined to assume it must have played some role on the dynamics of the response to lithospheric tearing. On a first glance it is clear that it may help explain why the Moho depth is shallower than expected; since the crust seems to be thinner under the graben than in areas not affected by Jurassic rifting, as outlined above.

The answer to which one, if any, of these two scenarios is correct will have to result from future work that takes into account the distribution of depocenters onshore and offshore, the morphology of the fold and thrust belts of the Caribbean Mountain System and Serrania del Interior, the timing of their uplift and, perhaps, physical and/or digital modeling that explores the possible effects of the Espino Graben on the advancing thrust fronts during the convergence period. Physical modeling has already shown that pre-existing basin morphology affects the geometry of fold and thrust belts [Macedo and
These experiences could provide a starting point for modeling the problem at hand.

1.8. Summary of conclusions

The crust under the SCDB is ~15 km thick, suggesting that the Venezuela Basin in our study area is floored by Caribbean Large Igneous Province crust, and that the CLIP extends to parts of the southeast Caribbean that were previously thought to be floored by normal oceanic crust. The Moho under the La Blanquilla High is ~25 km deep, and the average 1D P-wave velocity profile across this structure is similar to that observed at the Leeward Antilles Arc, and at the Aves Ridge, providing further support for Mann's [1999] hypothesis that these features form a single volcanic belt. The northwest trending Los Roques Canyon has a velocity structure that resembles that of the West Curacao Basin ~430 km to the west. The similar orientation and velocity structure of the two basins suggests that they might be the result of analogous processes, and the Los Roques Canyon may have resulted from rifting within the Venezuelan Antilles – Aves Ridge arc. The crustal structure between the Margarita Basin and the shoreline shows little variation, with a gently dipping Moho from 26.6 km depth under the northern edge of the Margarita basin, to a depth of 27.5 km under coast. We do not observe a sharp contrast in Moho depth and crustal velocities across the El Pilar fault zone as was expected from BOLIVAR work in other areas of the margin. The crustal block immediately south of the El Pilar Fault exhibits a Moho depth and a velocity structure similar to the Bonaire basin at longitude 67°W, interpreted as extended Caribbean island arc crust [Magnani et al., 2009], and it may represent an accreted terrane of Caribbean origin. We observe an
increase in crustal thickness at the southern edge of the Cariaco trough, i.e. the coastline, from \(~28\) km to \(~40\) km. Onshore, the thick sediment column that comprises deposits of the Espino Graben and Oriental Basin has a maximum thickness of \(11\) km. The contact between the Jurassic graben fill and the younger deposits is not evident in the velocity model. The Moho under the continental SA plate has a concave downwards shape and reaches a maximum depth of \(39.8\) km \(~100\) km inland. The continental crust is thinner along our profile than in other areas of the margin, possibly due to crustal thinning associated to the Jurassic rifting of the Espino Graben.
Chapter 2

Gravity inversion using seismically derived crustal density models and genetic algorithms: An application to the Caribbean – South American plate boundary

M. J. Bezada and C. Zelt

Abstract

Crustal density models derived from seismic velocity models by means of velocity-density conversions typically reproduce the main features of the observed gravity anomaly over the area but often show significant misfits. Given the uncertainty in the relationship between velocity and density, seismically derived density models should be regarded as an initial estimate of the true subsurface density structure. In this paper we present a method for estimating the adjustments necessary to a seismically derived density model to improve the fit to gravity data. The method combines the Genetic Algorithm paradigm with linear inversion as a way to approach the non-linear and linear aspects of the problem. The models are divided into three layers representing the sedimentary column, the crystalline crust and the lithospheric mantle; the depths of these layers are determined from the seismic velocity model. Each of the layers is divided into a number of provinces and a density adjustment ($\Delta\rho$) value is found for each province so that the residual gravity (difference between the observed gravity anomaly and the anomaly calculated for the seismically derived model) is minimized while keeping $\Delta\rho$ between predefined bounds. The preferred position of the province boundaries is found through the artificial evolution of a population of solutions. Given the stochastic nature of the algorithm and the non-uniqueness of the problem, different realizations can yield
different solutions. By performing multiple realizations we can analyze a set of solutions by taking their mean and standard deviation, providing not only an estimate of the $\Delta \rho$ distribution in the subsurface but also an estimate of the associated uncertainty. Synthetic tests prove the ability of the algorithm to accurately recover the location of province boundaries and the $\Delta \rho$ values for a known model when using noise-free synthetic data. When noise is added to the data, the algorithm broadly recovers the features that define the known model despite greater standard deviations of the solutions and the occurrence of artifacts in the mean solutions. The algorithm was applied to four profiles across the Caribbean – South America plate boundary. Some general patterns in the distribution of $\Delta \rho$ were observed consistently in the profiles and are correlated with the interpretations of the velocity models: Positive $\Delta \rho$ values in the sedimentary layer, negative $\Delta \rho$ values for island arc and extended island arc crust with an abrupt change to positive values in South American crust, and positive values in the mantle under the continent and island arc with a transition to negative values under the Caribbean oceanic crust.

2.1. Introduction

A powerful and widely used means of studying lithospheric structure and composition is the construction of seismic velocity models from earthquake and controlled source traveltime observations. Given the correlation between P-wave velocities ($V_p$) and densities ($\rho$) it is common practice to convert the velocity models obtained from seismic methods into density models in order to test their compatibility with gravity data. The conversion can be carried out with the many empirical and theoretical $V_p$-$\rho$ relations that are available in the literature, some of the most widely
used ones being the Nafe-Drake curve (polynomial regressions to this curve by Zelt [1989] and Brocher [2005]), relations derived specifically for sedimentary rocks [e.g. Gardner et al., 1974; Hamilton, 1978] and relations specific to crystalline crust and upper mantle rocks [e.g. Christensen & Mooney, 1995; Godfrey et al., 1997; Birch, 1961; Sobolev & Babeyko, 1994] (Figure 2.1). Researchers can choose between these formulas, or a combination of them, according to a priori geologic information and subjective preference. Density models thus derived often produce gravity anomalies that correlate well with the observations, but commonly have significant misfits, even when lithology-specific $V_p$-$\rho$ relations are used and careful measures are taken to account for the effects of pressure and temperature [e.g. Ravat et al., 1999; Korenaga et al., 2001]. This is to be expected, since the data on which these relations are based show significant scatter (Figure 2.1). The variability of the $V_p$-$\rho$ relation is caused by its dependence on a number of factors including mineralogy, composition, presence of melt, and porosity structure [Korenaga et al., 2001]. The effects of variations in these parameters can be so large as to call into question the usefulness of seismic velocities as effective constraints on densities [Barton, 1986]. Therefore, although the $V_p$-$\rho$ relations are certainly valuable, any density model derived using them must be regarded as an initial approximation and the need for subsequent model refinement should be expected. Site-specific $V_p$-$\rho$ relations will be found as a result of refining the density models to fit the observed gravity anomalies, which may help estimate or constrain some of the parameters that control the relation. Additionally, a spatially systematic variation of the $V_p$-$\rho$ relation could help define otherwise inconspicuous boundaries of tectonic units.
Figure 2.1: Nafe and Drake dataset (grey dots) and commonly used $V_p\rho$ relations. Thin black lines represent the Z89 curve $\pm 0.25 \text{Mg/m}^3$. References: Z89 - Zelt, 1989; G74 - Gardner et al., 1974; B05 - Brocher, 2005; G97 - Godfrey et al., 1997; B61 - Birch, 1961; CM95 - Christensen & Mooney, 1995.
Forward modeling is commonly used to refine the density models derived from seismic velocities [e.g. Horsefield et al., 1994; Barton & White, 1997; Hirsch et al., 2009; Klingelhofer et al., 2009; Sumanovac et al., 2009], and is sometimes guided by assumptions concerning the lateral variations in chemical composition, lithology, mineralogy and/or temperature [e.g. Ravat et al., 1999; Korenaga et al., 2001]. While certainly a valid and widely used approach, forward modeling has significant drawbacks when applied to gravity modeling, given the inherent non-uniqueness of the problem. This issue can be addressed by producing a set of different models that incorporate different assumptions. As an alternative to forward modeling, inversion schemes can be applied. However, in order to make the problem suitable for inversion techniques, restrictions are usually imposed on the distribution of the density adjustments [e.g. limited to a single layer: Kauahikaua et al., 2000; Korenaga et al., 2001]. Both of these approaches result in a limited exploration of model space and cannot provide a statistically valid estimate of the range of parameters that can fit the data adequately. Therefore, when using these methods, it is difficult to estimate the errors on the model parameters.

In this paper, we present an alternative approach to this problem based on the Genetic Algorithm (GA) paradigm, which aims to find suitable solutions to a problem by mimicking biological evolution [Vose, 1999]. GAs are a powerful and versatile tool for global optimization and are an efficient method for the exploration of model spaces with local minima. They have been shown to be useful in the analysis and modeling of gravity data both in 2D and 3D [e.g. Boschetti et al., 1997; Roy et al., 2002; Montesinos et al.,
2005; King, 2006; Chen et al., 2006]. The methodology we present here is based on a multi-realization GA that also incorporates linear inversion as part of the optimization process. The goal of this algorithm is to recover the density adjustments \( \Delta \rho \) required to improve the fit of a seismically-derived 2D crustal density model. We have applied this method to four active-source wide-angle profiles of the BOLIVAR project [Levander et al., 2006] across the Caribbean – South America plate boundary.

2.2. Description of the algorithm

2.2.1. Residual anomaly

A 2D initial density model is obtained from a 2D seismic velocity model by using a given \( V_p \)–\( \rho \) relation. In the examples presented in the following sections, we have used a polynomial regression to the Nafe and Drake curve [Zelt, 1989] hereafter referred to as Z89. To calculate the initial gravity anomaly we parameterize the density model as a series of trapezoidal blocks of constant density and use the method of Talwani et al. [1959]. In order to avoid edge effects, the models are padded on either end by laterally extrapolating the densities at the edges of the model for one model length. We obtain the residual anomaly by subtracting this initial calculated anomaly from the observations and removing the mean.
2.2.2. Parameterization

While the initial model can be arbitrarily complicated and can contain an arbitrary number of polygons distributed in any number of layers, in order to make the implementation of the GA feasible we simplify the problem as follows:

a) We divide the model into three layers representing the sedimentary cover, the crystalline crust and the upper mantle. We obtain the depths of the boundaries between these layers from the velocity model and assume these boundaries to be well constrained, and thus, fixed.

b) We assume that each of these layers can be divided laterally into a small number of “provinces” and that the provinces are separated by vertical boundaries. The number of provinces and the location of the boundaries in each layer are independent of the other layers.

c) We assume that $\Delta \rho$ is constant within each province and limited to a certain range ($\Delta \rho_{\text{min}} < \Delta \rho < \Delta \rho_{\text{max}}$).

By making these simplifications, the inversion process effectively results in static shifts in the model densities relative to the values in the initial model, with the shift being constant within a province. The question of the optimal number of provinces per layer will be discussed in section 3.2. The values of $\Delta \rho_{\text{max}}$ and $\Delta \rho_{\text{min}}$ were fixed in this implementation to ±0.25 Mg/m$^3$ based on the distribution of empirical data points about the Nafe-Drake curve (Figure 2.1).
2.2.3. Solution representation and evaluation

Since the general architecture of the GA method has been extensively described elsewhere [e.g. Goldberg 1989; Vose 1999; Sivanandam and Deepa, 2007], we will leave the details of our implementation to appendix A. In this section we will discuss the evaluation operator, which is the central part of the algorithm. Each density adjustment solution can be viewed as consisting of two parts: the geometry of the blocks (provinces) and the value of the density adjustment ($\Delta \rho$) in each block. Although the problem of calculating the gravity anomaly resulting from a 2D body has a non-linear dependence on the geometry of the body, it depends linearly on its density. Therefore, we approach the two parts of the problem in different ways. Since the layer boundaries are assumed to be fixed, the geometry of the blocks is determined by the set of parameters defining the province boundaries. It is these parameters that will represent a particular solution competing for survival in the GA. Using GA terminology, this can be thought of as the solution genotype. Due to the linear dependence of the gravity anomaly on the block's density, once we have defined the geometry of the blocks we can use the formulation of Talwani et al. [1959] to calculate a matrix $G$ that relates the $\Delta \rho$ values to the residual gravity anomaly they produce by $Gm = d_{est}$ where $m$ is a vector containing a given set of $\Delta \rho$ and $d_{est}$ is the corresponding residual gravity anomaly. Conversely, given the values of the residual anomaly $d_{res}$ and the matrix $G$ we can perform a linear inversion to obtain the values of $m$ that minimize $||Gm - d_{res}||$, thereby finding the set of density adjustments that would best explain (in a least squares sense) the observed gravity residual, given the predefined block geometry. Since we do not place constraints on the inversion, the
elements of \( \mathbf{m} \) are free to take values outside the permitted range, so we obtain the vector \( \mathbf{m}' \) by imposing minimum and maximum values (\( \Delta \rho_{\text{min}} \) and \( \Delta \rho_{\text{max}} \)) on the elements of \( \mathbf{m}' \).

The vector \( \mathbf{m}' \) can be thought of as the solution phenotype as it is the density distribution that results from the province geometry and the bounds placed on \( \Delta \rho \). We then calculate the gravity anomaly generated by the reset model by \( \mathbf{d}'_{\text{est}} = \mathbf{Gm}' \), and finally obtain a measure \( f \) of the fitness of the solution by calculating the RMS misfit between \( \mathbf{d}_{\text{res}} \) and \( \mathbf{d}'_{\text{est}} \). The fittest solutions are then those with the smallest \( f \) values, and we will refer to the fittest solution (minimum \( f \)) in a population as the \( \alpha \) solution. We let the solutions compete and evolve for a fixed number of generations and retrieve the \( \alpha \) solution at the end of the run, which we consider to be one realization. We note that, when applying inverse methods, it is often desirable not to minimize data misfit but to obtain a normalized data misfit of unity (misfit equivalent to data uncertainty; [Bevington, 1969]). However in the case of crustal scale 2D gravity modeling, uncorrelated noise stemming from observational error is negligible and other factors, difficult to estimate before the inversion, play a more important role as discussed in section 3.3. It is for this reason that the concept of normalized misfit is not used in our evaluation function.

2.2.4. Analysis of the results

Since the GA method is based on a guided but random process, after a set number of generations, it converges to a different \( \alpha \) solution in each run depending on the seed of the random number generator. Although GA’s are designed to avoid local minima, there is such a high level of non-uniqueness in the problem at hand that many different
combinations of model parameters can fit the data almost equally well. We speculate that if the process was allowed to run indefinitely, the global minimum might be found each time, independent of the random seed used. However, given the slow pace of convergence after a given number of generations we find it more practical and efficient to stop the convergence at a given point and repeat the process a number of times to obtain different approximations to the global minimum. There is an added benefit to this approach in that it allows us to examine the variability in the set of model parameters that fit the data at a given level.

The representation of each solution based on the location of the province boundaries does not allow for the direct comparison of two solutions (two significantly different sets of boundaries could result in very similar density distributions), so we resample the density distributions resulting from the α solutions emerging out of each realization into 1 km wide cells. We can then perform an elementary statistical analysis on the resampled density distributions to obtain the mean Δρ value and standard deviation for each 1 km wide cell. We will refer to this average as the mean α solution.

2.3. Synthetic test

In designing a synthetic model to test the algorithm, we used the geometry of the P-wave velocity model from BOLIVAR profile 67W [Magnani et al., 2009] as a template. The Moho interface was taken directly from the velocity model and the crystalline basement interface was constructed from the 5 km/s contour in the velocity model. We prescribed a series of density anomalies on each of the three layers as follows: We placed positive (0.10 Mg/m³) and negative (-0.17 Mg/m³) density anomalies in the
north and south of the sedimentary layer respectively, a negative density anomaly (-0.13 Mg/m^3) in the crystalline crustal layer near the center of the model and we divided the upper mantle layer into two zones of negative (-0.15 Mg/m^3) and positive (0.15 Mg/m^3) density anomalies in the north and south respectively, with a transition in three steps between the two zones near the center of the model (Figure 2.2). This results in four provinces in the sedimentary layer, three in the crystalline crustal layer and five in the uppermost mantle layer. This distribution of density anomalies was meant to pose a challenge to the methodology since negative and positive density anomalies in different layers of the model coincide laterally, producing competing effects on the residual gravity curve. The synthetic residual anomaly has a peak-to-trough amplitude of ~ 350 mgal and is clearly dominated by the long-wavelength effect of the upper mantle anomalies (Figure 2.2).

\[\Delta \rho \text{ (Mg/m}^3)\]

\[\text{N} \quad -200 \quad -150 \quad -100 \quad -50 \quad 0 \quad 50 \quad 100 \quad 150 \quad 200 \quad 250 \quad 300\]

\[\text{S} \quad 0 \quad 50 \quad 100 \quad 150 \quad 200 \quad 250 \quad 300\]

\(q_g\) (mgal)

**Figure 2.2:** Known model for synthetic tests (bottom) and associated residual gravity (top)
2.3.1. Noise-free residual gravity

For a first test, we use the synthetic residual gravity as input to the algorithm and we set the number of provinces to match those of the known model. We ran the algorithm for 150 generations per realization and 50 realizations. The results are satisfactory in that the mean $\alpha$ solution recovers the location and amplitude of most of the anomalies accurately (Figure 2.3). The distribution of the individual $\alpha$ solutions and the values of the standard deviations indicate that the level of constraint on the anomalies is variable (Figure 2.3). The density anomalies seem to be much better constrained in the south of the model (less variability in the individual $\alpha$ solutions and smaller standard deviations), where there are shorter wavelength lateral variations in the topography of the layer boundaries. The northern half of the sedimentary layer and the northernmost third of the igneous-metamorphic layer show the least constraint on $\Delta \rho$. This is likely due to their small thicknesses that make it easy to compensate a $\Delta \rho$ there with a laterally coincident, small change in the $\Delta \rho$ value of another layer. The RMS misfits of each of the fifty $\alpha$ solutions vary from 0.5 to 1.3 mgals with the mean being 0.8 mgal (Figure 2.4). Meanwhile, the RMS misfit of the mean $\alpha$ solution is 0.4 mgal, better than any of the individual $\alpha$ solutions. Another noteworthy result is that the areas where the anomalies are not recovered correctly are those where the uncertainty is greatest (Figure 2.3). Regarding the lateral boundaries of the anomalies, the northern boundaries of the negative sedimentary anomaly and the positive mantle anomaly, as well as the southern boundary of the crustal anomaly are recovered successfully. In contrast, the boundaries of the positive sedimentary anomaly and the northern boundary of the crustal anomaly are not recovered very well, although sharp gradients or small steps are observed in the mean
α solution at the location of the prescribed boundaries (Figure 2.3). The magnitudes of the mantle anomalies, the negative sedimentary anomaly and even the poorly laterally constrained positive sedimentary anomaly are recovered accurately, while the crustal anomaly is underestimated in the mean α solution by ~20%. The stepwise transition between the two mantle zones is not recovered in the mean α solution. Instead there is a zero anomaly zone that is partially in spatial agreement with the prescribed zero anomaly block, connected to the high and low density anomalies by gradients (Figure 2.3).

**Figure 2.3a:** Results of 50 realizations of the algorithm, using the residual gravity from the known model shown in figure 2.2 and the same number of provinces as the known model in each layer. From top to bottom: Observed (black crosses) and calculated (red line) residual gravity curves, RMS misfit value shown is for mean α solution; mean α solution with the boundaries of the known density anomalies superimposed in gray for reference, and plots of Δρ vs model distance for the sedimentary, crustal and mantle layers. In the 3 bottom plots each of the individual α solutions are shown as thin gray lines, the mean α solution is shown by the red line, the green shaded area shows one standard deviation above and below the mean and the known model is shown for references as the dashed blue line.
Figure 2.3b-e: Same as figure 2.3a but using one fewer province per layer with respect to known model (b), one additional province per layer (c), three additional provinces per layer (d) and five additional provinces per layer (e).
2.3.2. Incorrect number of provinces

This set of tests takes into account the fact that the number of provinces used to subdivide each layer is generally an unknown. It is therefore important to test the robustness of the algorithm with respect to this parameter. We consider four cases: The first one uses one fewer province in each layer with respect to the true model, the second one uses one additional province in each layer and the third and fourth use three and five additional provinces in each layer, respectively.

The results show that, for a fixed number of realizations, the ability of the algorithm to fit the data increases with the number of provinces used. As the number of provinces increases, we see an improvement of the RMS misfit of the mean $\alpha$ solution and, for numbers of provinces matching or exceeding those in the true model, a general reduction of the variability of the sets of solutions. In the case where we assumed one fewer province per layer with respect to the true model, the variability of the $\alpha$ solutions is very low, with the distribution of solutions seemingly bimodal in the sedimentary layer. This low variability is misleading since it gives the impression of a tight constraint even in areas where the true anomaly is not recovered well and results from the limitations imposed by the small number of provinces. In all cases, the RMS misfit of the mean solution is at least as small as the smallest RMS misfit of the individual $\alpha$ solutions (Figure 2.4). Overall, the mean $\alpha$ solutions for all cases contain the main features of the true model and the RMS fit is excellent (0.1-1.0 mgal). The northern boundary of the positive sedimentary anomaly and the magnitude and northern boundary of the crystalline crustal anomaly are the most difficult features to recover. These features become increasingly well resolved as the number of provinces increases and they are recovered.
accurately in the case with five additional provinces per layer. In that case, the northernmost ~50 km of the sedimentary and crystalline crustal layer (where these difficult to resolve features are located) show the largest standard deviations of \( \Delta \rho \), reflecting the fact that a variety of combinations of density anomalies in these two layers can explain the synthetic gravity values in this part of the model. The algorithm is unable to recover the stair-step transition between the low and high density mantle anomalies when more than three additional provinces per layer are added. Instead, the mean \( \alpha \) solution shows a density gradient connecting the two anomalies. A quantitative assessment of the success of the inversions is possible by calculating the L2 norm of the difference between vectors containing the resampled \( \Delta \rho \) values from the true model and each of the individual \( \alpha \) solutions as well as the mean \( \alpha \) solution for each of the cases considered. Figure 2.4 shows that the difference between the mean \( \alpha \) solution and the
true model decreases monotonically with increasing number of provinces per layer. The trends seen in the graphs presented in Figure 2.4 suggest that the addition of a greater number of additional provinces per layer would produce only incremental improvements in the RMS misfit and difference with the true model of the mean α solution.

We conclude that the algorithm can successfully recover a set of density anomalies if the key assumptions we have made are met (i.e. the layer boundaries are known accurately from the velocity model, the residual gravity is caused by static shifts in the $V_p$-$\rho$ conversion and these shifts are constant within a province). It is not necessary to know a priori the number of provinces that each layer should be divided into, as the results are robust with respect to this parameter and satisfactory solutions are found while using a number of provinces that differs from that of the true model. Moreover, it is desirable to use a greater number of provinces than is strictly needed, as this helps the algorithm converge to the correct solution.

**2.3.3. Residual gravity with added noise**

In the next set of tests we explore the robustness of the algorithm with respect to noise to the residual gravity curve. Due to the precision and accuracy of gravity meters, noise in gravity data is practically negligible in crustal scale studies. However, we have assumed that the layer boundaries that we define from the velocity model are accurate, but we should expect some degree of error depending on the quality of the seismic data. Our formulation also ignores 3D effects that could be locally significant, and the simple parameterization we use cannot represent small bodies with anomalous densities within a province. We added noise to the synthetic residual anomaly to simulate randomly
distributed density contrasts not accounted for by our parameterization. Since the intent is not to simulate uncorrelated noise, instead of adding a random value to each residual gravity point, we create a smooth noise curve by placing Gaussian random values with a standard deviation of 8 mgal at equal intervals of 20 km and resample them to match the observation interval of 5 km (Figure 2.5). We repeated the set of tests performed with the noise-free synthetic data set to gauge the effect of the number of tectonic provinces assumed on the set of solutions found. The results show that even after adding noise to the residual gravity curve, the algorithm is able to broadly recover the main features of the prescribed model (Figure 2.6).

As was the case for the noise-free synthetic test, we consider five cases where we varied the number of provinces per layer with respect to the true model from -1 to +5. In all of the five cases, we recovered the negative and positive density anomalies in the mantle. The magnitude of the negative mantle anomaly is more accurate and robust (smaller standard deviations), but that of the positive anomaly is generally adequate.
except in the case with five additional provinces per layer. The stepwise transition is not recovered well, but it is suggested by the mean $\alpha$ solution in the cases with the correct number of provinces and with one additional province per layer. The negative crystalline crustal anomaly is also present in the mean $\alpha$ solution in all the cases considered. The southern boundary of this anomaly is a very robust feature and is recovered accurately in all cases, while the northern boundary is only apparent in the results when fewer than three additional provinces per layer are used. The magnitude of the crystalline crustal anomaly is underestimated by 33 to 69%. The recovery of the anomaly magnitude improves as the number of provinces per layer with respect to the true model increases from -1 to 1, but no additional improvement is seen when a larger number of provinces per layer is used (Figure 2.6). In the sedimentary layer, the northern boundary of the negative anomaly is recovered accurately in all cases but the magnitude is severely underestimated. Only 50% of the magnitude of this anomaly is recovered when using the correct number of provinces, recovery worsens slightly when one additional province per layer is added and is very poor (~25% of the true value) in the cases with a larger number of additional provinces per layer. The positive sedimentary anomaly is generally overestimated (up to 100%). In this case, the estimate does improve consistently with the addition of more provinces per layer. The lateral boundaries of this anomaly are recovered only approximately, but within ~25 km of the corresponding boundaries in the true model (Figure 2.6). Additional, laterally small anomalies not present in the true model appear in the mean $\alpha$ solutions (e.g. at ~200 km, Figure 2.6) as a result of the noise. These artifacts become more pronounced with the addition of provinces.
Overall, although the fit to the data improves, the quality of the mean solution decreases if the number of additional provinces with respect to the true model is greater than one (Figure 2.7). Quantitatively, the norm of the vector difference between the mean \( \alpha \) solution and the true model is smallest when one additional province per layer is considered and increases for a larger number of additional provinces (Figure 2.7). While the additional provinces do not improve the estimate of the true model, they do produce a significant increase in the standard deviation of the individual \( \alpha \) solutions, making the interpretation of the set of solutions more difficult.

**Figure 2.6a:** Results of 50 realizations of the algorithm, using the noisy residual gravity from the known model shown in figure 2.2 and the same number of provinces as the known model in each layer. Panels as in figure 2.3a.
Figure 2.6b-e: Same as figure 2.6a but using one fewer province per layer with respect to known model (b), one additional province per layer (c), three additional provinces per layer (d) and five additional provinces per layer (e).
We can conclude that using a large number of provinces when applying the algorithm to real data degrades the quality of the results and leads to the strengthening of artifacts in the solutions. Our results suggest that many additional provinces do not result in better approximations to the true model and complicate the interpretation of the sets of solutions given the larger scatter produced. These tests suggest that, when using real data, this method would be most successful when the number of provinces assumed in the algorithm exceeds the number required or expected by one or two.

An encouraging result that arises from this set of tests is that by looking at the mean solution and the distribution of the solutions it is possible to discern what the main features of the true model are and what the level of uncertainty is. As elaborated above, the mean solution in all cases contains the transition from low to high density in the mantle, the low density anomaly in the crystalline crust near the center of the model and the high and low density anomalies in the north and south of the sedimentary layer respectively. These are all the features that define the true model, and although their magnitudes are not always recovered well, the polarities are correct and the recovered locations of their lateral boundaries are fairly accurate. In these tests, additional density anomalies not present in the true model are seen in the mean solutions. This implies that, as with most geophysical techniques, attention should be paid to the possible presence of artifacts in the solutions. Laterally small anomalies that occur in areas of the model otherwise known to be geologically homogenous should be considered potential artifacts; meaning not that they should be ignored but that their interpretation should be approached with caution and alternative explanations for the gravity residual they cause should be considered.
Figure 2.7: RMS misfit (a) and difference with respect to known model (b) as a function of the number of additional provinces used in the algorithm for the noise-free synthetic test. Vertical lines extend from the minimum to the maximum value of the individual α solutions, open circles indicate the mean of those values, diamonds show the values corresponding to the mean α solution.

2.4. Application to real data

In this section we describe the application of our procedure to 2D velocity models of four transects across the Caribbean-South American plate boundary produced as part of the BOLIVAR project [Levander et al., 2006]. The four boundary-normal profiles we will consider resulted from the analyses of first-arrival and reflected traveltimes from controlled-source seismic data. The profiles were named according to their approximate longitude: 70W [Guédez, 2007], 67W [Magnani et al., 2009], 65W [M. J. Bezada et al., 2010] and 64W [S A Clark et al., 2008]. The four profiles cross a series of tectonic provinces roughly perpendicularly. These provinces are, from north to south, the Venezuela basin over the Caribbean seafloor, the Southern Caribbean Deformed Belt, the Leeward Antilles extinct volcanic island arc and extended island arc, the strike slip fault
system, the coastal ranges (with the exception of profile 65W) and sedimentary basins on land (Figure 2.8).

The gravity anomaly values over the profiles were interpolated from a 3’ grid of land, ship and satellite measurements [C. Izarra et al., 2005] (Figure 2.8). The data represent Bouger anomaly on land and free-air anomaly offshore. For the initial gravity calculations, the models were parameterized as a set of trapezoids with constant density (Figure 2.9) and padded at the edges as described in section 2.1. The density of each trapezoidal block was determined from the block’s average velocity using the Z89 curve. The location of relative high and low gravity anomalies calculated from the initial models correlate well with the observations (Figure 2.9) but the level of misfit is significant, with original RMS misfits ranging from 23 to 69 mgal (average 51 mgal). Previous attempts to fit the gravity data for two of the profiles relied on forward modeling and constrained the density adjustments to laterally varying depth intervals to avoid the problem of tradeoffs between density anomalies at different depths. These modeling efforts reduced the RMS misfit to a minimum of 11 mgal [Magnani et al., 2009; Bezada et al., 2010].

Based on the experience with the noisy synthetic data set described in the previous section, and the number of tectonic provinces observed on the surface and interpreted from the seismic velocity models, we chose to use six provinces per layer when applying our algorithm to these data. As was the case for the synthetic models described above, we ran the algorithm for 150 generations and 50 realizations. The mean solutions achieve a significant reduction of the misfit with respect to the original density models, reaching RMS values of 4.5 to 7.7 mgal (Figure 2.10).
Figure 2.8: Gravity map of the southeast Caribbean showing the location of BOLIVAR profiles. Data from Izarra et al. (2005) shows Bouguer anomaly on land and free-air anomaly offshore. CAR – Caribbean Plate; SA – South American Plate; VB – Venezuela Basin; SCDB – Southern Caribbean Deformed Belt; LA – Leeward Antilles; SSFZ – Strike Slip Fault Zone.
Figure 2.9: Initial density models for four BOLIVAR profiles across the Caribbean – South American plate boundary: 64W (a), 65W (b), 67W (c) and 70W (d). The density models were derived from P-wave seismic velocity models using the Z89 curve. Observed gravity anomaly shown as black crosses, calculated anomaly as red line. RMS misfits are indicated in the figure.

As expected, the character of the sets of solutions is similar to what is seen in the noisy synthetic test, and there is significant variation in the individual $\alpha$ solutions (Figure 2.10). Based on the noisy synthetic results, we expect to encounter artifacts in the mean $\alpha$ solutions and will focus the analysis of the solutions on the polarity and the location of lateral changes in $\Delta \rho$. In general, in the mean $\alpha$ solutions for each of the four profiles we observe a good correlation between lateral changes in $\Delta \rho$ and the boundaries of tectonic provinces interpreted in the velocity models (Figure 2.11). We will describe the mean $\alpha$
Figure 2.10: Results of 50 realizations of the algorithm for the residual gravity and layer geometries from BOLIVAR profiles 64W (a), 65W (b), 67W (b) and 70W (d). Panels as in figure 2.3a.
Figure 2.11: Mean $\alpha$ solution for BOLIVAR profiles (from top to bottom) 64W, 65W, 67W, and 70W with the corresponding interpretation of the velocity model by Clark et al. (2008), Bezada et al. (2010), Magnani et al. (2009) and Guédez (2007) respectively. S.S.F.Z - Strike Slip Fault Zone; L.I.P - Large Igneous Province; V.I.A - Volcanic Island Arc; E.V.I.A. - Extended Volcanic Island Arc; SCDB - Southern Caribbean Deformed Belt.
solutions in terms of those tectonic provinces and will make some general observations based on the features that are roughly consistent across the different profiles.

In the mean $\alpha$ solution for each of the four profiles, mantle under the volcanic island arc, extended island arc and continent shows consistently slightly higher $\Delta\rho$ than the mantle north of the Southern Caribbean Deformed Belt (Figure 2.11). The transition between the two blocks occurs over a distance of 50 to 100 km and the magnitude of the difference in $\Delta\rho$ is 0.05 to 0.10 Mg/m$^3$ (Figure 2.10). This pattern seems counterintuitive from the perspective of the known variations in $\rho$ and $V_p$ as a function of mantle depletion. We would expect the mantle under the extinct island arc and the continent to be more depleted than that under the oceanic plate. Depletion of mantle peridotite (increase in the Mg#) does not produce significant variation in the P-wace velocity but produces a systematic decrease in density [Lee, 2003]. Therefore, if composition was the only contributing factor we would expect that, relative to a reference $V_p$-$\rho$ curve, depleted peridotite would have a lower density for any given $V_p$, the opposite of what we observe. This suggests that other factors influence the relationship between $V_p$ and $\rho$ in this area. In interpreting this spatial pattern in $\Delta\rho$ we must take into account that, in the seismic models, constraint on upper mantle velocities is limited to the shallowest depths, given the penetration of the Pn phase. It is therefore possible that variations in velocity (and density) at depth not captured in the original models are partially responsible for the $\Delta\rho$ pattern we observe.

The sedimentary layer generally exhibits positive $\Delta\rho$ in all of the four profiles. A notable exception occurs in the area directly above the volcanic island arc crust on profile
67W. Here, the sedimentary cover is very thin, and it is possible that this negative density anomaly results from the influence of the anomaly in the crystalline crust directly below it.

The distribution of $\Delta \rho$ in the igneous-metamorphic crust seems to divide it into three main provinces: The Caribbean oceanic crust, the Leeward Antilles volcanic island arc and extended island arc, and the continental South American crust. The Caribbean oceanic crust shows very large scatter in the individual solutions for all of the four profiles, with the mean $\alpha$ solution in most cases showing a weakly negative $\Delta \rho$ (Figure 2.10). Given the large standard deviations in the solutions and the relatively weak constraint on the northern edge of the velocity models, this set of results is inconclusive.

The extinct Leeward Antilles arc shows a generally negative $\Delta \rho$. It is worth noting that this area of the velocity models is well constrained and that the velocity structure of this volcanic belt was shown to be consistent laterally along the island chain [Magnani et al., 2009; Bezada et al., 2010], suggesting similar lithologies throughout the arc. The negative density anomaly ($\Delta \rho$ values of -0.10 to -0.17 Mg/m$^3$) is consistent in all of the profiles with the exception of 70W where it becomes very weakly positive near the center of the profile. On profile 67W the area covering the volcanic island arc has a stronger anomaly than the extended volcanic island arc crust and the boundaries show good agreement with the interpretation of the seismic velocity models (Figure 2.11). The $V_p$-$\rho$ relation that results from applying this negative $\Delta \rho$ to the Z89 curve differs significantly from empirical observations on magmatic arc rocks [Behn and Kelemen, 2006], producing smaller densities throughout the relevant $V_p$ interval (Figure 2.12). A thorough analysis and interpretation of these results in petrological terms is beyond the
scope of this paper, instead, we will briefly explore the possible causes of this anomaly. If we consider the density and P-wave velocity of some of the mineral components of basaltic and gabbroic rocks (Figure 2.12), we can speculate that a reduction in density could be produced by an above average proportion of quartz and feldspars. However, this would also tend to reduce the seismic velocity, while the Leeward Antilles arc exhibits relatively high velocities in the models [Magnani et al., 2009; Bezada et al., 2010]. In addition to a higher quartz and feldspar content, a shift to the magnesium end member in the orthopyroxene solid solution series might produce the results we observe, since it would decrease the proportion of the very dense ferrosilite and increase that of the relatively light and seismically fast enstatite. Therefore, these results suggest that the Leeward Antilles crustal rocks contain less iron and perhaps a greater percentage of quartz and feldspars than the average island arc crust.

![Figure 2.12: Z89 curve (solid black line) and the same curve shifted in density by -0.15 Mg/m$^3$ (dashed black line) to represent the velocities and densities of the island arc and extended island arc crust as determined by this study. Star represents a reference point on the Z89 curve with $V_p$=6.5 km/s. Velocities and densities of mineralogical components of basalt and gabbro from Bass (1995). ENS – Enstatite; FER – Ferrosilite; FSP – Feldspar (79% Orthoclase, 19% Albite); HOR – Hornblende; MUS – Muscovite; QZ – Quartz.](image-url)
In contrast to the Leeward Antilles chain, the South American crust exhibits a generally positive $\Delta \rho$ (0.05 to 0.10 Mg/m$^3$, with the exception of profile 70W, where the continental crust shows negative $\Delta \rho$ values). An interesting result is that the boundary between the positive $\Delta \rho$ values of the South American crust and the negative values of the volcanic island arc and extended volcanic island arc to the north occurs abruptly and, for profiles 67W and 64W coincides with the location of the strike slip fault system that represents the boundary between the Caribbean and South American plates. Based on differences in Moho depth and velocity structure in the north and south of the strike-slip fault system in these two profiles, the fault zone was interpreted to be the juxtaposition of two distinct types of crust [Magnani et al., 2009; Clark et al., 2008]. This analysis of the residual gravity supports this hypothesis as it shows different $V_p-\rho$ signatures on the opposite sides of the margin. On profile 65W, the boundary between the positive and negative $\Delta \rho$ zones occurs further inland, in spatial agreement with a deepening of the Moho south of the strike slip boundary. Unlike profiles 67W and 64W, profile 65W did not show significant variation of the velocity structure across the strike slip fault system [M. J. Bezada et al., 2010]. Again, the analysis of the residual gravity is consistent with the interpretation of the velocity model since the $\Delta \rho$ signature seems to be continuous across the strike-slip fault system and the step towards slightly positive values occurs ~40 km inland. As a final exercise we can take the mean $\alpha$ solution for each profile and add it to the corresponding seismically derived model to obtain absolute densities (figure 2.13). In some cases this results in unrealistic absolute densities in parts of the models (e.g. the upper crust in the center of profile 65W) meaning that the density adjustments may be too large. As the noisy synthetic tests showed, the amplitude of the density adjustments are
not always recovered correctly which is why we have focused our analysis on the polarity, location and lateral extent of the anomalies.

Figure 2.13: Absolute density models for four boundary-normal profiles obtained by applying the density adjustments from the mean $\alpha$ solutions to the corresponding seismically derived models. 64W (a), 65W (b), 67W (c) and 70W (d). The density models were derived from P-wave seismic velocity models using the Z89 curve. Observed gravity anomaly shown as black crosses, calculated anomaly as red line. RMS misfits are indicated in the figure and differ slightly from those shown in figure 2.10 due to differences in the parameterization.

2.5. Discussion and conclusions

Seismically derived density models should be regarded as initial approximations to the subsurface density structure as they often produce significant levels of misfit with
respect to gravity anomaly observations. The residual gravity (not explained by the seismically-derived model) can be viewed as resulting from a distribution of density anomalies or adjustments ($\Delta \rho$) in the subsurface. Here, we have presented a new method for the analysis of residual gravity curves. The method uses the Genetic Algorithm paradigm in combination with a linear inversion technique to find a simple distribution of density adjustments that will minimize the gravity anomaly misfit. The multi-realization nature of the algorithm allows it to provide a range of different solutions that fit the data at a similar level, which helps to assess the degree of non-uniqueness of the solution at different locations in the model. After multiple realizations have been completed, the mean and standard deviation of all the individual solutions can be taken as an estimate of the $\Delta \rho$ distribution in the subsurface and the associated uncertainty.

Tests showed excellent results when the noise-free synthetic residual gravity calculated from a known model was used as input to the algorithm. In those tests, the lateral boundaries as well as the magnitude of all the prescribed anomalies were recovered accurately. Additionally, the algorithm proved to be robust with respect to the number of provinces per layer used, achieving better results when the number exceeded that of the true model. When noise was added to the synthetic residual gravity to simulate a scenario where the model parameterization was inadequate, artifacts were observed in the mean solutions and there was not a major improvement associated with increasing the number of provinces per layer beyond one more than in the true model. However, even in this sub-optimal scenario, the main features of the true model were recovered. Specifically, the boundaries were recovered more accurately than the magnitude of the anomalies and the polarities of the anomalies were always recovered accurately.
To test the algorithm on real data, we applied it to velocity models and gravity data from the Caribbean-South American plate boundary. For all of the four profiles, a significant amount of scatter was seen in the sets of individual solutions. However, some consistent trends were observed in the mean solutions for each profile. Despite the remaining ambiguity, having multiple profiles across the same tectonic units allows us to make the following general observations: The sedimentary layer showed generally positive $\Delta \rho$ values, reflecting the fact that the $V_p-\rho$ curve used to obtain the initial model may underestimate the density of sedimentary rocks. The lithospheric mantle showed a transition from higher $\Delta \rho$ values in the south to lower values in the north occurring near the Southern Caribbean Deformed Belt near the northern end of the profiles. This observation is difficult to explain in terms of mantle depletion, suggesting that other factors must play an important role. The crust of the island arc and extended island arc parts of the profiles generally show negative $\Delta \rho$ values suggesting that the Leeward Antilles have lower iron content and possibly a higher proportion of quartz and feldspars than average island arc crust. Continental South American crust showed generally positive $\Delta \rho$ values. The transition between this positive $\Delta \rho$ regime and the negative $\Delta \rho$ regime to the north occurs abruptly and coincides with the strike slip plate margin for profiles 64W and 67W, supporting the interpretation of the velocity models that places different types of crust on either side of the fault zone [Magnani et al., 2009; Clark et al., 2008]. For profile 65W, the step to positive $\Delta \rho$ values occurs inland from the plate margin, also in accordance with the interpretation of the seismic velocity model [M. J. Bezada et al., 2010].
The application we have presented serves as an example of the usefulness of our algorithm in the analysis of residual gravity data. Due to the reliance of the algorithm on the layer boundary depths defined by the velocity model, the robustness of the results will improve with increasing constraint on the velocity models. Given the inherent non-uniqueness of gravity problems, we consider the ability of the algorithm to yield a range of different solutions its most important feature. In the application we present here we have used the mean of the set of individual realization solutions as an estimate of the subsurface distribution of $\Delta \rho$. This approach proved to be successful in the synthetic tests we carried out, but different alternatives are also valid. For example, an interpreter may choose to deviate from the mean solution when warranted by additional geologic information. The advantage of having a set of possible solutions is that they provide an idea of the acceptable model space from which to proceed. We acknowledge that many simplifications were made in order to make the algorithm work, but consider it a step in the right direction, and an improvement over the reliance on forward modeling or single-layer inversion schemes.
Chapter 3

Subduction in the Southern Caribbean: Images from finite-frequency P-wave tomography.

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Abstract

The eastern boundary of the Caribbean plate is marked by subduction of the Atlantic under the Lesser Antilles. The southeastern plate boundary is characterized by a strike-slip margin while different configurations of subduction of the southwest Caribbean under South America have been proposed. We investigate the slab geometry in the upper mantle using multiple-frequency, teleseismic P-wave tomography. Waveforms from P and PKIKP phases from 285 (Mb > 5.0) events occurring at epicentral distances from 30° to 90° and greater than 150° were bandpass filtered and cross-correlated to obtain up to three sets of delay times for each event. The delay times were inverted using approximate first Fresnel zone sensitivity kernels. Our results show the subducting Atlantic slab, as well as a second slab in the west of the study area that we interpret as a subducting fragment of the Caribbean plate. Both slabs have steep dips where imaged and can be traced to depths greater than 600 km. These results are consistent with transition zone boundary topography as determined by receiver function analysis. The Atlantic slab extends continent-ward south of the plate bounding strike-slip margin. We interpret this extension as continental margin lithospheric mantle that is detaching from beneath South America and subducting along with the oceanic Atlantic slab. The steep subduction of the Caribbean occurs ~500 km landward from the trench, implying an initial stage of shallow
subduction as far to the east as the Lake Maracaibo - Mérida Andes region, as has been inferred from intermediate depth seismicity.

3.1. Introduction

The Caribbean is a relatively small plate that originated in the Pacific, and now lies between the North and South American plates. It has a complicated tectonic history that includes episodes of voluminous magmatism in the Cretaceous that generated the Caribbean Large Igneous Province [Coffin and Eldholm, 1994], a subduction polarity reversal [e.g. Burke, 1988; Pindell and Dewey, 1982], clockwise rotation after collision with North America at ~55-60 Ma [Pindell and Kennan, 2007] and highly oblique collision with South America for the past ~55 Ma [Dewey and Pindell, 1986; Rosencrantz et al., 1988]. Motion of the Caribbean plate relative to a fixed South America has a magnitude of ~20 mm/yr in the southern Caribbean. The orientation of relative motion is E-W offshore eastern Venezuela while in the western Caribbean there is a small but significant (~5mm/yr at San Andrés island, Figure 3.1) southward component. As a result, NW-SE compression between the two plates occurs offshore the Colombian coast [Calais and Mann, 2009] (Figure 3.1). The nature of the Caribbean-South America plate boundary varies spatially according to the angle between the boundary and relative motion: The eastern plate boundary is defined by the Lesser Antilles subduction zone, the boundary between the Southeastern Caribbean and South America is characterized by strike-slip motion along the San Sebastian-El Pilar fault system, and in the southwest Caribbean
Figure 3.1: Tectonic setting of our study area (top): Red arrows show GPS vectors from a compilation by *Calais and P. Mann* [2009], plate boundaries from the UTIG plates database. SA – San Andrés island; LA – Leeward Antilles; LM – Lake Maracaibo; SS-EP – San Sebastián-El Pilar fault system; BF – Boconó fault system; BB – Barcelona Bay; OR – Orinoco River. Red asterisk shows location of Kick’em Jenny volcano. Blue rectangle shows the area of our focus and location of the bottom panel. Station distribution and seismicity (bottom): Symbols representing seismic stations and earthquake epicenters as described in the figure. BN – Bucaramanga nest; PC – Paria cluster; MA – Mérida Andes; PR – Perijá Range; SM – Santa Marta Massif; SF – Serranía de Falcón. Seismicity from the FUNVISIS catalog.
different configurations of shallow subduction of the Caribbean plate under South America have been proposed [e.g. van der Hilst and Mann, 1994; Kellogg and Bonini, 1982] (Figure 3.1)

The Atlantic (oceanic extension of the North and South American plates) subducts under the Caribbean at the Lesser Antilles island arc. The southernmost historically active volcano on the arc is the Kick’em Jenny, an underwater volcano located 8 km offshore northern Grenada and ~240 km from the Venezuelan coastline along the arc trend [Watlington et al., 2002] (Figure 3.1). The Benioff zone defined by intermediate depth seismicity along the arc reaches a depth of ~200 km, and its southern terminus is the Paria cluster of seismicity (Figure 3.1). This cluster has been interpreted as the site of lithospheric tearing [Clark et al., 2008], and as defining a steeply subducting slab [Russo et al., 1993]. VanDecar et al. [2003], using non-linear teleseismic tomography, imaged a landward continuation of the Atlantic slab onshore South America. Based on their images, they proposed a model for shear tearing of subducted passive margin lithosphere from the stable South American continental lithosphere. The landward continuation of the subducting slab, beyond the island arc had previously been inferred based on the outcropping of Plio-Pleistocene intrusive rocks onshore Venezuela [McMahon, 2001; Santamaria and Schubert, 1974] and the location of hot springs that may be associated with cooling of a shallow intrusive body [Urbani, 1989].

In a relatively small region at the southeast corner of the Caribbean, there is a transition from subduction to transform motion that has been interpreted as a Slab-Transform Edge Propagator (STEP) fault [Govers and Wortel, 2005]. Offshore eastern and central Venezuela, the plate boundary is clearly defined at the dextral San Sebastián-
El Pilar strike-slip fault system, with ~80% of the plate motion in eastern Venezuela being accommodated in an 80 km wide shear zone centered on those faults [Omar J Perez et al., 2001]. Further west, the plate boundary is more diffuse and is complicated by the presence of the Southern Caribbean Deformed Belt that extends from offshore western Colombia to offshore eastern Venezuela and the northeast trending Boconó strike-slip fault system (Figure 3.1). Gravity and seismic reflection studies have indicated large accumulations of sediments along the Southern Caribbean Deformed Belt, and the sedimentary sequence shows significant compressive deformation and faulting [Case et al., 1990; Magnani et al., 2009; Talwani et al., 1977]. The Southern Caribbean Deformed Belt has been interpreted as the boundary between the Caribbean and South American plates and the site of Caribbean subduction under South America [Kellogg and Bonini, 1982]. Under northwestern Colombia, several authors have used intermediate depth seismicity to outline a subducting fragment of the Caribbean plate. This subduction is thought to have a shallow dip of 20° to 30° towards the ESE [Malave and Suarez, 1995; Ojeda and Havskov, 2001; Pennington, 1981] or SE [Kellogg and Bonini, 1982] with the subduction angle possibly steepening at depths greater than 150 km [Pennington, 1981]. This shallow subduction of the Caribbean plate is thought to be the driving force behind Laramide-style deformation and uplift in the overriding South American plate, specifically in the Santa Marta Massif, Perijá Ranges and Mérida Andes [Kellogg and Bonini, 1982] possibly forming an orogenic float [Audemard and Audemard, 2002]. A different interpretation of Caribbean subduction was put forth by van der Hilst and Mann [1994] on the basis of seismic tomography images. They proposed a shallower subduction of 17° toward the SSE with the Caribbean subducting beneath the northern
Columbia and northwestern Venezuela to a depth of 225 km. Thus, there are two views of subduction of the Caribbean under northwestern South America; one view calls for shallow subduction from the west, the other calls for shallow subduction from the north.

In this paper, we investigate the slab geometry in the upper mantle using finite-frequency P-wave tomography. The data consist of teleseismic delays obtained mainly from the BOLIVAR (Broadband Ocean-Land Investigation of Venezuela and the Antilles arc Region [Levander et al., 2006]) passive seismic array and supplemented with data collected more recently in the west of the region. The study area covers the southeast Caribbean margin and northernmost South America from the Guayana Shield in the south to the Leeward Antilles in the north and from eastern Venezuela to the Venezuela - Colombia border in the west. Our results show the subducting Atlantic slab, as well as a second slab in the west of the study area that we interpret as a subducting fragment of the Caribbean plate. Both slabs have steep dips where imaged and can be traced to depths greater than 600 km. These results are consistent with transition zone boundary topography as determined by receiver function analysis.

3.2. Data

The data were collected by the BOLIVAR seismic array which included the 39 stations of the Venezuelan National Seismological Network as well as a temporary (~17 month) deployment of 35 IRIS-PASSCAL, 8 Rice broadband stations and data from a year long deployment of 11 OBSIP ocean bottom seismographs (Figure 3.1). The array covers the eastern part of Venezuela with typical station spacing of ~50-100 km while coverage is sparser in the west of the study area. There, we rely on the stations of the
Venezuelan permanent network and a later deployment of 8 PASSCAL and Rice instruments along a north-south profile at ~70°W. The array covers an area of ~900,000 km² from south of the Orinoco river at latitude of ~6°N, along the northern Guayana Shield, to the Caribbean Sea at latitudes of up to 14°N; and from eastern Venezuela at a longitude of 61.5°W to the Perijá range in western Venezuela at a longitude of 72°W (Figure 3.1).

Events with Mb ≥ 5.5 and epicentral distances between 30-90° and 150-180° were initially examined for this study. To fill under-sampled backazimuths, smaller (5 ≤ Mb < 5.5) events were examined, and those with consistent waveforms across the array were selected and included in the analysis. Delay times from P and PKIKP phases from 285 events were included in the final inversion. The event distribution provided good azimuthal coverage and there is a notable abundance of near-antipodal events from the Sumatra-Andaman region (Figure 3.2). Waveforms from the selected events were bandpass filtered at three narrow frequency bands centered on 1.0, 0.5 and 0.3 Hz. Traveltime delays were obtained by cross-correlation of the filtered waveforms so that up to three sets of delays were obtained for each event, bringing the final number of delay times used in the inversion to 7273 (2484, 2453 and 2436 delay times for the 1.0 Hz, 0.5 Hz and 0.3 Hz bands; respectively).

3.3. Method

The multiple-frequency tomography we use differs from purely ray-theoretical tomography in that it approximately accounts for shifts in arrival times caused by velocity perturbations located off the geometrical ray path. This approach allows more
Figure 3.2: Location of events used for this study in polar projection centered on the array (left) and on the array antipode (right). The epicentral distance from the array is labeled on the figure. Dashed lines represent plate boundaries.
thorough use of high-quality broadband seismograms and imposes physically based smoothness criteria on the inversion thereby reducing the need for ad hoc regularization of the inverse problem. In this study, we use the method of Schmandt and Humphreys [2010] that is based on the single scattering “Banana-Doughnut” formulation of Dahlen et al. [2000] but only considers the sensitivity within the first Fresnel zone and applies ray-normal smoothing to address ray-location uncertainty.

The model is defined as an irregular rectangular mesh. Cell size is variable to accommodate the loss of resolution with model depth as well as towards the edges of the array. Horizontal mesh node spacing is smallest inside the array footprint, where it is 42km, and increases stepwise until it reaches 56 km 400 km outside the array. Cell height is also variable from 24km for the shallowest layer of cells (36 to 60 km depth) to 60 km for the deepest layer (685 to 745 km depth).

To avoid mapping the effect of crustal velocity structure into the mantle, we apply a crustal correction to the delay times. Ray-theoretical correction times are calculated using a crustal model derived from 2D active-source P-wave velocity models, 3D receiver function studies and 3D gravity data. The model consists of linear velocity gradients from the surface to the igneous-metamorphic basement, from the basement to the Moho discontinuity and from the Moho to a depth of 50 km. Basement depths were estimated using gravity anomaly data [Izarra et al., 2005] and controlled with active source seismic profiles [Christeson et al., 2008; Clark et al., 2008; Guédez, 2007; Magnani et al., 2009; Schmitz et al., 2002; Bezada et al., 2010] and published estimates of basement depth based on potential field data [Feo-Codecido et al., 1984]. Moho depth was estimated by interpolating the depths determined by active source seismic profiles
and selected Moho depths obtained through a receiver function study [Niu et al., 2007; Bezada et al., 2006].

In addition to model slownesses, the inversion solves for station and event terms. The purpose of the station terms is to absorb the effect of shallow velocity anomalies improperly accounted for by the crustal model. In order to prevent it from also absorbing the effect of our target mantle heterogeneities we introduce a damping parameter. The event terms aim to remove the dependence of an event's mean arrival time on the set of stations that recorded it. We regularize the inversion by including model norm damping as well as smoothing operators. Model norm damping drives the inversion towards the minimum structure solution, minimizing the occurrence of anomalies not strongly required by the data. The smoothing operator we apply is depth dependent, so that it is strictest in the plane normal to the mean ray path at each depth.

The cost function we minimize is then [Schmandt and Humphreys, 2010]:

\[ c = \| A m - d \|^2 + k_1 \| L m \|^2 + k_2 \| m \|^2 \]

Where \( A \) is the partial derivative matrix that relates the model parameters \( m \) (velocity perturbations in each voxel, event and station terms) to the observed delay times \( d \), \( L \) is the matrix that defines the depth dependent smoothing operators, and \( k_1 \) and \( k_2 \) are free parameters defining the relative weight of the smoothness and norm damping constraints, respectively. A solution \( m \) that minimizes \( c \) is found using the iterative LSQR method [Paige and Saunders, 1982].

The selection of the values of the free parameters is somewhat subjective, but was guided by the well-known L-curve that shows the trade-off between model norm and variance reduction with varying damping coefficients \( k_2 \). The addition of the
smoothness constraint means that an L-curve can be obtained for each smoothness parameter ($k_1$).

Figure 3.3 shows three of these curves that behave as expected (more smoothing results in slightly larger model norms and smaller variance reductions). We chose a moderate value of $k_1$ and a value of $k_2$ near the turning point of the corresponding L-curve. Different values of these free parameters did not significantly modify the shape or amplitude of the important and well resolved anomalies in the model (Figure 3.4).

We use the AK135 [Kennett et al., 1995] 1D velocity model as our reference model and to calculate the ray paths and sensitivity kernels.

Figure 3.3: Trade-off between variance reduction and model norm for $k_2$ values ranging from 2 to 24 at an interval of 2 and different $k_1$ values as indicated in the figure. The larger symbol corresponds to the values used in the inversion: $k_1=15$ and $k_2=12$. See text for a description of the $k_1$ and $k_2$ parameters.
Figure 3.4: Effect of free parameters on the model. Model slices at 95 km depth. The preferred model is shown in the center panel.
Figure 3.4 (continued): Effect of free parameters on the model. Model slices at 375 km depth. The preferred model is shown in the center panel.
Figure 3.4 (continued): Effect of free parameters on the model. Model slices at 685 km depth. The preferred model is shown in the center panel.
3.4. Model

Here we describe the location, shape and amplitude of the significant velocity anomalies we observe, as well as their correlation with tectonic features. At lithospheric depths, we observe laterally small (~100 km in diameter) and spatially consistent anomalies. Deeper in the mantle we observe the two most striking features of the velocity model: Large, steeply dipping high velocity anomalies (2.5 to 4%) that we interpret as subducted Atlantic and Caribbean slabs.

In the shallowest 200 km of the model, we find a clear pattern of low velocity anomalies under the Caribbean plate, except under the islands of the Leeward Antilles which show very shallow (upper 100 km) neutral to positive (1%) velocity anomalies (Figure 3.5). North-South oriented shallow low velocity anomalies (-1.5 to -2.5%) also characterize the area directly east of Lake Maracaibo, east of the Mérida Andes, and beneath the uplifted Serranía de Falcón. This anomaly extends to ~200 km depth. The most prominent feature in western Venezuela is a high velocity anomaly (up to ~2.5%) in the very northwest of the study area, near the Perijá range (Figure 3.5) that is concentrated beneath the westernmost station in our array but spreads out beneath Lake Maracaibo with depth.

A negative anomaly is observed in central Venezuela south of the Barcelona Bay extending as far south as the Orinoco River. South of the Orinoco and east of 64°W the Guayana Shield is characterized by a small (~1%) positive velocity anomaly that decreases in lateral extent with depth and reaches depths of ~245 km (Figure 3.5). The western part of the shield (south of the Orinoco and longitudes 68°W to 64°W) shows no significant velocity anomalies. Given the distribution of stations along the northern edge
Figure 3.5: Velocity anomalies in the shallowest 245 km of the model. The color scale is saturated at ±1.5% to highlight the Guayana Shield anomaly.
of the shield, the anomaly pattern we recover suggests that the northern limit of the cratonic root under the eastern part of the shield reaches at least as far north as the Orinoco River, and possibly ~100 km further. In contrast, in the western part of the shield, the cratonic root does not seem to extend as far north as the Orinoco River. The pattern of shallow mantle anomalies we present, and particularly the lack of a high velocity anomaly under the northern edge of the western Guayana Shield, is consistent with a recent S-wave velocity model derived from Rayleigh wave tomography [Miller et al., 2009]. The latter model lacks the low velocity anomalies directly east of Lake Maracaibo, but suffers from poor resolution in that part of the study area. All of the anomalies described above become insignificant at depths greater than ~245 km.

The two main features of the model, and the ones we will focus on for the remainder of the paper, are vertically continuous positive anomalies located in the east and west of the study area. In the east, we image the subducting Atlantic slab while in the west we image a steeply dipping slab that we interpret as a subducted fragment of the southernmost Caribbean plate (Figure 3.6). The Atlantic slab is visible from shallow depths at the eastern end of the study area and at 95 km depth follows the trend of the Lesser Antilles Arc. The northern edge of the anomaly is 100 km SSW of the island of Grenada and the southern end is located onshore, 110 km southwest of the El Pilar fault. The Atlantic slab anomaly widens with depth from ~95 km at 115 km depth to ~150 km at 460 km depth, possibly as a result of decreased resolution. At depths greater than 285 km, the anomaly has a distinct “L” shape, with one arm trending Northeast-Southwest along the subducting Atlantic plate and the other trending East-West, parallel to the plate boundary faults. The anomaly dips towards the west at an angle of ~70°. The magnitude
Figure 3.6: Horizontal slices through the model at depths indicated in the figure (top two rows), vertical slices through the model at location indicated in top right panel (third row), and 3D image of the subducted slabs (1.5% isoanomaly surface) viewed from the south. ATL – Atlantic slab; CAR – Caribbean Slab.
of the anomaly reaches 3.2 to 4.2% at its core with the E-W arm of the anomaly consistently showing amplitudes ~1% smaller than the NE-SW arm.

The second important positive velocity perturbation occurs in the west of the study area. It becomes well resolved at a depth of ~200 km and appears connected to the shallower anomaly surfacing beneath the Perijá range. This feature dips steeply towards the ESE and is clearly visible to a depth of over 600 km, though it weakens and vanishes by a depth of 700 km, most likely due to lack of resolution as it is near one side of our seismograph array. The anomaly has an approximately SSW-ENE trend, but its shape is not very well defined given the poor station coverage in the area. This Caribbean slab anomaly has smaller peak amplitude, compared to the Atlantic slab, of 2% to 2.7%.

### 3.5. Model Assessment

To gain insight into the resolution of our model, we performed a number of synthetic tests in which we prescribed a velocity anomaly corresponding to a predefined slab geometry. Synthetic delay times calculated from these models were used as input to our inversion scheme and the recovered model was used to evaluate the success of the inversion. The object of these tests is to assess the robustness of the geometry of the features in our model, specifically their lateral extent and the maximum depth.

In our model, at a depth of 95 km, the northern edge of the Atlantic slab occurs ~100 km south of the island of Grenada, coinciding with the northern limit of our array. Obviously, we would expect the slab to continue northward under the island chain. To verify that this boundary results from the array aperture, we carried out a synthetic test where the input model had the geometry of the Atlantic slab. The model contained a
westward dipping (70°) +4% velocity anomaly following the trend of the island arc and continuing landward under the continent. Synthetic delay times through the model were calculated using our receiver geometry and, after adding random noise with a standard deviation of 200 ms, these times were inverted using the same free parameters as in our final inversion. The result shows that, at a depth of 95 km, the positive anomaly is only recovered as far north as 11.75°N, consistent with our image (Figure 3.7). We performed a similar test to investigate if horizontal smearing could be responsible for the landward continuation of the Atlantic slab. In this case, the positive anomaly in the prescribed model only extended as far south as the strike slip fault system (10.5°N). After adding noise and inverting the synthetic delay times, we recovered the southern edge of the anomaly at its prescribed location. No significant horizontal smearing of the anomaly was observed (Figure 3.7).

The southwestern end of the study area is the site of one of the more interesting features of the model, yet it is sparsely covered by stations. Given the sparse station coverage, we ran a synthetic test in order to explore how well we could expect to recover the geometry of a large anomaly in this part of the study area. Outside of our station coverage, the input model for this test was based on the slab depths inferred from seismicity by Malave and Suarez [1995] and was defined as follows: The 3% positive anomaly dips towards the ESE and the dip increases stepwise from 20° at the trench offshore western Colombia to 30° and finally 80° at depths greater than 245 km, under the western edge of our array. The width of the synthetic subducting slab extends farther to the northeast and southwest than the anomaly recovered in our model. We find that the inversion successfully recovers the northeastern extension of the prescribed slab while its
Figure 3.7: synthetic tests of northward (top two rows) and southward (bottom two rows) extension of the Atlantic slab and comparison with the velocity model. The northern limit of the recovered anomaly is determined by the station distribution. Southward extension of the slab is recovered accurately, no significant horizontal smearing is observed.
southwestern extension seems to be determined by station coverage; as the prescribed anomaly was not adequately recovered beyond the southern limit of the anomaly in our model (Figure 3.7).

We test the prospect of vertical smearing on our velocity anomalies by attempting to recover synthetic anomalies meant to replicate the subducting slabs in our model but truncated at a depth of 450 km. After adding noise and inverting the delay times we obtain models that resolve the maximum depth of the anomaly to within ~50 km for both anomalies, suggesting that significant vertical smearing of the anomalies is not likely (Figure 3.9).

Finally, the conspicuous absence of a Caribbean slab shallowly dipping southward beneath northwestern Venezuela prompted us to conduct tests to see if such a feature could be resolved with our data set and method. In this case, the input model was meant to replicate the geometry of the Caribbean slab as interpreted by van der Hilst and Mann [1994]; dipping to the SSE at an angle of 17° and reaching a depth of 250 km, with a velocity anomaly of +3%. The inversion of the synthetic delay times with added noise does not adequately recover the prescribed anomaly. Instead, we see a pattern of small positive anomalies in the area above the prescribed slab and small negative anomalies to the east of the slab (Figure 3.10). The shallow dip of the slab may be the main reason why our experiment geometry seems to be insensitive to such a feature.

It is worth noting that resolution tests such as the ones we have conducted are inherently optimistic. The formulation used in the inversion process assumes the subsurface consists of elastic, isotropic media, and that ray paths are accurately known. Because the sensitivity kernels are used to produce the synthetic delay times, these
Figure 3.8: Synthetic test of lateral extension of the Caribbean slab and comparison with the velocity model. The northeastward extension of the anomaly is well resolved but resolution is lost to the southwest of Lake Maracaibo.
Figure 3.9: Synthetic test of vertical smearing of slab anomalies and comparison with velocity model. The maximum depth of the synthetic anomalies is recovered well, no significant vertical smearing of the anomalies is observed.
Figure 3.10: Synthetic test of data sensitivity to a shallow slab such as described by Van der Hilst and Mann [1994] and comparison with velocity model. The prescribed velocity anomaly is not recovered well in this case. The positive anomaly seen in the velocity model at depths greater than 300 km is the Caribbean slab intersecting the C-C' plane from the west.
assumptions are met in the tests. However, anisotropy has been documented in the upper mantle in the Caribbean region [Growdon et al., 2009; Masy et al., 2009; Piñero-Feliciangeli and Kendall, 2008; R. M. Russo et al., 1996], some degree of anelasticity is likely and there is uncertainty in the ray paths stemming from hypocenter mislocations and structure that is out of the volume we analyze. Therefore, these tests represent a best-case scenario and place an upper bound on the expected resolution. As a result, the most robust conclusions we can draw about these tests are those pertaining to which features are irresolvable with our experiment geometry. With the above stated caveat, we conclude that the velocity anomalies central to our analysis are well resolved both vertically and laterally to the degree required by our interpretation. The loss of resolution in certain areas of the model demonstrated in the synthetic tests was taken into account in the interpretation process as discussed in the following section.

3.6. Discussion

The new images of mantle structure we present here warrant reinterpretation of subduction processes in both the southeastern and southwestern Caribbean. In this section we discuss our results in the context of previous studies and explore the implications of the subduction geometries we define for regional tectonic models.

3.6.1. Absence of southward dipping slab

In the previous section we established that our experiment geometry is largely insensitive to a very shallowly southward dipping Caribbean slab such as the one described by van der Hilst and Mann [1994]. The signature of a shallowly dipping slab
would be weak positive anomalies in the area above the slab and weak negative anomalies elsewhere (Figure 3.10), whereas, our model shows low P velocity anomalies at shallow depths under northwestern Venezuela, above the presumed location of the shallow slab (Figure 3.10). While inconclusive in this regard, our study does not support the existence of shallow subduction from the north. Although an early analysis of a small subset of BOLIVAR active source data seemed to be consistent with the presence of a shallowly subducted slab in northwestern Venezuela [Bezada et al., 2008], the full analysis of these data [Guédez, 2007] as well as those of other boundary-normal profiles [Clark et al., 2008; Magnani et al., 2009] showed a continuous Moho across the Southern Caribbean Deformed Belt and beneath the Venezuelan archipelago, but no features that would indicate a continuous Caribbean Plate subducting beneath South America from the north. Similarly, a PdS receiver functions study found no evidence for such subduction [Niu et al., 2007]. Lastly Rayleigh wave tomography shows low to intermediate upper mantle shear velocities in this region, relative to further north in the Caribbean, and further east beneath South America from the Moho to ~130 km depth [Miller et al., 2009]. We conclude that the evidence does not support a continuous slab subducting southward beneath northwestern Venezuela.

### 3.6.2. Atlantic subduction

A very robust feature of this model is the landward continuation of the Atlantic slab south of the island arc. This feature had been initially imaged by VanDecar et al. [2003], and had been inferred since the 1970s due to the presence of igneous bodies onshore eastern Venezuela. The Carúpano Rhyolites ~15 km inland from the eastern
Venezuelan coast are Plio-Pleistocene igneous rocks with an age of ~5Ma whose geochemical signature is consistent with bulk mixing of a slab derived source and average continental crust [McMahon, 2001; Santamaria and Schubert, 1974]. Even more recent magmatism is inferred from the presence of thermal springs near the El Pilar fault system, south of where the Carúpano Rhyolites outcrop. The ongoing cooling of a shallow magmatic body is thought to be the heat source for the hot springs [Urbani, 1989]. Additional seismic evidence for the landward continuation of the slab comes from a recent surface wave study. That S-wave velocity model shows a positive anomaly at 80-125 km depth that correlates spatially with our image of the Atlantic slab, although it was originally interpreted as part of the South American lithosphere [Miller et al., 2009].

Clark et al. [2008] proposed that the strike slip plate boundary produced a clean tear through the entire lithosphere, a process that would be currently concentrated at the Paria cluster of seismicity. This tear would separate subducting Atlantic lithosphere from the buoyant South American lithosphere and mark the southern end of the subduction. The distribution of intermediate depth seismicity was cited as strongly supporting this hypothesis, as it is absent south of the strike-slip boundary. The body wave tomography data showing the slab south of the strike slip fault in the tear region suggest a more complicated mechanism for detaching the Atlantic lithosphere from the South American lithosphere than a clean tear.

The Paria cluster is likely the result of water released by the subducting oceanic plate, and high strain rates. Noting that traditional fault mechanics do not predict brittle faulting below 50 km depth except under very high strain rates, in the last decade several publications have pointed to dehydration embrittlement as a main cause for intermediate
depth seismicity [e.g. Jung et al., 2004]. It is therefore possible that the absence of intermediate depth seismicity south of the Paria cluster results from a lack of hydrous minerals in the subducting lithosphere there and deformation occurring by mechanisms in the ductile rather than brittle regime. Adding to the concentration of seismicity in the Paria cluster could be a concentration of deformation and resulting elevated strain rates at this point in the slab, where oceanic and continental margin lithosphere meet.

To explain the L shaped anomaly we can consider northward subduction of South American passive margin lithosphere as suggested by [VanDecar et al., 2003]. However, if the passive margin of South America had subducted towards the north we would expect the E-W arm of the “L” shaped anomaly in our model to extend further west along most of the Venezuelan coast. Furthermore, we must consider that the South American passive margin is indeed part of the Atlantic slab and it cannot have subducted simultaneously towards the north and towards the west under the Lesser Antilles, as this creates a geometric problem.

We propose an alternative explanation where the strike slip fault system marks the southern end of subduction for the Atlantic oceanic crust, but part of the northern South American lithospheric mantle is torn off from beneath the plate and subducts along with the Atlantic slab (Figure 3.11). Rifting formed a passive margin along northern Venezuela during the opening of the South Atlantic and proto-Caribbean. This margin has been tectonically active in the Cenozoic and therefore we presume it is not underlain by a depleted, buoyant, cratonic mantle keel that would resist deformation, but rather by a more mobile mantle capable of subduction. If this hypothesis were correct, the lithospheric mantle subducting south of the strike slip fault system would be of a
continental margin nature and it would be reasonable to assume that it did not contain hydrous minerals. As outlined above, this would presumably prevent the occurrence of intermediate depth seismic events and explain the absence of Benioff seismicity south of the Paria cluster. If the mantle lithosphere indeed detached from the crust in northern Venezuela, one might expect the upwelling of asthenospheric material to replace it. This upwelling material could undergo decompression melting and basaltic magmatism would be likely along northern Venezuela, but no such magmatism has been reported. We suggest that the lithospheric mantle is only partially removed, and decompression melting is not significant enough to produce intrusive bodies that reach the surface. We note that [Miller et al., 2009] image a low shear velocity anomaly centered on Barcelona Bay at depths greater than ~70 km and interpret it as upwelling mantle material.

Based on this hypothesis, and the southwestward extension of the slab in our model, we expect the northernmost ~100-200 km of the Venezuelan mainland to be
underlain by thinned lithosphere, which would leave the boundary susceptible to further
defformation and suggests isostatic rebound may play a role in the uplift of the coastal
ranges. An important corollary of our hypothesis is that the view of the San Sebastián -
El Pilar Fault system as a planar fault on a lithospheric scale [Burke, 1988; Clark et al.,
2008] needs to be revised to include more complicated plastic deformation and
downwelling of the lower lithosphere.

3.6.3. Caribbean Subduction

We begin the analysis of our image of Caribbean subduction by noting that the
steeply dipping Caribbean slab resolved in our model at depths greater than 200 km is
~500 km inboard from the trench (we assume the slab subducts under the Southern
Caribbean Deformed Belt offshore northwestern Colombia in an east-southeasterly
direction, roughly paralleling the GPS observations on San Andrés island, Figure 3.1).
This would imply an initial, shallow, stage of subduction with a dip of ~22° (Figure
3.13), consistent with the seismicity-derived estimates of Malave and Suarez [1995]
(25°), Pennington [1981] (20°) Ojeda and Havskov [2001] (27°), and the “northern slab”
of Zarifi et al. [2007] (25°). A stage of shallow subduction has been shown to produce
dehydration of the subducting slab at depths and temperature conditions unfavorable to
the production of dehydration-induced melting in the surrounding material [English et al.,
2003] which would explain the lack of volcanism at the site of the steepening of the slab
near the Mérida Andes. Shallow subduction of the Caribbean plate is widely thought to
play an important role in driving deformation and shortening in the Santa Marta Massif,
Perijá ranges and Mérida Andes on the overriding South American plate [Audemard and
Audemard, 2002; Kellogg and Bonini, 1982]. Our model is consistent with that hypothesis and additionally shows that initial shallow subduction is followed by an abrupt steepening of the slab and that the steeply dipping slab reaches depths greater than 600 km.

Since only part of the Caribbean plate seems to be subducting under South America, the question arises as to where the subducting section detaches or tears from the stable section. Presumably, this would occur west of our study area, however there is no clear indication from bathymetry, potential field data or seismicity of what the location of this tear is. We speculate that an ancient fracture zone (obscured by the large volume magmatism that generated the Caribbean Large Igneous Province) could provide a zone of weakness that would facilitate such tearing, so that no significant seismicity is produced.

An interesting problem associated with the Caribbean subduction we image is the relationship between the down going slab and the Bucaramanga seismic nest (Figure 3.12). Seismicity studies have suggested that the seismic nest occurs within (or possibly at the edge) of the subducting Caribbean slab [Malave and Suarez, 1995; Ojeda and Havskov, 2001; Pennington, 1981; Zarifi et al., 2007]. Our resolution tests show that our data do not effectively constrain the southwestward extension of the slab, and even though we do not image the anomaly extending beyond the Colombia-Venezuela border this is a consequence of our station distribution rather than the true slab geometry (Figure 3.8). In the future, efforts should be made to improve the image of the mantle structure in the west of our study area and extend it westwards to include the shallow subduction segment of the Caribbean slab as well as the area surrounding the Bucaramanga nest to
Figure 3.12: Bucaramanga Nest and 1.5% isovelocity contour of the Caribbean slab anomaly. Map view on the left, 3D view from SE on the right. Earthquake hypocenters with depths greater than 100 km are shown. Note that the Bucaramanga Nest occurs SW of the Caribbean slab anomaly beyond the resolved area of our model.
Figure 3.13: Cross section A-A’ through the velocity model (location on Figure 3.9) with projected earthquake hypocenters from EHB bulletin [International Seismological Centre, 2009] (circles) and Malavé and Suarez, [1995] (squares) (bottom) and corresponding elevation profile (top). SM – Santa Marta Massif; PR – Perijá Range; LM – Lake Maracaibo; MA – Mérida Andes.
gain insight into the origin of the remarkable seismic activity in this area.

3.6.4. Correlation with transition zone topography

Due to the nature of the mineral phase transitions that generate the 410-km and 660-km seismic discontinuities, we expect the depth at which these transitions occur to be affected by the subduction of cold material. Global and regional studies have consistently shown a thickening of the transition zone caused by uplift of the 410-km and deepening of the 660-km associated with subducting slabs [Vidale and Benz, 1992; Collier and Helffrich, 1997; Flanagan and Shearer, 1998; Niu et al., 2005]. Since the slabs in our model pass through the transition zone, we correlate our slab anomalies with the transition zone discontinuities as imaged by PdS receiver functions [Huang et al., 2010]. The receiver function study used data from the BOLIVAR array and the IASP91 [Kennett and Engdahl, 1991] velocity model to produce common conversion point images of 410-km and 660-km topography. Here, we present a brief joint analysis of the two results.

Topography of the 410-km discontinuity is consistent with our image of the Atlantic slab, as 20-35 km of uplift are observed for the NE-SW arm of the Atlantic slab anomaly, and a smaller 7-15 km for the E-W arm (Figure 3.14). The continent-ward extension of the slab (E-W arm) seems to produce a less significant uplift. As discussed above, this part of the anomaly has a ~1% smaller magnitude with respect to the oceanic NE-SW oriented part of the slab. This anomaly amplitude difference implies a temperature difference of ~170 K [Isaak, 1992]. We can compare this value to the temperature difference estimated from the difference in 410-km discontinuity uplift which gives us 100 to 150 k (using a Calpeyron slope of 4Mpa for the phase transition.
(Katsura et al., 2004)) and see that both estimates are roughly consistent. Considering that the temperature at the base of the crust is higher under continents than under the oceans by ~300 K (McKenzie et al., 2005) and the proposed nature of the E-W arm (continental margin lithospheric mantle), a 100-170 K temperature difference between it and the oceanic part of the slab seems plausible. We acknowledge there is a problem with this explanation as the difference in the magnitudes of the two arms of the anomaly is only seen at depths greater than ~250 km, and if it is caused by a difference in the original temperature it should be particularly clear at shallower depths. An additional factor we must consider is the significant effect of water on transition zone discontinuities (Higo et al., 2001; Litasov, 2005; Wood, 1995). The different response to the part of the slab we interpret as being continental margin lithospheric mantle may be a consequence of less water there than in the oceanic part of the slab. In the case of 660-km discontinuity, there is a gap in the data in the area where we image the slab. However, in the data around a slab the rim of a depression is apparent (Figure 3.14). This simple analysis is very encouraging in that it shows consistency across two completely independent data sets.

The response of the transition zone discontinuities to the subducting Caribbean is even clearer, as a prominent uplift of the 410-km discontinuity of up to 44 km and an equally important depression of the 660-km discontinuity (20-25 km) are observed with good spatial agreement with the subducted slab (Figure 3.14). The greater response of the discontinuities to the subducting Caribbean slab seems paradoxical given the smaller amplitude of its velocity anomaly. However, in synthetic tests, the amplitude of the
Figure 3.14: Subducted slabs and topography of transition zone discontinuities. 410-km (top), 660-km (middle) and 3D view from the northeast (bottom). ATL – Atlantic slab; CAR – Caribbean Slab.
Caribbean slab was underestimated by our inversion to a greater degree than that of the Atlantic slab (Figure 3.9), so it is likely that the real velocity anomaly produced by the Caribbean subduction is greater than the one in our model and it may be greater than the Atlantic slab anomaly.

3.7. Conclusions

We find the lithospheric mantle under the Caribbean plate to be relatively slow except beneath the Leeward Antilles. Shallow slow velocity anomalies also occur east of Lake Maracaibo and south of Barcelona Bay in central Venezuela. The Guayana Shield shows a small positive velocity anomaly. The distribution of this anomaly suggests that the cratonic root of the shield extends at least as far north as the Orinoco River in the east of the shield, but does not reach as far north as the river in the west, although the exposed crustal rocks are cratonic.

We find no evidence to support shallow southward subduction of the Caribbean plate beneath northwestern Venezuela. However, we cannot rule out the existence of this subduction given the lack of sensitivity of our data set and method to such a feature.

We image the steep subduction of the Atlantic slab in the east of our study area as well as the subduction of a southernmost section of the Caribbean plate in the west. Both of these subducting slabs go through the transition zone as confirmed by transition zone topography. The Atlantic slab extends continent-ward south of the surface location of the main strike slip fault system. We interpret this southward extension of the slab to be continental margin mantle lithosphere subducting with the oceanic lithosphere of the Atlantic. This hypothesis would imply that the southward extension of the slab is warmer
than the main (oceanic) part of the slab, consistent with observations of a smaller velocity anomaly and smaller associated topography of the 410-km discontinuity. Alternatively, the difference in 410-km topography could also be caused by a lack of water in the continent-ward extension of the slab. The Paria cluster of seismicity does not mark the southern boundary of the subducting Atlantic slab. The absence of intermediate depth seismicity south of the cluster may be caused by a lack of hydrated minerals in the slab there or could be a result of concentration of strain and higher strain rates at the site of the cluster. Partial removal of the lithosphere beneath the coastal areas of Venezuela may help drive uplift of the coastal ranges and will leave the boundary susceptible to future tectonic deformation.

The steep Caribbean subduction we image occurs ~600 km landward of the trench, which implies a stage of shallow subduction west of our study area consistent with previous studies. The relationship between this Caribbean slab and the Bucaramanga nest of seismicity remains unclear as the seismic nest lies beyond the resolvable area in our model, but it is possible that the seismicity occurs within the slab as has been suggested. The response of the transition zone discontinuities to the Caribbean subduction is greater than that for the Atlantic subduction in spite of the former showing a smaller velocity anomaly than the latter. It is likely that our velocity model underestimates the amplitude of the Caribbean Slab anomaly.
Appendix A: Details of GA implementation

In this appendix we describe how each new generation of solutions is created from the previous generation. For clarity, we will make a distinction between the set of parent and offspring solutions, since not all of the offspring solutions will pass on to the next generation and become potential parents (this is analogous to the fact that not all individuals in a biological population reach sexual maturity). We will refer to the individuals in generation $n$ as $parents_n$ and the solutions produced by combining them as $offspring_n$. The next generation of solutions $parents_{n+1}$ is created by replacing a portion of $parents_n$ with an equal number of individuals from $offspring_n$ as described in the following paragraphs. Also for clarity, we will add a superscript to denote the fitness of the individual, with $parents_n^1$ being the fittest individual solution in generation $n$ (referred to in the text as the $\alpha$ solution), $parents_n^2$ the second fittest and so on.

Recall that each solution is represented by a vector whose components are real numbers indicating the lateral locations of province boundaries. Each boundary location (vector component) is termed a gene. The value of each gene is limited to the range defined by the northern and southern extremes of the model.

The application of three different operators is necessary to create generation $n+1$ from the solutions of known fitness in generation $n$. These are: mating, mutation and replacement.

The mating operation is performed $P$ times with $P$ being the population size. It begins by selecting two individuals from $parents_n$ to mate. An elitism factor $E_s$ (with a value ranging from 0 to 1) is introduced so that the $(E_s \times P)$ fittest solutions in the
Parents population of solutions automatically become selected as one parent with their mating partner being selected at random from the population. For the remaining \((1 - E_s) \times P\) of the matings, both parents are selected at random from within the population. For example, for \(P = 100\) and \(E_s = 0.3\), 100 mating operations will take place in each generation; for the first 30 mating operations \(parents,^1\) through \(parents,^{30}\) will be chosen as one of the parents with the second parent in each case being selected at random from \(parents_n\). In the remaining 70 mating operations both parents will be selected randomly from \(parents_n\).

From the two parent solutions selected, the mating operator produces one offspring solution by uniform crossover: The genes in the offspring solution take the value of the corresponding gene in one of the parent solutions, with an equal probability of choosing either parent for each gene.

Once the mating has been completed, the offspring individuals undergo mutation by random replacement: a certain percentage of all the genes is replaced by a different value picked at random from within the range specified. The percentage of genes replaced is termed mutation rate \(mr\). This completes the process of creating the offspring population of solutions. This offspring population is then evaluated to measure the fitness of the individual solutions.

The evaluation operator has been described in the text. Here we note that the linear inversion of the system of equations \(Gm=d\) (where \(G\), \(m\) and \(d\) are as defined in the text) is given by \(m=(G^T G)^{-1}G^T d\) yielding the vector \(m\) that minimizes \(\|Gm-d\|\) in a least squares sense.
Finally, generation \( n+1 \) is created by replacing a number of individuals from \( parents_n \) with an equal number of individuals from \( offspring_n \). Here too we introduce an elitism factor \( E_r \) (with a value ranging from 0 to 1), so that fittest \( (E_r \times P) \) individuals of generation \( n \) proceed unchanged to generation \( n+1 \), and the \( ((1- E_r) \times P) \) least fit individuals of generation \( n \) are replaced by the \( ((1- E_r) \times P) \) fittest individuals of the offspring population. For example, for a \( P = 100 \), and \( E_r = 0.2 \):

\[
parents_{n+1} = parents_n^1, parents_n^2, ..., parents_n^{20}, offspring_n^1, offspring_n^2, ..., offspring_n^{50}.
\]

This completes the process and forms the population \( parents_{n+1} \) of known fitness from which \( parents_{n+2} \) can be calculated by repeating the steps described above.

The process is repeated for a set number of generations \( maxg \).

Clearly, an initial population of known fitness is required for the first generation. The initial population \( parents_0 \) is produced by choosing randomly generated values within the range define by the model’s extremes.

There are a number of free parameters that can be adjusted such as the population size \( P \), the mutation rate \( mr \), the elitism factors in the selection phase of the mating operator \( E_s \) and the replacement operator \( E_r \). The values we have used were selected through trial and error using the noise-free synthetic dataset and are shown in Table 1. High elitism factors tend to cause the algorithm to converge prematurely, whereas for small values convergence slows down and the algorithm requires a greater number of generations to achieve a given level of misfit. Variations in the mutation rate have a similar effect; exceedingly small values may cause premature convergence while high
values tend to delay convergence. With respect to population size, a greater value implies a more expanded search of model space in each generation and helps the algorithm converge to a given misfit level on fewer generations but increases the CPU time per generation. Finally, the user specifies a number of realizations. Here again a compromise must be made, taking into account CPU time. The number of realizations should be large enough to create a statistically significant sample of the space of adequate solutions. We have settled on 50 realizations for all the runs presented in this paper based on tests performed with the noise-free synthetic data set, as it is difficult to estimate a priori what the level of variation in the sets of solutions will be.

It is important to note that while it is difficult to estimate the optimum values of the free parameters for a given data set, and to evaluate the effect of the interaction between the different free parameters, the tests we conducted with synthetic data produced comparable results over a range of different values for each of these parameters. This allows us to conclude that although trial and error is necessary to optimize the values of free parameters, the algorithm is robust with respect to these values.
<table>
<thead>
<tr>
<th>Description</th>
<th>Parameter</th>
<th>Value</th>
</tr>
</thead>
<tbody>
<tr>
<td>Number of realizations</td>
<td>maxr</td>
<td>50</td>
</tr>
<tr>
<td>Number of generations</td>
<td>maxg</td>
<td>150</td>
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<tr>
<td>Populations size</td>
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<tr>
<td>Elistism factor for selection</td>
<td>E_s</td>
<td>0.3</td>
</tr>
<tr>
<td>Mutation rate</td>
<td>mr</td>
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</tr>
<tr>
<td>Elitism factor for replacement</td>
<td>E_r</td>
<td>0.2</td>
</tr>
</tbody>
</table>

**Table 1:** Parameter values used in the algorithm for all tests and real data applications in the text. These values were selected by trial and error with the noise-free synthetic data set described in the text.
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