Variation in Upper Plate Crustal and Lithospheric Mantle Structure in the Greater
 and Lesser Antilles from Ambient Noise Tomography

D. Schlaphorst^{1,2}, N. Harmon³, J.M. Kendall⁴, C.A. Rychert³, J. Collier⁵, A. Rietbrock⁶, S.
 Goes⁵, and the VoiLA Team[†]

- 4 Goes, and the vollA rea
- 5 †The VoiLA Team:
- 6 R.W. Allen⁵, L. Bie⁶, J.D. Blundy⁴, B. Chichester³, G.F. Cooper⁷, J.P. Davidson⁸, R.G. Davy⁵,
- 7 T.J. Henstock³, S.P. Hicks⁵, C.G. Macpherson⁸, B. Maunder⁵, J. Prytulak⁸, J. van Hunen⁸, J.J.
- 8 Wilkinson^{5,9}, M. Wilson¹⁰
- 9
- ¹⁰ ¹Instituto Dom Luiz (IDL), Universidade de Lisboa, Portugal.
- ¹¹ ²School of Earth Sciences, University of Bristol, UK.
- ¹² ³Ocean and Earth Science, University of Southampton, National Oceanography Centre,
- 13 Southampton, UK.
- ⁴Department of Earth Sciences, University of Oxford, UK.
- ⁵Department of Earth Science and Engineering, Imperial College London, London, UK.
- ¹⁶ ⁶Geophysical Institute (GPI), Karlsruhe Institute of Technology, Karlsruhe, Germany.
- ¹⁷ ⁷School of Earth and Environment, Cardiff University, Cardiff, UK
- ¹⁸ ⁸Department of Earth Sciences, Durham University, Durham, UK.
- ⁹Department of Earth Sciences, Natural History Museum, London, UK.
- ²⁰ ¹⁰School of Earth and Environment, University of Leeds, Leeds, UK.
- 21
- 22 Corresponding author: David Schlaphorst (dschlaphorst@fc.ul.pt)

23 Key Points:

- With land and ocean bottom seismometers, we map significant variations in sediment/crustal thickness in fore- and back-arc in new detail.
- Sediments thickens towards the southern arc, whereas crustal thickness increases towards
 the north.
- Volatile input and pathways depend on slab seafloor morphology, 3D flow and/or melt
 ponding beneath the upper plate redistributing melt.

31 Abstract

The crust and upper mantle structure of the Greater and Lesser Antilles Arc provides insights 32 into key subduction zone processes in a unique region of slow convergence of old slow-33 spreading oceanic lithosphere. We use ambient noise tomography gathered from island 34 broadband seismic stations and the temporary ocean bottom seismometer network installed as 35 part of the VoiLA experiment to map crustal and upper mantle shear-wave velocity of the eastern 36 Greater Antilles and the Lesser Antilles Arc. Taking the depth to the 2.0 km/s contour as a proxy, 37 we find sediment thickness up to 15 km in the south in the Grenada and Tobago basins and 38 thinner sediments near the arc and to the north. We observe thicker crust, based on the depth to 39 the 4.0 km/s velocity contour, beneath the arc platforms with the greatest crustal thickness of 40 around 30 km, likely related to crustal addition from arc volcanism through time. There are 41 distinct low velocity zones (4.2-4.4 km/s) in the mantle wedge (30-50 km depth), beneath the 42 43 Mona Passage, Guadeloupe-Martinique, and the Grenadines. The Mona passage mantle anomaly may be related to ongoing extension there, while the Guadeloupe-Martinique and Grenadine 44 anomalies are likely related to fluid flux, upwelling, and/or partial melt related to nearby slab 45 features. The location of the Guadeloupe-Martinique anomaly is slightly to the south of the 46 obliquely subducted fracture zones. This feature could be explained by either three-dimensional 47 mantle flow, a gap in the slab, variable slab hydration, and/or melt dynamics including ponding 48 49 and interactions with the upper plate.

50 1 Introduction

Subduction zones play a key role in driving plate tectonics, delivering slabs and their 51 volatile elements to the mantle. Slab dehydration during the subduction process results in melt 52 generation, which rises and eventually leads to the formation of volcanic island arcs. Hydration 53 may vary along strike depending on slab crustal structure, fracture zones and sediment packages 54 (Stern, 2003). However, the relative importance of these delivery mechanisms and the exact 55 pathways of hydration and melt to surface volcanism are not well known. Imaging variations in 56 the crust and lithospheric mantle in these regions offers insights into subduction processes and 57 the formation of continental crust through arc volcanism. Seismic methods offer the best 58 potential to image the arc lithosphere at the scale of tens of kilometres required. However, many 59 60 studies of island arcs are limited by the restriction of land stations to mostly small arc islands (noisy) in a linear alignment or sparsity in linear offshore active-source seismic profiles. A 61 further limitation on our understanding of subduction processes is that until recently, seismic 62 studies have focussed mostly on Pacific subduction zones (see Melekhova et al. (2019) for lists 63 of references). The Volatile Recycling in the Lesser Antilles (VoiLA) experiment is focussed on 64 the Lesser Antilles Arc (LAA), where slow-spread lithosphere are subducted (Goes et al., 2019). 65

Many subduction zones show complex crust and mantle structure in the upper plate, with 66 67 along-arc heterogeneities, such as mantle xenolith mineral composition and discontinuity depths (e.g., Boynton et al., 1979; Kodaira et al., 2007; Melekhova et al., 2019; Schlaphorst et al., 2018; 68 Shillington et al., 2004). These are likely related to factors including variation in slab age, 69 convergence rate and direction. In addition, it is likely that oceanic plates are more hydrated at 70 slow spreading centres that at fast spreading centres (e.g., Davy et al., 2019) and the distribution 71 of hydration can vary significantly along the arc (e.g., Schlaphorst et al., 2016). Related to that, 72 erupted magmas from Guadeloupe and Dominica show a high δ^{11} B anomaly that is indicative of 73

dewatering of serpentinized oceanic crust (Cooper et al., 2020). In recent years, differences in the nature and intensity of the aforementioned slab heterogeneities between regions of subduction of fast spreading plates (mostly located in the Pacific) and those of slow spreading plates (found in the Atlantic) have become increasingly clear (Bie et al., 2020; Cooper et al., 2020; Melekhova et

⁷⁸ al., 2019; Schlaphorst et al., 2018).

79 Here, we image the LAA and the eastern Greater Antilles (GA), a slow-subduction zone located at the eastern boundary of the Caribbean plate, using ambient noise tomography (ANT). 80 ANT has been used in a recent study to map shear-wave velocities in the Caribbean (Arnaiz-81 Rodríguez et. al, 2020). In our study, data are recorded by a dense network of permanent land 82 stations and additionally an array of oceanic bottom seismometers deployed during the VoiLA 83 experiment (VoiLA; Goes et al., 2019), which increases resolution in the fore- and back-arc. 84 Previous studies have investigated the crustal structure in this area via active source imaging and 85 receiver functions (e.g., Allen et al., 2019; Arnaíz-Rodroguez et al., 2016; Boynton et al., 1979; 86 Chichester et al., 2020; Christeson et al., 2008; Dorel et al, 1974; Kopp et al., 2011; Melekhova 87 88 et al., 2019; Pauldron et al., 2020; Schlaphorst et al., 2018; Sevilla et al., 2012) but the interisland subsurface is poorly studied so far. Our aim is to image the shallow mantle and crust, 89 90 investigating variations in sediment accumulation, crustal thickness and the velocity structure in 91 the uppermost mantle across the arcs.

93 2 Tectonic Setting



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Figure 1: Map of the LAA and the GA, including bathymetry and the locations of seismic island 95 stations (purple triangles) and OBS stations (red inverted triangles). Station details can be found 96 in Supplementary Table S1. The global location of the area is shown in the inset. The islands of 97 the LAA are coloured in vellow (Volcanic Caribbees) and brown (Limestone Carribees), and the 98 islands of the GA are white. The location of the outer arc is shown as a thick grey line with the 99 southern buried part depicted in a lighter grey (Allen et al., 2019). Thin black dashed lines 100 101 indicate fracture zones and the purple show the boundary between the proto-Caribbean (pC) and equatorial Atlantic (eA) seafloor (Cooper et al., 2020). TR indicates the start of the Tiburon 102 Ridge. Thick black hashed lines indicate the paths of the cross-correlated data from island-103 island and OBS–OBS station pairs for the examples shown in Fig. 3. Red stars show the 104 locations of the individual results (Fig. 8). Thin solid lines indicate positions of the cross-105 sections (Fig. 10) for the GA (blue) and LAA (red). 106

The eastern GA and LAA (Fig. 1) are located on the northern and eastern margin of LIPthickened (large igneous province) oceanic crust of the Caribbean plate (Mauffret & Leroy,
1997). The origin and evolution of the Caribbean plate since the Cretaceous can be inferred from
its crustal structure. Over the past 100 My the Caribbean plate has been moving eastwards

relative to both the North and South American plates. This motion has been accommodated by

subduction, and strike-slip boundaries, resulting in large-scale shear zones through

113 continental/oceanic crust and arc volcanism at various times and places on the plate margin

114 (Boschman 2014). Subsequently, sedimentation and tectonism have obscured many of these

115 features, therefore making it difficult to unravel a complex history.

The eastern GA area includes the islands of Hispaniola, Puerto Rico and the Virgin 116 Islands, which together form the Northern Caribbean Plate Boundary Zone. The islands are 117 located on microplates, which override the North American Plate and the thickened Caribbean 118 Plate in two subduction zones to the north and to the south (Dolan et al., 1998). Dominated by a 119 left-lateral, east-west strike-slip motion, the complex tectonic setting of the area includes oblique 120 convergence of the North American Plate, as well as a pull-apart basin in the Mona Passage 121 between Hispaniola and Puerto Rico (Dolan et al., 1998; Masson & Scanlon, 1991; Ten Brink, 122 2005). Furthermore, a potential slab gap east of Puerto Rico has been suggested in several 123 studies, based on bathymetry, seismic reflection profiles and gravity measurements (Ten Brink, 124 2005), as well as shear-wave splitting (Meighan et al., 2013; Schlaphorst et al., 2017). 125

The volcanic islands of the LAA have been formed by slow (18 to 20 mm/yr; DeMets et 126 al., 2000), westward subduction of the North and South American plates beneath the Caribbean 127 Plate (Wadge & Shepherd, 1984). The entire arc extends about 800 km in a mostly north-south 128 orientation. To the north and south it is bound by the GA and the South American continent, 129 respectively. To the west it neighbours the Grenada Basin and the Aves Ridge (Boynton et al., 130 1979; Christeson et al., 2008), an abandoned, Mesozoic volcanic arc. The Grenada Basin forms a 131 depression characterised by a flat bottom in the south with depths down to 3 km that spans a 132 region of roughly 150 by 600 km along the curved outline of the LAA and has formed due to 133 north-south striking back-arc spreading while the so-called outer arc (OA; Fig. 1) was active 134 (Allen et al., 2019; Bouysse, 1988; Padron et al., 2020). 135

The surface expression of the present-day arc comprises eleven major islands and the 136 smaller Grenadine islands between St. Vincent and Grenada. Due to a spatial and temporal 137 variations in magmatic output, the size and spacing of the volcanic islands varies. Around the 138 island of Dominica, the LAA bifurcates to the north: the western active limb (the Volcanic 139 Caribbees) includes the western part of Guadeloupe, as well as Montserrat, St. Kitts, St. 140 Eustatius and Saba; the inactive eastern limb (the Limestone Caribbees) includes the eastern part 141 of Guadeloupe, as well as Antigua, Barbuda and St. Martin (Bouysse et al., 1990). The 142 Limestone Caribbees are part of the old OA that continues to the south buried under sediments 143 from the Barbados accretionary prism (Allen et al., 2019). Due to the curvature of the present-144 day arc, a systematic change in subduction style can be observed, with increased strike-slip 145 movement towards the northern end (DeMets et al., 1994, 2000). A further variation along the 146 arc includes the sedimentary load of the incoming plate. Northward extent of sediment on the 147 incoming plate sourced from the South America continent is blocked by the presence of the east-148 west trending Tiburon Ridge and other submarine highs leading to thicker sediment cover in the 149 south relative to the north. However, it is possible that a major part of these clastic sediments is 150 151 accreted rather than subducted, forming the Barbados accretionary prism (e.g., Faugères et al., 1993). The orientation of the Wadati-Benioff zone along the arc also shows a significant change 152 and thickens towards the south. The top of the subducting slab can be inferred from seismicity, 153 lying at a depth of roughly 120 km beneath the islands of the LAA (Bie et al., 2020). Finally, 154 155 there is a change in the age and origin of the plate along strike, with the proto-Caribbean lithosphere subducting ahead of the South American oceanic lithosphere south of Dominica and 156

- 157 lithosphere generated at the Mid-Atlantic Ridge (MAR) subducting to the North (Cooper et al.,
- 158 2020).
- 159





Figure 2: Station pairs and nodal parameterisation. a) Map showing the station pairs used, 162 including locations of seismic island stations (purple triangles) and OBS stations (red inverted 163 triangles). Note that due to the addition of OBS stations the tomography can be carried out on a 164 broader range in the LAA. Moreover, we did not include station pairs between the GA and the 165 LAA, since no crossing paths were possible. b) Nodal parameterisation used in this study with 166 nodes being indicated by the white circles. The spacing is 0.5° by 0.5°. The outside nodes are 167 placed at a 1° distance and are included to dampen any potential heterogeneity outside the 168 169 region of interest. Nodes exist across the Caribbean Sea where we do not expect data recovery owing to the setup of the code. As there is no sensitivity to these nodes due to lack of station-to-170 station paths, this does not have a negative effect on the resolution. 171



173 Figure 3: Examples of 1-year symmetric vertical-vertical component cross-correlations of island

stations (a) and temporary VoiLA OBS (b). The additional peak in (a) around 0 s is likely a

175 processing artifact as a result of data gaps in the land station data (whereas OBS data are

176 *continuous without gaps) and does not affect the tomography results. For the corresponding*

177 station–station paths see Figure 1.



180 Figure 4: see following page for caption. (Please be aware that this is a version with compressed

181 quality for the Word document.)

182 *Figure 4: Symmetric cross-correlations of the vertical-vertical component of all station pairs*

183 ordered by station distance for the island stations (a) and the VoiLA OBS (b). The data are

184 bandpass filtered using a 4th-order Butterworth bandpass filter between 1.5 s and 50 s. The

185 Rayleigh waves can be seen with linear moveouts. The cross-correlation of the island stations

- introduces large peaks at around 0 s (a) due to data gaps in the land station data (whereas the
 OBS data are continuous without gaps). The cross-correlated data of the OBS shows a higher
- amount of noise in the codas (further away from 0 s than the Rayleigh wave onsets; b). However,
- 189 these features do not affect the measurements, since we are focusing on onsets in a window
- around the Rayleigh wave arrivals using a search window from 3.0 km/s to 4.8 km/s. The search
- 191 window is indicated by grey lines beyond a distance of 600 km. A version focusing on the first
- 192 300 km can be found in Supplementary Figure S1.

193 We use ambient noise cross correlations to perform Rayleigh wave tomography across the region, following the technique laid out by Sabra et al. (2005) and Shapiro et al. (2005). We 194 use continuous data recorded at 32 OBS stations from the VoiLA project and 39 permanent land 195 stations (Fig. 2a; Suppl. Tab. S1). The OBS were deployed in 2016 and operational for a time 196 span of approximately 14 months, although deployment times vary from station to station 197 198 (Collier, 2017). We use data from land stations covering the same time span as the OBS network. Additional land station data were used from an interval of equal length in the years 2012/2013 to 199 test for temporal stability. The method follows the basic processing scheme of Bensen et al. 200 (2007) and in particular the scheme presented by Harmon and Rychert (2016). 201

The data are processed in 1-day windows, which allows for faster computation. Here, we 202 concentrate on the vertical components, which show Rayleigh waves most clearly and offer the 203 highest signal-to-noise ratio (e.g., Tanimoto, 2006; Webb, 2007; Mordret et al., 2013a). Before 204 the noise cross-correlation functions (NCF) are generated, the data are downsampled to 1 Hz. 205 Any mean or trend is removed from the raw traces and amplitudes are normalised by a running 206 average filter. Spectral whitening is applied in a frequency band from 0.01 to 0.33 Hz by using 207 the magnitudes of the smoothed Fourier coefficients of the amplitude spectrum as inverse 208 209 weights for the complex spectrum, acting as a spectral normalisation (Bensen et al., 2007; Harmon et al., 2008). At this frequency band, the noise field is predominantly generated by 210 oceanic waves and double-frequency microseism (Webb, 1998), which shows a relatively 211 random source distribution (e.g., Harmon et al., 2008; Mordret et al., 2013b; Yang & Ritzwoller, 212 2008). 213

Station pairings for the NCFs are chosen so that their minimum distance is larger than 214 three times the wavelength, which counteracts near-field effects, while simultaneously allowing 215 for phases to become distinguishable (Bensen et al., 2007; Harmon et al., 2008). The interstation 216 distance range is between 30 km and 1000 km (Figs. 3, 4). In this study we were not able to use 217 cross-correlation of OBS data with permanent station data due to back scattering effects, 218 219 multipathing and mode interference at the islands, an issue observed and discussed in previous studies (Harmon et al., 2012; Tian & Ritzwoller, 2015; see also Suppl. Fig. S2). However, both 220 221 cross-correlation datasets are combined afterwards, resulting in a total number of 1906 pairs (496 OBS pairs and 1410 permanent station pairs; Fig. 2a). The 1-day NCFs for a particular station 222

pairing are then stacked. Due to the linearity of cross-correlation the result is the equivalent of a
 longer timeseries (Bensen et al., 2007).

The waveforms in the positive ("causal") and negative ("acausal") parts of the resulting 225 vertical-vertical (ZZ) component of the NCF are equivalent to waves travelling in both directions 226 between the stations. Summing the positive and the time-reversed negative parts results in a 227 causal, symmetric NCF that represents an average of both parts (Bensen et al., 2007; Harmon et 228 al., 2010; see also Figs. 3, 4). The slight asymmetric behaviour of the traces can be explained by 229 inhomogeneous distribution of noise sources (Bensen et al., 2007; Mordret et al., 2013b), though 230 at low frequencies this is expected to be a minor problem. We window the symmetric NCF using 231 a Tukey window with a 40 s falloff before unwrapping the phase to determine individual cycle 232 ambiguities (Harmon & Rychert, 2016). 233

To estimate the average phase velocity structure of the region we approximate the real 234 component of the Fourier transformed NCF using Bessel functions of the first kind, representing 235 2-dimensional noise source distribution (Aki, 1957). Harmon et al. (2010) show that this 236 approximation is valid at the longer periods used in this study (>7 s). The minimum variance 237 determines the best fit and thus the best amplitude/phase velocity pair. We search for phase 238 velocities in an interval of expected values (3.0 km/s to 4.8 km/s) and select periods ranging 239 from 6 s to 50 s. To investigate the lateral phase velocity structure at different depths we 240 introduce nodal parametrisation (Fig. 2b) and use 2d sensitivity kernels (Zhou et al., 2004) from 241 the average dispersion as the background model, allowing for perturbations from the uniform 242 model in our tomography method (Harmon & Rychert, 2016). The sensitivity kernels are 243 averaged at the nodes using an 80 km Gaussian smoothing window for the inversion 244 parameterization. The same smoothing window is used to perform distance weighted 245 interpolation from the nodes after the inversion to the higher resolution grids $(0.1^{\circ} \text{ spacing})$ 246 shown in the paper. In the damped iterative inversion scheme, we use a damping of 0.2 km/s for 247 the phase velocity at each node. After the inversion we propagate the error out onto the finer grid 248 using the full covariance matrix. 249

Shear-wave velocity inversions from pseudo-dispersion curves are conducted on a 0.1° by 250 0.1° grid throughout the region of interest. We use an iterative non-linear inversion. The starting 251 model is built of 54 layers; the first represents a water layer with changing thickness (Becker et 252 al., 2009); the second represents a sediment layer, again with changing thickness based on a 5x5 253 arc-minute global sediment thickness compilation (Divins, 2003); the remaining layers have 254 fixed thicknesses (two 1 km layers at the top and 5 km below). All layers (except for the water 255 layer, which has $v_s=0.0$ and $v_p=1.5$. km/s) have a v_p/v_s ratio of 1.8 as an average of the ratio that 256 varies laterally and with depth (Schlaphorst et al., 2018). The v_s values are: 0.7 km/s for the 257 sediments (1.5 km/s for sediment thicknesses greater than 1 km), 3 to 4 km/s in even increments 258 in the crust, and 4.5 km/s in the mantle. The low velocity zone in the asthenospheric mantle 259 260 wedge beneath 70 km improves the fit to the longer frequencies, although the change is not significant due to the low sensitivity at that depth. A low velocity zone in this region would be 261 262 consistent with higher fluid and/or melt concentrations as suggested by receiver function imaging of a sharp velocity drop with depth and geochemical indicators of serpentine (Chichester et al., 263 2019; Cooper et al., 2020). Densities are calculated using relationships for crustal (Christensen & 264 Salisbury, 1975) and mantle layers (Birch, 1961). We replace the water layer with the sediment 265 266 values for regions above sea-level. Due to the marine environment of the region we expect low

shear-wave velocities in the upper layers, which are prone to be water saturated (Mordret et al.,

268 2014). An upper limit for S-wave velocity of 4.5 km/s in the crust and 4.85 km/s in the mantle is

fixed in the inversion. We use a normalized chi-squared value as our objective function for the

inversion. In most regions, we are able to reach a normalized chi-squared value of < 1, indicated we are fitting the data within its error. This criterion is used subsequently to mask out regions

we are fitting the data within its erfor. This criterion is used subsequently to mask out regions without sufficient misfit reduction. The shear-wave phase velocity uncertainties are the formal

errors in the model parameters from the last iteration of the linearised least squares inversion.

We examine the resolution of the phase velocity map inversion using a recovery test (Fig. 5). As input we use anomalies placed with 1° distance for 14 and 20 s periods and 1.5° distance for 30 and 45 s periods. These periods were chosen to cover the entire range used. The anomaly strength is set to $\pm 1\%$ around the average velocity for each period, which is a good proxy for the strength of anomalies that can be expected in this study. The synthetic phase is calculated using the finite frequency kernels from each period. We also include random noise to the synthetic data, equivalent to the amount of noise in the real data.

281 Checkerboard tests reveal a good recovery of features along the LAA for all periods 282 especially inside the network, due to the dense station distribution. The resolution in the western 283 GA is poorer due to the confined distribution of station locations along the islands (in the LAA

this is greatly improved due to the inclusion of off-arc OBS data).



Figure 5: Recovery tests for periods shown in Figure 7. The features are placed with 1° distance for the shorter periods and 1.5° distance for the longer periods. For every period, the average velocities (here called ψ) match the ones obtained as averages of the respective periods in the real data. Note that random noise is added to the synthetic phase data, which is equivalent to the amount of noise in the real data. Further recovery tests can be found in Supplementary Figures S3–S6.

293 **4 Results**



294

Figure 6: Measured average phase velocities and their uncertainties. Note that OBS data are systematically faster due to the different crustal properties at oceanic OBS locations.



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Figure 7: Surface wave tomography maps. Regions of strong low and high velocity anomalies
are labelled and are discussed in the text. Phase velocity uncertainties can be found in
Supplementary Figure S7.

The average phase velocities determined from the OBS and land stations (Fig. 6) show 302 303 good agreement at most periods, with the dispersion recorded by the OBS being slightly faster (up to 0.05 km/s). Both dispersion curves show a steeper velocity gradient at shorter periods, 304 reaching a plateau at longer periods that are sensitive to the upper mantle. At periods shorter than 305 14 s the agreement is worse at many individual grid points. This is likely due to the OBS data 306 307 sensing inherently different structure from both the fore- and back-arc and rapid variation in water depth. We therefore exclude these periods in the following analysis and concentrate on the 308 309 longer periods from 14 s to 45 s. To keep the approach unbiased, we keep this range for every point of the study area. 310

Phase velocity maps for different periods generated from the combined dataset of 311 permanent and OBS stations are shown in Figure 7. Note that the smaller distance to the nodal 312 313 border in the south does not affect the recovery of the phase velocity. At 14 s the average Rayleigh wave phase velocity for the GA and the LAA is around 3.3 km/s (Figs. 6, 7a). Two 314 strong low-velocity anomalies can be observed in the southern Grenada Basin (L_{GB}; 2.56 ± 0.08 315 km/) and northeast of the northern end of the LAA (L_{NE} ; 2.78 ± 0.09 km/s), which are influenced 316 by the recovery in those regions. Furthermore, a large region of high velocities of up to $3.71 \pm$ 317 0.06 km/s show in the northern part of the Grenada Basin (H_{GB}). 318

At 20 s the average Rayleigh wave phase velocity for the GA and the LAA is around 3.6 km/s (Figs. 6, 7b). An elongate high-velocity anomaly can be identified, which spans the area

- north of the Grenada Basin, a continuation of the high-velocity anomalies observed at 14 s, and
- meets the LAA around St. Lucia and St. Vincent, where it reaches phase velocities of up to 3.75
- ± 0.07 km/s (H_{GB}). This creates a tripartite pattern along the arc with higher velocities in the centre and significantly lower values to the north (Martinique to Saba) and to the south (the
- 325 Grenadines and Grenada). A further smaller high-velocity anomaly is located north of the Virgin
- Islands, with phase velocities of up to 3.72 ± 0.10 km/s (H_{VI}). Three major low-velocity
- anomalies can be identified. One is located at the eastern end of Hispaniola at the beginning of
- the Mona Passage and shows phase velocities as low as 3.33 ± 0.08 km/s (L_{MP}). Two further
- 329 minor low-velocity anomalies with phase velocity values of around 3.4 km/s are visible in the
- 330 back-arc corner between the GA and the LAA (L_{Co} ; 3.51 ± 0.08 km/s) and the southern LAA
 - 331 back-arc (L_{GB}; 3.56 ± 0.08 km/s).

At 30 s the average phase velocity value is 3.8 km/s (Figs. 6, 7c). The strong low-velocity anomaly beneath eastern Hispaniola also stands out at this period (L_{MP} ; 3.68 ± 0.07 km/s). Along the LAA, a patchy pattern is observable with low-velocity anomalies in the north around St Eustatius (L_{StE} ; 3.77 ± 0.07 km/s), as well as the centre around Dominica (3.78 ± 0.04 km/s) and in the south around Grenada (3.78 ± 0.08 km/s), the latter two being connected via the Grenada Basin (L_{GB}). These three lows are intercepted by areas of high-velocity anomalies around Montserrat to Guadeloupe (H_{MG} ; 3.89 ± 0.05 km/s), as well as Martinique to St. Vincent (H_{MStV} ;

339 3.87 ± 0.08 km/s).

At 45 s the average phase velocity value is 3.9 km/s around the LAA with slightly higher values of 4.0 km/s in the GA (Figs. 6, 7d). A round area of high velocity can be observed in the north-eastern part of the LAA (H_{NE} ; 4.08 ± 0.05 km/s). In the west of the LAA a low-velocity anomaly is located in the southern two-thirds of the Grenada Basin (L_{GB} ; 3.83 ± 0.06 km/s). The feature is in a similar location as the low velocity region at 30 s. The area north of Guadeloupe consistently shows higher average shear-wave velocities under the arc with the feature extending

346 to western Puerto Rico.



Figure 8: Dispersion results for 6 locations around the GA (Mona Passage, Puerto Rico) and
LAA (Northern Back-arc, Guadeloupe, Martinique, Southern Back-arc). The locations are
indicated in Figure 1. (a) Model of shear velocity dispersion curve (line) based on 1-D phase
velocity estimates (black circles with error bars). (b) Resulting best fit shear velocity model. (c)
Rayleigh wave sensitivity kernels with depth at different periods with c representing the model

354 phase velocity (Zhou et al., 2004). Note that the depth scale is condensed below 60 km.



Figure 9: S-wave velocity depth slice maps in the GA and the LAA. The red-to-light blue scale

358 indicates S-wave velocities commonly associated with crustal values, whereas the yellow-to-

dark-blue scale indicates values associated with mantle values. White areas depict regions
 without results, due to either a high misfit in the inversion or location with great distance from

the station pair paths (see Fig. 2a). Thin black hashed lines indicate fracture zones (Cooper et

- *al.*, 2020). Note that based on our resolution tests, we do not interpret features smaller than 50
- 363 *km*, which are influenced by small-scale variations in water depth and sediment thickness.





Figure 10: Velocity-depth transects along the GA (a), across the LAA (b,c), and along the LAA 367 (d), including b-values along the arc (e). The transect crossing positions are marked by white 368 vertical bars. See Figure 1 for line locations. The black dots in (b,c) show all seismicity in a 30 369 km distance to the transect, based on the catalogue of the Institut de Physique du Globe de Paris 370 371 (IPGP) from the years 1996 to 2012 and events located with the VoiLA OBS network (Bie et al., 2020). The black dashed line show the location of the subducting slab based on slab2 (Hayes et 372 al., 2018). The black inverted triangles show the positions of the LAA and the OA as the 373 Limestone Caribbees in the north (b) and the buried OA in the south (c). The black vertical 374 dotted lines with black horizontal bars indicate the Moho depth estimates based on Melekhova et 375 al. (2019, thick bars), and Schlaphorst et al. (2018, thin bars) at stations where the former study 376 did not provide results (Saba) or where the difference between both studies is greater than 1 km 377 (St. Eustatius, Dominica). The approximate location of two major subducted fracture zones (see 378 Figs. 1, 9) is bounded by vertical white dotted lines; uncertainties arise due to the oblique 379 subduction angle. Lack of data results in the grey-shaded black under the western part of the 380 crosscutting transects. The two along-arc b-value profiles show two different seismic catalogues 381 (OVS – Observatoires Volcanologiques et Sismologiques of the Institut de Physique du Globe de 382 Paris; SRC – Seismic Research Centre of The University of the West Indies) and are taken from 383 Schlaphorst et al. (2016), see that study for further calculation details. 384

385 Individual dispersion curves and shear velocity depth-velocity profiles (Fig. 8a,b) are highly variable depending on location. Below the arc, the shear-wave velocity generally 386 387 increases from 2.5 km/s at depths shallower than 5 km to values around 4.5 km/s at depth of 25 -35 km. In contrast, below the fore- and back-arc, the velocities are slower velocities down to a 388 depth of approximately 15 km. Below this depth however, velocities in the back-arc exceed the 389 arc values, already reaching 4.5 km/s, the average value at the bottom of the model (150 km), at 390 391 this depth in the southern back-arc. The pattern changes in the Mona Passage, where the shape of the velocity increase is more gradual throughout. However, the fact that the corresponding 392 dispersion curve does not match the phase velocities within error, likely disturbed at the short-393 period end (14 s), points to greater and perhaps unresolved complexity in the region (Fig. 8a). 394

Shear velocity depth slices (Fig. 9), and along- and across-arc transects (Fig. 10) show
significant changes in the crust and upper mantle. The scale used (red – light blue and afterwards
yellow – green – dark blue) indicates the values typically associated with crustal (up to 3.9 km/s)
and mantle (greater than 4.0 km/s) shear-wave velocity values with a light yellow layer in
between to make the separation visible.

400 The crust is thicker beneath the GA and the LAA and thinner in the fore- and back-arc regions (Fig. 9b). Furthermore, low-velocity anomalies at 20 km can also be found beneath the 401 Aves Ridge, the remnant volcanic arc to the west of the back-arc region behind the Grenada 402 Basin (see Fig. 1). In the southern basin, shear-wave velocity values are among the lowest of the 403 404 entire study area (down to 2.58 ± 0.07 km/s) at 10 km depth but increase rapidly with depth to values among the highest at 20 km depth (up to 4.85 ± 0.04 km/s), also observable in the 405 406 individual profiles (Fig. 8b). Around 30 km depth, the GA and the northern LAA show significantly lower velocity values compared to the southern LAA (4.0 - 4.2 km/s against 4.1 - 4.2 km/s407

408 4.5 km/s; Fig. 9c). At greater depths, the area around the central LAA (roughly Guadeloupe to
 409 Martinique) shows decreased velocities to the surrounding LAA (Fig. 9d–f).

This north-south division of the LAA is also apparent in the along-arc transect, with the island of Dominica roughly marking the change (Fig. 10d). The arc crust is on average 5 km thinner in the south (20 - 25 km vs 25 - 30 km). At 10 km depth, lower velocities can be found beneath the islands in the north (Saba, St. Eustatius, St. Kitts) and the south (Grenada). In contrast, the lowest velocities at around 30 to 40 km depth can be found beneath the islands towards the centre of the arc (especially Dominica, but also Guadeloupe; Fig. 10d).

Transects crosscutting the LAA (Fig. 10b,c) generally show thicker crust beneath the arc. In the north around Montserrat the thickened crust persists many tens of kilometres into the foreand back-arc (Fig. 10b), whereas in the central arc around Martinique the thinning occurs more abruptly (Fig. 10c). There, a low velocity region ($v_s < 4.3$ km/s) is apparent below 70 km creating the mantle wedge. In the cross sections the majority of crustal seismicity above the slab matches the crustal thickness derived from the tomography values (Fig. 10b,c). Furthermore,

towards the forearc seismic velocities at depths between 30 km and 55 km are lower in the

423 central arc (4.2 - 4.4 km/s) than in the northern part (4.3 - 4.5 km/s).

424 In the eastern part of the GA (Fig. 10a) the depth-velocity profile resembles that in the northern

LAA. Mantle values around 40 km depth are higher beneath Puerto Rico, comparable to the

northern and southern LAA, whereas to the west (Mona Passage, Hispaniola) and the east

427 (Virgin Islands) the subsurface structure profile resembles more the pattern observed in the

428 central part of the LAA.



Figure 11: see next page for caption.

433 Figure 11: Equivelocity surfaces to map sediment layer depth and Moho depth. a) Depth of 2.0

434 *km/s surface as an estimate of sediment limit. Circles represent depth estimates from previous*

- 435 study (Chichester et al., 2020), the size represent quality of the results. b) Pointwise comparison
- 436 between previous and current studies, also showing the correlation coefficient ($r_{2.0}$). c) Depth of
- 437 4.0 km/s surface as an estimate of Moho depth. Diamonds, circles and squares represent depths
- 438 from previous studies (Schlaphorst et al., 2018; Melekhova et al., 2019). d) Pointwise
- 439 comparison between previous and current studies, also showing the correlation coefficient ($r_{4.0}$).
- 440 Note that for two measurements Moho depth estimates between both previous studies differ
- 441 significantly. In those cases both symbols are plotted and connected with grey hashed line. The
- 442 *depths of the 4.3 km/s surface at the same points are indicated by smaller blue symbols,*
- 443 connected by blue dotted lines and their correlation coefficient is shown in blue ($r_{4.3}$). e)
- 444 *Crystalline crustal thickness, calculated from difference between depths of 4.0 km/s and 2.0 km/s* 445 *surfaces.*

Our result may also be used to specifically map sediment and crustal thickness (Fig. 11). 446 We choose average proxy values that are typically associated with the base of the sediment layer 447 (2.0 km/s; Fig. 11a) and the location of the Moho (4.0 km/s; Fig. 11c). We acknowledge that 448 449 slightly different choices, such as gradients or other nominal velocity values are possible, but our choices compare well with estimates from previous studies of sediment thickness (Chichester et 450 al. 2020; circles in Figs. 11a,b) and the Moho depth (Schlaphorst et al., 2018; Melekhova et al., 451 2019; symbols in Fig. 11c,d). Resolution in the shallowest parts of the model is diminished (Fig. 452 453 8c) and may partially explain the good correlation with sediment structure that was an input of the starting model. However, we do observe changes from the start model at these depths that 454 improve the fit to the data. 455

The depth to the 2.0 km/s velocity contour shows thick sedimentary loads (Fig. 11a) in 456 the area around the southern LAA in the fore-arc (Tobago Basin) and back-arc regions (Grenada 457 Basin) with values exceeding 7 km. In the back-arc the extent of the deepened 2.0 km/s velocity 458 contour in the Grenada Basin can be observed as far north as the region west of Guadeloupe. 459 460 There however, the sedimentary thickness decreases to values of around 4 km. The northern part of the LAA, as well as the entire GA show thinner sedimentary thicknesses, generally not 461 exceeding 2 km. Thinner sedimentary thicknesses of less than 1 km are observed right on the 462 Aves Ridge, the LAA and the continuation into the GA. 463

The depth to the 4.0km/s velocity contour shows the deepest values on the LAA and the GA (Fig. 11c) with values exceeding 25 km. The LAA shows consistently deeper values in the northern part (north of Dominica). Together with Hispaniola and the Mona Passage, the depths can be observed to reach 30 km there. In contrast, the fore- and back-arc basins show significantly shallower depths at values below 15 km, especially in the southern and central Grenada Basin.

Differencing the 4.0 km/s and the 2.0 km/s surfaces, we can calculate an estimate for the crustal thickness (Fig. 11e). The resulting map clearly shows the thicker crustal areas of the LAA, the GA and the Aves Ridge and the thinner crustal areas with minimum values reaching 10 km in the fore- and back-arc with the Grenada Basin again being the most pronounced. 474 Maximum values reach 30 km in the northern LAA and parts of the GA and are slightly lower at 475 around 25 km in the southern LAA and the Aves Ridge.

476 **5 Discussion**

The shear-wave velocity structures of the eastern GA and the LAA observed in this study reveal the variations in sediment and crustal thickness across the region. These results yield insights into the construction of the arc and adjacent basins, especially its tectonic history in terms of arc jumps and the influence of fracture zones and sediment subduction on the along-arc heterogeneity.

The sediment thickness we infer from the 2 km/s contour map in this study is thicker in 482 the south in both the Grenada Basin and Tobago Basin near the South American Margin (Fig. 483 11a). The thickness is in general agreement with point measurements derived from scattered 484 phases and refraction work in the region (Allen et al., 2019; Chichester et al., 2020; Padron et al., 485 2020), showing a correlation coefficient of r = 0.80, which gives confidence in the choice of this 486 velocity as our estimated proxy (Fig. 11b). The thickening of sediment load is consistent with 487 high sediment output from the Orinoco river and active back-arc spreading in the Paleogene 488 (Allen et al., 2019). 489

490 In a similar way, there is general agreement between the 4.0 km/s contour depth and the 491 crustal thickness values derived from receiver function studies and petrological constraints (Fig. 11c; Schlaphorst et al., 2018; Melekhova et al., 2019). The southern sub-arc crust is thinner than 492 493 that in the north in all models (Melekhova et al. 2019; Schlaphorst et al., 2018). The trend in our result is more pronounced than that of the receiver function study, and the absolute depths of our 494 study, based on the 4.0 km/s contour, is smaller than those from receiver functions (Fig. 11d). 495 496 This is likely explained by the difference in calculation method. Receiver functions are based on 497 relative velocity contrasts at a single point measured with phases with almost vertical incidence angles, whereas ambient noise tomography is averaged over a larger inter-station area, thus 498 499 having broader lateral sensitivity. In addition, the absolute depth of the surface wave derived Moho is not well resolved owing to trade-off with velocities above and below the contour. Still, 500 501 the correlation coefficient of r = 0.63 hints to a positive correlation. We also compare previous results to the 4.3 km/s contour, which increases our inferred Moho depth, but also decreases the 502 correlation coefficient (Fig. 11d). In addition, the thicker crust in the Aves Ridge and thinner 503 crust in the Grenada and Tobago Basins (22 km vs < 10 km) agree with values found by an 504 505 active source wide angle seismic profile experiment across the region (Christeson et al., 2008).

The variations in crustal thickness are likely related to the complex tectonic setting with 506 multiple subduction fronts at the edges of the microplates in the GA, with active and relict parts 507 of the arc in near proximity (Boschman et al, 2014). Thicker crust beneath the LAA and GA arc 508 platforms, in comparison to areas surrounding the arc, is consistent with the notion of magmatic 509 addition to the top and base of the Caribbean crust due to arc volcanism (Fig. 11c,e). Variation 510 within the arc platform, e.g. the northern LAA is thicker than the Southern LAA, is likely due to 511 512 the presence of the OA, which was active 55 to 30 Ma when the Grenada Basin was opening in the south (Allen et al., 2019). In other words, the arc platform in the northern LAA is thicker due 513 to more sustained, but laterally varying arc volcanism (Wadge, 1984). The crystalline crustal 514 thickness inferred by differencing the depth between the 2.0 km/s and 4.0 km/s contour in the 515

southern Grenada and Tobago Basins implies southward thinning to < 8 km (Fig 11e). This is

517 consistent with past back-arc extension until ~8 Ma (Allen et al., 2019). The northern part of the

518 Grenada Basin appears to be made of stretched arc material with no back arc spreading, leading

to the bulbous shape of the OA in the south LAA and larger distances between the LAA and the

520 OA. This pinning of the arc position in the north was due to differential boundary forces with the

521 American plates (Figs. 1,10B,C; Allen et al., 2019; Padron et al., 2020).

Velocity variations at 30 to 60 km depth, which we interpret to be the upper mantle, are in general agreement with previous results. The pronounced section of low velocity (< 4.4 km/s) beneath Guadeloupe to Martinique is consistent with noise tomography using the permanent land stations (Arnaiz-Rodríguez et al., 2020) and also the teleseismic study using the same data set used here (Cooper et al., 2020).

Variations in upper plate mantle velocities also reveal additional insight into the 527 dynamics of the system and in particular the relationship between the wedge and arc volcanism. 528 Slow velocities beneath the north-central arc (Guadeloupe, Dominica and Martinique) at 30 to 60 529 530 km depth, and to a lesser extent beneath the Grenadines and Grenada at 30 to 40 km are all located in regions of thickened crust, interpreted based on the 4.0 km/s velocity contour. These 531 anomalies could be caused by upwelling of hot mantle material and/or a small amount of partial 532 melt and fluids, potentially caused by flux melting and/or decompression melting. The along-arc 533 heterogeneities are observed on small length scales (tens of kilometres), which is also observed 534 in changes of characteristics such as mantle xenolith mineral composition (Melekhova et al., 535 2019). 536

The low-velocity zone in the mantle wedge beneath Guadeloupe and Martinique is 537 bounded by two major fracture zones to the north, which also lie near the boundary of the proto-538 Caribbean/equatorial Atlantic seafloor boundary (Cooper et al., 2020; Harmon et al., 2019). The 539 b-values of slab earthquakes (Schlaphorst et al., 2016) are also higher in the forearc offshore of 540 Dominica and Martinique, in close proximity to our seismic anomaly, suggesting greater fluid 541 release in this part of the subduction zone. The higher *b*-values also coincide with a region of 542 lower seismic velocity on the slab from local earthquake and active source tomography (Paulatto 543 et al., 2017). The high δ^{11} B anomaly in erupted magmas from Guadeloupe and Dominica, 544 indicative of dewatering of serpentinized oceanic crust, occurs just north of that region, closer to 545 the fracture zones on the slab in the north that bisect Guadeloupe and Martinique (Cooper et al., 546 547 2020). In addition, pre-existing faults from the formation at the ridge can open via bending during the subduction process, thus allowing for further hydration. However, the lack of 548 intraslab, trench-parallel normal fault focal mechanisms from earthquakes in the forearc and 549 outer rise makes this process unlikely to have a dominant role. An alternative explanation is that 550 there is three-dimensional flow owing to the curvature of the arc or gaps in the slab (e.g., Harris 551 et al., 2018; Schlaphorst et al., 2017), which can have an effect on the lateral placement and 552 extension of the anomaly to the south of the fracture zones. Similarly, ponding melt could 553 redistribute fluids and/or melt between the slab and the arc (Ha et al., 2020). Linked to that, the 554 most productive volcanoes over the last 100 kyr can be found on Dominica (Wadge, 1984). 555 Another possibility is that along-strike variability in the morphology of the subducted seafloor 556 itself can have an impact on the velocity structure of the mantle wedge and the upper plate. For 557 instance, oceanic core complexes, that are formed in slow- and intermediate-spreading 558 environments, at times when magma supply is limited (Smith et al., 2008) may cause crustal 559

alteration, based on slow seismic velocities outboard on the incoming plate of our study area

561 (Davy et al., 2019).

562 Further north, the lower velocities throughout the upper mantle between Hispaniola and 563 Puerto Rico (Fig. 10a) occur in a region with moderate interpreted crustal thickness (close to 30 564 km). The anomaly is interesting and potentially different to the rest of the arc where low upper 565 mantle velocities are associated with thickened crust and an active volcanic arc. This anomaly 566 could arise due to the ongoing opening of the Mona Passage (Jansma et al., 2005). In other 567 words, the extensional tectonics may cause upwelling of hotter, potentially more hydrated mantle 568 material from depth, possibly also associated with decompression melting.

569 6 Conclusions

In this study we used ambient noise tomography to map crustal and upper mantle shear-570 wave velocity of the eastern GA and the LAA using data from permanent land stations and from 571 ocean bottom seismometers collected as part of VoiLA experiment in 2016/17. We observe 572 significant along-arc heterogeneities on small length scales (~50 km) that agree with 573 heterogeneities of a variety of characteristics such as mantle xenolith mineral composition and 574 discontinuity structure (including crustal thickness) variation, changes in sediment and water 575 subduction, as well as melt generation distribution observed by previous studies. Our estimates 576 of sedimentary basin and crustal thickness agree with point measurements and profiles 577 throughout the LAA from previous modelling of receiver functions, petrology, and active source 578 studies. We observe increased sediment thickness (> 7 km) in the southern fore- (Tobago Basin) 579 and back-arc regions (Grenada Basin). This can be explained by sediment supply from the 580 Orinoco and active back-arc spreading in the Paleogene. The crust is thicker beneath the GA and 581 the northern LAA islands, reaching values of around 30 km. The LAA south of Dominica shows 582 smaller crustal thickness values of 20 to 25 km. In the fore-arc, where the older arc has been 583 buried under the thick Barbados sediments (Allen et al., 2019), and in the back-arc regions in the 584 Grenada Basin, crustal thickness are as low as 10 km. Along the arc, shear-wave velocities in the 585 upper mantle are significantly lower beneath the Mona Passage, as well as beneath the central 586 LAA from Guadeloupe to Martinique (4.2 - 4.4 km/s). This can be explained by upwelling of 587 hot mantle material or a small amount of partial melt, or both, and is likely caused by flux or 588 decompression melting or a combination of these. The subduction of the proto-589 Caribbean/equatorial Atlantic seafloor boundary at two major fracture zones are located just 590 north of the anomaly that runs from Guadeloupe to Martinique. Therefore, three dimensional 591 flow because of slab curvature and/or a gap in the slab, seafloor morphology such as oceanic 592 core complexes, and/or melt ponding with potential influences from the overriding plate (or a 593 combination of those) are likely to have a dominant impact on the location and degree of arc 594 volcanism in the area. In general, slow spreading ridges, in our case the MAR, lead to high 595 amount of hydration with high along-arc variability. This results in the slow subduction of highly 596 597 heterogeneous slab hydration patterns that facilitate the variable crustal and upper mantle structure we observed. 598

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- 615 <u>516a01b6af3f</u>), PR (<u>https://doi.org/10.7914/SN/PR</u>), TR (No Doi), WI
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