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Seismic imaging of the Upper Mantle structure and dynamics beneath the Southern Caribbean plate boundary and Venezuela

by

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A THESIS SUBMITTED IN PARTIAL FULFILLMENT OF THE REQUIREMENTS FOR THE DEGREE

Doctor of Philosophy

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HOUSTON, TEXAS
August 2013
ABSTRACT

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The Caribbean-South America plate boundary has a complicated tectonic history that has been matter of debate and the focus of many studies for decades, yet many questions remain unanswered. The aim of this work, developed within the framework of the BOLIVAR (USA) and GEODINOS (Venezuela) projects, is to use different seismological techniques to study the lithospheric structure under the southern Caribbean and Venezuela, in order to understand some aspects of the present structure and its tectonic evolution. A shear wave splitting analysis in northwestern Venezuela revealed three areas with different deformation mechanisms: (1) Islands and coastal regions have large splitting times (~2-3 s) and a fast polarization direction parallel to the direction of the relative plate motion of the Caribbean plate respect to South America, which can be explained by a strong eastward flow confined at the CAR-SA plate boundary. (2) The stable South America plate showed weak seismic anisotropy with an origin likely in the asthenosphere. (3) Large splitting times and a ~NE-SW fast direction are observed at stations deployed along the Mérida Andes range, suggesting that the subcontinental mantle is also deformed beneath the range. It is likely the lithospheric mantle played a
major, if not dominant, role in the formation of the Mérida Andes. The upper mantle structure of the area was obtained by combining three types of seismic data: Ps and Sp receiver functions and Rayleigh wave tomography. Results reveal the presence of the Moho of the subducting Caribbean Plate beneath the northwestern part of the Maracaibo Block. Tomographic images indicate that the subducting Atlantic slab appears to be attached to the continental South American lithosphere, pulling it down and removing the continental lithospheric mantle beneath the Serrania del Interior. A lithospheric thickness map was also obtained. The lithosphere asthenosphere boundary shows significant variations and seems to correlate well with major tectonic provinces in the region. Finally ambient noise cross-correlations between station pairs yields to Empirical Greens Function as waveform data input for the adjoint tomography based on spectral element methods. The adjoint tomography utilizes a more accurate full wave finite-frequency theory compared to the previous ray theory, and will iteratively refine the initial smooth 3D model to achieve more detailed high-resolution images of the upper most mantle structure of eastern Venezuela. Low velocity anomalies correspond to the major sedimentary basins and high velocity anomalies correspond to the stable craton
Acknowledgments

First and foremost I would like to thank God. For being next to me and guiding me throughout my career. I could never have done this without the faith I have in you.

I would also like to thank my family; they have always been a source of motivation. They have always been supportive and proud of my accomplishments. My Mom (mami) deserves an especial mention as she provided all the tools and values that I ever needed to advance in my education and to become the person I am today. Thank you Mami.

This research would not have been possible without my advisor and mentor. Dr. Fenglin Niu from whom I learnt a lot especially how to be a scientist. Thank you so much for sharing you excitement and knowledge with me. I would also like to express my gratitude to Dr. Alan Levander, for being a second advisor, for including me in his group, and for the positive encouragement of my work. Thanks to Dr Julia Morgan and Dr David Alexander for being part of my thesis committee.

I also thank Dr. Min Chen for teaching me how to “kill CPU’s” and for being my guide during my last project.

I would like to express my gratitude to FUNVISIS personnel, who helped collect the data I work with during this project. I’m especially grateful to Dr. Michael Schmitz who always pushed me to work hard, and also pushed me to pursue a doctoral degree.
The Rice University Earth Science Department has been a wonderful community over the past few years. Thanks to the administrative staff for managing all the things that we so often take for granted, and doing so with a warm smile and friendly disposition.

The support and help of my friends allowed me to stay motivated throughout this process. In particular the STAR group, Huafeng Liu, Rachel Margolis, Imma Palomeras and Sally Thurner, for their constructive criticism and help. Even though Jacob Siegel, is not part of the STAR group, he deserves to be mentioned as he supported me since my first day as a graduate student. I will also want to thank Maximiliano Bezada for is comments and the discussions about our research, and for being my first contact when I started the graduate program. Ricardo Gallardo, thanks for being there as a friend, as a classmate, as a neighbor, for taking care of me when I was sick, for your great support, for everything. Dani Dani (la mini creti) Thank you for everything, cooking for me during Ike, to letting me stay at your place when I needed it. I also give thanks to Mariano Arnaiz for all his help and support from Venezuela and China.

This work and the broader BOLIVAR and GEODINOS projects were made possible by funding from the NSF Continental Dynamics program (grant EAR0003572 and EAR0607801), with support from FONACIT (grant G-2002000478) and PDVSA-INTEVEP-FUNVIS (project 04-141) in Venezuela. Also this work was supported in part by the Data Analysis and Visualization Cyberinfrastructure funded by NSF under grant OCI-0959097.
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Chapter 1

Introduction

The Caribbean - South American plate boundary is a broad deformation zone resulting from prolonged continuous interactions between the South America (SA) and Caribbean (CAR) plates. Approximately 55 Ma ago, the southern edge of the CAR began colliding obliquely along the northern edge of SA (Figure 1-1), starting in northwestern SA (northern Colombia and northwestern Venezuela) and migrating eastward progressively (relative to SA) (e.g. Kellogg and Bonini, 1982; Meschede and Frisch, 1998; Pindell et al., 2005). These interactions resulted in the development of fold and thrust belt systems and foreland basins in different stages of development along a right lateral transpressional zone that extends ~1000 km from northern Colombia to the Gulf of Paria in northeastern Venezuela. Deformation occurs in a broad zone of ~300 km wide in eastern Venezuela and up to ~600 km wide in western Venezuela and eastern Colombia (Meschede and Frisch, 1998; Audemard and Audemard, 2002).
Figure 1-1 Tectonic evolution of the CAR-SA plate boundary. Modified from Pindell (2006)

GPS measurements indicate that the CAR is currently moving at a speed of approximately 2 cm/yr relative to SA, roughly along the strike of the El Pilar fault system in the east. The relative motion in the west shows a component of oblique convergence (Perez et al., 2001; Weber et al., 2001), which leads to the subduction, and underthrusting of the southern edge of the CAR beneath northwestern SA (Figure 1-2).

Intermediate depth seismicity and tomographic images suggest that the CAR plate subducts at a very low angle beneath northwestern SA offshore of the Santa Marta Massif (Van der Hilst and Mann, 1994; Malave and Suarez, 1995; Taboada et al., 2000). The CAR plate then dips steeply to the ESE under Lake Maracaibo and the Mérida Andes (Pennington, 1981; Taboada et al., 2000; Bezada et al, 2010a). The
transpressional regime in western Venezuela has also divided the area into several tectonic blocks that move independently from the surrounding plates. The most noticeable one is the Maracaibo Block (MB), which is bounded by the left lateral strike slip Santa Marta Bucaramanga Fault (SMB) to the southwest, the right lateral Bocono Fault to the southeast, and the right lateral Oca-Ancon (OA) fault to the north. The direction of motion of the MB is northeast relative to SA, as if is escaping from the continent (Figure 1-2). Intracontinental deformation in northern Colombia and westernmost Venezuela is revealed by the presence of the Santa Marta Massif in Colombia, the Perija Range in the Venezuelan-Colombian border, and the Mérida Andes in western Venezuela (Kellogg and Bonini, 1982). Since the timing of the uplift (~10 Ma) of these mountain ranges coincided with the Panama-SA collision, there are suggestions that the uplift is likely to be the consequence of the late Miocene-Pliocene collision (Molnar and Sykes, 1969). On the other hand, Kellogg and Bonini (1982) and Bezada et al. (2010a) believed that the flat subduction beneath northern Colombia and westernmost Venezuela is the most likely cause for the uplift and deformation in the area.

Plate motion is accommodated by transpression and transtension along the right lateral San Sebastian-El Pilar strike-slip fault system in central and eastern Venezuela (Perez et al 2001; Audemard et al., 2005). The fault system marks the northern edge of coastal thrust belts, such as Cordillera de la Costa and Serrania del Interior, and their associated foreland basins (e.g. Guarico and Maturin Basin).
Seismicity along the plate boundary is mostly shallow (< 50 km; Palma and Romero, 2011), and is concentrated along the major strike slip fault systems. Seismicity distributes more diffusively on the thrust faults in the Serrania del Interior, the Cordillera de la Costa, and the Mérida Andes (Figure 1-2), and is almost absent within the rest of Venezuela.

Transform motion ends at the eastern end of the strike slip El Pilar fault system, at the southernmost corner of the Antilles subduction zone, where the oceanic lithosphere of the SA plate (Atlantic seafloor) subducts beneath the Lesser Antilles (Molnar and Sykes, 1969; Russo and Speed 1992). The top of the subducting slab has been inferred using intermediate-depth earthquakes up to ~200 km deep (VanDecar et al., 2003). Intermediate depth seismicity ceases abruptly beneath the Gulf of Paria, where there is a distinct cluster of intermediate depth earthquakes (Figure 1-2). This earthquake cluster has been interpreted as an expression of lithospheric tearing (e.g. Russo et al., 1993; Clark et al., 2008a, 2008b). Tomographic images reveal that the Atlantic slab is steeply subducting toward the west, together with the presence of a continuous high-velocity anomaly beneath the Serrania del Interior (VanDecar et al., 2003; Bezada et al., 2010a). This high-velocity anomaly was interpreted as an aseismically descending lithosphere due to either convective removal of the SA continent (Bezada et al., 2010a), or northward subduction of the SA passive margin lithosphere (VanDecar et al., 2003).
Figure 1-2 (a) Caribbean tectonics, with GPS vectors in the southern Lesser Antilles Arc and on San Andreas (SA) Island in the western Caribbean. (b) Broadband seismic stations in Venezuela used in the investigation of the LAB. Red triangles indicate permanent stations of the National Seismic Network of Venezuela. Blue and pink triangles represent temporary deployments under the first and second phases of the BOLIVAR project. White triangles are OBS stations deployed during the first phase of the BOLIVAR project. Seismicity at different depths is shown with color-coded circles. Major active faults are indicated solid black lines. MB: Maracaibo Block, BF: Bocono Fault, OAF: Oca-Ancon Fault; SMF: Santa Marta–Bucaramanga Fault; SMM: Santa Marta Massif; SS-EP: San Sebastian -El Pilar Fault; BAB: Barinas-Apure Basin; CC: Cordillera de la Costa; SI: Serrania del Interior; OR: Orinoco River; LA: Leeward Antilles. GS: Guayana Shield. GP: Gulf of Paria; EG: Espino Graben. CB: Cariaco Basin. T&T: Trinidad and Tobago.
The Guayana Shield is a part of the Amazonian Craton and is located in the southeastern and south-central Venezuela (Gonzalez de Juana, 1980). The Guayana Shield is composed mainly of Archean and Proterozoic rocks, and is usually divided into four provinces: the Archean Imataca province and three Proterozoic provinces, Pastora, Cuchivero and Roraima (Yoris and Ostos, 1997).

The Eastern and western Venezuelan are known as one of the world’s major petroleum production regions. Many geological and active seismic studies have been conducted to study the stratigraphy of the basins in the area (e.g., Gonzales de Juana, 1980; Di Croce, 1995; Passalaqua et al., 1995; Parnaud et al., 1995; Jacome et al., 2008). However, most of these studies are limited in local scale. Meanwhile previous passive seismic studies of the mantle structure of the Caribbean-South American plate boundary have been largely limited to either broad regional, or very local studies [Malave and Suarez, 1995; Russo et al., 1992, 1993; van der Hilst and Mann, 1994; VanDecar et al., 2003]. In order to better understand the geological and seismic structure, especially the deep crustal and upper mantle structure, both active and passive seismic experiments were conducted by scientists from the US and Venezuela under the BOLIVAR (Broadband Ocean-Land Investigations of Venezuela and the Antilles arc Region) and GEODINOS (Geodinamica Recient del Limite Norte de la Placa Sudamericana) Projects. While the main goal of these projects is to understand the evolution and current state of the margin, and to evaluate island arc accretion as a mechanism for continental crust growth (Levander et al., 2006), the high quality active and passive seismic data collected by
the projects have permitted to image seismic structure of the deep crust and upper mantle and study the relevant dynamic processes at regional scales.

Throughout the two projects, crustal structure beneath Venezuela and its surrounding area have been determined to unprecedented fine level using a variety of seismic data and techniques (Schmitz et al., 2002; 2005; Niu et al., 2007; Guedez, 2007; Clark et al., 2008c; Bezada et al., 2008; Bezada et al., 2010b; Magnani et al., 2010). Meanwhile, finite-frequency tomography using teleseismic P-wave traveltime data yield a high-resolution 3D P-wave velocity model of the region down to ~800 km, which clearly reveals the subducting Atlantic slab in the east and the Caribbean slab in the west (Bezada et al., 2010a). The temperature variations at the base of the upper mantle derived from the P-wave velocity model agree well with those estimated from the transition zone thickness map (Huang et al., 2010). Miller et al. (2009) studied the upper mantle velocity structure using ballistic surface wave dispersion data with a primary goal of understanding the structure and evolution of the upper mantle beneath the complex plate boundary.

As discussed above, a variety type of deformation styles are observed in this region, and so far they have been largely attributed to shallow crustal processes (e.g. Kellogg and Bonini, 1982; Audemard and Audemard, 2002; Jacome et al., 2003; Chacin et al., 2005). It is unclear whether and how the lithospheric mantle is involved, largely because of the lack of seismic images of the lithosphere, especially the thickness of the lithosphere. My thesis focuses on mapping lithospheric structure and relevant processes by integrating various types of seismic data.
collected by the BOLIVAR project using the state-of-the-art seismic imaging techniques. More specifically, we first combined surface wave, Ps and Sp receiver function data to constrain a map of lithosphere thickness beneath the region. We then measured seismic anisotropy using shear wave splitting data to investigate how coherently deformation is distributed vertically within the lithosphere.

This dissertation is composed of slightly modified versions of three manuscripts in different stages of the publication process that describe the lithospheric structure under the southern Caribbean and Venezuela at different depths by utilizing different types of seismic data and techniques.

The lithosphere is the mechanically strong layer that acts as a rigid plate. It includes the crust and the portion of the upper mantle that behaves elastically on long time scales. Knowledge of the composition and structure of the lithosphere is necessary to understand continental lithospheric evolution. However, direct sampling (especially at depths greater than 10 km) is hardly feasible. Xenoliths and outcrops can be used to infer the lithospheric composition of a particular area, but whether these local samples are representative of large regions is always a concern. Only indirect techniques, such as seismology, allow us to study the physical properties across large sections of the lithosphere. The characteristics of seismic waves are a function of the earth’s elastic properties, and inferences about physical properties such as temperature, pressure, mineralogy and composition can be derived.
The first Chapter reports new findings on various mantle deformation mechanisms near the boundary between the southwestern Caribbean and South America plate.

Information about mantle deformation can be obtained through the study of seismic anisotropy. In the general sense anisotropy is a term used to describe a medium whose properties are functions of orientation. In that way seismic anisotropy occurs when the elastic waves travel with different speeds depending on their propagations and polarization directions through a medium (Savage, 1999). In the mantle, seismic anisotropy is usually caused by the strain-induced, preferred orientation of mantle minerals (such as olivine), therefore shear wave splitting, or birefringence, is one of the most effective methods to characterize the depth and extent of the mantle strain field (Savage, 1999; Silver 1996).

Russo et al. (1996) characterized the relationship between the plate motion, crustal deformation and upper mantle deformation along the Caribbean plate boundary. They found that stations located in northeastern Venezuela have a nearly east - west fast polarization direction with unusually large splitting times (~2 s), which are approximately twice as large as the global average of ~1 s (Silver, 1996). Pinero-Feliciangeli and Kendall (2008) using both core (SKS/SKKS) and local phases, obtained similar results near the plate boundary. They also observed smaller splitting times with polarization directions parallel to the local structures suggesting vertically coherent deformation of the crust and the upper mantle in continental SA. Using stations deployed during the BOLIVAR project in eastern
Venezuela, Growdon et al. (2009) measured waveform splitting of the core phases SKS/SKKS. Their results showed very large splitting times in stations located at the northeastern Venezuelan coast, consistent with previous studies, which they interpreted to be caused of strong mantle flow around the southern edge of the Atlantic plate. They also found that the splitting time is considerably smaller towards the interior of SA.

With an improved multi-event stacking method, we measured seismic anisotropy from broadband stations in northwestern Venezuela, and found a large and systematic variation in seismic anisotropy. The splitting times estimated from island and coastal stations are close to the largest values observed globally, and suggest the presence of a strong mantle flow at the plate boundary. The subducting Caribbean plate imaged by tomographic studies may play an important role in directing and focusing the flow. Inside the Barinas-Apure basin, delay times are small with an east-west orientation, which is consistent with the absolute motion direction of the South American plate. Along the southeastern edge of the MB the split measurements suggest that the entire lithosphere is deformed coherently parallel to the Bocono Fault, and lithospheric mantle played a major, if not dominant, role in the formation of the Mérida Andes.

The second chapter presents the upper mantle velocity structure beneath Venezuela and southern Caribbean using three types of seismic data: Ps and Sp receiver functions and Rayleigh wave tomography.
Receiver-function analysis is a commonly used technique to estimate crustal thickness (depth of the Moho discontinuity). It is also used to image seismic discontinuities in the mantle. The technique is designed to estimate discontinuity structure directly beneath the receiver by deconvolving one component of a teleseismic recording from another to remove source signature from seismograms (Ammon, 1991). A teleseismic record can be considered as the combination of the direct arrival and a series of conversions and reflections at boundaries below the recording station. Given a velocity model, the time differences between converted waves and the direct arrival can be used to estimate the depths of major discontinuities in the mantle, such as Moho and the Lithosphere-Asthenosphere Boundary (LAB).

Using teleseismic data recorded by the BOLIVAR array, Niu et al. (2007) used the primary Ps conversion and crustal reverberations to estimate crustal thickness and average Poisson’s ratio. They observed that Moho depth varies from 16 km beneath the southern Caribbean plate to 52 km beneath the Maturin Basin and Merida Andes, and found good correlation between crustal structure and tectonic provinces. Huang et al., (2010) also used Ps receiver functions to study the mantle transition zone beneath the southern Caribbean Plate boundary and Venezuela. Their results suggest that the topography of the 410 km and 660 discontinuities is consistent with subduction processes occurring at east and west edges of the boundary.
Analysis of fundamental mode Rayleigh waves is a standard way to determine the three-dimensional shear wave velocity structure of a region. The two-step inversion technique developed by Yang and Forsyth (2006a) accounts for effects due to multipathing of the incoming wavefield. The first stage involves computing 2-D phase velocities using the so-called two plane wave method (Forsyth & Li, 2005; Yang & Forsyth, 2006a, 2006b). Finite-frequency amplitude and phase sensitivity kernels can be used to improve lateral resolution (Yang & Forsyth, 2006b). The second stage is the inversion of dispersion curves to obtain the shear velocity structure.

Miller et al. (2009) measured the shear velocity structure of the crust and upper mantle of the CAR-SA boundary region by analysis of fundamental mode Rayleigh waves in the 20- to 100-s period band recorded at the BOLIVAR array. However they did not include finite-frequency kernels to determine phase velocities. Their resulting model shows lateral variations that correspond to tectonic provinces and boundaries. Velocity changes delimited by the San Sebastian –El Pilar fault system, mark the difference between the South American continental lithosphere and the Caribbean oceanic lithosphere.

It can be challenging to identify the LAB only from receiver-function data when there are either multiple peaks or a lack of distinct peaks in the stacked traces, which can happen if the LAB is a gradual boundary that generates primarily low-frequency S-to-P or P-to-S converted energy. Thus, combining the Rayleigh wave
tomography model and the receiver function data (Ps and Sp) can lead to more accurate estimations of the lithospheric thickness (LAB).

Our results reveal the presence of the Moho of the Caribbean Plate beneath the northwestern part of the Maracaibo Block. Tomographic images indicate that the subducting Atlantic slab appears to be attached to the continental South American lithosphere, pulling it down and removing the continental lithospheric mantle beneath the Serrania del Interior. The resulting lithospheric thickness map shows significant variations and correlates well with major tectonic provinces in the region. Changes in lithospheric thickness from east to west lead us to believe that the lithospheric mantle beneath continental South America was partially removed during the subduction of the Atlantic, and this processes appeared to be occurring over the last 55 Ma.

The third chapter presents the crustal shear velocity structure beneath eastern Venezuela and southeastern Caribbean using ambient noise data as an input for waveform tomography.

Cross-correlation functions (CCF) computed from a pair of recordings of ambient noise data, can be used to retrieve the Green’s function between two receivers, with one of the receivers approximated as an impulsive source (Weaver and Lobkis, 2001). Coherent surface wave signals can be extracted from the CCFs (Shapiro and Campillo, 2004), which can be used to perform surface wave tomography at local scales.
Traditional surface wave tomography, based on teleseismic data, provides fundamental information about the Earth’s interior and its three-dimensional structure. These measurements made from ballistic surface wave have some limitations. First, resolution is somewhat limited to the distribution of sources and receivers, resulting in blank areas that are not sample by the data. Second, inversion processes often require some information about the source that is not usually known accurately. Third, measurements made with teleseismic surface waves are average values over large areas, because of this, is difficult to make short-period measurements, as this may result in simultaneous arrivals of waves with different raypaths (Shapiro and Campillo, 2004). CCFs obtained from ambient noise data contain waves propagating in all possible directions, and could significantly improve local resolution. They are also able to provide with higher frequency signals (10 – 50s) that helps to resolve small-scale structures within the crust.

Assuming that the noise is spatially uncorrelated, we calculated Empirical Green’s functions (EGF) from the CCFs for all possible stations pairs from the BOLIVAR array in eastern Venezuela. Using the spectral element method and adjoint tomography, we computed banana-doughnut kernels for all station pairs, and used them to construct misfit kernels, which serve as the gradients required in the conjugate gradient algorithm to refine the existing 3D shear velocity model of the crust.

Our results provide a detailed 3D crustal shear velocity structure for eastern Venezuela. Due to the opening of the Atlantic on the Jurassic, extensional stresses
promoted the formation of grabens along the edges of the Atlantic Ocean. In Venezuela the most important graben is the Espino Graben. Our results reveal low velocities along the axis of the graben, indicating the graben may have different thermal or compositional structure from the neighboring basins (Guarico and Maturin Basins). On the other hand, the Guayana Shield shows little lateral variations. 3D variations of the Moho are observed within the study region, and seem to correlate well with major tectonic provinces in eastern Venezuela. These results are consistent with results from the wide-angle active seismic data collected by the BOLIVAR project (Bezada et al., 2010; Clark et al., 2008; Magnani et al., 2009).
Chapter 2

Mantle flow beneath northwestern Venezuela: seismic evidence for a deep origin of the Merida Andes

We measured shear wave splitting from SKS data recorded by the national seismic network of Venezuela and a linear broadband PASSCAL/Rice seismic array across the Mérida Andes. The linear array was installed in the second phase of the passive seismic component of the BOLIVAR project to better understand the complicated regional tectonics in western Venezuela. Using the method proposed by Wolfe and Silver (1998), SKS waveforms from 2 to 36 earthquakes, mostly from the

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Tonga subduction zone, were selected for each of the 23 stations in the region in order to do a splitting analysis. The fast polarization directions can be divided into 3 zones, all in agreement with local GPS data: The first zone comprises the stations north of the dextral strike-slip Oca-Ancon fault. These stations show the largest split times (1.6-3.2 s), oriented in a roughly EW direction, and are similar to splitting observations made further to the east along the strike slip plate boundary (Growdon et al., 2009). We attribute this to trench-parallel mantle flow that passes around the northwest corner of the subducting Caribbean plate and along the northern edge of South America as proposed by Russo and Silver (1994), forming an eastward flow beneath the southern Caribbean plate. Zone two is the Mérida Andes, with the right lateral Bocono fault in the center, where split orientations are at ~N45°E, suggesting that the observed seismic anisotropy is likely caused by lithospheric deformation parallel to the Bocono fault. Zone three is east of the Bocono fault inside the Barinas-Apure Basin, where the measured split times are smaller (~0.8 s) with an EW fast direction that is consistent with those observed at the Guarico Basin, Maturin Basin and the Guayana Shield in the east, and are interpreted as orientation with the motion of the continent.

2.1. Introduction

The boundary between the Caribbean (CAR) and South American (SA) plates in western Venezuela is a wide area where active faults and extensive deformation are observed (e.g. De Toni and Kellog, 1993; Audemard and Audemard, 2002). Transpressional faulting has divided the boundary region into several distinct
tectonic blocks that move independently of the surrounding plates. For example, the triangular Maracaibo Block (MB), bounded by the left-lateral strike-slip Santa Marta-Bucaramanga (SMB) Fault to the southwest, by the right-lateral strike-slip Bocono Fault to the southeast, and by the right-lateral strike-slip Oca-Ancon (OA) Fault to the north, is moving northeastward relative to the SA plate as if it is escaping from the continent (Figure 2-1). Intracontinental deformation within northern Colombia and northwestern Venezuela is revealed by the presence of the northern end of the Andean belt, known as the Eastern Cordillera. The Cordillera consists of three recently uplifted (<10 Myr) mountain chains: the Mérida Andes in Venezuela, the Perija Range, straddling the Venezuelan-Colombian border, and the Santa Marta block in Colombia (e.g., Kellogg and Bonini, 1982; Audemard and Audemard, 2002). Although it is generally believed that timing of the uplift is related to the late Miocene-Pliocene collision of the Panama arc with the South America (Molnar and Sykes, 1969), the uplift is likely the result of flat slab subduction of the southern edge of CAR under northern South America (Bonini and Kellogg, 1982; Bezada et al., 2010).

GPS measurements clearly indicate oblique convergence between the western part of the CAR and the SA plate (Figure 2-1, Weber et al., 2001). Tomographic studies suggest the CAR plate starts to subduct beneath SA offshore of the Santa Marta Massif with a very low angle (Van der Hilst and Mann, 1994; Malave and Suarez, 1995; Taboada et al., 2000). It then dips steeply to the ESE under the Mérida Andes (Taboada et al., 2000; Bezada et al, 2010). This subduction is likely responsible for the uplift and deformation observed in the region. Beside the three
major uplifts, moderate deformation and seismicity are observed in the Serrania de Falcon located the northeastern Maracaibo block, as well as under the central Maracaibo block. Several models have been proposed to explain the observed formations, and most invoke shallow processes (for example type A, continent-continent, subduction) for the orogeny (e.g., Kellogg and Bonini, 1982; Audemard and Audemard, 2002). Thus far, there is little data, especially deep seismic data, with which to evaluate these models.

Seismic anisotropy provides essential information about the style and geometry of mantle deformation. Seismic waves in an anisotropic media travel with different speeds depending on their propagation and polarization directions. The major upper mantle mineral, olivine, has a highly anisotropic crystal structure, with up to 25% variation in P- and S-wave velocity. These anisotropic minerals can be aligned preferentially through mantle deformation, resulting in seismic anisotropy with a reduced magnitude of a few percent (e.g. Silver, 1996). In areas where the lithosphere is undergoing shortening, the fast directions tend to parallel the strike of the orogenic belts. Meanwhile, seismic anisotropy also seems to have a close correspondence to horizontal mantle flow developed under a simple shear setting beneath the station (e.g., Silver, 1996). The close relationship between the stress/strain field and seismic anisotropy thus can be used to map upper-mantle deformation associated with a wide range of tectonic processes.
Measuring shear wave splitting, or birefringence, is one of the most effective methods to characterize seismic anisotropy in the upper mantle (e.g., Silver, 1996; Savage, 1999). The polarization direction of the fast shear wave, $\phi$, and the delay time between the fast and slow waves, $\delta t$, are used to parameterize seismic anisotropy. Russo et al. (1996) found stations located at the southeastern CAR-SA plate boundary exhibit a nearly east-west fast direction with unusually large splitting times of ~2 s. This is approximately twice as large as the global average of ~1 s (Silver, 1996). Growdon et al. (2009) measured waveform splitting of the core phases, SKS/SKKS, recorded by the broadband seismic stations deployed under the BOLIVAR (Broadband Onshore-Offshore Lithospheric Investigation of Venezuela and the Antilles Arc Region), and obtained very large splitting times from stations deployed at the northeastern coast of Venezuela, consistent with the results of Russo et al. (1996). They also found that the splitting time drops quickly towards the interior of SA. Most of the stations located at the Guayana shield, Maturin basin, show a splitting time < 1 s. They interpreted the large splitting times observed in northeastern Venezuela as caused by a strong mantle flow associated with a slab tear. The Atlantic slab is tearing from continental South America in response to the eastward retreat of the Antilles trench. In addition to the SKS splitting data, the tear was also suggested by a variety of seismic data, such as local seismicity and active source reflection/refraction data (Clark et al., 2008ab), surface wave (Miller et al., 2009), body wave tomography (Bezada et al., 2010), and receiver function data (Niu et al., 2007; Huang et al., 2010), and seems to play a major, if not dominant, role in the regional tectonics.
Most of the broadband seismic stations were installed in eastern Venezuela during the main phase deployment of the BOLIVAR project, leaving it difficult to address important tectonic issues along the CAR-SA plate boundary in northwestern Venezuela, such as the slab geometry of the Caribbean plate subducted under northwestern SA, and its role in uplift of the Mérida Andes, the Perija and Santa Marta mountain ranges, and development of the Maracaibo block generally. During the second phase of BOLIVAR, we installed a linear broadband array across the northeastern end of the Mérida Andes. In this chapter we present a summary of seismic anisotropic structure beneath western Venezuela estimated from data recorded by the linear array as well as by the National Seismic Network of Venezuela. Our goal is to characterize the relationships between plate motions, and the style of crustal and mantle deformation for the study region. Results are consistent with previous studies but also provide new constraints on mantle deformation along the mountain belts.

2.2. Data and Analysis

We used SKS/SKKS waveform data recorded at 23 broadband seismic stations deployed in western Venezuela (Figure 2-1b). 14 stations belong to the permanent National Seismic Network of Venezuela operated by the Fundación Venezolana de Investigaciones Sismológicas (FUNVISIS). We installed 2 Rice stations on the islands Curaçao and Aruba during the first phase of the BOLIVAR deployment and we had two- and five-years of data from these two stations, respectively. The linear seismic array across the Mérida Andes consisted of 6
broadband stations that operated from October 2008 to November 2009. We also analyzed data recorded by the GSN (Global Seismic Network) station SDV from earthquakes occurring between 1994 and 2000.

We manually checked all the SKS/SKKS data recorded at epicentral distances between 85° and 120° from earthquakes with a magnitude > 5.6, and selected 253 SKS/SKKS seismograms with good signal-to-noise ratio (SNR). No good data were acquired by station MRP3, but the other 22 stations have at least three good SKS/SKKS records (Table 2-1). A total of 122 events were used, among which 57 were recorded by the GSN station SDV. The remaining 65 events occurred between December 2003 and August 2009. Most of the earthquakes are from the southwest Pacific, the northwest Pacific, and the Mediterranean Sea regions, clustering into three back azimuthal directions that are roughly perpendicular or parallel to each other (Figure 2-2). The FUNVISIS and Rice stations were recorded with a sampling rate of 100 samples per second (sps). They were decimated to 40 sps to be consistent with the sampling rate of the BOLIVAR linear array stations. We further filtered the data with a bandpass filter of 0.05-0.5 Hz before measuring the fast direction and splitting time.

To better constrain the splitting parameters ($\phi$, $\delta t$), we have applied a multiple event stacking approach similar to the one developed by Wolfe and Silver (1998). Instead of estimating individual ($\phi$, $\delta t$) sequentially for multiple earthquakes, the multi-event stacking method solves a pair of $\phi$ and $\delta t$ that minimizes either the summed energy in the transverse component
or the summed second eigenvalue $\lambda_2$ of the covariance matrix of the corrected particle motion (Li and Niu, 2010)

\[
\Lambda_2(\varphi, \delta t) = \left( \sum_{i=1}^{N} w_i \lambda_2(i, \varphi, \delta t) \right) / \sum_{i=1}^{N} w_i.
\]  

Here $E_{Ti}(\varphi, \delta t)$ and $\lambda_2(i, \varphi, \delta t)$ are, respectively, the transverse energy and the smaller eigenvalue of the covariance matrix of the $i^{th}$ event, computed after correcting wave propagation effects in an anisotropic medium with a fast polarization direction of $\varphi$ and delay time of $\delta t$. $w_i$ is the weight of the $i^{th}$ event and is taken as the averaged signal-to-noise (SNR) of the two horizontal components, and $N$ is the total number of the events. To compute the SNR, we chose a noise time window before the SKS arrival with the same length as the SKS signal. We also used the total SKS energy recorded at the two horizontal components to normalize the traces before computing the transverse energy $E_{Ti}(\varphi, \delta t)$ in order to ensure each event has the same contribution to the total energy. We varied $\varphi$ in the range of $0^\circ$ to $180^\circ$ with an increment of $1^\circ$, and $\delta t$ from 1.0 to 4.0 s in increments of 0.05 s. With the measured $(\varphi, \delta t)$, we further computed the polarization directions of the SKS arrivals to make sure that they are consistent with the geometric back azimuths. Differences between the calculated polarization directions and geometric back azimuths usually vary from $-15^\circ$ to $15^\circ$ with a mean close to zero. Since the calculated polarization could be affected by sensor misorientation, we further used
the particle motions of the direct P waves to calibrate the sensor orientations (Niu and Li, 2011). We found that all the stations used in this study were properly aligned with an error < 10°.

Table 2-1 SKS splitting Parameters

<table>
<thead>
<tr>
<th>Station</th>
<th>Lon. (°)</th>
<th>Lat. (°)</th>
<th>% of events</th>
<th>% of azimuth</th>
<th>ϕ (°)</th>
<th>δt (s)</th>
</tr>
</thead>
<tbody>
<tr>
<td>ARUB</td>
<td>-70.001</td>
<td>12.509</td>
<td>3</td>
<td>2</td>
<td>84±9</td>
<td>2.0±0.3</td>
</tr>
<tr>
<td>IMOV</td>
<td>-70.902</td>
<td>12.358</td>
<td>3</td>
<td>2</td>
<td>82±5</td>
<td>1.9±0.3</td>
</tr>
<tr>
<td>CURA</td>
<td>-68.959</td>
<td>12.180</td>
<td>6</td>
<td>4</td>
<td>90±10</td>
<td>2.1±0.3</td>
</tr>
<tr>
<td>MONV</td>
<td>-69.970</td>
<td>11.955</td>
<td>3</td>
<td>1</td>
<td>73±4</td>
<td>2.0±0.7</td>
</tr>
<tr>
<td>SRP1</td>
<td>-69.900</td>
<td>11.318</td>
<td>4</td>
<td>2</td>
<td>90±12</td>
<td>1.7±0.3</td>
</tr>
<tr>
<td>JACV</td>
<td>-68.830</td>
<td>11.087</td>
<td>5</td>
<td>3</td>
<td>80±6</td>
<td>2.1±0.5</td>
</tr>
<tr>
<td>DABV</td>
<td>-70.636</td>
<td>10.922</td>
<td>20</td>
<td>3</td>
<td>84±7</td>
<td>1.3±0.3</td>
</tr>
<tr>
<td>CCP2</td>
<td>-69.833</td>
<td>10.879</td>
<td>8</td>
<td>2</td>
<td>87±5</td>
<td>3.2±0.7</td>
</tr>
<tr>
<td>SIQV</td>
<td>-69.808</td>
<td>10.649</td>
<td>24</td>
<td>3</td>
<td>76±4</td>
<td>2.0±0.3</td>
</tr>
<tr>
<td>VIRV</td>
<td>-72.406</td>
<td>10.503</td>
<td>22</td>
<td>3</td>
<td>52±8</td>
<td>0.5±0.1</td>
</tr>
<tr>
<td>QARV</td>
<td>-70.524</td>
<td>10.207</td>
<td>7</td>
<td>2</td>
<td>78±3</td>
<td>0.5±0.2</td>
</tr>
<tr>
<td>CURV</td>
<td>-69.961</td>
<td>10.013</td>
<td>7</td>
<td>2</td>
<td>75±14</td>
<td>1.0±1.0</td>
</tr>
<tr>
<td>TERV</td>
<td>-69.192</td>
<td>9.964</td>
<td>13</td>
<td>4</td>
<td>32±7</td>
<td>1.1±0.3</td>
</tr>
<tr>
<td>CRP4</td>
<td>-69.583</td>
<td>9.788</td>
<td>10</td>
<td>4</td>
<td>50±5</td>
<td>1.3±0.4</td>
</tr>
<tr>
<td>SANV</td>
<td>-69.536</td>
<td>9.501</td>
<td>10</td>
<td>4</td>
<td>54±7</td>
<td>1.0±0.4</td>
</tr>
<tr>
<td>PPP6</td>
<td>-69.460</td>
<td>8.941</td>
<td>5</td>
<td>4</td>
<td>71±10</td>
<td>0.7±0.4</td>
</tr>
<tr>
<td>SDV</td>
<td>-70.633</td>
<td>8.879</td>
<td>57</td>
<td>3</td>
<td>54±5</td>
<td>1.2±0.4</td>
</tr>
<tr>
<td>VIGV</td>
<td>-71.364</td>
<td>8.840</td>
<td>11</td>
<td>2</td>
<td>65±6</td>
<td>1.3±0.4</td>
</tr>
<tr>
<td>SOCV</td>
<td>-70.857</td>
<td>8.284</td>
<td>8</td>
<td>3</td>
<td>55±5</td>
<td>1.4±0.3</td>
</tr>
<tr>
<td>PNP7</td>
<td>-69.302</td>
<td>8.074</td>
<td>3</td>
<td>1</td>
<td>85±8</td>
<td>1.0±0.4</td>
</tr>
<tr>
<td>CAPV</td>
<td>-72.314</td>
<td>7.865</td>
<td>13</td>
<td>2</td>
<td>43±9</td>
<td>0.8±0.2</td>
</tr>
<tr>
<td>ELOV</td>
<td>-69.483</td>
<td>7.001</td>
<td>11</td>
<td>3</td>
<td>76±8</td>
<td>0.7±0.3</td>
</tr>
</tbody>
</table>

We also used the method of Li and Niu (2010) to compute errors in measuring the splitting parameters (ϕ, δt):
\[
\frac{E_T(\phi, \delta t)}{E_{T,\text{noise}}} \leq 1 + \frac{k}{n-k} f_{k,n-\alpha}(1-\alpha)
\]  

(3)

Here \( n \) is the number of degrees of freedom, which was calculated based on the method of Silver and Chan (1991). \( \alpha \) is the confidence level, \( k=2 \) is the number of parameters, and \( f \) represents the F-distribution. \( E_{T,\text{noise}} \) is the noise energy and was calculated by averaging the noise levels of the two horizontal components.

Figure 2.2 (a) Locations (red dots) of the 122 events located at 85*-120* distance from the 22 stations used in this study. Note most of earthquakes occurred in the Tonga subduction (SW), the Japan-Kurile-Kamchatka-Aleutian Arcs (NE), and the Mediterranean region (NE). Only three earthquakes are from the Mid-Atlantic Ridge and one from the East Pacific Rise (SSE and SSW, respectively). At (b) Back azimuthal distribution of the 122 events, showing the number of events in each 5-degree bin. Note that the source regions have back azimuths approximately 90 degree apart, making it difficult to use them to investigate azimuthal dependence of the splitting parameters.

In general, the joint solution determined from the summed transverse energy agrees with individual solutions of each event and falls into a narrowly defined region into the \((\phi, \delta t)\) domain (Figure 2-3a-i). This can be further demonstrated by overlapping individual solutions on the \((\phi, \delta t)\) plane. For each event, we assigned a
unit value to a grid if the grid is situated in the solution region (red area in Figures 2-3a-h) defined by (3). Zero value is given otherwise. That is

\[ n_i(\varphi, \delta t) = \begin{cases} 
1, & \text{if } (\varphi, \delta t) \text{ falls in the 95\% confidence region} \\
0, & \text{otherwise} 
\end{cases} \quad (4) \]

These individual solution regions were then overlaid and grid values were averaged:

\[ n(\varphi, \delta t) = \frac{1}{N} \sum_{i=1}^{N} n_i(\varphi, \delta t) \quad (5) \]

The grids with an average value of one (red area in Figure 2-3j) were considered to be possible solutions and we took the average \((\varphi, \delta t)\) as the final solution. In general the optimum \((\varphi, \delta t)\) estimated from the two methods agree with each other, but the overlapping method yields a more tightened solution region (Figure 2-3j) than that constrained by the multi-event stacking method (Figure 2-3i).

### 2.3. Results and Discussion

We applied the above stacking method to all the 22 stations that have high SNR SKS/SKKS data. To ensure that our measurements are robust, we used the following criteria in the splitting analysis: (1) energy on the transverse component is significantly reduced after the correction of anisotropy, (2) difference in the fast-axis direction measured from minimizing transverse energy and second
eigenvalue is small (within the error), and (3) the measurements are not affected by
the selection of time window length.

Among the 22 stations, CCP2 had the largest splitting time, 3.2±1.0s, in the
region (Figures 2-3i and 2-3j). This unusual large splitting time is consistent with
single-event based measurements (Figures 2-3a to 2-3h), and is nearly insensitive to
the selection of SKS/SKKS time windows. Figure 2-4a shows the uncorrected eight
SKS/SKKS waveforms recorded at the station. The SKS/SKKS phase is clearly shown
in the transverse component with amplitudes comparable to the radial ones (Figure
2-4a). Most of them exhibited an elliptical particle motion (Figure 2-4b). After the
correction of seismic anisotropy, the transverse component shows virtually no
SKS/SKKS energy (Figure 2-4c), and their particle motions are almost linear (Figure
2-4d). Silver and Chan (1988) showed that the radial and transverse components
\( u_R(t) \) and \( u_T(t) \) are related to the source waveform \( s(t) \) by the following two
equations:

\[
\begin{align*}
    u_R(t) &= s(t) \cos^2 \phi + s(t - \delta t) \sin^2 \phi \\
    u_T(t) &= [s(t) - s(t - \delta t)] \sin \phi \cos \phi
\end{align*}
\]

Here \( \phi \) is the angle between the fast and radial directions. When \( \delta t \) is small
compared to the dominant period of \( s(t) \), the transverse component, \( u_T(t) \), is
approximately proportional to the time derivative of the radial component, \( u_R(t) \).
The uncorrected seismograms in Figure 2-4a clearly show this feature.
Figure 2-3 Estimates of the two splitting parameters (ϕ, δt). (a-h) Color contour plot of $E_{Ti}(ϕ, δt)$ for the 8 events used in estimating the optimum (ϕ, δt). White plus and red area represent the minimum value and the 95% confidence region, respectively. Green star indicates the joint solution of the events. The two vertical white lines indicate the polarization direction of the incoming SKS/SKKS wave and its perpendicular direction (I and j). Color scales used in plotting the contours are shown below the plots. (i) Color contour plot of summed transverse energy $E_T(ϕ, δt)$. Green star and red area indicate the minimum value and the 95% confidence region. Vertical white lines indicate the back azimuths of the 8 events and their perpendicular directions. (j) Color contour plot of averaged grid count $n(ϕ, δt)$ defined by equation (5). Red area indicates 95% confidence region constrained by all the events. Green star is the average of 95% confidence region.
Figure 2-4 (a) Original radial and transverse components of the 8 events. Their corresponding splitting measurements are shown in Figure 2-3. Note the large SKS/SKKS amplitude in the transverse component. (b) Particle motion of the radial and transverse components. (c) Corrected radial and transverse components. Note that in the corrected seismogram, the transverse component shows little or no SKS/SKKS energy. (d) Particle motion of the corrected radial and transverse components, note the change from an elliptical to linear particle motions.
The estimated splitting parameters \((\varphi, \delta t)\) for each station are listed in Table 2-1 and are shown in Figure 2-5. Uncertainties in \(\varphi\) and \(\delta t\) were estimated by the overlapping method. We also listed the numbers of events and back azimuths used in the measurements. Among the 22 stations measured, 20 have events arriving from at least two back azimuthal directions (Table 2-1, Figure 2-2). For all the 22 stations, we were able to find a common solution that falls into the individual solution regions of all the events. For a homogenous anisotropic layer with a horizontal symmetry axis, \(\varphi\) and \(\delta t\) measured from waves arriving from different azimuths are expected to have the same values. When a dipping symmetry axis or a depth varying anisotropy is present (e.g., Silver and Savage, 1994; Schulte-Pelkum and Blackman, 2003), \(\varphi\) and \(\delta t\) would show a periodic azimuthal variation with a period of 90 degrees. As mentioned above, most of the stations have only 2 to 3 back azimuths that are approximately 90 degrees apart (Figure 2-2b). The poor distribution in azimuth of the data here does not allow us to examine depth variation of the anisotropic structure beneath the study region. Our interpretation will be focused on the large lateral variations seen in the area.

We have 9 stations located at three islands and along the coast north to the now largely inactive dextral strike-slip Oca-Ancon fault (white box in Figure 2-5). All the stations show a fast direction roughly parallel to the EW direction. The splitting times vary from 1.3 s to 3.2 s with an average of 2.0 s, which is consistent with the large delay times observed from stations deployed along the eastern section of the CAR-SA plate boundary adjacent to the active plate boundary strike-slip faults (Russo et al., 1996; Growdon et al., 2009). As mentioned above, we paid special
attention to consistency between different events in determining the splitting parameters; it is unlikely that the large delay times observed here are caused by contamination from some particular ray paths. Also, measured splitting parameters for raypaths originating in Tonga and in the Japan-Kurile-Kamchatka-Aleutian regions were similar, implying that the splitting is accumulated along the upper mantle part of the raypath beneath the station. In principle, either a thick or a strong anisotropic layer can result in a large delay time. For example, a 5% S-wave anisotropy is sufficient to give a delay time of 2 s in a 200-km thick upper-mantle layer. Thus to explain the largest 3.2 s delay time observed here, an S-wave anisotropy of 7-8% is required if we fixed the anisotropic thickness to the ~200 km. Mainprice and Silver (1993) calculated S-wave anisotropy of natural xenolith and ophiolite samples and found that the ophiolite samples are twice as anisotropic (maximum anisotropy of 8-9%) as the kimberlite nodules (maximum anisotropy of ~3.7%). Thus it is still possible to produce a ~2-3 s delay time within a boundary where subduction of oceanic lithosphere is involved.

Russo and Silver (1994) proposed a large-scale trench-parallel mantle flow beneath the Nazca slab based on a compilation of seismic anisotropy measurements made along the west coast of South America. Although the Nazca plate is subducting beneath the SA plate, the Nazca trench is retreating progressively to the west. The combination of the two motions is known as retrograde motion. If the slab is decoupled from the mantle below, the retrograde motion of the slab exerts stress on the asthenosphere beneath it, leading to the development of a trench-parallel flow below the slab. Based on subduction geometry, Russo and Silver (1994) further
hypothesized that the flow is initiated at the coast of the central Andes and moves north and south separately along the coast. Both branches of the flow are diverted eastward when they clear the edge of the subducting plates beneath the Caribbean and Scotia Sea. The eastward flow beneath the CAR, coupled with the eastward translation of the Atlantic slab relative to SA created a strong vertically coherent shear along the strike-slip El Pilar fault systems between the southeastern CAR and the SA plates. Russo et al. (1996) and Growdon et al. (2009) attributed the observed large splitting times to a vertical shear directly beneath the plate boundary. Growdon et al. (2009) also found that the estimated delay times decrease quickly toward the south in eastern Venezuela, indicating that the flow induced strain must drop rapidly away from the boundary.

We also consider the large splitting time observed in the western coast of Venezuela to be associated with the large-scale retrograde flow proposed by Russo and Silver (1994) (Figure 2-6). As mentioned before, tomography images (Van der Hilst and Mann, 1994; Bezada et al., 2010) showed that the southern edge of the CAR plate is subducting beneath the Santa Marta and Perija ranges along a WNW-ESE direction. The P-wave tomography images of Bezada et al. (2010) showed that the Caribbean plate starts subduction beneath northern Colombia at a very low angle and then dips almost vertically beneath the eastern flank of the Perija Range and Lake Maracaibo. A slab tear was hypothesized to separate the steeply dipping slab from the Caribbean to the north (Taboada et al., 2000) (Figure 2-6). We speculated that the retrograde flow passes around the northwestern corner of the SA plate at a location far west of the subducting Caribbean slab, as we observed
minor seismic anisotropy (0.0-0.5s) at VIRV located at the east flank of the Perija Range. The subducting slab thus may act like a barrier that prevents the retrograde flow from entering below the SA plate and directs the flow toward the northern edge of the slab below the hypothesized tear. To produce the observed 2 to 3 s delay time, the flow has to have a vertical dimension of ~200 km if we assume an S-wave anisotropy of 8%.

![Diagram](image)

**Figure 2-5** Measured splitting parameters are shown together with the topography of the study region. The orientation of the black solid lines indicates the orientation of the fast direction. In general the study area can be divided in 3 different zones. The first one is north of the Oca-Ancon dextral strike slip fault, where splitting times are the largest (1.6 – 3.2 s) with an east-west orientation. The second zone, southeast of the Maracaibo Block, shows a splitting direction N45E, parallel to Bocono Fault and Mérida Andes. In the third zone (inside Barinas-Apure Basin) also with an east-west orientation, the smallest delay times are observed (~0.8 s).

The 3 stations inside the Barinas-Apure basin situated at the southeast side of the Bocono Fault (white ellipse in Figure 2-5) recorded the lowest splitting times,
~0.8 s among all the stations (except for VIRV and QARV, Table 2-1). The estimated fast directions average to 77° or 257° East of North, consistent with the absolute motion (~262° clockwise from north) of the SA plate (Gripp and Gordon, 2002). The estimated fast directions and delay times are also consistent with those observed in the Maturin basin and the Guayana shield in eastern Venezuela (Growdon et al., 2009). We thus interpret the observed seismic anisotropy to be caused by a simple asthenospheric flow parallel to plate motion. While stations along the coast and within the Barinas-Apure basin exhibited an ~EW fast direction, the 8 stations in the Mérida Andes (yellow rectangle in Figure 2-5) showed a NE-SW fast direction, parallel to the strike of the Mérida Andes. Moderate delay times were obtained from these stations with amplitude varying from 1 to 1.4 s. Pinero-Feliciangeli and Kendall (2008) measured (\(\phi, \delta t\)) at the GSN station SDV and found similar alignment between the fast polarization direction and geological strike. As described above, seismic anisotropy measured from SKS waveform splitting is closely related to the deformation style and geometry in the upper mantle. Thus the good coherence between the geologic fabric and the fast direction observed here suggests that the crust and the upper mantle of the Mérida Andes deforms coherently (Figure 2-6) in response to compressional forces related to uplift of the mountains. Assuming a maximum S-wave anisotropy of 4% (Mainprice and Silver, 1993), we estimated that the deformed subcontinental mantle extends to a depth of ~150-200 km in order to generate a 1-1.4 s splitting time if we assume anisotropy is evenly distributed across the lithosphere.
Figure 2-6 Model of the subducting CAR plate with a hypothesized slab tear near the coast of western Venezuela that separates the subducting from the non-subducting Caribbean plate. Red arrow represents the large-scale mantle flow around the subducting plate, similar to that proposed by Russo and Silver (1994). The eastward flow produces a strong shear zone in the upper mantle along the CAR-SA plate boundary, which is observed as large splitting times in the coastal region adjacent to and north of the plate boundary. Along the Mérida Andes the split measurements suggest the entire lithosphere is deformed coherently parallel to the Bocono Fault (see the inset cartoon, which was modified from Silver (1996)), where F indicates the strike of the main fault and the arrows represent the direction of the shear stress field across the fault). Inside the Barinas-Apure Basin delay times are small with in a direction consistent with the absolute motion direction of the South American plate.

Deformation associated with the uplift of the Mérida Andes has been assumed to be confined within the crust and to be related to type-A (continent-continent) subduction, although it is still debated as to whether the subduction is SE- (e.g., Kellog and Bonini, 1982; Burke et al., 1988; Colletta et al., 1997) or NW-directed (e.g., Audermard and Audermard, 2002). Our measurements of splitting times here, however, suggest that beneath the mountain range a vertically coherent deformation extends much deeper than the crust-mantle boundary, which is estimated to be ~50 km deep (Niu et al., 2007). This is because: 1) if the thick crust is the main cause of the observed 1-1.4 s splitting time, then an unreasonably high
crustal anisotropy (~10%) is required, as for a 50 km thick crust with an average S-wave velocity of 3.5 km/s, the travel time is approximately 14 s for a vertically propagating wave; 2) Niu et al. (2007) found that at the GSN station SDV the arrival time and amplitude of the Moho Ps conversion depend on the directions of the incoming P waves. The azimuthal variation, however, doesn't exhibit a \( \pi \)-periodicity, which is an indicator of crustal anisotropy. Rather, the variation can be better explained by a dipping Moho structure. Thus it is likely that the formation of the Mérida Andes has been accompanied by shortening of the entire lithosphere. As the surface deformation is highly localized in the Mérida Andes, we speculate that a weak and thin lithosphere may have been present under the mountain range before its recent uplift. There are geological and crustal seismic studies (e.g., Molnar and Sykes, 1969; Kellog and Bonini, 1982; Colletta et al., 1997; Audermard and Audermard, 2002; Duerto et al., 2006) suggesting the occurrence of active rifting in the study area during the Jurassic. Rift related Jurassic grabens have been found in outcrops and have been identified in seismic sections across the Mérida Andes.

2.4. Conclusions

We measured shear wave splitting at 22 stations in northeastern Venezuela with a multi-event stacking method. With the limited data, we found no apparent azimuthal variation of the splitting parameters and employed a one-layer anisotropic model in interpreting the measurements. We found three distinct areas with different orientations and magnitudes, and inferred three different deformation mechanisms across western Venezuela: (1) Stations located on north of
the Caribbean-South American plate boundary on islands and in the coastal area north to the Oca-Ancon fault have the largest splitting times, ~2-3 s, and a fast direction parallel to the EW CAR plate motion direction, which can be explained by a strong eastward flow confined at the CAR-SA plate boundary. (2) Seismic anisotropy estimated from stations located within the stable SA plate is weak, is parallel to SA plate motion, and has an origin likely in the asthenosphere. (3) Intermediate splitting times, ~1.0-1.5 s, and a ~NE-SW fast direction are observed at stations deployed in the Mérida Andes, suggesting that the deformation of the crust and subcontinental mantle contributes to the anisotropy. It is likely that the lithospheric mantle plays a major, and possibly dominant, role in the formation of the Mérida Andes.
Chapter 3

Lithospheric expressions of the Cenozoic subduction, Mesozoic Rifting and the Precambrian Shield in Venezuela

We have combined surface wave tomography with Ps and Sp receiver-function images based on common-conversion-point (CCP) stacking to study the upper mantle velocity structure, particularly the lithosphere-asthenosphere boundary (LAB), beneath eastern and central Venezuela. Rayleigh phase velocities in the frequency range of 0.01-0.05 Hz (20-100 s in period) were measured using the two-plane-wave method and finite-frequency kernels, and then inverted on a 0.5°×0.5° grid. The phase velocity dispersion data at each gridpoint were inverted for 1D shear velocity profiles using initial crust-mantle velocity models constructed from previous studies. The 3D velocity model and receiver-function images were interpreted jointly to determine the depth of the LAB and other upper mantle
features. The tomographic images revealed two high velocity anomalies extending to more than \(\sim 200\) km depth. One corresponds to the top of the subducting Atlantic plate beneath the Serrania del Interior. The other anomaly is a highly localized feature beneath the Maturin Basin, which we speculate is a lithospheric drip from the South America (SA) plate. The LAB depth varies significantly in the study region. It is located at \(\sim 110\) km depth beneath the Guayana Shield, and reaches \(\sim 130\) km at the northern edge of the Maturin Basin, which may be related to the downward flexural bending due to thrust loading of the Caribbean plate by the load from Atlantic subduction. Immediately to the west, lithosphere is thin, \(\sim 50-60\) km, extending NE-SW along the Jurassic Espino Graben from the Cariaco basin offshore to the Orinoco river at the edge of the craton. The LAB in this region is the top of a pronounced low velocity zone. Westward the lithosphere deepens to \(\sim 80\) km depth beneath the Barinas Apure Basin, and to \(\sim 90\) km beneath the Neogene Merida Andes and Maracaibo block. Both the upper mantle velocity structure and lithosphere thickness correlate well with surface geology and are consistent with northern South American tectonics.

### 3.1. Introduction

The Southern Caribbean plate boundary has been formed by interactions between the South America (SA) and Caribbean (CAR) plates since collision of northwestern Venezuela with the Caribbean \(\sim 55\) Ma (e.g. Kellogg and Bonini, 1982; Meschede and Frisch, 1998; Pindell et al., 2005). GPS measurements indicate that the CAR is currently moving approximately 2 cm/yr relative to SA, parallel to the
strike slip fault system in the east, with oblique convergence in the west (Perez et al., 2001; Weber et al., 2001; Bilham et al., 2013), the latter results in subduction and underthrusting of the southern edge of the Caribbean beneath northwestern South America (Figure 3-1). Although in eastern and central Venezuela almost all CAR-SA plate motion is taken up along the El Pilar-San Sebastian strike-slip faults, deformation occurs in a broader zone, ~300 km wide in eastern Venezuela and up to ~600 km wide in western Venezuela and eastern Colombia (e.g. Meschede and Frisch, 1998; Audemard and Audemard, 2002; De Toni and Kellogg, 1993). The plate boundary in western Venezuela is complicated by the motion of the triangular Maracaibo block (MB), a deforming lithospheric block that is escaping northward relative to SA along the Bocono and Santa Marta strike-slip fault systems (Figure 3-1). The ranges bounding the MB are the Merida Andes on the east, and the Perija Range and the Santa Marta Massif on the west (Kellogg and Bonini, 1982; Audemard and Audemard, 2002), forming the Eastern Cordillera of the northernmost Andes.

In central and eastern Venezuela, plate motion is accommodated by transpression and transtension along the right lateral San Sebastian-El Pilar strike-slip fault system (Perez et al 2001; Audemard et al., 2005), a strike-slip system of similar scale to the San Andreas Fault in California. The San Sebastian-El Pilar system marks the northern edge of coastal thrust belts (such as Cordillera de la Costa and Serrania del Interior) and their associated foreland basins (e.g. Guarico and Maturin Basin). Seismicity along the strike slip boundary is mostly shallow (< 50 km; Palma and Romero, 2011), and concentrated along the major strike slip fault systems with more diffuse seismicity on the thrust faults in the Serrania del Interior,
the Cordillera de la Costa, and the Mérida Andes (Figure 3-1). At its eastern end, the transform motion ends at the southernmost corner of the Antilles subduction zone, where the Atlantic plates (oceanic lithosphere of the north and South America plates) are subducting beneath the Greater and Lesser Antilles (Molnar and Sykes, 1969; Russo and Speed 1992). Intermediate depth seismicity has been used to infer the top of the subducting slab to ~200 km depth, which ceases abruptly beneath the Gulf of Paria (VanDecar et al., 2003). The southernmost intermediate depth seismicity cluster offshore of the Paria Peninsula has been interpreted as the expression of lithospheric tearing (e.g. Russo et al., 1993; Clark et al., 2008a, 2008b), with the transition from subduction zone to strike slip system characterized as a Subduction-Transform Edge Propagator (STEP) boundary (Govers and Wortel, 2005). Tomographic images show a steeply subducting slab, and also an aseismic landward continuation of descending lithosphere beneath the Serrania del Interior (VanDecar et al., 2003; Bezada et al., 2010a). This was interpreted as either convective removal of lithospheric mantle from the continental SA near the subducting Atlantic slab (Bezada et al., 2010a), or northward subduction of the SA passive margin lithosphere (VanDecar et al., 2003). Outcropping of intrusive (~5Ma) rocks onshore Venezuela (Santamaria and Schubert, 1974; McMahon, 2001), and the location of geothermal systems in eastern Venezuela associated with a shallow intrusive body (Urbani, 1989) are interpreted as geologic indicators of the presence of the edge of the slab beneath continental SA.
Figure 3-1 (a) Caribbean tectonics, with GPS vectors in the southern Lesser Antilles Arc and on San Andreas (SA) Island in the western Caribbean. Red box indicates the study region shown in (b). Blue box indicates study region for the Rayleigh wave tomography. (b) Broadband seismic stations in Venezuela used in the investigation of the LAB. Red triangles indicate permanent stations of the National Seismic Network of Venezuela. Blue and pink triangles represent temporal deployments under the first and second phases of the BOLIVAR project. White triangles are OBS deployed during the first phase of the BOLIVAR project. Seismicity at different depths is shown with color-coded circles. Major active faults are indicated solid black lines. MB: Maracaibo Block, BF: Bocono Fault, OAF: Oca-Ancon Fault; SMF: Santa Marta-Bucaramanga Fault; SS-EP: San Sebastian -El Pilar Fault; BAB: Barinas-Apure Basin; CC: Cordillera de la Costa; SI: Serrania del Interior; OR: Orinoco River; LA: Leeward Antilles. GS: Guayana Shield. GP: Gulf of Paria; EG: Espino Graben.
In western Venezuela and eastern Colombia, intermediate depth seismicity and tomographic images suggest that the CAR plate subducts beneath SA offshore of the Santa Marta Massif with a very low angle (Van der Hilst and Mann, 1994; Malave and Suarez, 1995; Taboada et al., 2000). It then dips steeply to ESE under Lake Maracaibo and the Mérida Andes (Pennington, 1981; Taboada et al., 2000; Bezada et al., 2010a), reaching transition zone depths. The flat subduction beneath northern Colombia and westernmost Venezuela is thought to be responsible for the uplift and deformation of the Santa Marta Massif, the Perija and the Mérida Andes (Kellogg and Bonini, 1982; Bezada et al., 2010a). Another viewpoint holds that these ranges are the result of the late Miocene-Pliocene collision of the Panama arc with SA (Molnar and Sykes, 1969).

The Archean-Proterozoic Guayana Shield, part of the Amazonian Craton, underlies southeastern and south-central Venezuela (Gonzalez de Juana, 1980). The Guayana Shield is composed mainly of Archean and Proterozoic rocks, and is usually divided into four provinces (Yoris and Ostos, 1997). The Archean Imataca province (3700-3400 My) lies in the northern end of the shield just south of the Orinoco River. It is bounded by two Proterozoic provinces: Pastora (2300-1900 My) and Cuchivero (1900 – 1400 My), to the southeast and southwest, respectively. A fourth Proterozoic province, Roraima (1800 – 1600 My), lies further south of the Pastora and Cuchivero provinces beyond the study area.

In the Cenozoic, northern Venezuela and the continental interior have been affected by a range of geodynamic processes: Subduction, transpression, and
transtension have formed the modern plate boundary. The crustal structure of much of Venezuela has been determined from a variety of seismic probes (Schmitz et al., 2002; 2005; Niu et al., 2007; Guedez, 2007; Clark et al., 2008c; Bezada et al., 2008; Bezada et al., 2010b; Magnani et al., 2010), and the deeper structure through the transition zone has been determined from regional and global teleseismic tomography (Bezada et al., 2010a; Bijwaard et al., 1998). The present study is designed to image the structure of the lithosphere and the lithosphere-asthenosphere boundary (LAB) using Rayleigh wave tomography, Ps and Sp receiver functions. This study is part of the BOLIVAR (Broadband Onshore-Offshore Lithospheric Investigation of Venezuela and the Antilles Arc Region) and GEODINOS (Geodinamica Reciente del Limite Norte de la Placa Sudamericana) projects, a joint U.S. - Venezuelan investigation of the structure and evolution of the southeastern and central CAR-SA plate boundary (Levander et al., 2006). The data used here consist of teleseismic events recorded by a broadband seismic array with 93 elements including both land and ocean-bottom seismometers extending from the Caribbean basin to the Guayana shield in the Venezuelan interior (Figure 3-1).

3.2. Data, Methods and Results

3.2.1. BOLIVAR data set

The data used in this study were recorded by the BOLIVAR array; the first phase of the deployment included a temporary network (~18 months) of 27 stations deployed by IRIS-PASSCAL, a yearlong deployment of 13 OBSIP ocean
bottom seismographs, and 8 Rice broadband stations. The second phase consisted of 7 broadband stations installed for 13 months in western Venezuela. The portable array data are complemented by data from 37 stations of the Venezuelan National Seismological Network operated by the Fundación Venezolana de Investigaciones Sismológicas (FUNVISIS) and one broadband station operated by the Global Seismic Network (GSN) (Figure 3-1b). The 93 stations cover an area of ~1200 km by ~600 km. In eastern Venezuela, station spacing varies from ~10 km to ~100 km, with an average of ~50 km. Stations in western Venezuela (> ~68°W) are primarily from the FUNVISIS network and the 7 stations of the second phase of deployment. The station coverage thus is much sparser in the west.

3.2.2. 3D Initial velocity model

Initial crustal velocity models including constraints on Moho depths are important for inverting for shear velocity from surface wave dispersion data and in estimating depth of mantle discontinuities using receiver function data. We first constructed a 3D crustal model from refraction and reflection/wide-angle seismic data (Schmitz et al., 2002; 2005; Guedez, 2007; Clark et al., 2008c; Bezada et al., 2010b; Magnani et al., 2010). We used the Vp/Vs relationship of Brocher et al. (2010) to estimate crustal shear wave velocity structure. Crustal thickness was compiled from the active wide-angle seismic data cited above and passive receiver function data (Niu et al., 2007). We also included 3D mantle velocity structure obtained from finite-frequency P-wave tomography (Bezada et al., 2010a) in our
initial model. S-wave velocity in the mantle was computed using the Vp/Vs ratio of the AK-135 model (Kennett et al., 1995).

3.2.3. Receiver function analysis

Receiver functions have been widely used to estimate Moho depth and to image seismic discontinuities in the mantle. In a layered medium, a teleseismic record can be considered as the summation of the direct arrival and a series of conversions and reflections at boundaries below the station. Receiver functions are an attempt to approximate a Green’s function associated with structure beneath the receiver by deconvolving one component of a teleseismic signal from another to remove source signals from seismograms (Ammon, 1991). For instance, deconvolving the vertical (or longitudinal) component from the radial (or in plane shear) component helps to isolate the P-to-S converted waves, generating a Ps receiver function. Given a velocity model, one can use the time differences between converted waves and the direct arrival to estimate the depths of the Moho and deeper boundaries in the mantle, such as the LAB. To improve the signal-to-noise ratio (SNR) of the P-to-S converted arrivals, we employed the common-conversion-point (CCP) stacking method (Dueker and Sheehan, 1997), and here used the method described in Levander et al. (2011). We selected a total of 48 events with Mw ≥ 6 and epicentral distances between 35° and 90° recorded by the BOLIVAR array to generate Ps receiver functions (Figure 3-2).
Figure 3-2 (a) Location of the 48 events located at 35°-90° used for the Ps receiver function (red dots) and of the 42 events located at 55°-85° used for the Sp receiver function (blue dots). (b) Location of the 45 events located at > 30° used for the Rayleigh wave tomography (red dots). (c) Example of raypath coverage for Rayleigh waves with a period at ~45 s. Triangles indicate the broadband stations in the study area.
The procedure to generate Sp receiver functions is similar to the method of calculating Ps receiver functions. In this case, the in-plane S wave and its precursors are deconvolved from the longitudinal component to isolate S-to-P converted arrivals before the direct S (Yuan et al., 2006). In the Ps receiver function data, P-to-S converted arrivals from deep structures usually suffer from interference of multiple reflections from shallow interfaces. In particular, Moho reverberations can mask the P-to-S conversion at LAB depths when the two interfaces fall within certain depth ranges. This interference sometime makes it difficult to image the LAB with Ps receiver function data. Since the S-to-P conversions arrive before, and multiple reflections arrive after the direct S wave, such interference does not exist in the Sp receiver function data. The lower frequencies are also more appropriate for imaging gradual velocity transitions such as a thermally controlled LAB. On the other hand, the spatial resolution of Sp receiver function data is about 5 times coarser than that of the Ps receiver functions (Levander et al., 2011) due to the lower frequency content of S-wave signals (Yuan et al., 2006). We chose a total of 42 events with Mw > 5.7 and epicentral distances between 55° to 85° during the period of 2003-2009 to generate Sp receiver functions (Figure 3-2a).

The Ps receiver functions we present in this study have a shaping filter with high corner frequencies of 0.5 Hz and 1Hz, while the Sp receiver functions have 0.2 and 0.1 Hz shaping filters. Receiver functions were calculated using both iterative deconvolution (Ligorria and Ammon, 1999) and water level deconvolution (Ammon, 1991). We found that there are no significant differences in the receiver functions computed with different methods.
3.2.4. Finite Frequency Rayleigh wave tomography

In order to determine the three-dimensional shear wave velocity structure within the study area, we used the two-step inversion technique developed by Yang and Forsyth (2006a). The first stage involves computing 2-D phase velocities using the two plane wave method (Forsyth & Li, 2005; Yang & Forsyth, 2006a, 2006b). This method accounts for effects due to multipathing of the incoming wavefield. Finite-frequency amplitude and phase sensitivity kernels were used to improve lateral resolution (Yang & Forsyth, 2006b). This method of calculating phase velocities has proved to be effective in calculating phase velocity using different data sets from around the world (Yang & Forsyth, 2006a; Schutt et al, 2008; Yang and Ritzwoller, 2008; Liu et al., 2011; Liu et al., 2012). A previous investigation of this area using the same dataset (Miller et al., 2009) did not make use of the finite-frequency kernels to determine phase velocities.

The two-plane wave method assumes that the region is small and distant from the sources, such that waveform complexity is simple enough to be adequately represented as the sum of two interfering plane waves, a primary signal and a secondary multi-pathed signal. If the dimensions of the area analyzed are small compared to the distance to the source, the phase and amplitude of the source can be assumed to be constant. Considering that the station distribution is denser in eastern Venezuela than in the rest of the country, we re-parameterized the original study area (12° x 8°) used by Miller et al. (2009), into a smaller area of (7° x 5°) in eastern Venezuela, between 61°W to 68°W longitude and 6.5° to 11.5° latitude.
(Figure 3-1), with a grid spacing of 0.5° and surrounded by a larger grid of 2 more sets of grid nodes, in order to absorb travel time residuals that cannot be modeled by the two plane wave representation. Within this area there are 53 stations from the combined FUNVISIS/BOLIVAR array (Figure 3-1). We analyzed data from a total of 45 earthquakes within a distance of 30° to 120°, and magnitude greater than 5.7 that occurred between December 2003 and May 2005 (Figure 3-2b). As shown in Figure 2c, we have moderately dense raypath coverage (Figure 3-2c, Table 3-1).

Table 3-1 Phase velocity inversion data

<table>
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<th>Freq. (mHz)</th>
<th>Period (s)</th>
<th>Average Phase velocity (km/s)</th>
<th>Number of raypaths</th>
<th>Number of events</th>
<th>Wavelength (km)</th>
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Figure 3-3 (a) Average Rayleigh wave phase velocities at 14 periods ranging from 20 to 100 s. Error bars represent two standard deviations. (b) Model norm is shown as a function of the dispersion residual norm for various damping parameters. The optimum value of the damping parameter is given at the elbow of the L-shaped curve. (c) Model resolution kernels for layers at depths of 15, 55, 105, 155, and 205 km. Resolution decreases with depth.

We inverted for isotropic phase velocities at each 0.5°×0.5° grid point at periods ranging from 20 to 100 s (Table 3-1). We found that the average phase velocities within the study area vary from 3.488 km/s at 20s to 4.122 km/s at 100s (Figure 3-
3a, Table 3-1). The next step was inversion of dispersion curves at each grid point for 1D Vs models, for which we used the DISPER80 code (Saito, 1988). The phase velocities at each grid node were inverted individually, using the 3D starting velocity model described above. We tested various damping parameters and calculated an L curve, which is a plot of the model norm versus the norm of dispersion residuals, and found an optimum value of 0.25 (Figure 3-3b). The optimum value of the damping parameter agrees with those used in previous studies that employed the same inversion method (Schutt et al., 2008; Liu et al., 2011). Considering that the greatest depth sensitivity to shear velocity is at approximately 1/3 of the wavelength for Rayleigh waves (Aki and Richards, 2002), shorter periods (up to 25 s) sample crustal structures and longer periods sample upper mantle structure as illustrated by the resolution kernels in Figure 3-3c. The resulting 1D shear velocity profiles at each grid node were combined to form a pseudo-3D velocity volume that extends to a depth of ~200 km (Figure 3-5).

In Figure 3-4, we show phase velocity variations in the study area observed at six different frequencies (periods). At frequencies corresponding to crustal structures (Figure 3-4a and 3-4b), we observe a very slow anomaly beneath the Serrania del Interior that extends east toward the Trinidad, and west to the Cariaco basin. At 34 s there is a high velocity anomaly at ~63°W beneath the Guayana Shield (Figure 3-4c). At almost all periods and therefore depths, the Guayana Shield shows a persistent higher-than-average velocity compared to the surrounding regions (Figures 3-4a to 3-4f).
Figure 3-4 Phase-velocity perturbations at periods of 20, 25, 34, 50, 67, and 91 s. Phase velocity inversions were conducted with a grid spacing of 0.5°x0.5°. The thick and thin dashed lines indicate the San Sebastian-El Pilar Fault, and the thrust front of the coastal cordillera.
Figure 3-5 Lateral variations of shear wave velocity at depths of 15, 30, 60, 80, 100, 120, 150, and 200 km are shown in (a), (b), (c), (d), (e), (f), (g) and (h), respectively. SS-EP: San Sebastian-El Pilar Fault system; CC: Cordillera de la Costa; SI: Serrania del Interior; EG: Espino Graben; CB: Cariaco Basin. Note a low velocity zone (LVZ) extending to ~100 km beneath the Cariaco Basin, and two deep high-velocity anomalies beneath the SI and the southern part of the Maturin Basin.
The Vs model (Figure 3-5) shows rapid variations in shear velocity between the different tectonic provinces to ~80 km depth, which we illustrate further below with cross sections. The upper mantle exhibits a number of persistent features with depth. In eastern Venezuela a complex high velocity feature extends to 200 km. We associate the northeastern fast anomaly (up to 4.70 km/s) lying beneath the Serrania del Interior with the southern edge of the subducting Atlantic plate. We identify a second deep anomaly beneath the Maturin Basin (Figures 3-5c-3-5h). High velocities (4.55-4.60 km/s) persist beneath parts of the Guayana shield to 200 km. In central Venezuela below ~80 km depth, the upper mantle has little lateral variation and relatively low (4.40-4.45 km/s) velocities.

### 3.3. Interpretation

#### 3.3.1. Crustal structure

In this section we compare the Rayleigh wave tomography and receiver functions to infer crustal structure. In general our derived Moho topography is consistent with the Moho previously derived from wide-angle and receiver function data of the BOLIVAR project (Schmitz et al., 2002; Niu et al., 2007; Guedez, 2007; Clark et al., 2008c; Magnani et al., 2009; Bezada et al., 2010b), although with some notable differences at the northern edge of the study area. These differences are mainly related to the lack of good quality data from the OBS stations to constrain the CCP stacked image offshore.
In western Venezuela, beneath the Paraguana Peninsula, CCP stacked images reveal the presence of a second P to S conversion at 40 to 60 km, dipping at \( \sim 15^\circ \) towards the continent (Figure 3-6b), which we interpret as the Moho of the CAR plate underthrusting beneath SA. It is important to notice that there is limited data at the northern end of the profile; only two stations on the Paraguana Peninsula are used to constrain this feature.

Beneath the Maracaibo Block, there is also a second P to S conversion at \( \sim 60 \) km, dipping at \( \sim 17^\circ \) towards the continent (Figure 3-7), which we interpret as the Moho of the subducting CAR plate. The observed dip angle here is expected to be lower than the true dip because CCP stacking does not restore dips to the proper angle. It, however, agrees reasonably well with values derived from the Wadati-Benioff seismicity beneath northern Colombia, i.e., 20° by Pennington (1981) and 25° by Malave and Suarez (1995). Like Profile A, only limited data are employed in constraining this structure.

In central Venezuela beneath the Cordillera de la Costa, we find a second P to S conversion at \( \sim 35 \) km depth that extends from the Bocono fault to the Serrania del Interior (Figures 3-7 and 3-8). Although crustal structures are not so clear on the Sp receiver functions due to the lower frequency content, the sections show complications in the locations of the Ps events. The shear velocity profile from surface waves (Figure 3-8c) shows velocities of \( \sim 4.0 \) km/s below the Moho. These low velocities may suggest the presence of crustal material from the CAR plate due to underthrusting.
Figure 3-6: (a) Map showing the location of the 6 profiles (A)-(E) that are shown in Figures 3-6 to 3-11. CCP stacked images of the Ps (b) and Sp (c) receiver functions of profile (A) are shown together for comparison. Red and blue colors indicate positive and negative velocity jumps, respectively. Black and white solid lines indicate the Moho and LAB picks from the images, and the dotted line represents the Moho depth estimated from active source data along the 70° west line (A.S.) (Guedez, 2007; Bezada et al., 2008). The second black line labeled with “CAR?” at the left side of the Ps image in (b) is interpreted as the subducted Caribbean plate. Black dashed line represents the top of the subducted Caribbean plate interpreted by Bezada et al., 2008). Seismicity within 50 km along the profile is denoted by small black dots.
Figure 3-7 (a) and (b) are similar to Figure 3-6b and Figure 3-6c, respectively, but for the profile B. (c) S-wave velocity inverted from Rayleigh wave dispersion data. The two solid black lines in (c) are the input Moho and the picked LAB, respectively. The long white dashed line in (a) and (b) denotes the LAB location derived from the surface wave data. Active-source estimates of the Moho at 70°W (Guedez, 2007; Bezada et al., 2008), 67°W (Magnani et al., 2009), 65°W (Bezada et al., 2010b), and 64°W (Clark et al., 2008c) are also shown in (a) and (b). The shower event beneath the Cordillera de la Costa in (a) is interpreted as the detachment base of the overthrust Caribbean plate. The deep tilted event beneath the Maracaibo Block in (a) is attributed to the subducting Caribbean plate. Black dashed line represents the top of the subducted Caribbean plate interpreted by Bezada et al., 2010a).
Figure 3-8 Same as Figure 3-7 except for the profile C. The shallow solid line marked with "Detachment?" in (a) is interpreted as the detached base of the overthrust Caribbean plate. Moho estimates from active-source data are the 67° west line analyzed by Magnani et al. (2009).
Figure 3-9 Same as Figure 3-7 except for the profile C. Active-source estimates of the Moho along the profile (64°W) are taken from Clark et al. (2008c).
Figure 3-10 Same as Figure 3-7 except for the profile E.
Figure 3-11 Same as Figure 3-7 except for the profile F.
In eastern Venezuela, beneath the Maturin Basin, where crustal thickness is ~50 km, Rayleigh wave tomography indicates the presence of high velocity (4.25 km/s) in the lower crust (Figure 3-9). Schmitz et al. (2008) also observed high velocities in the lower crust of this area. South to the Maturin Basin, the crust is around 46 km thick beneath the Guayana Shield (Figure 3-10), consistent with results from active source data (Schmitz et al., 2002).

3.3.2. Rayleigh Wave Tomography: Uppermost Mantle Structure

Beneath eastern Venezuela the surface wave tomography shows some interesting features in the uppermost mantle. Low velocities (~4.45 km/s) extending from ~100-200 km depth beneath the Guarico basin, aligned with the direction of the Jurassic Espino Graben (Figures 3-5c-3-5d, 3-8c and 3-11).

We associate the distinct high velocity anomaly from 60-200 km depth under the Serrania del Interior (Figure 3-5e to 3-5g, and 3-9) with the subducting Atlantic slab beneath the Caribbean. As shown in previous tomography images (VanDecar et al., 2003; Bezada et al., 2010a) the anomaly crosses the strike-slip margin and extends ~100 km further south to the southern edge of the Serrania thrust front. It is difficult to interpret the high velocity anomaly beneath the South America continent with the current subduction geometry. VanDecar et al. (2003) suggested that the high velocity anomaly beneath the continent is due to a northward subduction, Bezada et al. (2010a), on the other hand, argued that a simultaneous subduction toward the north under the Caribbean and toward the west under the Lesser Antilles is geometrically difficult to achieve. They proposed that the high velocity
anomaly extending south of the strike-slip system results from convective removal of the South American continental margin lithospheric mantle by the descending Atlantic slab (Figure 11 in Bezada et al., 2010a).

Beneath the Maturin Basin, north of the Orinoco River there is a second high velocity body extending to ~200 km depth (Figure 3-5h and Figure 3-8). We speculate that it is a drip of continental mantle lithosphere that has been destabilized by a combination of the Atlantic subduction and the formation of the foreland basin.

Beneath the Cariaco basin there is a pronounced low velocity zone (LVZ, ~4.3km/s) that extends to ~150 km depth (Figures 3-5, 3-7 and 3-12). Miller et al. (2009) associated this LVZ with asthenospheric flow around the southern edge of the Atlantic plate resulting from slab rollback. Our results do not support this interpretation, since the flow seems to come from underneath of the CAR plate instead. It is, however, unclear what dynamic processes have been involved in generating this LVZ (Figure 3-12).

3.3.3. Lithosphere-Asthenosphere Boundary (LAB)

The lithosphere is defined as the mechanically strong layer that acts as a rigid plate (e.g. Fischer et al., 2010; Lee et al., 2011); the LAB thus, is a rheological boundary separating the rigid lithospheric lid from the mechanically weaker asthenosphere (Fischer et al., 2010). Vertical gradients of geophysical properties have been used as a proxy to identify the LAB, for example, a negative seismic
velocity gradient, a negative seismic impedance gradient, or a rapid change in temperature gradient (Fischer at al., 2010).

We used the Rayleigh wave tomography model and the receiver function data together to make two separate, largely consistent, LAB depth estimates. It can be challenging to identify the LAB from receiver function data when there are either multiple peaks or essentially no distinct peaks in the stacked traces. The latter occurs if the LAB is such a gradual boundary that it generates little S-to-P or P-to-S converted energy (Yuan and Romanowicz, 2010). Thus, the additional constraints provided by the surface wave tomography model have been used with the receiver functions. For the CCP stacked Ps and Sp receiver function traces, we associate negative peaks at depth range between 50 and 200 km as a potential LAB. From the surface wave tomography shear velocity model, following Fischer et al. (2010), we define the LAB as the center of the shallowest negative velocity gradient beneath the Moho in each velocity profile in the 3D grid (Figure 3-13).

We choose an LAB that is consistent between the datasets, for instance, when several peaks are present in the receiver function data, we select one that has the closest value to the surface wave data. If there are no distinct peaks in the receiver function data, we take the value from the surface wave data. Lateral variations of the LAB depth estimated from receiver function and surface wave data are shown in Figures 3-14a and 3-14b, respectively, with an error estimate of ±10 km.
Figure 3-12 3D Visualization of the lithospheric structure and the Low Velocity Zone (LVZ) from the Rayleigh wave tomography results. Orange dots and red squares denote seismicity and seismic stations in the area, respectively. (a) View from the east; (b) view from the north; (c) view from the west. GS: Guayana Shield. ES: Espino Graben. MB: Maturin Basin. CC: Cordillera de la Costa. SI: Serrania del Interior. T&T: Trinidad and Tobago CB: Cariaco Basin. Figure generated with ParaView.
In general the LAB depths estimated from the two types of data agree reasonably well (< 7% difference on average), while the map resulting from surface wave data appears to have better lateral resolution than the receiver function map. In western Venezuela where there is no overlap between the 2 data sets, we estimated LAB depth from the receiver functions only.

Figure 3-13 Inverted shear-wave velocity models (blue) at 8 locations are shown together with the initial models (black). Map in the middle shows the location of the profiles.
We observed a factor of three in LAB depth variation across the study area (Figure 3-14). In places lithosphere thickness changes dramatically over very short distances, for example from the Espino Graben in the Guarico Basin to the adjacent Maturin Basin. There is good correlation between the observed LAB depth and surface geological and tectonic terranes. Beneath the Guayana Shield, the LAB depth varies between 110 and 120 km, and is roughly consistent with the depth extent of a fast velocity anomaly beneath the same region revealed by the finite-frequency body wave tomography of Bezada et al. (2010a). The Archean Imataca province appears to have a slightly deeper LAB (~120 km) than the two Proterozoic provinces ~110 km). The LAB deepens northward, reaching approximately 130 km depth at the northern edge of the Maturin Basin, likely resulting from flexure of the SA lithosphere in response to the loads on the end of the SA plate provided by the sedimentary basin, the thrust belts of the Serrania del Interior and the subducting Atlantic Plate.

Immediately to the west, we find a shallow LAB at ~50 km extending southwest from the offshore Cariaco basin and thickening slightly to 60 km along the axis of the Jurassic Espino Graben (Figure 3-11). To the southwest, the LAB beneath the Jurassic Apure Mantecal Graben (Figure 3-14) is also only ~50-60 km deep. Further to the west, the lithosphere is slightly thicker, ~80 km thick beneath the Barinas Apure Basin. It thickens westward to ~90 km beneath the Neogene Merida Andes and the Maracaibo block located above the steeply dipping subducting Caribbean plate.
Figure 3-14 Depth variations of the LAB estimated from receiver function images (a) and surface wave tomography. SA: South America; OAF: Oca-Ancon Fault; EPF: El Pilar Fault; SSF: San Sebastian Fault; T&T: Trinidad and Tobago. Blue box in (a) indicates the Rayleigh wave tomography study region.
3.4. Geodynamic Implications

3.4.1. Caribbean Underthrusting and Subduction under South America

South of the Southern Caribbean Deformed Belt, beneath the Paraguana Peninsula in northwestern Venezuela (Figure 3-6, Profile A) Bezada et al. (2008) observed a clear arrival in a long offset active-source profile, which they interpreted as reflections from the Moho of the subducting Caribbean plate. The profile they modeled is almost identical to profile A in Figure 3-6, but extends ~2° further north into the Caribbean. The geometry of the slab and its southward extent could not be determined by their data. The CCP stacked images along our profile A (Figure 3-6) show a southeastward dipping feature, between 15 and 20 km beneath the SA Moho, which we interpret as the Moho of the underthrust Caribbean plate beneath the Paraguana Peninsula at the northeastern corner of the Maracaibo Block.

On the west side of the Maracaibo Block, beneath the Perija Range, there is also evidence of a dipping structure, which disappears beneath Lake Maracaibo at ~70 km depth just west of the Merida Andes (Figure 3-7), where finite-frequency P wave tomography (Bezada et al., 2010a) shows that the Caribbean slab starts to subduct steeply. We interpret this as the Moho of the CAR plate that is subducting from west to east beneath the SA plate.

Bezada et al. (2010a) and Masy et al. (2011) proposed a tear in the Caribbean plate separating the underthrusting segment in the northern Maracaibo Block and the subducting segment beneath the western Maracaibo Block. The tear is located in the vicinity of the Oca Fault zone, but has not been imaged directly. Profiles A and B
(Figures 3-6 and 3-7) are consistent with slab tearing, as the slab is present at shallow depth north to the Oca-Ancon Fault (OAF), but is found only on the west side of Lake Maracaibo consistent with the top of the subducting Caribbean found in the teleseismic tomography (Bezada et al., 2010a).

The areas north and south of the putative tear in the slab show distinctly different deformation styles: The NE region is characterized by broad uplift of the entire region: Serrania de Falcon, widespread faulting, and diffuse seismicity. The southwestern region is characterized by two narrow Laramide-style basement faulted uplifts: the Merida Andes and the Perija range.

The shallow slab geometry proposed by Van der Hilst and Mann (1994) suggest that the CAR slab extends continuously from the Caribbean basin beneath the entire Maracaibo Block southward to the Bucaramanga nest of seismicity in Colombia, and east of the Merida Andes into the Barinas Apure Basin. Our interpretation of a torn slab does not agree with this scenario.

In central Venezuela the oblique convergence of the Caribbean plate is accommodated mainly by dextral shear along the San Sebastian – El Pilar Fault system and by shortening observed in the Cordillera de la Costa thrust belt, as the arc terrenes thrust onto the South American margin (Jacome et al., 2003). In that sense we interpret the second Ps conversion, observed in our CCP stacked images under Cordillera de la Costa, as a detachment surface beneath the arc terranes of the CAR thrust onto the South America margin, during transpression.
3.4.2. Mesozoic Rifting and the Guayana Shield

In Venezuela, the opening of the Atlantic Ocean during the Jurassic produced a series of NE-SW oriented grabens, such as the Espino Graben and the Apure Mantecal Graben (Burke, 1977). These grabens have played an important role in the evolution of some of the sedimentary basins within the country (Yoris and Ostos, 1997; Salazar, 2006). Our results show a thin lithosphere beneath these grabens (Figures 3-11 and 3-14). The thin lithosphere extends northeastward to the offshore Cariaco basin, delimiting the western edge of the Venezuelan Craton (Guayana Shield).

As mentioned above, the Guayana Shield is composed mainly of Archean and Proterozoic rocks, which can be further divided into four major provinces (Yoris and Ostos, 1997, Schmitz et al., 2002). Our estimates of the lithospheric thickness within the Shield indicate that changes in thickness correlate well with the boundaries of the provinces (Figure 3-14). The Rayleigh wave tomography reveals high velocities extending north of the Orinoco River (Figure 3-5) to the Maturin Basin, and beneath the Serrania del Interior. We speculate that this could represent evidence of the continuation of the igneous-metamorphic basement of the Guayana Shield up to the Maturin Basin, as suggested by previous studies (e.g., Yoris and Ostos, 1997).
3.4.3. Atlantic Subduction and Lithospheric Drips

Since the 1970’s, it has been suggested that the geodynamic processes onshore eastern Venezuela are more complicated than a simple transition from transform motion to subduction (e.g. Molnar and Sykes, 1969), defined by a clean tear through the whole lithosphere (Clark et al., 2008a; Govers and Wortel, 2005) at the Paria Peninsula.

Igneous rocks of ~5My intrude the crust of the Araya – Paria Peninsula (Santamaria and Schubert, 1974; McMannon, 2001). Urbani (1989) found more recent signs of magmatism near the El Pilar fault system, south of the Carupano rhyolites. He associated the geothermal systems in eastern Venezuela with shallow intrusive bodies. This geologic evidence has been interpreted as an indicator of the presence of the edge of the subducting slab beneath SA.

The deeper part (>120 km) of the high velocity anomaly associated with Atlantic subduction (Figure 3-15) is continuous across the plate bounding El Pilar Fault zone, and continues to the southern edge of the Serrania del Interior. This region lacks intermediate depth seismicity, which stops at the Paria cluster just north of the El Pilar fault system. Bezada et al. (2010a) argued that the strike slip system marks the southern end of the Atlantic oceanic lithosphere and that the deep high velocity anomalies south of the faults indicate continental South American mantle lithosphere being convectively removed by the Atlantic subduction.
Figure 3-15 3D Visualization of the lithospheric structure from the Rayleigh wave tomography results. Orange dots and red squares denote seismicity and seismic stations in the area, respectively. (a) 3D perspective view of the P-wave finite-frequency tomography model of Bezada et al. (2010). 3D perspective views of the S-wave Rayleigh wave tomography model: (b) view from the east; (c) view from the west; (d) view from the north. GS: Guayana Shield. ES: Espino Graben. MB: Maturin Basin. CC: Cordillera de la Costa. SI: Serrania del Interior. T&T: Trinidad and Tobago. Figure generated with ParaView.
We also speculated that the second high velocity beneath the Maturin Basin is a drip of continental South America (Figures 3-5 and 21), although the dynamic processes involved are unclear.

On Figure 3-15d we observe a thick lithosphere in the east (~130 km), a relatively thin lithosphere in most of central Venezuela (~80 km), and a slight increase under the northwestern edge of the Maracaibo block (90 – 100 km, Figure 3-14). This large change in lithospheric thickness suggests to us that the convective removal of the base of the South American continental lithosphere during subduction of the Atlantic slab is a continous process, which has occurred progressively, from west to east starting ~55 Ma ago. If this is the case, the thin lithosphere observed in western and central Venezuela is the remnant of this process.

3.5. Conclusions

We used Ps and Sp receiver functions and Rayleigh wave tomography to investigate the crust and upper-mantle velocity structure beneath Venezuela and southern Caribbean. The LAB shows significant variations, which are well correlated with different parts of the plate boundary interior. The stable Guayana Shield has the thickest lithosphere (~110-120 km) with some variation in depth beneath its Proterozoic and Archean blocks. The lithosphere beneath the pull-apart Cariaco basin off the coast and the Jurassic Espino Graben, on the other hand, is only ~50-60 km thick. Most of central and western Venezuela is underlain by a relatively thin
lithosphere (~80-90 km). We image the top 200 km of a deep (>600 km) anomaly in northeastern Venezuela identified in a previous body wave tomography study. This anomaly, adjacent to the Antiles subduction zone, extends south of the plate bounding El Pilar fault beneath the Serrania del Interior to the northern edge of the Maturin Basin. The northern part of the anomaly is the southernmost edge of the subducting Atlantic slab. The Ps receiver function images reveal a southeastward dipping reflector, which we interpret as the Moho of the Caribbean plate beneath the northwestern part of the Maracaibo Block.
High-resolution Adjoint Tomography of the crust beneath Eastern Venezuela using Empirical Greens function data from Ambient Noise

A broad zone of diffuse deformation and faulting defines the Caribbean and South American plate boundary in Eastern Venezuela. To obtain a high-resolution 3D shear velocity model for the crust, we used adjoint tomography to iteratively minimize the misfit between synthetic seismograms and the empirical Green's functions derived from ambient noise cross-correlation data between station pairs using 42 stations of the Venezuelan seismic network and from the BOLIVAR project. This dataset provides good path coverage of eastern Venezuela with higher frequency seismic signals (10 – 50 s) than analysis of ballistic Rayleigh waves to resolve small-scale structures. The resulting 3D velocity model reveals low velocities distributed along the axis of the Espino Graben, indicating that the crustal
composition of the graben differs from the rest of the Eastern Venezuelan Basin. Moho depth varies within the study region from 20 km beneath the Cariaco Basin to 50 km beneath the Maturin Basin, in good agreement with results from active seismic studies.

4.1. Introduction

The boundary between the Caribbean (CAR) and South America (SA) plates in eastern Venezuela is a broad zone of deformation (~300 km) resulting from oblique collision of SA-CAR starting at ~55 Myr in northwestern SA and progressing eastward as the Americas have swept past the CAR. Fold and thrust belts and associated foreland basins have developed on the SA mainland along the right lateral plate boundary as it has lengthened along the northern SA margin (e.g. Meschede and Frisch, 1998; Audemard et al., 2005). Modern plate motion is accommodated by dextral shear along the San Sebastian –El Pilar Fault system at a rate of 2 cm/yr east (Perez et al., 2001; Weber et al., 2001), accompanied by NW-SE shortening in the Cordillera de la Costa and Serrania del Interior Thrust belts (Perez et al 2001; Audemard et al., 2005; Jacome et al., 2003) (Figure 4-1). The plate boundary exhibits both transpressional and transtensional features, such as the pull-apart Cariaco Basin (Escalona et al., 2011).

The Eastern Venezuelan basin (figure 4-1; Yoris and Ostos, 1997) is divided into the eastern Guarico Basin south of the Cordillera de la Costa, the Espino Graben south of the offshore Cariaco basin, and the Maturin Basin south of the Serrania del
Interior. Prior to formation of the fold and thrust belts the geodynamic evolution of the Basin can be divided into three major episodes (Parnaud et al., 1995; Di Croce, 1995): 1) A pre-rift phase in the Paleozoic characterized by shallow marine sediment deposition. 2) A Jurassic and earliest Cretaceous rifting and drifting phase during opening of the central Atlantic characterized by the formation of SW-NE grabens, e.g. the Espino Graben. 3) Cretaceous-Paleogene formation of the Proto-Caribbean passive margin. In central and eastern Venezuela the final phase of basin formation resulted from the aforementioned west to east migrating oblique collision of CAR and SA during the Neogene and Quaternary. This collision telescoped elements of the Proto-Caribbean passive margin and the Leeward Antilles arc terranes onto northern South America, creating the fold belts in the north, the two associated foredeeps and a southern platform zone (Ysaccis, 1997; Hung, 1997; Ostos et al., 2005). At the end of the Miocene the oblique convergence between the CAR and SA ended, and the plate boundary evolved to its present configuration of a nearly purely strike slip margin (Pindell et al., 2005; Ysaccis, 1997).

The Guayana Shield, lying south of the Eastern Venezuelan Basin (Gonzalez de Juana, 1980) is composed mainly of Archean and Proterozoic rocks, divided into four provinces; the Archean Imataca province and the three Proterozoic provinces, Pastora, Cuchivero and Roraima (Yoris and Ostos, 1997).
Figure 4-1 (a) Caribbean tectonics, with GPS vectors in the southern Lesser Antilles Arc and on San Andreas (SA) Island in the western Caribbean. Red box indicates the study region shown in (b). (b) Broadband seismic stations in eastern Venezuela used to estimate the Empirical Green’s function between station pairs. Red triangles indicate permanent stations of the National Seismic Network of Venezuela. Blue triangles represent temporal deployments under the BOLIVAR project. Major active faults are indicated dashed black lines. MB: Maracaibo Block, BF: Bocono Fault, OAF: Oca-Ancon Fault; SMF: Santa Marta–Bucaramanga Fault; SS-EP: San Sebastian –El Pilar Fault; BAB: Barinas-Apure Basin; CC: Cordillera de la Costa; SI: Serrania del Interior; OR: Orinoco River; LA: Leeward Antilles. GS: Guayana Shield, CB: Cariaco Basin.
The Eastern Venezuelan Basin is known globally as a major petroleum-producing province, and has therefore been the site of many stratigraphic studies (e.g. Gonzales de Juana, 1980; Parnaud et al., 1995; Di Croce, 1995; Passalaqua et al., 1995; Jacome et al., 2008). Only recently have 2D active source seismic experiments provided information about the velocity crustal structure of the crust and uppermost mantle (Schmitz et al., 2002; Schmitz et al., 2005; Levander et al., 2006; Clark et al., 2008ab; Guédez, 2008; Magnani et al., 2009; Bezada et al., 2010). Miller et al. (2009) and Masy et al. (2013) used ballistic teleseismic Rayleigh wave tomography to determine the shear velocity structure of this region, however, they did not constrain the upper to mid-crustal structures, as the periods they measured were 20s and greater.

Ballistic teleseismic Rayleigh wave tomography provides fundamental information about the Earth’s interior and its three-dimensional structure. However these measurements have some limitations resulting from the distribution of sources and receivers, which can leave some areas unsampled. Moreover, information about the middle and upper crust is often lacking because signals with periods shorter than 20s can be difficult to analyze due to severe multi-pathing. In contrast cross-correlation functions between local stations computed from ambient noise data eliminate the multi-pathing effects experienced on long paths from the earthquake source, as well as provide higher frequency signals (10 – 50s) to better resolve crustal features.
Here we present a different approach to ambient noise tomography (Chen et al., 2013 submitted). Empirical Green's functions (EGF) extracted from CCFs can be used as input for adjoint waveform tomography based on the spectral element method to refine an existing 3D velocity model (Chen et al, 2103 submitted; Tromp et al., 2010; Komatitsch et al., 2004). For this method it is not necessary to measure surface wave dispersion curves and create phase or group velocity maps required by the traditional ambient noise tomography, as it directly accounts for the interactions between the EGFs and the shear velocity structure. We apply this method to the broadband data acquired in eastern Venezuela as part of the BOLIVAR (Broadband Onshore-Offshore Lithospheric Investigation of Venezuela and the Antilles Arc Region) and GEODINOS (Geodinamica Reciente del Limite Norte de la Placa Sudamericana) projects (Levander et al., 2006). We characterize the three-dimensional shear structure of the crust of the southernmost Caribbean, i.e., the Leeward Antilles arc, and eastern Venezuela as far south as the Guayana Shield, providing results consistent with previous work as well as providing new constraints on the three dimensional shear velocity of the middle and upper crust.

4.2. Data and Methods

4.2.1. Empirical Green’s function Calculation

We collected a total of 677 days of continuous waveform data recorded at 42 stations in eastern Venezuela; 14 stations of the Venezuelan National Network operated by the Fundacion Venezolana de Investigaciones Sismologicas (FUNVISIS),
28 stations from the temporary BOLIVAR broadband network, which were in operation from December 2003 to June 2005. We cut the continuous records into daily segments and removed the instrument response from the raw records to obtain displacements records.

We first use these daily records to calculate the CCFs for a total of 457 stations pairs using a running time-window method. If a gap was present within the running window of one of the seismograms, the CCF of that time window was not computed. For each time window any linear trend and the data mean were removed from the seismograms. In order to enhance the ambient noise signals, we applied spectral whitening by setting amplitudes in the frequency band between 0.02 - 1 Hz to 1, and 0 elsewhere while the phased remained unchanged. Finally, daily cross-correlations functions were summed to increase the signal to noise ratio. The resulting cross-correlation for a station pair is a two-side time function with both causal (positive time) and acausal (negative time) signals (Bensen et al., 2007). If the sources of ambient noise are evenly distributed around two stations, the CCF is expected to be symmetric. However, we observe considerable asymmetry, indicating differences in the noise source process and the distance to the noise source. The South American coast lying north and east of the study area, and the mountain belts in the north (Figure 4-1) are likely the cause of the observed asymmetry. In order to approximate the study region to an area with spatially uncorrelated noise sources, we compressed the two-side signal into a one-sided signal by averaging the causal and acausal signals of the CCF (Figure 4-2) (Bensen et al., 2007; Tromp et al., 2010).
Figure 4-2: Example of Surface waves obtained from ambient noise cross-correlations sorted by interstation distance. Moveout with increasing interstation distance can be clearly observed. (a) Seismograms with a pass band filter between 10 – 20 s (b) 15-30 s (c) 20 – 40 s (d) 25 – 50 s
For a given station pair A and B, the relationship between the CCF, $C(t)$, the Empirical Green's Function (EGF), $\Gamma(t)$, and the real Green's function, $G(t)$, can be represented as (Yao et al., 2006):

$$\frac{dC_{AB}(t)}{dt} = -\Gamma_{AB}(t) + \Gamma_{BA}(-t) = -G_{AB}(t) + G_{BA}(-t)$$

(1)

where $\Gamma_{AB}(t)$ is the causal part and $\Gamma_{BA}(-t)$ is the acausal part of the EGF at A for a fictitious point source at B. This empirical relationship is an approximation of the exact Green's function that depends on the frequency amplitude correction and errors due to the uneven source distribution of the ambient noise (Roux et al., 2005; Yao et al., 2006; Yao et al., 2009).

4.2.2. 3D Initial velocity model and forward modeling

In order to apply the waveform inversion successfully, it is important to have an initial velocity model that is sufficiently accurate, which will allow the misfit function to converge to the global minimum. If the misfit between synthetic waveforms and the observed data is more than half-cycle a local minimum may be encountered leading to incorrect results (Virieux and Operto, 2009). Masy et al. (2013, submitted) generated a regional 3D shear wave velocity model based on Rayleigh wave tomography. This model shows rapid variations in shear velocity between the different tectonic provinces to ~200 km depth. However, the minimum period used was 20 s, which provides little constraint on the upper to mid-crustal velocities. We took this model as the starting Vs model, and used the empirical
Vp/Vs relationships of Brocher et al. (2010) to estimate P-wave velocity structure and density.

Synthetic Green’s functions (SGF) were computed with the spectral element code SPECFEM3D (Komatitsch et al., 2004; www.geodynamics.org/cig/software). For each master station we assumed a source located at 1 km depth. The source was an approximation of a delta source-time function (a Gaussian function with a half duration of 2.5 s).

The simulation domain is a box 7° (EW) by 5° (NS) and 120 km in the vertical direction, from 61°W to 68°W longitude and 6.5° to 11.5° latitude (Figure 4-1). The mesh was created with a grid spacing of ~ 2.5 km in each coordinate. The mesh included topography and bathymetry.

The stations are numbered from 1 to N (N is total number of stations, 42 for this study). Forward simulations are done from 1 to N-1 stations for each master station. Stations used previously as a master station are excluded from the correlation station list as SGF between station pairs to avoid redundancy.

4.2.3. Data selection

Following Chen et al. (2013), we first filter the EGFs and SGFs to four frequency passbands, 10-20 s, 15-30 s, 20-40 s and 25-50 s. For each band, we normalize the EGFs with the maximum amplitude of their corresponding SGFs. We then use the automatic windowing code, FLEXWIN, developed by Maggi et al. (2009) to select windows for misfit measurement. Only windows with certain misfit and
below are chosen as the data of the inversion. We also keep the distance between two stations to be at least twice as large as the average wavelength of each passband (Figure 4-3). For the initial velocity model (m00) a total of 510 windows is selected. As the velocity model is updated more windows are added to the inversion, the final model (m09) has a total of 980 windows (Figure 4-4).

4.2.4. Misfit measurements and 3D sensitivity kernels

According to finite-frequency theory, the sensitivity to velocity perturbations around the ray path is defined by a ‘banana-doughnut’ kernel (Dahlen et al., 2000; Hung et al., 2000), and the sum of all the station-pair kernels determines the array sensitivity to variations in the velocity model.

For each selected window, we made frequency dependent traveltime misfit measurements using a multi-taper method (Zhou et al., 2004; Chen et al., 2013 submitted). The adjoint sources for the selected windows are constructed from the time derivative of the time reversed SGF’s weighted by the misfit measurement of that particular window (Figure 4-3). We use SPECFEM3D for the adjoint simulation by sending back the adjoint sources at all the secondary stations. The interaction of the adjoint and the forward wavefield from the primary generates misfit kernels that are sensitive to the shear velocity, bulk sound speed and density structure along the great circles between the primary station and the secondary stations. These kernels are summed and smoothed with a Gaussian function of a radius of 40 km in the horizontal plane and 10 km in the vertical plane, then they are
preconditioned with an approximate Hessian term to get the final sensitivity kernels (Chen et al., 2013, submitted).

Figure 4-3 Example of misfit comparison between the EGF’s data (black line), and the synthetic data generated with the initial model (blue line) and the final model (solid red line). Black dashed boxes are time windows selected by FLEXWIN for final model (m09). (a) Seismograms with a pass band filter between 10 – 20 s (b) 15-30 s (c) 20 – 40 s (d) 25 – 50 s
4.2.5. Model update

We use the initial model described above as a starting point (m00). The discrepancies in phase between the SGF and the EGF suggest that the model can be improved. We assume that the updated model is represented as the starting model (or previous model) plus a perturbation derived from the ‘banana-doughnut’ kernels. To minimize the misfit of the objection function $\chi$, we use a conjugate gradient method (Chen et al., 2013, submitted):

$$
\chi = \frac{1}{N_c} \sum_c \chi_c
$$

$$
\chi_c = \sum_{pc} \int \frac{1}{N_{pc}} \left( \Delta \tau_{pc}(\omega) \right)^2 \sigma_{pc}(\omega) d\omega
$$

where $\chi$ is the total misfit function for all passbands ($N_c = 1-4$, one for each passband), and $\chi_c$ is the misfit contribution for each passband ($N_{pc} = \ldots$ with $p$ the index for the measurements in each pass band), $\Delta \tau_{pc}$ is the traveltime misfit at frequency $\omega$ for measurement $p$ in passband $c$, and $\sigma_{pc}$ is the uncertainty in the measurement of $\Delta \tau_{pc}$.

The search direction for the next minimum is given by the kernels modified by the dot product between the adjoint wavefield and backward forward wavefield (Hessian term) (Chen et al., 2013, submitted). For each iteration, the step length was determined using a line search, in which the minimum of the misfit function is chosen to update the previous model (Figure 4-5). The relationship between the data and the model is nonlinear, in the sense that the inversion needs to be iterated...
several times to converge to the minimum of the misfit function (Virieux and Operto, 2009).

![Figure 4-4 Histogram summarizing the statistics of the timeshift distribution for the initial model m00 (gray bars) and the final model m09 (blue bars) for all the selected windows](image)

This minimum misfit was reached after 9 (m09) iterations (figure 4-5). There is a ‘jump’ in the misfit evolution curve from model 3 to model 4, due to the inclusion of more EGF’s being added to the inversion process as model misfit decreases, and hence more data pass the window selection criteria in FLEXWIN.
Figure 4-5 (a) Line search example using travel time misfits to choose next model, in this case from the 6th iteration, i.e. using m05 as the starting model and selecting the best misfit for m06. (b) Traveltime misfit evolution curve. The overall traveltime misfit decreases with increasing number of iterations until iteration m09. The increase in misfit between the models 3 and 4 results from the addition of windows to the inversion. The blue circles from m00 to m03 represent the same set of windows used for the inversion. Pink circles from m04 to m06 indicate the same set of windows used during the inversion. Red circles indicate the same set of windows used from m07 to m09.

Figure 4-4 shows the histogram of travel time misfits for the initial (m00) and final (m09) models. The timeshift distribution for the initial model peaks at 5 seconds,
indicating that the initial predicted SGF’s are overall faster than the EGF’s. The histogram from the final model peaks at 0 seconds, indicating that the final model predicts the data well.

Figure 4-6 Maps of the Hessian preconditioner at 10 km, 20 km, 30 km, 40 km, 50 km and 60 km. The Hessian term is a good proxy of ray coverage.
The EGF’s in this study are primarily sensitive to the shear velocity structure of the crust to about 40 km depth, with considerably less sensitivity at 50 km based on maps through the Hessian preconditioner (Figure 4-6; Chen et al., 2013, submitted).

**4.3. Results and discussion**

In Figure 4-7, we show five depth slices at 10, 20, 30, 40, and 50 km depth of the model m09, in which velocity perturbations are shown together with the average shear velocity. The average velocity is computed from all the mesh grids including those shaded areas, which are not resolved by our data and may have extreme values that affect the resulting average velocities. For example the velocity perturbations at 30 km depth (Figure 4-7c) computed from the box average is centered rather to the negative side. However, the average velocities are only used as references in computing the velocity perturbations, and the range of perturbations should be barely affected by the reference. The top three depths (Figure 4-7a-4-7c) are expected to be inside the crust except some small offshore areas in the north; their velocity perturbations fall roughly between approximately -20% and 15%. The deeper two depths (Figure 4-7d and 4-7e) include both crustal and mantle rocks, thus velocity perturbations appear to be slightly larger, between approximately -30% and 20%.

At the depth of 10 km (Figure 4-7a), the most substantial lower velocity anomaly is located inside the Maturin Basin, where the sediments are nearly 12 km thick
(Schmitz et al., 2005). Velocity inside the Guarico Basin, however, is close to normal. In general, shear wavespeeds inside the Guayana Shield are higher than those in the coastal regions. Among the three provinces inside the shield, the Archean Imataca province stands out as a distinct high velocity body. P-wave velocity estimated by Schmitz et al. (2002) from active source data also shows the same feature. Niu et al. (2007) found that the Imataca province is also featured by a distinct low Vp/Vs ratio inside the shield. All these observations suggest that rocks inside the Archean terrane may be compositional different from the rest of the shield.

At 20 km deep (Figure 4-7b), the distinct low velocity anomalies align with the front of thrust belts. The SW-NE trending Espino Graben, on the other hand, starts to stand out as a distinct low velocity anomaly. This feature reaches to the highest level at 30 km (Figure 4-7c), where the velocity of graben is approximately 20% lower than that of the Maturin Basin in the east and Guarico Basin in the west. During the opening of the Atlantic Ocean, a series of grabens and half-grabens were developed along the coast of SA (Burke, 1977). The Espino Graben is one of these extensional structures in eastern Venezuela. This Graben is filled with continental sediments (red layers), volcanic material, and shallow water shales (Feo-Codecido et al., 1984; Parnaud et al., 1995; Ostos et al., 2005), while the two adjacent basins (Guarico and Maturin Basins) are filled with passive margin sediments with oceanic origins (Yoris and Ostos, 1997). This difference might be the cause of the shear velocity contrast observed here.
Figure 4-7 Lateral variations of shear wave velocity at depths of 10, 20, 30, 40 and 50 km are shown in (a), (b), (c), (d) and (e) respectively. Grey areas represent regions with no resolution. SS-EP: San Sebastian-El Pilar Fault system; CC: Cordillera de la Costa; SI: Serrania del Interior; EG: Espino Graben; CB: Cariaco Basin.
Figure 4-8 Absolute shear velocities at depths of 10, 20, 30, 40 and 50 km are shown in (a), (b), (c), (d) and (e) respectively. Grey areas represent regions with no resolution. SS-EP: San Sebastian-El Pilar Fault system; CC: Cordillera de la Costa; SI: Serrania del Interior; EG: Espino Graben; CB: Cariaco Basin.
Figure 4-9 Upper left: Map showing the location of 7 profiles (A)-(G). (a) Profile A. (b) Profile B. (c) Profile C. (d) Profile D. (e) Profile E. (f) Profile F. (g) Profile G. Grey areas represent regions with no resolution. Solid line is the Moho derived from this study. Dashed lines are the Moho derived from wide-angle experiments (Magnani et al., 2009; Bezada et al., 2010; Clark et al., 2008; Schmitz et al., 2002) SS-EP: San Sebastian-El Pilar Fault system. CB: Cariaco Basin.
As mentioned above, the large velocity variations observed at 40 and 50 km deep (Figures 4-7d and 4-7e) are mainly caused by velocity contrasts between crustal and mantle rocks. Moho depth is ~35-40 km inside the Espino Graben and Guayana, and varies from 40 to 45 km beneath most part of the Guarico and Maturin basins. In the northeastern most of the Maturin Basin the crust is close to 50 km thick (Niu et al., 2007). Thus velocity inside the Maturin Basin is substantial lower than that in the other regions.

In Figure 4-9, we also show seven depth profiles along different tectonic regions. Profile A (Figure 4-9a) crosses the Cordillera de la Costa and the Guarico Basin up to the northern edge of the Guayana Shield. Moho depth varies from ~45 km beneath the Cordillera de la Costa, to ~38 km beneath the Guarico Basin, to 42 km beneath the Guayana Shield. These results are consistent with the results obtained by Magnani et al. (2009) and Schmitz et al. (2002).

Profile B (Figure 4-9b) shows the transition in crustal thickness from 20 km beneath the Cariaco Basin, to ~45 km beneath the Maturin Basin to ~40 km in the Guayana Shield. Bezada et al. (2008) and Schmitz et al. (2002) presented similar results. Very slow velocity (2.6 km/s) indicative of sediments reaches to nearly 10 km deep beneath the Cariaco Basin and ~5 km at the western edge of the Maturin Basin (Figure 4-8).

Profile C (Figure 4-9c) runs across the Serrania del Interior and the Maturin Basin, where there is a drastic transition in Moho depth, changing from ~30 km beneath the Serrania del Interior to ~50 km Moho beneath the Maturin Basin. This
trend is consistent with results of Clark et al. (2008) derived active seismic data. Sediments at the depocenter inside the Maturin Basin reach to nearly 18 km based on a characteristic shear velocity of ~2.6 km/s (Figure 4-9c and 4-9g). Flexural modeling for eastern Venezuela (Jacome et al., 2003) shows that the thrusting of the Serrania del Interior is not sufficient to explain the thick sediment column observed in the Maturin Basin. A sublithospheric pull associated with the Atlantic subducting slab is necessary to produce the extra subsidence. Jacome et al. (2008) showed that the same scenario does not apply for the Guarico Basin; the flexural response of the Cordillera de la Costa thrust sheet is enough to produce the observed stratigraphy and geometry of the basin.

Profile D (Figure 4-9d) is along the San Sebastian - El Pilar Fault system. The observed variations on Moho depth are consistent with results from previous studies under the BOLIVAR project (Bezada et al., 2008; Clark et al., 2008; Magnani et al., 2009). The observed Moho beneath the Cordillera de la Costa is shallower than it is under the Sierra del Interior, which extends to almost 50 km.

Profile E (Figure 4-9e) is located at the northern edge of the Guayana Shield, and is featured by a higher velocity as compared to other the profiles. The upper crust appears to be rather homogenous (Figure 4-8). There is a distinct high velocity anomaly in the lower crust beneath the Archean Imataca Province (Figure 4-8).

Profile F (Figure 4-9f) extends roughly along the axis of the Espino Graben, the overall velocity of the profile is lower than that of the others. Profile G runs across the Guarico Basin and the Maturin Basin, which show a clear difference
between the two basins, which may indicate that the two foreland basins are in different evolution stages.

4.4. Conclusions

We used empirical Green’s functions derived from ambient noise cross correlation data as input for waveform adjoint tomography to investigate the crust velocity structure beneath eastern Venezuela. The 3D refined crustal shear velocity model shows lateral variations that are well correlated with the geologic structures within the Eastern Venezuela Basin and the Guayana Shield. Low velocities along the axis of the Espino Graben may indicate that the sedimentary filling of the basin is different from the rest of the Eastern Venezuela Basin. Crustal velocity beneath the Archean Imataca province is distinctly higher than that of Proterozoic provinces. The observed variations in Moho depth are consistent with those estimated from previous wide-angle profiles and flexural modeling.
Chapter 5

Conclusions

The BOLIVAR and GEODINOS projects were designed to understand the interactions and evolution of the CAR-SA plate boundary. Throughout the development of these projects, a variety of seismic experiments have allowed determining the crustal structure under Venezuela and the southern CAR. This study focused on determining the lithospheric structure and deformation styles under the CAR-SA plate boundary and Venezuela by using different types of seismic data: shear wave splitting, Ps and Sp receiver functions, teleseismic surface waves and ambient noise.

Shear wave-splitting measurements in northwestern Venezuela revealed three distinct areas with different polarization directions and magnitudes of delay times. Each of these areas can be characterized with a unique deformation mechanism: (1) stations located north of the Oca-Ancon fault, along the CAR-SA plate boundary, have the largest splitting times (> 2 s), and a E-W polarization
direction, which can be explained by a strong mantle flow travelling eastward along the CAR-SA plate boundary; (2) stations located within the Barinas Apure Basin, in stable SA, have weak seismic anisotropy parallel to the plate motion of SA, with an origin to be likely in the asthenosphere; (3) stations along the Merida Andes have intermediate splitting times (~1-1.5 s) with a NE-SW fast direction parallel to the mountain range, suggesting coherent deformation of the crust and the lithosphere resulting from the transpressive stresses related to the uplift of the cordillera.

By integrating these results with previous measurements of seismic anisotropy based on the first phase of the BOLIVAR data (Figure 5-1, Growdon et al., 2009), we found strong seismic anisotropy along the entire the CAR-SA plate boundary with a fast polarization direction roughly parallel to the E-W strike of the boundary. Growdon et al. (2009) attributed this strong anisotropy in eastern Venezuela to a mantle flow, which originates behind the Atlantic plate and flows around its southern edge, and emerges beneath the continental SA due to the eastward rollback of the Atlantic slab (Figure 5-2).

In general, the magnitude of seismic anisotropy estimated from the splitting time between the fast and slow arrivals decreases rapidly towards the south, and the fast polarization directions also vary slightly, which are parallel to local structures in general, suggesting a vertically coherent deformation between the crust and upper mantle (Figure 5-1). Thus we speculate that the lithospheric mantle plays an important role in the formation of the mountain ranges within the plate boundary and Venezuela.
Figure 5-1 Measured splitting parameters from the BOLIVAR Project. Blue triangles are from this study and red triangles are from Growdon et al. (2009). The orientation of the black solid lines indicates the orientation of the fast direction. The length of the line indicates the magnitude of the splitting time. Red arrows are GPS measurements relative to SA.
Using receiver-function data we measured the depth of the lithosphere-asthenosphere boundary (LAB). The LAB depth shows significant variations across the study region, but shows a good correlation with different tectonic blocks. The stable Guayana Shield has a thick lithosphere (~110-120 km) with some little variations in depth (~10 km). The lithosphere beneath the Cariaco basin off the coast and the Jurassic Espino Graben, on the other hand, is only ~50-60 km thick. In eastern Venezuela, beneath the Maturin Basin, the lithosphere is thick (~130 km), while most of the central and western part of the country is underlain by a relatively thin lithosphere (~80-90 km). This change in lithospheric thickness might indicate that convective removal of the base of the SA lithosphere associated with the subduction of the Atlantic slab is a continuous process, ongoing as the subduction zone of the Atlantic slab progressed along the plate boundary from west to east starting ~55 Ma ago (Figure 5-2).

Teleseismic Rayleigh wave tomography and adjoint tomography using ambient noise data provided three-dimensional shear velocity structure beneath eastern Venezuela. Lateral variations of Vs correlate well with the geologic provinces. A high velocity anomaly is shown at depth range between 60 and 200 km beneath the Serrania del Interior. It runs across the strike-slip fault system, and nearly reaches to the northern edge of the Maturin Basin. Based on the change in lithosphere thickness in the area, we speculate that the high velocity anomaly is likely a chunk of the continental lithosphere of the SA being sinking together with the subducting Atlantic slab due to convective removal (Figure 5-2).
Finally, the Ps receiver-function images reveal a southeastward dipping reflector in northwestern Venezuela, located at 15 to 20 km deep beneath the Moho of the SA. By combining with the P-wave tomography results, we interpret the reflector as the Moho of the Caribbean plate subducting beneath the northwestern part of the Maracaibo Block.

Figure 5-2 Schematic representations of the geodynamic processes occurring along the CAR-SA plate boundary. The Atlantic tears off SA in a highly localized zone. In northern SA, CAR tears beneath the MB. Both subducting slabs penetrate the transition zone (Bezada et al., 2010a). Seismic anisotropy along the northern edge of the MB indicates the presence of a trench parallel mantle flow that passes around the NW corner of the subducting CAR plate. Similarly, the same interpretation could be made for the subduction Atlantic plate. Red stars indicate possible plate tears. SS-EP: San Sebastian –El Pilar fault system. MB: Maracaibo Block. SMF: Santa Marta Fault. CAR: Caribbean Plate.

The main phase of the BOLIVAR broadband array was deployed mainly in eastern Venezuela and the seafloor of the southeastern CAR, leaving the western Venezuela largely uncovered by instruments. In the second phase of the project, we
installed 7 stations in the western Venezuela for 12 months. The limited data suggest that the CAR under northwestern SA has separated into a subducting segment on the south and an under-thrusting segment on the north. Whether this is a slab tear, separation occurs at an unrecognized transform fault, or is a contortion in the slab is unclear. Deploying a denser seismic array across northern Colombia and western Venezuela (Serrania de Falcon, Maracaibo Block, Merida Andes, Perija and Santa Marta Ranges, and the Barinas- Apure Basin) would be very important to examine: (1) the slab geometry of the subducting CAR plate beneath northern Colombia and its role involved in the formation of the Maracaibo Block and (2) the uplift of the surrounding mountain ranges (Merida Andes, Perija Range and Santa Marta Massif). Seismic studies that incorporate high-resolution data sets from both western and eastern Venezuela will be ideal to understand the overall evolution of the southern CAR plate boundary. Some suggestions for further work in this region include:

1. Detailed seismic anisotropy measurements using teleseismic core phases (SKS/SKKS) in western Venezuela. The data can be used to constrain the location of the possible tear beneath the Maracaibo Block. Because it is difficult to determine the depth of the anisotropy using SKS phases, the use of intermediate depth local earthquakes and Ps and Sp receiver functions may help to constrain the depth distribution of seismic anisotropy in the area.

2. Surface wave tomography and ambient noise tomography of western Venezuela will allow to image 3D shear velocity structure, resulting in an integrated picture of the entire southern CAR plate and Venezuela.
3. High-resolution finite frequency body wave tomography. A detailed tomographic model will provide a better understanding of the relationship between the lithospheric mantle and the crustal structures.

4. New velocity models resulting from tomography can be used as reference model to reprocess and incorporate new receiver functions. This will allow imaging the top of the CAR plate and its variations along strike. The improved receiver function images would provide a better constraint on the topography of major discontinuities such as the Moho discontinuity and the LAB.
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Appendix A

Hit count maps of the Ps receiver function CCP images at 10 km, 20 km, 30 km, 50 km, 70 km, 90 km, 110 km, 130 km, 150 km. These maps are a representation of the quality and resolution of the results.
Appendix B

Hit count map of the Sp receiver function CCP images at 30 km, 50 km, 70 km, 100 km, 120 km, 1500 km. These maps are a representation of the quality and resolution of the results. Images look more diffuse than the Ps maps due to the lower frequency content of the Sp receiver functions.
Appendix C

Histogram summarizing the statistics of the timeshift distribution for the initial model m00 (gray bars) and the final model m09 (blue bars) for the selected windows for each band pass during the ambient noise waveform tomography.
Appendix D

Ray coverage distribution of each band pass for ambient noise tomography

bp10-20

bp15-30

bp20-40

bp25-50
Appendix E

Vertical cross-sections of the Hessian preconditioner at 63W, 64.5W and 66W longitude, 7.5N, 9N and 10.5 N latitude. The Hessian term is a good proxy of ray coverage. Thus is a good indicator of the resolution.