RESEARCH ARTICLE

10.1002/2017JB014688

Key Points:

- The magmatic signature of the subducting Nazca plate is mapped at midcrustal depths under the central Cordillera
- Asthenospheric magmas might have breached the Caribbean flat slab through a preexisting vertical tear
- Fluid migration on major faults might also explain observed low-velocity anomalies under the Lower Magdalena Basin

Supporting Information:

- Supporting Information S1
- Data Set S1
- Data Set S2

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Citation:

Poveda, E., Julià, J., Schimmel, M., & Perez-Garcia, N. (2018). Upper and middle crustal velocity structure of the Colombian Andes from ambient noise tomography: Investigating subduction-related magmatism in the overriding plate. *Journal of Geophysical Research: Solid Earth*, 123. <https://doi.org/10.1002/2017JB014688>

Received 9 JUL 2017

Accepted 13 JAN 2018

Accepted article online 23 JAN 2018

Upper and Middle Crustal Velocity Structure of the Colombian Andes From Ambient Noise Tomography: Investigating Subduction-Related Magmatism in the Overriding Plate

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Abstract New maps of *S* velocity variation for the upper and middle crust making up the northwestern most corner of South America have been developed from cross correlation of ambient seismic noise at 52 broadband stations in the region. Over 1,300 empirical Green's functions, reconstructing the Rayleigh wave portion of the seismic wavefield, were obtained after time and frequency-domain normalization of the ambient noise recordings and stacking of 48 months of normalized data. Interstation phase and group velocity curves were then measured in the 6–38 s period range and tomographically inverted to produce maps of phase and group velocity variation in a 0.5° × 0.5° grid. Velocity-depth profiles were developed for each node after simultaneously inverting phase and group velocity curves and combined to produce 3-D maps of *S* velocity variation for the region. The *S* velocity models reveal a ~7 km thick sedimentary cover in the Caribbean region, the Magdalena Valley, and the Cordillera Oriental, as well as crustal thicknesses in the Pacific and Caribbean region under ~35 km, consistent with previous studies. They also display zones of slow velocity at 25–35 km depth under regions of both active and inactive volcanism, suggesting the presence of melts that carry the signature of segmented subduction into the overriding plate. A low-velocity zone in the same depth range is imaged under the Lower Magdalena Basin in the Caribbean region, which may represent either sublithospheric melts ponding at midcrustal levels after breaching through a fractured Caribbean flat slab or fluid migration through major faults within the Caribbean crust.

1. Introduction

The recent tectonic evolution of northwestern South America has been shaped by the complex interactions of up to three different lithospheric plates: South America, Caribbean, and Nazca. It is accepted that this corner of the continent maintains a record of Grenvillian basement, originated during the amalgamation and disassembly of western Pangaea (Spikings et al., 2015). The subsequent collision of the Caribbean igneous province created new crust, which was added to the Pacific margin during Early and Late Cretaceous times (Escalona & Mann, 2011), and emplaced the Panama arc in western Colombia during Middle Miocene times (Cediel et al., 2003; Restrepo & Toussaint, 1988; Taboada et al., 2000). These terrains were later reworked and altered by convergence-related processes. The end result is the present configuration of the Colombian Andes (north of the Ecuadorian border) into three main ranges—the Eastern Cordillera (EC), the Central Cordillera (CC), and the Western Cordillera (WC)—and two intervening valleys—the Middle Magdalena Valley (MMV) and the Cauca-Patía Valley—separating the ranges (Figure 1). The origin and composition of these three main ranges are essentially different, each of them deriving from distinct tectonic processes driven by plate convergence (Cediel et al., 2003; Taboada et al., 2000). The composition of the South American plate is presently being altered by subduction-related volcanism, which reaches as far north as ~6°N latitude. Volcanism is active along the CC and inactive along the EC and is offset at 5°N latitude by ~270 km in the EW direction, following the pattern delineated by intermediate, subduction-related seismicity (e.g., Pennington, 1981; Vargas & Mann, 2013). The NW corner of South America thus constitutes a unique tectonic setting to investigate how subduction-related processes deform and alter the composition of the overriding plate.

Passive-source seismic studies in the region have focused mainly on the relative geometry of the subducting Nazca and Caribbean plates under South America (Chiarabba et al., 2016; Pennington, 1981; Syracuse et al., 2016; van der Hilst & Mann, 1994; Vargas & Mann, 2013). An early study by Pennington (1981), for instance,

Figure 1. Tectonic and simplified geologic setting of the northern Andes. Geologic provinces from Gómez et al. (2007) and major faults from Vellozo et al. (2012). Plate motions relative to South America from Trenkamp et al. (2002). Gray thick line, Envelops of North Andean Block (NAB); area with vertical lines, Panamá arc; red triangles, Active volcanoes; blue triangles, Inactive volcanoes. The different provinces are the following: SMM, Santa Marta Massif; Mab, Maracaibo block; SMBF, Santa Marta-Bucaramanga fault; SM, Santander Massif; MA, Merida Andes; SSJB, Sinú and San Jacinto Basin; CRB, César-Ranchería Basin; PB, Plató Basin; LMB, Lower Magdalena Basin; SJB, San Jorge Basin; SLR, San Lucas Range; AB, Antioqueño Batholith; BR, Baudo Range; MMV, Middle Magdalena Valley; RS, Romeral suture; Panamá arc; AB, Antioqueño batholith; CC, central Cordillera; EC, Eastern Cordillera; LIB, Llanos Basin; LFS, Llanos faults system; WC, Western Cordillera; CV, Cauca Valley; IB, Ibagué batholith; SMV, Upper Magdalena Valley; QM, Quetame Massif; GM, Garzón Massif; CV, Cauca Valley; MR, Macarena ranges; TB, Túmaco Basin; PSS, Panama South America suture.

used patterns of seismicity and focal mechanisms to define two separate segments of subducted lithosphere—Cauca and Bucaramanga—that were associated to the Nazca and Caribbean plates, respectively. The segments were found to be dipping at different angles, thus defining a slab window within the South American mantle that explained the offset of intermediate-depth seismicity and surface volcanism. The detailed geometry of the slab window was later refined through seismic tomography, suggesting either superposition of the Nazca and Caribbean plates (van der Hilst & Mann, 1994), tearing of the Caribbean slab (Vargas & Mann, 2013), tearing of the Nazca slab (Chiarabba et al., 2016), or even multiple tears segmenting the subducting plates (Corredor, 2003; Syracuse et al., 2016). Fewer studies have focused on the structure of the overriding South American plate. Refraction surveys, for instance, were performed in southern Colombia

as part of the Nariño Project (Aldrich et al., 1973; Meissnar et al., 1976; Meyer et al., 1976; Mooney et al., 1979; Ocola et al., 1975), producing a crustal cross section for the region; and a 1-D velocity model for the Colombian crust was developed by Ojeda and Havskov (2001), seeking to improve accuracy in location of crustal seismicity. On a more regional scale, crustal thickness was surveyed by Poveda et al. (2015), based on receiver functions analysis and a published compilation of crustal thickness observations; and images of lateral *P* wave and *S* wave velocity variation between 12.5 and 155 km depth were developed by Syracuse et al. (2016), although with no tectonic interpretation at crustal levels. In spite of these efforts, therefore, a regional-scale investigation of lateral seismic velocity variation of the overriding South American plate is still missing.

Surface-wave tomography has been successfully utilized to image the continental lithosphere at several scales (e.g., Feng et al., 2004; Ritsema et al., 2004). Tomographic images from naturally occurring surface waves, however, rarely provide information at periods shorter than 20 s, which are required to probe into the upper and middle crust (Levshin & Ritzwoller, 2001). Fortunately, the modal representation of ambient seismic noise for a diffuse wavefield with a homogeneous source distribution has demonstrated that surface wave imaging at local to regional scales can be performed with ambient seismic noise (Sabra et al., 2005; Snieder, 2004).

Indeed, an ever increasing number of ambient seismic noise studies have been produced during the last decade mapping different parts of the globe (e.g., Abdetetal et al., 2015; Bensen et al., 2007; Boué et al., 2014; Haned et al., 2016; Lin et al., 2007; Mottaghi et al., 2013; Saygin & Kennett, 2010; Shapiro et al., 2005; Stehly et al., 2006; Villaseñor et al., 2007; Ward et al., 2013; Yang et al., 2007; Yao et al., 2006). These studies have conclusively demonstrated that empirical Green's function recovery from average cross correlations of ambient noise between pairs of seismic stations is capable of providing tight constraints on upper and middle crustal structure.

In this study, we use information derived from cross correlations of ambient seismic noise between pairs of permanent seismic stations in NW South America to reconstruct the surface wave portion of the empirical Green's functions and develop surface wave tomographic images at periods sampling upper and midcrustal levels. More specifically, we utilize seismic stations deployed by the Colombian Geological Survey between 2012 and 2015 and around the Andean region to reconstruct the empirical Green's functions for all possible pairs of stations and develop high-quality, Rayleigh-wave group and phase velocity tomographic images at periods between 6 and 38 s. Our tomographic study includes more than ~780 interstation raypaths that sample the region, derived from ~1,300 empirical Green's functions imaging the upper and middle crust at an unprecedented resolution in the region. The group and phase velocity maps are further inverted for isotropic *S* wave velocity structure, building a detailed 3-D *S* velocity model for the middle and upper crust under NW South America. The resulting images show an excellent correlation with surface geology at shallow depths and reveal low-velocity patterns at midcrustal depths (25–35 km) that correlate with active and inactive surface volcanism in the central portion of the EC and the southern flank of the CC, respectively. These low-velocity anomalies are interpreted as magmas derived from the mantle wedge above the Nazca and Caribbean plates. More strikingly, a similar low-velocity anomaly is also observed under the Lower Magdalena Basin (LMB), where flat subduction of the Caribbean plate has prevented the occurrence of surface volcanism. We argue that this low-velocity region represents either the signature of fluid expulsion near crustal faults or upwelling magmas from the underlying asthenosphere, which would have breached the flat-subducting Caribbean plate through a preexisting fracture and pond at midcrustal levels within the overriding plate.

2. Geology and Crustal Structure

2.1. Geology and Tectonic Setting

NW South America is broadly divided into the northern Andes, the Caribbean region, and the Eastern Llanos. The northern Andes comprise the Andean regions of Ecuador, Colombia, and Venezuela, which are structured into three main ranges (EC, CC, and WC) and two intervening valleys (MMV and Cauca-Patia Valley). Precambrian and Paleozoic rocks form the basement of the EC and the CC, immediately west of the Guyana Shield. The EC is covered by sequences of Cretaceous sedimentary rocks, while the CC is dotted by instances of Meso-Cenozoic batholiths related to subduction of the Nazca plate. Cenozoic andesitic

volcanism dominates the southern portion of the CC (Figure 1). The WC, on the other hand, consists of oceanic rocks that accreted during Cretaceous and early Cenozoic times and are covered by sedimentary rocks. The division between the WC and the CC is the Cauca Valley and the Romeral Fault System, which separate the accreted oceanic part from the provinces of continental affinity next to the Guyana shield (Figure 1). The separation between the central and the WCs is marked by the presence of the MMV, which is covered by sequences of Cretaceous, Paleogene, Neogene, and Quaternary sedimentary rocks from the EC and the CC. In the northernmost corner of the northern Andes outcrops the Santa Marta Massif (SMM), which includes Jurassic volcanics and Paleozoic metamorphic rocks, along with volcano-sedimentary Cretaceous materials (Cardona et al., 2011).

The Caribbean region can be geologically described as an Oligocene-to-Recent forearc Basin—the LMB—overlying the shallowly dipping Caribbean slab (Bernal-Olaya, Mann, & Escalona, 2015; Mora-Bohórquez et al., 2017). The LMB is in turn composed of several subduction-related basins, such as the Plató, San Jorge, and Sinú-San Jacinto Basins, and is bounded to the west by Sinú-San Jacinto fold-and-thrust belt. The Plató and San Jorge Basins consist of Paleogene sediments overlying the Precambrian and Paleozoic metamorphic basement rocks, while the Sinú-San Jacinto fold-and-thrust belt is regarded as an accretionary wedge that includes rocks of Cretaceous age as well as Tertiary and Quaternary lacustrine sediments. The sedimentary prism making up the Sinú-San Jacinto Basin extends offshore along the margin of the northern South American Plate, with thicknesses of up to 10 km (Bernal-Olaya, Mann, & Escalona, 2015; Flinch, 2003; Lara et al., 2013; Montes et al., 2010; Toto & Kellogg, 1992). The Eastern Llanos represents an extensive fore-land basin associated to the Andean orogen that overlays the Guyana craton (e.g., Farris et al., 2011).

The northern Andes and the Caribbean region behave as a coherent tectonic unit known as the North Andean Block (NAB), which can thus be regarded as a collage of several terrains that accreted against the Guyana Shield at different stages through Earth's history (Cediel et al., 2003; Restrepo & Toussaint, 1988). The latest addition to the NAB is the Panamá arc, which includes the Baudó Range and represents an oceanic, basic, and ultramafic magmatic arc covered by oceanic sediments (Case et al., 1971; Cediel et al., 2003; Restrepo & Toussaint, 1988; Taboada et al., 2000). Flueh et al. (1981) reported sedimentary sequences with thicknesses between 7.5 and 10 km under the Panama arc. The Panama arc is thus seen as an exotic terrain within the NAB (Duque-Caro, 1990; Taboada et al., 2000).

The present geology of northwestern South America is partly the result of a complex interplay between the South American (overriding) plate and the (subducting) Nazca and Caribbean plates. An initial model of this interplay was proposed by Pennington (1981). Through careful selection of hypocentral locations of earthquakes and focal mechanisms in the region, the author identified two distinct segments of subducted lithosphere beneath the Colombian Andes: the Bucaramanga segment, related to subduction of the Caribbean plate, and the Cauca segment, related to subduction of the Nazca plate. These two segments would have resulted from the collision of the Panamá isthmus against South America and buoyancy of partially subducted bathymetric features, implicitly implying the existence of a slab tear at depth separating the postulated segments. The Bucaramanga segment would be subducting in the N109°E direction and dipping at 20°–25°, while the Cauca segment would be subducting in the N35°E direction with a 35° dip.

Although several studies have been conducted to investigate the precise contact geometry between the Nazca and Caribbean plates, this issue still remains a debate. Tomographic images developed by van der Hilst and Mann (1994) imaged oblique convergence of the Caribbean and Nazca slabs (including the Panama isthmus segment), showing an overlap between latitudes 5.2°N and 7°N with both plates being simultaneously overridden by the South American plate. This interpretation was adopted in Cortés and Angelier (2005) and Taboada et al. (2000). Vargas and Mann (2013), on the other hand, return to the slab tear model of Pennington (1981). From additional earthquake locations accumulated during the intervening years, coda-Q tomography and regional magnetic and gravity data, they propose the existence of a lithospheric, ~240 km wide, east-west trending slab tear at 5°N latitude—the Caldas tear—that separates the shallowly dipping Panama arc indenter from the more steeply dipping normal oceanic crust of the Nazca plate. Similar to Pennington (1981), they further propose that the tear would have resulted from the collision of the Panamá arc and buoyancy of a subducted oceanic ridge. Tomographic imaging by Chiarabba et al. (2016) places the tear entirely within the Nazca plate, pushing the boundary of this plate north and indicating an EW trending slab tear separating a steeply subducting segment to the south from a flat subducting

segment to the north. Finally, Syracuse et al. (2016) presented 3-D models of *P* velocity and *S* velocity variation for the mantle and hypocentral relocations based on the joint inversion of local body wave arrivals, surface wave dispersion measurements, and gravity data. They imaged up to three separate slab fragments, separated by two slab windows: one at about 5°N, coinciding with the termination of active volcanism, and one at 7.5°N, marking the location of the Bucaramanga seismic nest. The region at ~5.5°N (northern termination of volcanism) is interpreted as a nonoverlapping slab tear of the Nazca plate, where intermediate-depth seismicity is shifted ~200 km toward the east, beneath EC. A similar model had been previously proposed by Corredor (2003) from a three-dimensional model of CMT (Centroid-Moment-Tensor solutions and other seismicity in northwestern South America.

In spite of the many interpretations from tomographic imaging, all models seem to agree that the Nazca plate is subducting with an average angle of 35° in a nearly eastward direction, although with high variability along the subduction front and, therefore, presenting segmentation (Gutscher et al., 1999). The Caribbean plate, on the other hand, would be subducting at shallower dip angle <10° in a WNW-ESE direction (Bernal-Olaya, Sanchez et al., 2015; Mora-Bohórquez et al., 2017), with no apparent signs of segmentation. All models also agree that the shift in intermediate-depth seismicity and surface volcanism would be the result of deep versus shallow subduction along the Colombian trench.

2.2. Crustal Structure

The average crustal structure of the NW Andean region can be obtained from continental-scale models. Chulick et al. (2013), for instance, presented contour maps and statistical analyses of the seismic structure of the crust and upper mantle, based on a large seismic compilation of multiple data sets. They reported an average crustal thickness of 38 km for South America, with a maximum crustal thickness estimate of 60 km for Colombia. Feng et al. (2004, 2007) and Lloyd et al. (2010), on the other hand, produced tomographic images from surface wave dispersion with ~200 km resolution in the continent; although ray coverage is low for northern Colombia, crustal thickness is reported to be in the 40–50 km range. These values are consistent with thickness up to 45 km for the Colombian Andes and 33 km for the LMB reported from analysis of satellite gravity measurements from the Gravity field and steady state Ocean Circulation Explorer mission by van der Meijde et al. (2013).

The most complete survey of crustal structure for the NW Andean region is the crustal thickness map presented in Poveda et al. (2015) from receiver function analysis, which has been updated here following the procedures described in their study. We include 11 new stations (Figure 2) that improve the sampling of the Caribbean region and the WC and CC. A close inspection of the figure reveals that the crust in the southern Pacific Coast is about 22 km thick, increasing to about 30 km under the Panama arc. The WC is characterized by thicknesses between 30 km south of 4°N and 40 km north of this latitude, similar to those observed under the Cauca Valley. Crustal thickness increases to 52–58 km beneath the northern CC and to almost 60 km beneath some of the volcanoes of the Southern Cordilleran system; in the EC, crustal thickness is between 55 and 60 km. In the MMV, separating the CC from the EC, crustal thickness is around 45 km. In the Caribbean region, the crust is ~30 km thick under the Sinú-San Jacinto Basin and ~35 km thick under LMB, the Plató Basin, and the San Jorge Basin. These variations may be due to the shallowly dipping Caribbean slab (Bernal-Olaya, Sanchez et al., 2015). Beneath the SMM, the updated crustal thickness map displays a crustal thickness of ~40 km, decreasing to ~30 km under the Guajira Peninsula.

Detailed velocity structure was developed from active-source profiling in southern Colombia as part of Nariño Project (Aldrich et al., 1973; Meyer et al., 1976; Ocola et al., 1975), where the crust between 1°N and 4°N latitude, and 82°W and 76°W longitude was mapped with seismic refraction lines (Meissnar et al., 1976). In the offshore portion, results revealed *P* wave velocities of ~3.2 km/s for the upper crust (top 5 km) and 6.8 km/s for the lower crust (down to 15 km); onshore, *P* wave velocities ranged between 6.0 and 6.8 km/s for the upper and lower crusts and were found to be ~8.0 km/s in the underlying mantle. Crustal thicknesses ranged between 30 and 40 km in the region, with upper and lower crustal thicknesses of 11–19 km and 15–25 km, respectively. Bowland and Rosencratz (1988) investigated the upper crustal structure of the Caribbean Sea region from multichannel seismic refraction and found that oceanic crust is between 10 and 15 km thick, with *P* velocities around 4.5 km/s in the upper crust, 7.0 km/s in the lower crust, and 8.2 km/s in the uppermost mantle.

Figure 2. Update of the interpolated crustal thickness map for NW South America from Poveda et al. (2015).

From relocalization of hypocenters, Ojeda and Havskov (2001) developed a 1-D crustal velocity model that improved location accuracy of local (crustal) seismicity, and Ojeda and Ottemoller (2002) developed a local QLg tomography that defined several crustal regions according to attenuation values. Ojeda and Havskov (2001) developed a regional model with three layers for the crust (32 km), with P wave velocities of 4.8, 6.6, and 7.0 km/s with thicknesses of 4, 21, and 7.5 km, respectively, and an upper mantle velocity between 8.0 and 8.1 km/s. A constant V_p/V_s ratio of 1.78 was determined for the entire data set. Ojeda and Ottemoller (2002), on the other hand, subdivided the central portion of the study area into seven zones, according to attenuation characteristics of QLg. The principal regions with high attenuation correspond to active volcanic areas and the northeastern cordillera, which is characterized by young sedimentary rocks and lacustrine sediments, and high water content in the Bogotá altiplano.

Finally, as mentioned above, images of lateral P wave and S wave velocity variation for depths ranging between 12.5 and 155 km were produced in the recent tomographic study of Syracuse et al. (2016). As the

Figure 3. Broadband seismic stations used in this work. The color coded represents recording time window duration in each station.

crustal slices in that study are directly comparable to our velocity models, we postpone the discussion of those images until section 6, where they will be compared to the results developed here from ambient noise tomography.

3. Data and Data Processing

3.1. The Data Set

The data set used in this study consists of the continuous recordings of the 50 seismic stations of the *Red Sismológica Nacional de Colombia* (RSNC), which is operated by the Colombian Geological Survey, and three nearby Global Seismograph Network stations in Ecuador, Venezuela, and Panamá. In total, it makes a combined network of 53 permanent broadband stations that recorded continuously from April 2012 to December 2015. The equipment is quite heterogeneous and consists of a combination of Streckeisen STS-2, Güralp CMG-3 T, Güralp CMG3-ESP, Trillium 120p, Reftek 151-120A, and Reftek 152B-120 broadband sensors (flat response in velocity down to 120 s) feeding high-gain digitizers (Quanterra Q330, Güralp

Table 1
Seismic Station Parameters (Geographical Location and Recording Time Window)

Station	Latitude (deg)	Longitude (deg)	Recording time used (dd/mm/yy)	
ANIL	4.491	−75.403	04/05/2012	31/10/2015
ARGC	9.858	−74.246	26/04/2013	31/12/2015
BBAC	2.021	−77.247	27/04/2013	31/12/2015
BCIP	9.166	−79.837	01/01/2013	31/12/2015
BRR	7.107	−73.712	19/04/2012	03/12/2015
CAP2	8.646	−77.359	19/04/2012	03/07/2015
CBOC	5.864	−76.012	08/08/2013	31/12/2015
CHI	4.629	−73.318	19/04/2012	31/12/2015
COD	9.935	−73.444	19/04/2012	17/06/2014
CRJC	11.021	−72.881	27/11/2014	31/12/2015
FLO2	1.583	−75.653	19/04/2012	31/12/2015
GARC	2.187	−75.493	24/02/2014	31/12/2015
GCUF	1.226	−77.345	30/08/2012	31/10/2015
GUY2	5.223	−75.371	27/04/2013	31/12/2015
HEL	6.191	−75.529	19/04/2012	31/12/2015
LCBC	8.857	−76.368	19/11/2013	31/12/2015
MACC	2.145	−73.848	28/05/2013	31/12/2015
MAP	4.004	−81.606	19/04/2012	14/11/2015
MARA	2.822	−75.954	19/04/2012	17/08/2013
MON	8.778	−75.665	19/04/2012	31/12/2015
NOR	5.564	−74.869	06/04/2013	31/12/2015
NUQC	5.709	−77.277	18/08/2013	06/11/2013
OCA	8.238	−73.319	02/06/2012	31/12/2015
ORTC	3.909	−75.426	20/06/2013	31/12/2015
OTAV	0.237	−78.451	01/01/2013	31/12/2015
PAL	4.906	−76.283	19/04/2012	31/12/2015
PCON	2.327	−76.397	19/04/2012	03/06/2014
PGA1	3.747	−71.571	03/04/2014	31/12/2015
PGA2	3.901	−71.561	04/04/2014	16/06/2014
PIZC	4.965	−77.361	10/06/2014	31/12/2015
POP2	2.541	−76.676	19/04/2012	31/12/2015
PRA	3.714	−74.886	19/04/2012	31/12/2015
PRV	13.376	−81.364	31/05/2012	31/12/2014
PTA	7.147	−77.809	01/09/2012	31/12/2015
PTB	6.539	−74.456	19/04/2012	31/12/2015
PTGC	4.199	−72.134	09/10/2013	31/12/2015
PTLC	−0.171	−74.797	02/02/2013	31/12/2015
PUAC	0.549	−76.571	29/07/2014	19/05/2015
ROSC	4.844	−74.321	19/04/2012	31/10/2015
RREF	4.901	−75.347	19/04/2012	23/06/2015
RUS	5.893	−73.083	19/04/2012	31/12/2015
SDV	8.883	−70.634	01/01/2013	31/12/2015
SJC	9.897	−75.181	02/06/2012	31/12/2015
SMAR	11.163	−74.224	21/09/2012	31/12/2015
SML	8.801	−74.071	02/11/2012	31/12/2015
SOTA	2.135	−76.609	19/04/2012	23/05/2014
SPBC	5.652	−74.072	13/07/2013	31/12/2015
TAM	6.435	−71.791	28/07/2012	31/12/2015
TUM	1.824	−78.727	19/04/2012	31/12/2015
URE	7.752	−75.533	24/06/2012	31/12/2015
URI	11.702	−71.993	19/04/2012	31/12/2015
YOT	3.983	−76.345	19/04/2012	31/12/2015
ZAR	7.492	−74.858	19/04/2012	31/12/2015

DM24, and Taurus) sampling continuously at 100 samples per second. We considered all accessible vertical component data within the recording time window from all the permanent stations. Figure 3 displays the location of the broadband stations, color-coded by recording time window duration, while Table 1 lists station coordinates and recording time windows. Realize that, because the network was gradually built, the amount of data available among the stations may vary. The final data set offers good spatial and azimuthal coverage for the Colombian Andes and most of the Caribbean region of Colombia, with an average interstation spacing of approximately 400 km, and minimum and maximum interstation distances of 20 km and 1300 km, respectively.

3.2. Empirical Green's Functions

To retrieve the empirical Green's functions by cross correlation, we followed the approach described in Shapiro and Campillo (2004), Sabra et al. (2005), and Bensen et al. (2007), with modifications. Data processing can be divided into three stages: (i) preprocessing, (ii) cross correlation, and (iii) stacking. Only vertical-component data were considered, which implies that only the Rayleigh wave portion of the empirical Green's function was reconstructed by cross correlation. The aim of the preprocessing stage is to decrease the influence of earthquake signals and instrumental irregularities from the time series prior to cross correlation. It starts with cutting the continuous recordings into nonoverlapping, 1 day long segments, which are demeaned, detrended, and tapered with a 5% cosine window. Instrumental responses are then deconvolved from the daily time series to be band-pass filtered between 0.01 and 0.2 Hz and decimated to 1 sample per second. Time- and frequency-domain normalization is then applied to remove the signature of seismicity within the frequency band of interest. Temporal normalization is implemented through 1 bit normalization, preserving the phase and neglecting amplitude variations within the time series (Bensen et al., 2007; Larose, 2004); frequency-domain normalization or "spectral whitening" normalizes spectral amplitudes to unit value without altering the phase, effectively broadening the frequency band of the ambient noise recordings (e.g., Bensen et al., 2007). After time-frequency normalization is applied to all 1 day long segments, cross correlations are computed in the frequency domain for all station pairs for frequencies between 0.01 and 0.2 Hz.

Stacking of the 1 day long ambient-noise cross correlations is achieved through the time-frequency phase-weighted stack (tf-PWS) of Schimmel et al. (2011). The tf-PWS makes use of the S-transform (Stockwell et al., 1996) to obtain a time-frequency representation for each cross correlogram, so that the stacking is controlled with frequency-dependent resolution while maintaining a close relationship between the time averaging and the Fourier spectrum (Stockwell et al., 1996).

The time-frequency-dependent phase coherence attenuates incoherent noise, because it allows for the determination of weak amplitude signals that are more phase coherent than background noise. Signals obtained with this strategy have a signal-to-noise ratio (SNR) greater than those obtained with conventional methods (e.g., the classical correlation and stack method; Bensen et al., 2007; Lin, et al. 2008; Villaseñor et al., 2007; Yang et al., 2007).

Figure 4. Extracted Rayleigh waves from ambient noise for the all stations with RMS > 20. All traces are band passed between 10 and 50 s and are normalized. The dashed lines show an average velocity between 2.1 and 3.2 km/s.

A total of 1,286 cross-correlation stacks were constructed from the assembled data set. A record section of cross correlations is displayed in Figure 4, where only those cross correlations with SNR > 20 are included for displaying purposes. The empirical Green's functions clearly emerge from the ambient seismic noise for distances ranging between about 50 km and over 1,100 km. Moreover, traveltimes correspond to propagation velocities between ~2.1 and ~3.2 km/s, which are consistent with short-period, fundamental-mode, Rayleigh wave propagation.

Selection of stable empirical Green's function was achieved through analysis of the coherence functional (Schimmel et al., 2011), which measures the similarity of two time series. First, waveform similarity was calculated using the zero-lag cross correlation between a reference trace (correlogram with maximum stack size) and a subset of correlograms that correspond to a fixed number of days, which are selected randomly with 30 days increments to the maximum stack length. Then, a plot of similarity versus number of days was built, so that signal emergence and minimum number of days could be assessed for each Green's function for several interstation raypaths and azimuthal directions. Figures 5 and 6 illustrate the process through select examples corresponding to two different interstation raypaths. The first raypath (Figure 5e) travels through the central portion of the Colombian Andes in the N-S direction and has an interstation distance of 423 km, while the second raypath (Figure 6e) samples the eastern portion of the Caribbean region and the CC and has an interstation distance of 723 km. The figures reveal that, for distances greater than 700 km, about 450 days (Figures 6a and 6b) are necessary for convergence of the empirical Green's function; while for shorter raypaths of ~400 km, about 250 days, seem to be enough (Figures 5a and 5b). This relatively large number of days required to achieve convergence contrasts with other studies in the stable portion of the continent, where just a few weeks were enough to achieve convergence (e.g., Dias et al., 2015). We think that the observed rate of convergence is probably due to the intense

seismic activity in the country, especially around the Bucaramanga nest (Ojeda & Havskov, 2001; Prieto et al., 2012; Zarifi et al., 2007), which could not be fully mitigated through the standard time-frequency normalization process. This might be especially true for the URI and HEL interstation path (Figure 6a), which intersects the Bucaramanga nest. Note how stacking of just 100 days is not enough to recover both the causal and acausal portions of the empirical Green's function, which is eventually recovered when adding additional 1 day correlograms to the stack.

4. Phase and Group Velocity Tomography

4.1. Measuring Group and Phase Velocity

The empirical Green's functions extracted in the previous section are dominated by surface waves. Group velocity can thus be measured by applying the time-frequency analysis (FTAN) of Levshin et al. (1972) and Levshin and Ritzwoller (2001), which is based on the application of a series of narrow band-pass filters to the signal. The FTAN method also implements a phase-matched filter that cleans dispersion curves from multipathed signals (Herrin & Goforth, 1977), so that the whole procedure can be easily automated (Levshin & Ritzwoller, 2001). Phase velocity can be measured from the empirical Green's function as (Lin et al., 2008)

$$\varphi(t) = kr + \omega t + \lambda + 2\pi N + \frac{\pi}{2} - \frac{\pi}{4}, \quad (1)$$

where k is the wave number, r the interstation distance, ω is the instantaneous frequency, t is the observed group-delay time, λ is the source phase or initial phase, and $2\pi N$ represents the phase ambiguity term (with $N = 0, \pm 1, \pm 2, \dots$). The phase corrections at the end of equation (1) are specific to empirical Green's functions

Figure 5. Analysis of the signal stability and convergence for HEL-POP2; (a) Empirical Green's function (EGF) convergence for different amount stack data, selected from random form; (b) similarity value in function of random days number that show the convergence evolution. In this case the paths converge in ~250 days; (c) dispersion curves for Rayleigh group and phase velocities, each color is associated to EGF in Figure 5a. Uncertainties from 3 month stack is shown; (d) example of frequency-time analysis (FTAN) diagram showing Rayleigh wave group velocity with phase match filter; and (e) path between HEL and POP2 stations (423 km).

from ambient noise cross correlations. The $\pi/2$ represents the phase shift from the negative time derivative, and the $\pi/4$ arises from interference of the nonstationary terms in a homogeneous noise source distribution (Lin et al., 2008; Snieder, 2004). The phase ambiguity ($2\pi N$) is assessed by comparing the long-period phase velocity measurements to a reference curve, which here is based on the preliminary reference Earth model global Earth model (Dziewonski & Anderson, 1981). Sometimes, FTAN fails to measure the dispersion curve due to abrupt jumps or spectral gaps. To overcome this shortcoming, the period window is reduced sequentially in bands from 5–100 s down to 5–30 s, in steps of 10 s.

Figure 6. Similar to Figure 5. Analysis of the signal stability and convergence for HEL-URI.

To measure dispersion velocity, we first reject empirical Green's functions with $\text{SNR} < 7$ at each period. The tf-PWS technique of Schimmel et al. (2011) is quite successful in attenuating incoherent noise in the stacked cross correlations, so the SNR is improved dramatically after increasing the number of available daily cross correlations. Following Bensen et al. (2007), we also eliminate dispersion measurements on cross correlations with interstation distances less than three wavelengths. To estimate uncertainties, we applied the repeatability of measurements through seasonal variability. First, we selected 12 overlapping, 3 month long time series for each interstation path for a time window between 2012 and 2015. The first 3 month stack will include data from January through March for each year between 2012 and 2015, the second 3 month stack will include data from February through April for each year between 2012 and 2015, and so on. After the dispersion curve for group and phase velocity is measured for all the 3 month long cross correlations, the standard deviation is calculated if at least half of the twelve 3 month stacks have a SNR greater than 7. If this criterion is not satisfied, the standard deviation is taken as the average value for all the raypaths.

Figure 7. Statistics for the estimated group and phase speed maps to the observations taken after all stages of data rejection; (a) velocity uncertainty for group and phase versus period; (b) number of paths versus period; (c) traveltime uncertainty versus period used for tomography inversion; and (d) average of path length versus period.

Figures 5 and 6 also illustrate the measuring of dispersion velocities through two select empirical Green's functions. The figures reveal that the stability of the dispersion curves is slightly dependent on interstation distance, especially for periods longer than 20 s, where microseismic energy decays abruptly (Bromirski, 2009). For the raypath through the central part of the Colombian Andes (Figure 5), the dispersion curve reveals normal dispersion between 7 and 22 s with an average group velocity of ~ 3.2 km/s. The standard deviations, calculated from seasonal variability, are around 0.05 km/s and 0.03 km/s for group and phase velocity, respectively (Figure 5c). For the raypath sampling the eastern portion of the LMB and the CC (Figure 6), we observe a gradual increase in both phase and group velocity between 7 and 40 s, with standard deviations of ~ 0.07 km/s for group velocity and ~ 0.02 km/s for phase velocity (Figure 6c). The final results of FTAN processing are shown in Figures 5d and 6d, where the dispersion curves for phase and group velocity are clean of spurious and multipathed signals. Also note how for the longer raypath (Figure 6c), the dispersion curve obtained from the empirical Green's function is not stable at long periods when stacking only 100 days, in accordance with the lack of convergence of the empirical Green's function noted in the previous section.

The average group and phase velocity uncertainties, number of measurements that passed the quality control, average arrival time uncertainty, and average raypath length for each period are displayed in Figure 7. Figure 7a shows that for periods between 6 and 30 s the average group velocity uncertainties are less than 0.1 km/s, dropping significantly for intermediate periods to ~ 0.05 km/s. For phase velocities, the standard deviation is much smaller, ranging from ~ 0.06 to ~ 0.1 km/s, with a minimum of 0.04 km/s at 14–22 s. Our uncertainties are similar to those reported in independent studies with similar data sets (e.g., Dias et al., 2015; Lin et al., 2007; Mottaghi et al., 2013). The greatest uncertainties are for periods > 30 s, which we believe are due to the decrease in amplitude of the microseismic noise spectrum at those periods that make it difficult to recover the empirical Green's function. Traveltime uncertainties as a function of period are displayed in Figure 7c, which will be used to weight the data sets in the tomographic inversion (section 4.2). Figures 7b and 7d show the number of raypaths and the average path length in each period for the selected observations, respectively, demonstrating that the optimal coverage is in the intermediate period range of 12–30 s.

4.2. Tomographic Inversion

The interstation dispersion curves developed in the previous section are next inverted tomographically to produce phase and group velocity maps of NW South America at different periods. The Fast Marching

Figure 8. L curve for 22 s group (black) and phase (red) velocity tomography; (a) RMS misfit with model roughness, selected values are 10 and 2 for group and phase, respectively; (b) RMS misfit and model variance, selected values are 9 and 3 for group and phase, respectively; (c) RMS variation with number of iterations, note that with ~4 iterations the convergence is reached.

Surface Tomography package of Rawlinson and Sambridge (2005) is utilized for the inversion. The formulation of the forward problem is based on the solution of the eikonal equation by finite differentiation of the phase or group delays, thus mapping the evolution in time of the first-arriving wavefronts across the study area. This eikonal solver, known as the Fast Marching Method, is described in detail elsewhere (Sethian, 1996; Sethian & Popovici, 1999), and its migration to the surface wave tomography problem can be found in Rawlinson and Sambridge (2005). This methodology is advantageous when the stations coverage is sparse, providing stable, robust solutions for wave propagation in heterogeneous media (Rawlinson & Sambridge, 2004). The inversion procedure follows an iterative, nonlinear scheme known as the subspace method (Kennett et al., 1988), which acts by projecting the quadratic approximation of the objective function of the inversion into a multidimensional subspace. Therefore, at each iteration, the optimization functional $S(m)$ is projected into a subspace of dimension smaller than the original parameter space in order to reduce the computational effort in the local linearization. More details of the subspace method can be found in Rawlinson and Sambridge (2004). Because of the size of the study area and the amount of data, we found that it is not necessary to project in greater than one dimension. The optimization functional is defined as

$$S(m) = (d - g(m))^T C_d^{-1} (d - g(m)) + \epsilon (m - m_0)^T C_m^{-1} (m - m_0) + \eta m^T D^T D m \quad (2)$$

where m_0 is the reference model, C_m is the a priori model covariance matrix, C_d is the data covariance matrix, and D is the second derivative smoothing operator. The parameters $\epsilon (\geq 0)$ and $\eta (\geq 0)$ are the damping and smoothing regularization parameters, respectively. The smoothing parameter controls the trade-off between data misfit and model smoothness, while the damping parameter controls the trade-off between data misfit and proximity of the inverted model to a reference model m_0 (Rawlinson & Sambridge, 2005).

A discretized grid consisting of $0.5^\circ \times 0.5^\circ$ cells was defined throughout the study area, with a constant initial velocity taken from the average observed group or phase velocity at each period. Uncertainties from 3 month stacks (Figure 7c) were then included in the objective function (equation (2)) to define the covariance matrices. Appropriate regularization parameters, ϵ and η , in the objective function were determined through L curve analysis (Aster et al., 2012). We explored several combinations of regularization parameters by varying the damping and smoothing from 0 to 100 in steps of 0.01, 0.1, 1.0, and 10 for all periods. Figure 8 displays sample trade-off curves for both group and phase velocity tomographic inversions at 22 s, demonstrating values of $\epsilon = 9$ and $\eta = 10$ for group velocities and $\epsilon = 3$ and $\eta = 2$ for phase velocities. Parameters for tomographic inversions at other periods are similar, ranging between 0.3 and 10 for smoothness (ϵ) and between 0.3 and 9.0 for damping (η). Furthermore, the tomographic inversions converge after six iterations for the selected regularization values (Figure 8c). Figure 8b also shows that model variance decreases to an average of $\sim 0.015 \text{ km}^2/\text{s}^2$ (group) and $\sim 0.01 \text{ km}^2/\text{s}^2$ (phase) with our selected parameters or, equivalently, to average standard deviations of $\sim 0.12 \text{ km/s}$ and $\sim 0.1 \text{ km/s}$, respectively. These values are within the confidence bounds of the observed dispersion curves obtained from the 3 month stacks and are sufficiently high to indicate that the data are not overfitted.

The robustness of the tomographic inversions depends mostly on the azimuthal coverage of the seismic raypaths. Figure 7b already showed that path density is best for periods between 12 and 30 s, with over 800 measurements for both group and phase velocities. However, the azimuthal raypath distribution is not uniform in the study region, being highest for the central part of Colombia and gradually decreasing

Figure 9. Checkerboard resolution tests and recovery for group velocities for 7, 14, 22, 30, 34, and 38 s (large maps). The black polygon encompasses the region used in modeling crustal V_p , and the number in the upper left shows mean group velocity obtained from the real data set and number of paths. Small maps show input synthetic checkerboard test for each period.

toward the edges of the study area. To assess the bias due to poor raypath coverage, we used standard checkerboard tests at each period. The tests were performed using input models of alternating fast and slow velocity defined through a grid of nodes with B-spline cubic interpolation with the same grid size, same raypath geometry, and same regularization values as for the real model. Each input model had

velocities varying ± 0.5 km/s around a background velocity for each dispersion map. Additionally, Gaussian noise was added to the synthetic group and phase arrival times for all station pairs. A series of checkerboard tests mimicking the raypath coverage of all inversion periods (i.e., 7 s, 14 s, 22 s, 30 s, 34, and 38 s) are shown in Figure 9. The recovery is very good in the center of the study area, especially in the Andean region, but the recovery worsens in the northern part of the Caribbean region (Guajira Peninsula), the Pacific coast bordering with Ecuador, and also toward the edges, especially at long periods. Nonetheless, the recovery of the velocity contrasts is excellent. Similar results hold for checkerboard tests using phase velocities (not shown).

4.3. Phase and Group Velocity Images

Figure 10 displays the tomographic results for group and phase velocity for periods of 7, 22, and 34 s; the remaining group and phase velocity maps can be found in the supporting information section (Figures S1 and S2). The group and phase velocity maps are organized in order of increasing period, therefore increasing penetration depth into the crust. Group velocities, for instance, sample down to a depth approximately equal to one third of their wavelength; thus, short periods (7–10 s) are a good indicator of sedimentary structure, while periods above 30 s are primarily sensitive to the *S* wave velocity structure at ~ 30 km of depth. The tomographic maps at short periods (7 s) show an excellent correlation with surface geology (recall Figure 1). The main sedimentary basins, such as the LMB and MMV, the Túmaco Basin along the Pacific coast, the Maracaibo Basin in Venezuela, and the sediments covering the Panamá arc, are clearly depicted by slow phase and group velocities, as well as the EC between latitudes 4°N and 6°N , which is geologically expressed as a Cretaceous sedimentary basin. In contrast, the provinces composed of igneous-metamorphic complexes, such as the SMM, the Santander Massif (SM), the CC—which include the Antioquia and Ibagué batholiths (AB, IB)—the WC, and the southern segment of the EC—which encompasses the Garzon Massif—are mapped with higher-than-average velocities.

At intermediate periods of 22 s a similar pattern is observed, especially for group velocities. This is because some of the sedimentary structure leaks into the longer periods, which keep some sensitivity to shallow structure as well as upper crustal structure. This is especially true for the accretionary wedge making up the Caribbean in the Sinú-San Jacinto fold belt (SSJB) and the LMB regions along the Atlantic coast of Colombia, which consist of ~ 10 km thick deposits (Lara et al., 2013; Montes et al., 2010; Toto & Kellogg, 1992; Vernet et al., 1992), and for the sedimentary basins along the Pacific coast that possess a similar accretionary wedge, although not as thick as in the Caribbean. Phase velocities, on the other hand, seem to be sampling deeper structure when compared to the corresponding group velocities. Slow velocities are still mapped within the Caribbean basins, but slow velocities also correlate with active volcanic regions along the CC and EC. In contrast high-speed anomalies are observed along the Pacific coast and Eastern Llanos, perhaps resulting from a thinner crust under those regions. A similar interpretation seems to hold at periods of 34 s for both group and phase velocities, indicating that group velocities are no longer sensitive to shallow sedimentary structure, although slow velocities in the Panamá arc and the LMB might still be related to thick sedimentary depositions (recall section 2). Finally, faster phase velocities are observed in the Maracaibo and Caribbean basins, perhaps sensing mantle velocities under a thin crust. The correlation of slow velocities under the CC with active volcanics is, nonetheless, remarkable.

5. Shear Wave Velocity Models

5.1. Inverting Dispersion for *S* Velocity

From the tomographic maps shown in Figure 10, local dispersion curves can be extracted from each grid point at successive periods ranging between 6 and 38 s, and the resulting phase and group velocity dispersion curves can then be jointly inverted to develop *S* wave velocity-depth profiles for each grid point. Inversion of group and phase velocities has been performed through the Neighborhood Algorithm (NA) of Sambridge (1999), as implemented in the software package Dinver (<http://www.geopsy.org>) of Wathelet (2008). This technique has been applied successfully by several authors (e.g., Behr et al., 2010; Kao et al., 2013; Köhler et al., 2015; Köhler et al., 2012) to estimate isotropic layered models from dispersion data. The implementation combines computation of dispersion curves (forward problem) through the Dunkin (1965) formulation and an improved version of the original NA. The implementation is based on a powerful geometric sampling with Voronoi cells, which explores the model space in a quasi-random, sequential, and

Figure 10. Rayleigh-wave group and phase velocity maps for selected periods of 7, 22, and 34 s. Major faults (black lines) are superimposed from Velloza et al. (2012), the same as Figure 1. The blue polygon encompasses the region used for the V_s inversion (section 4.2). The red and blue triangles show the active and inactive volcanoes, respectively.

nonuniform manner, taking into account previous estimations of the cost function to generate new and better families of solutions. The misfit function is optimized in a least squares sense (L2-norm) between the dispersion curve predicted by the model and that constructed from the surface wave maps. The velocity search in the NA is performed by varying 20% the velocity in each layer.

Figure 11. The 1-D profiles of the V_s resulting from inversion. The colors illustrate the density distributions of models with misfits smaller than 0.1. The white line shows the observed data. (a) Profile next to Santa Marta Massif (SMM), (b) profile under Western Cordillera (WC), and (c) profile south of CC.

We considered two starting S velocity models in order to assess the variability of the inversion results. The first starting model is a modified IASP91 model (Kennett & Engdahl, 1991) with Moho depths constrained from crustal thickness estimates developed by Poveda et al. (2015). The V_p/V_s ratio is constrained to be between 1.60 and 1.85, consistent with typical values of the rocks in the continental crust (Christensen, 1996); and density is kept to the values from IASP91 with a 10% perturbation. Wathelet et al. (2004) demonstrated that influence of density in the dispersion curve is minimal. The second model is based on a crust of uniform shear wave velocity of 4 km/s with layer thickness of 4 km, with no a priori constraints on crustal thickness. V_p/V_s varies in the same way as for the first model. Finally, the density varies 20% around a starting density of 3.0 g/cm³ for each layer.

The NA search in the model space starts with 100 models, and the 50 best Voronoi cells in the model space are then selected to generate 100 new models during several iterations. The inversion is stopped after 250 iterations, resulting in 25,000 shear velocity models being evaluated at each grid point. From the set of 25,000 models examined during the NA process, we select and average the 500 best fitting models to obtain a representative model for each grid point. Figure 11 displays inversion examples at three select locations starting from the constrained IASP91 velocity model, where a good fit between observed and predicted dispersion curves can be observed. Optimal misfit values are around 0.06 km/s, with the corresponding best fitting models being concentrated within a narrow portion of the model

space. The select models are located in the SMM (Figure 11a), the WC (Figure 11b), and between the WC and CC (Figure 11c). Tests of depth resolution for the dispersion curves are given in the supporting information (Figure S3), showing that negative, delta-like anomalies interspaced at depths of 10, 20, 30, and 40 km, respectively, are recovered at the right depths for spikes shallower than 40 km. Our *S* wave models are thus only reliable for depths down to ~35 km.

The uniqueness of the results is assessed from the similarity of the velocity models developed from the two starting models (constrained IASP91 and unconstrained, uniform crustal velocity). A detailed comparison is provided in the supporting information (Figure S4), showing histograms of root-mean-square (RMS) between the inverted *S* velocity models developed from the two starting models that reveal differences are small, especially in the upper crust. Somewhat larger differences begin to appear after 30 km, concluding that the dependence with the starting model is minimal. For interpretation, we will take the results from the IASP91 starting model with crustal thickness constraints. Maps of average misfit distribution between observed and predicted dispersion velocities are given in the supporting information (Figure S5). The misfit varies between 0.01 and 0.15 km/s, with overall better fits in the Andean Region (misfit <0.06 km/s) than in the Caribbean and Pacific coasts (< 0.1 km/s, with local occurrences of 0.1–0.15 km/s).

Finally, to check the robustness of some crustal anomalies—specifically, those under thick sedimentary layers such as the LMB—we performed two synthetic tests (Figure S6). The first test consists of a spike test for the 30 s group velocity tomography, concluding that anomalies for this period are well recovered. The second test consists of an inversion of synthetic “data” generated after assuming a 30 km thick crust with a 10 km thick sedimentary layer. This second test demonstrates that, although sedimentary velocities are smeared into the shallower basement (down to ~15 km depth), they do not pervade the entire crustal column.

5.2. Models of *S* Velocity Variation

The suite of 1-D *S* velocity profiles were assembled to build maps of lateral *S* velocity variation for depths ranging between 5 and 35 km. Solutions generated from the IASP91 model with Moho depth constraints were selected for the construction of the maps. The minimum curvature interpolation method of Smith and Wessel (1990), with a tension factor of $T = 0.25$, was considered for this purpose. A number of velocity slices are presented in Figure 12 at depth intervals of 5 km. The *S* velocity maps show features similar to the dispersion velocity maps, with shallow slices (5–10 km depth) mostly correlating with surface geology and deeper slices (25–35 km depth) mostly correlating with the location of active surface volcanism.

More in detail, at 5 km depth (Figure 12a), we find marked slow *S* velocity values on the northwestern side of the Sinú and San Jacinto belts (~2.4 km/s), Maracaibo Basin (~2.7 km/s), the EC between 4°N and 5°N (~2.7 km/s), and the Pacific coastal basin (~2.5 km/s), all of them correlating with the presence of sedimentary rocks (Figure 1). High velocities of ~3.6 km/s characterize the northern CC, especially in the Antioqueño Plateau and the Antioqueño Batholith. This region is characterized by the intrusion of Mesozoic quartz diorite batholiths and plutons (Restrepo-Moreno et al., 2009), with metamorphic basement of Paleozoic and Precambrian ages (Case et al., 1971). Batholiths of Jurassic age (Cardona et al., 2011; Tschanz et al., 1974) present in the northernmost part of Colombia, such as the SMM, also correlate with relatively high velocities (~3.2 km/s). Finally, high-velocity values of ~3.3 km/s are observed in the EC, MMV, and CC south of ~4°N, correlating with mapped outcrops such as the Garzón Massif, the Quetame Massif, and with Jurassic plutonism under the CC.

Similar patterns are observed at 10 km depth (Figure 12b), although some slow-velocity features start to fade away. This is the case of the Panamá arc, where velocity increases from 2.8 km/s to 3.1 km/s at 5 and 10 km, respectively, and the central segment of the EC, where the velocity estimated is ~3.0 km/s. High velocities of ~3.6 km/s north of the CC, nonetheless, remain.

At 20 km depth, low-velocity anomalies (~3.0 km/s) are observed throughout the Caribbean region (SSJB, PB, and SJB) and the central segment of the EC (~3.3 km/s). In contrast, the northern segment of the CC, the MMV and the SM (EC northernmost flank), and SMM display high velocities of ~3.7 km/s with little lateral variation among them. This high-velocity pattern extends farther down to 25 and 30 km depth. At 25 km depth under the Pacific coast, south of 4°N, high velocities (> ~4.0 km/s) are also observed. More interestingly, within the same depth range, low velocities of ~3.4 km/s are found under the volcanic complexes, especially under the volcanoes near the border with Ecuador and those located between 3°N and 5.4°N (Figure 12d).

Figure 12. Shear wave velocity maps for selected depths. The red and blue triangles indicate the active and inactive volcanoes, respectively, and major faults (blue lines) are superimposed from Veloza et al. (2012). The thick black lines shows locations of profiles in Figure 13. The dashed red lines shows proposed Caldas tear (Vargas & Mann, 2013), and the dashed blue line represents the Romeral suture (RS) on LMB. Other features are labeled as in Figure 1, and depth is indicated in the lower right corner of each panel.

At 30–35 km depth, we find high velocities (> 4.1 km/s) along the Pacific coast (Tumaco Basin and Panama arc), part of the WC, LMB, and the Llanos Basin, probably reflecting thin crust. In the CC and CE slow velocities around 3.7 km/s dominate the regions, consistent with the presence of a crustal root. Slow velocities (< 3.5 km/s) along the southern CC still correlate with active volcanic systems located on the surface, and similar slow velocities are also observed in the central part of the EC, approximately coinciding with the location of inactive volcanism. Slow velocities are also observed at these depths in the southern flank of the LMB under the San Jorge Basin. A possible origin for these anomalies is discussed in the next section.

6. Implications for Magmatism and (Magmatic) Overprint in the Upper Plate

The velocity slices described in the previous section show a generally good agreement with previous studies (Flueh et al., 1981; Meyer et al., 1976; Mooney et al., 1979) and, at shallow depths, correlate well with major surface geological features (Gómez et al., 2007; Veloza et al., 2012). Our results are directly comparable to those produced by Syracuse et al. (2016). From the joint inversion of local P wave and S wave arrival times reported by the RSNC, regional surface wave dispersion in the 10–70 s period range, and gravity data from the Earth Gravitational Model 2008 (EGM2008; Pavlis et al., 2012), Syracuse et al. (2016) produced images of lateral P wave and S wave velocity variation down to 155 km depth for the Colombian Andes and adjacent regions. The S velocity depth slices at 12.5 km and 32.5 km (see Figure 6 in Syracuse et al., 2016) are—at long wavelengths—similar to our S velocity maps at 10 and 35 km depth, respectively (Figures 12b and 12f), displaying similar patterns of fast and slow S velocity throughout the area. At shallow crustal depths, for instance, both studies show slow velocities in the LMB and the Eastern Llanos, separated by faster velocities in the intervening cordilleras; at intermediate crustal depths, both studies display slow velocities in the LMB and under the volcanic areas of the EC and CC. Due to the inclusion of shorter periods in the development of our velocity models, however, the resulting velocity images have better resolution at short wavelengths. Our images, for instance, clearly depict the fast velocities characterizing the Antioqueño batholith at shallow depths (Figures 12a and 12b) and successfully image the shift in volcanism following the trend marked by the Caldas tear at intermediate depths (Figures 12d and 12e), which were not resolved by Syracuse et al. (2016).

6.1. Magmatic Overprinting

Figures 12a and 12b show that the signature of sedimentary basins and accretionary wedges can be tracked through slow S velocities (< 3.0 km/s) at shallow depths (5–10 km). In the LMB, slow S velocities are observed along the SSJB, Plató-San Jorge Basin, and Cesar-Rancheria Basin, where sediment thickness can reach values of 3–12 km (Lara et al., 2013; Montes et al., 2010; Sanchez & Mann, 2015; Toto & Kellogg, 1992; Vernet et al., 1992); the Cesar-Rancheria Basin, however, is not clearly discriminated in the shear wave velocity maps because the dimensions (20–30 km) are lower than the resolution length (~ 50 km) of the velocity maps. Similarly, the Plató and San Jorge Basins—also characterized by thick sedimentary sequences—display slow S velocities. And slow velocities are also found in the Maracaibo Basin in Venezuela, where sediment thickness varies between 3 and 9 km (Mann et al., 2006). Slow velocities (2.8–3.0 km/s) are also observed under the central flank of the EC (4°N to 6.5°N, Figures 12a and 12b), which is considered a fold and thrust belt with Cretaceous sedimentary cover units (Horton et al., 2015; Mora & Parra, 2008), and the neighboring MMV, which is covered by Quaternary-Neogene sedimentary rocks (Mora et al., 2006; Parra et al., 2009). These anomalies do not show continuity at larger depths, as sedimentary basins are shallow features that are confined within the shallow crust.

Some shallow, high-velocity anomalies also lack continuity at depth. High velocities are, for instance, observed in the Ibagué Batholith and paralleling the volcanic line under the CC; however, these anomalies are replaced by low-velocity anomalies at larger depths of 20–35 km (Figures 12c–12f). As argued in the next section, we believe that velocity anomalies below those depths are dominated by the signature of subduction-related volcanism, implying that the geological formations observed at surface levels, which attest to the accretionary and deformational history of the northeastern Andean region, are being overprinted at depth by younger magmatism.

In contrast, some high-velocity anomalies do have continuity at depth. High velocities are observed in the SSM and SM down to ~ 30 km depth (Figures 12a–12e), reflecting the presence of an outcropping crystalline basement, Jurassic Plutonism, and Precambrian metamorphic rocks (Cardona et al., 2011); fast velocities are also observed along the CC, which can be traced down to 35 km depth north of 5°N latitude and extending

laterally into the MMV (Figures 12a–12f). According to Montes et al. (2005), this whole region represents a unique crustal block with a more rigid behavior and less deformation than the surrounding EC, consistent with persistently fast velocities.

This would conversely attest to the lack of magmatic overprinting in these areas due to the absence of volcanic activity in this region. Magmatic overprinting at depth of surface geological features in the Colombian Andes seems thus to be dictated by the geometry and segmentation of the subducting Nazca and Caribbean slabs.

6.2. Subduction-Related Volcanism

Active volcanism in Colombia is presently found along the CC, south of $\sim 5^{\circ}\text{N}$ latitude (Figure 1), and has been related to the subduction of the Nazca plate along the Pacific margin of South America. This volcanism is generally described as mainly explosive in character and andesitic in composition (e.g., Marín-Ceron et al., 2010). The strong correlation between slow S velocities at intermediate depths (25–35 km) and surface volcanism in the CC makes us think that those velocities are related to magma production along the subduction front. The magmatic evolution of the Colombian Andes was investigated in Weber et al. (2002) from the analysis of deep crustal xenolith suites. They proposed a model for the deep Andean crust in which basaltic components (e.g., oceanic plateau) and minor sediment are incorporated into the subduction-accretion complex, with areas of possible melt formation within the mantle wedge, mantle-crust transition, and accreted lower crust. Considering that our observed S velocities display a similar S velocity range of 3.4–3.6 km/s within the entire 25–35 km depth range, which exclude a purely thermal origin, and that the xenolith suite does not display evidence of pervasive crustal melting (Weber et al., 2002), we conclude that the observed slow S velocities likely correspond to subduction-related magmas ponding at midcrustal depths.

More enigmatic is the presence of even slower S velocities (3.2–3.6 km/s) in the same depth range under the inactive volcanoes of the EC (Figures 12d–12f). Volcanism in the EC is concentrated in the Paipa-Iza volcanic complex, which is believed to have formed during the late Pliocene when the Cordillera reached its present elevation (Pardo et al., 2005). Recent thermochronological and geochronological constraints on the eruption history of this complex developed from zircon fission-track analysis and U-Pb data have bracketed the time of this volcanic activity between 5.9 Ma and (at least) 1.8 Ma (Bernet et al., 2016). Interestingly, the study argues that hydrothermal activity, with water temperatures up to 73°C , CO_2 gas seepages associated with regional faults, and a strong local geothermal gradient of $74^{\circ}\text{C}/\text{km}$, supports the existence of a magma chamber at shallow depths. If this inference were correct, an interpretation of the observed slow S velocities under the volcanic complex as deriving from partial melts—similar to the interpretation under the active volcanism of the CC—would be plausible. Indeed, bulk V_p/V_s ratios in the area are high, around 1.82–1.86 (Poveda et al., 2015), as well as L_g attenuation (Ojeda and Ottemoller, 2002), consistent with the presence of partial melts in the crust. Our results therefore support Bernet et al.'s (2016) conclusion that volcanism under the EC might be inactive but not necessarily extinct. Moreover, magmatic activity in the region might help explain the nonisostatic residual topography of the region reported by Yarce et al. (2014), as the thermally perturbed asthenospheric wedge expected in the region (Chiarabba et al., 2016) might provide the necessary buoyancy to explain the elevations.

Alternatively, low-velocities may also be an indication of aligned anisotropic minerals. High V_p/V_s ratios might be related to high pore fluid pressure and crack anisotropy (Wang et al., 2012), and recent local S -splitting estimations (Idárraga-García et al., 2016) display high anisotropy values of 3.09% and 3.43% for the EC crust. Moreover, it is well known that the EC is a deformed NE trending fold-and-thrust belt, and several authors have documented the crustal shortening, thickening, and uplift that occurred in this region (Colletta et al., 1990; Cortés et al., 2006; Dengo & Covey, 1993; Mora et al., 2010) during late Miocene (Egbue et al., 2014; Taboada et al., 2000). The detachment surface depth under EC fold-and-thrust belt is postulated to be between 20 and 30 km depth (Mora & Parra, 2008), which correlates well with the termination of slow S velocities at depth (Figure 12).

Even more enigmatic is the presence of similarly slow S velocities under the LMB, where slower-than-average S velocities are found for almost the entire depth range (Figures 12a–12e). S velocities are 2.4–2.8 km/s down to ~ 10 km depth, which have been found consistent with the accretionary prism reported earlier (section 6.1); S velocities of 3.0–3.4 km/s at 20–30 km, however, are harder to explain. To help clarify this issue, a NW-SE

Figure 13. Cross section on LMB and Andean regions (see Figure 12a), and associated topographic profiles, showing several provinces (see Figure 1). Segmented white lines are the top of the Caribbean slab and sediment thickness from Bernal-Olaya, Mann, and Vargas (2015) and Mora-Bohórquez et al. (2017).

trending cross section crossing the LMB and EC regions is displayed in Figure 13a. The figure shows that the Caribbean region, like the EC, consists of a shallow layer of slow (2.4–2.8 km/s) S velocities overlying an also slow (3.0–3.4 km/s) upper middle crust. Moreover, also like the EC, bulk V_p/V_s ratios in the Caribbean region are in the 1.82–1.86 range (Poveda et al., 2015), which would again be consistent with the presence of partial melts. Strikingly, this region is regarded as amagmatic.

We argue, nonetheless, that partial melts might be present within the LMB. A recent study of field, geochemical, geochronological, biostratigraphical, and sedimentary provenance of basaltic and associated sediments in northern Colombia has revealed the existence of Middle Miocene (13–14 Ma) mafic volcanism within the Caribbean continental margin (Lara et al., 2013). In particular, geochemical analysis showed that basalts were formed by asthenospheric melting at shallow depths mixed with some additional slab input and, together with biostratigraphic results, demonstrated that they were deposited in a platform setting. Considering the low-angle subduction of the Caribbean plate and the corresponding absence of a mantle wedge (recall section 2.1), the authors proposed the presence of a vertical tear in the subducting plate through which asthenospheric melts would have found their way through while acquiring the observed slab input. Vertical tears might result from along-strike extension related to the strong curvature of the subduction as it passes around the NW corner of South America, and this type of extension was invoked in Bernal-Olaya, Sanchez et al. (2015) to explain the origin of the Plato and San Jorge Basins within the LMB. Assuming a speed of 20 mm/yr for the Caribbean plate relative to the South American plate, and a position of the tear under the continental platform at 13–14 Ma, it is easy to calculate that the tear would be presently located 260–280 km farther inland under the South American plate. This distance range is consistent with the location of the Caribbean low S velocity region (Figures 12c–12e), suggesting that the Caribbean slab might have been breached by asthenospheric melts that are now ponding at mid-crustal depths within the overriding plate.

As stated before (recall section 5.1 and Figure S6), we are confident that the midcrustal, low-velocity anomaly under the LMB is not an artifact from the inversion procedure. Moreover, the trace of the Romeral Suture—which is regarded as a major fault separating the oceanic crustal rocks of the WC from the continental-affinity basement to the east (Cediel et al., 2003; Restrepo & Toussaint, 1988; Taboada et al., 2000)—is very consistent with the velocity contrast separating the CC from the sedimentary units described above at shallow depths (Figures 12a and 12b), suggesting a continuation of this fault into the LMB (Figure 12a) (Bernal-Olaya, Mann, & Escalona, 2015; Taboada et al., 2000; Toro et al., 2004). This consistency degrades at larger depths (Figures 12d–12f), suggesting that the low-velocity anomaly under the LMB at midcrustal depths is not related to sedimentary structure.

However, we do acknowledge that the large lateral extent of the anomaly is at odds with the lack of surface volcanism associated with it, and that alternative mechanisms might be at play. Low-velocity zones may also be related to fluid migration along major faults zones, especially when coincident with large V_p/V_s ratios (Thurber et al., 2003; Zhao et al., 1996). Such relationships have been recently invoked, for instance, to explain large V_p/V_s ratios within the western Hispaniola crust (Corbeau et al., 2017) and, perhaps more importantly, to explain low-velocity anomalies along the Sinú and Romeral Fault system (Bernal-Olaya, Mann, & Vargas, 2015). In the case of the LMB anomaly, the Romeral fault system—which, as described above, might actually be extending under the basin—could provide the paths for fluids emanating from the subducting Caribbean plate to infiltrate the surrounding continental mantle and reach midcrustal depths. Fluid migration along the Romeral fault system would then explain both the low velocities reported in this study and the high bulk V_p/V_s ratios reported in the receiver function study of Poveda et al. (2015). Moreover, fluid expulsion might also provide an alternative explanation for the midcrustal, low-velocity anomaly under the EC, as Bernal-Olaya, Mann, and Vargas (2015) have related low velocities at depths >80 km near the Bucaramanga earthquake nest with such phenomenon.

Finally, we note that our velocity slices clearly image the signature of the “Caldas tear” postulated by Vargas and Mann (2013) at crustal depths within the overriding plate. As mentioned in section 2, the Caldas tear is an E-W trending feature that extends for ~ 240 km and separates two slab segments that subduct at different angles. The projection of the Caldas tear is superimposed in our S velocity model (Figures 12d and 12e), showing it can be traced down to 20–35 km depth.

Acknowledgments

The authors want to express their gratitude to the Colombian Geological Survey (Servicio Geológico Colombiano, SGC) for generously providing the data for conducting this study, in particular to Marta Calvache, technical director of geological hazards of Colombian Geological Survey, and Viviana Dionicio, coordinator of the Seismological Network (Red Sismológica Nacional de Colombia, RSNC). Raw data are freely available and accessible to the scientific community through written request and justification addressed to the Colombian Geological Survey (sismologo@sgc.gov.co). E. P. acknowledges support from the Coordenação de Aperfeiçoamento de Pessoal de Nível Superior (CAPES) through a 4 year scholarship to complete his Ph.D. J. J. thanks CNPq for his research fellowship (CNPq, grant 308171/2012-8). M. S. acknowledges support by Brazilian Science Without Border Program, grant 40.2174/2012-7. N. P. G. thanks the Colombian Geological Survey, contract 393-2016. Maps were produced using Generic Mapping Tools (GMT) software (Smith & Wessel, 1990). We also thank the Editor Martha Savage, the reviewer Paul Mann, and an anonymous reviewer for their help in improving the final manuscript.

7. Conclusions

We have presented the first ambient noise tomography of the Colombian Andes and surrounding regions from recordings of 53 broadband seismic stations belonging to the *Red Sismológica Nacional de Colombia* (RSNC). A total of 1,300 empirical Green’s functions—identified as fundamental-mode, Rayleigh waves—were recovered, covering much of the country and furnishing new constraints in areas where previous information was limited. Our results resolve geological features with high accuracy at shallow levels, as well as patterns associated with magmatism and plate segmentation at midcrustal depths. More specifically, slow S velocities under the CC have been attributed to accumulation of magmas in the middle crust generated in the underlying mantle wedge, and evidence of magmatic activity has been reported under the inactive volcanoes of the EC. Slow S velocities in the Caribbean middle crust have been related to asthenospheric melts that would have breached the Caribbean flat slab through a vertical tear. Alternatively, migration of fluids along faults from expulsion of nearby subducting slabs could provide a feasible explanation for such velocity anomalies.

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