This document contains further information on the acquisition and processing of geophysical data as well as the geochemical/geochronology studies discussed in the paper “The role of arc migration in the development of the Lesser Antilles: A new tectonic model for the Cenozoic evolution of the eastern Caribbean” by Allen et al..

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1. MAGNETICS

Our new magnetic anomaly grid is created from a dataset of 335 cruises covering the time period from 1960-present. A region much larger than the target study area of the Lesser Antilles was chosen in order to maximise the treatment of historical cruises which transit
across the study area. By including comparatively simple, well studied areas of oceanic crust (e.g. Atlantic crust of the North American plate) with well documented magnetic anomalies we are able to provide a qualitative benchmark for the data quality of the grid as a whole. The accurate capture of seafloor spreading anomalies (Figure 2) in our grid allows us to interpret similar scale magnetic features in the more tectonically complex Caribbean interior region with confidence.

An initial dataset of 347 cruises was compiled across the region. The bulk of these ship tracks (327) came from the compilation of Quesnel et al. (2009), a cleaned version of NGDC (National Geophysical Data Centre) magnetic survey data supplemented with a further 18 IFREMER cruises requested from the SeaDataNet database and two cruises (JC133 and JC149) from our own VoiLA (Volatile Recycling in the Lesser Antilles) research project. These additional cruises were independently screened for data spikes and navigational errors. IFREMER cruise data was converted from MGD77 format to useable .gmt files with the MGD77 supplement to the generic mapping tools (Wessel and Chandler, 2007). Data coverage across the study area is generally excellent with the final grid containing more than 1.3 million individual magnetic anomaly points (Figure S1). Coverage is understandably poor in the very shallow water of the Bahamas Platform, South American passive margin and the Nicaragua Rise. There are also regions of comparatively low data density on the incoming plate to the north and northwest of the study area. However this has not had any great impact on our ability to resolve seafloor spreading anomalies in these areas.

Magnetic anomalies for all cruises were calculated by subtracting the IGRF12 global reference field. Cross-over analysis (Chandler and Wessel, 2012; Hsu, 1995; Wessel, 2010) was used to provide a quantitative assessment of overall dataset quality by comparing the differences between magnetic anomaly values at intersecting ship-tracks. Initial cross-over
analysis of all cruises gave a mean cross-over error of 46.0 nT (from 18207 ship track interactions).

Diurnal corrections were applied to all cruises based on data from the three nearest permanent geomagnetic observatories (TEO: Mexico, SJG: Puerto Rico, KOU: French Guiana, Figure S1). Data was downloaded from the INTERMAGNET online database. Annual mean diurnal variation curves, sampled at 1 minute intervals, were calculated for each observatory from all available data up to the present day (TEO: 10 years, SJG: 21 years, KOU: 21 years, Figure S2). These three curves were then used to apply a diurnal adjustment to each individual magnetic anomaly measurement, weighted depending on the linear distance from each of the 3 observatories. This correction only led to a relatively minor reduction in the whole dataset mean xover error to 45.5 nT.

This small improvement was not unexpected given (i) the relatively poor geographical coverage of geomagnetic observatories (ii) the limited recording window (the observatories are only active for about 40% of the time covered by the ship-tracks used and even within that contain extended periods of absent data), (iii) the annual variability of the Earth’s magnetic field measurement (curves for individual years regularly deviate from the mean by +/-20 nT, Figure S2), and (iv) the accuracy of applying this curve back through time. A regional levelling process was subsequently applied to the whole dataset in order to more accurately reduce the impact of these poorly captured magnetic field variations.

Before further processing the worst quality cruises (12 cruises, those with mean xover errors of greater than 100 nT) were removed, leaving a final dataset of 335 cruises. This reduced the mean xover error for the dataset to 41.7 (from 17811 interactions) with little impact on the data coverage.
The dataset was levelled regionally using the same methodology as Quesnel et al. (2009) to reduce the cross-over error between intersecting tracks. This process corrects the ship-tracks on a regional scale for non-geological variability resulting from factors such as positioning errors as well as outstanding temporal field variations.

In the first step of the levelling process the differences between a given magnetic measurement and all other readings within 10km (ignoring those along the same ship-track) were calculated. Each of those differences was then given a weight \( w \) by distance (equation 1, \( r \) is distance and \( r_0 \) is a constant with a value of 2 km) and summed to calculate an overall regional difference value for each reading, with the closest readings being the most heavily weighted. These difference values were then filtered along each ship-track using a 50 km Gaussian filter. This prevents the applied difference changing too rapidly along a single ship-track, for instance if it enters a region of particularly low data density, or is strongly influenced by a particularly anomalous reading(s) from another ship track. Filtered differences were then added back on to the original track measurements to give the final levelled magnetic anomaly value.

Equation 1:

\[
w = \left[ \frac{r_0^2}{(r_0^2 + r^2)} \right]^2 \quad (1)
\]

Following this process the absolute mean cross-over error was reduced to 22.5 nT (for the final total of 17811 cross-overs). This improvement in the cross-over error is clearly demonstrated in Figure S3.

The levelled data were gridded in GMT with a single cruise weighting applied based on the track xover errors such that the highest quality cruises would be counted most strongly in the final grid (Smith and Wessel, 1990). Our final marine magnetic chart for the region is
gridded at 2 arc minutes (3.5 km x 3.7 km assuming 15 degrees latitude i.e. ~13 km²). The
final processing step was a Reduction to Pole carried out in GMT using a window of 0.5
degrees and a filter of 25 coefficients (Figure S4). Inclination and Declination were taken
from the IGRF12 grid.

2. SEISMIC

a) Introduction

The seismic data used for this study was acquired by the R.R.S. James Cook in the period
April-May 2017. The survey consisted of a series of 8 MCS and Wide-angle seismic profiles
largely orientated (N-S) throughout the Grenada back-arc Basin (Figure S5). The results from
profiles 4, 6 and 5+8 are given here. The data were acquired using a 13 gun seismic array of
nominal 4800 cu in (79 L) volume and 2000 psi (14 MPa) pressure and towed streamer (3000
m long for profiles 4 and 5, 300 m long for profiles 6 and 8). All data acquisition was done at
a speed of 5 knots with a 60 second shot interval leading to an approximate shot spacing of
150m. Shots were recorded by a network of 34 OBS stations (10 Scripps, 24 DEPAS
instruments) throughout the Antilles region. Note that there are two breaks in the R.R.S.
James Cook shooting tracks due to permitting issues. This was partly compensated by the
recording of shots during the GARANTI experiment by our seabed network. These shots
(profile gar1 – see figure S5) have been used to supplement our own set of picks for the wide-
angle modelling along profiles 5 and 8.

The location of the new profiles relative to previous seismic experiments in the area is
shown in figure S5. Profile 4, in the Southern Grenada Basin, overlaps with previous studies
conducted by Christeson et al. (2008) and Boynton et al. (1979) with the more recent study
finding oceanic like crust underlying the southern Grenada Basin. In comparison the northern
Grenada Basin is much less well studied, with Kopp et al. (2011), off Guadeloupe, the only
previous active source seismic experiment close to this region of the back arc.
b) MCS Processing

Individual profiles were processed using a standardised processing flow. After geometry assignment all traces were first filtered using an Ormsby bandpass filter (3-8-100-200 Hz) for noise reduction. A minimum phase predictive deconvolution with a 200 ms operator was then applied. A spherical divergence correction with a time-power constant of 2.3 (i.e. t*2.3) was applied prior to NMO (Normal Moveout Correction). NMO semblance velocity analysis was used for the lines which were shot using the long (3000 m) streamer (lines 4 & 5). For processing of the short (300 m) streamer lines (6 and 8) a constant velocity depth function was used (1480 m/s to 2200 ms, 1500 m/s to 3500 ms, 2000 ms^-1 to end). Traces were summed using a mean method to produce the final stacks.

c) Wide Angle Modelling

Most wide-angle OBS stations recorded arrivals at distances of up to 80km with the longest picked offsets coming from >180 km.

As the acquisition geometry was not a purpose designed refraction survey the OBS do not lie directly along acquisition lines. For modelling therefore stations were projected onto the nearest point along the shot line. Shots locations were then relocated by the same amount along the line, to maintain the source-receiver offset. The biggest such relocation used in the study was ~15 km. As a result the models produced are not a perfect representation of the crust along the line, but should instead be thought of as regional “type” lines, representative of an average crustal structure within the vicinity of the line. Occasionally two stations would project onto similar places despite being separated geographically by several 10s of km. In some cases this made it impossible to achieve a working model due to significant changes in crustal structure outside of the plane of the profile (e.g. stations 3 and 14, which are separated by 60 km geographically, but project within ~10 km of each other along profile 4). In such an
event the closer station was chosen (unless distances were similar and one station possessed notably higher data quality).

All profiles were modelled as a 2D profiles in the ray-tracing software RAYINVR (Zelt and Smith, 1992), which uses a Runge-Kutta method to solve the ray tracing equations in 2D. Lines 5 and 8 were modelled as a continuous combined 2D profile (this combined profile is referred to as line 58 from here onwards) using arrivals from 6 stations (18, 19, 21, 23, 26, 30). Shots from the French R/V L’Atalante cruise line gar-l were used to further supplement the pick set in this area and contributed greatly to bridging the gap between lines 5 and 8. However, basement depth is poorly constrained in the ~90 km gap between the two lines where no MCS data was acquired. Lines 4 and 6 were modelled as stand-alone velocity profiles using picked arrivals from 3 OBS (1, 2 and 3) and 5 OBS (24, 25, 26, 27, 28) respectively. For lines 4 (235 km) and 58 (325 km) velocity nodes were 10 km apart. For line 6 which is shorter (135 km) a 5 km spacing was used.

Seafloor and top basement surfaces were picked from the MCS data for each of the profiles. In some regions the presence of a strong seafloor multiple made the top basement hard to interpret. These picks were modelled as reflected phases at 10 km intervals along each of the profiles. A water column velocity of 1.52 km s\(^{-1}\) was assumed for all lines, while sediment velocities were determined by modelling of picked arrivals from the sediment column as turning rays and constrained by the top basement travel time picks from the reflection profiles.

In the southern part of the Grenada Basin, there is very little seafloor topography and a relatively flat top-basement surface. Clear Pg and Pn arrivals along with reflected phases from the top basement were identified (Figure S6A-C). The thick sedimentary section in the southern basin also meant that turning rays in the deepest sedimentary package were
recognisable to exceedingly long travel times (almost 20 s) and offsets (60 km). On all 3 instruments there is a clear ~8 km s$^{-1}$ (Pn) refractor that is clearly the result of the back-arc Moho.

In the northern region of the survey, where the seafloor and crustal topography are more heterogeneous it proved difficult to pick consistent reflected phases (Figure S6A). As a result the velocity models for lines 58 and 6 are entirely built from P-wave first arrival picks separated into distinct upper and lower crustal units. The very thick crustal section (up to 30 km in some places) limited the number of recorded arrivals from below the Moho, which is poorly constrained compared to the southern profile, with arrivals from the mantle only appearing at comparatively long offsets of over 100 km.

The final models shown were produced using the software RAYINVR (Zelt and Smith, 1992) through a process of iterative forward modelling until a best overall model fit was achieved, whilst trying to avoid the addition of very extreme lateral velocity variations at depth. Starting models were based on the few existing studies of the Grenada Basin region, with a typical oceanic crustal structure in the southern basin (Christeson et al., 2008, Figure S5C) and a thick 2 layer crustal structure in the northern arc, based on the outcomes of the studies of Kopp et al. (2011, Figure S5B) and Sevilla et al. (2010). The thick sediments in the southern Grenada Basin were modelled as 3 layers based on the interpretation of Aitken et al. (2011). The final models shown in Figure S8 (Main text Figure 4) have normalized chi-squared values of 0.720 (Line 4 - RMS: 0.085 s), 1.421 (Line 58 - RMS: 0.119 s) and 1.427 (Line 6 - RMS: 0.119 s) respectively, with all picks assigned an uncertainty of 0.1 s.

3. TIMING OF IGNEOUS ACTIVITY IN THE ANTILLES REGION
The new tectonic model for the Lesser Antilles region was formulated within the known constraints from the ages of igneous and volcanogenic rocks in the region and their tectono-magmatic affinity. These constraints are summarised in Figures S8, S9 and Table A1.

a) Late Cretaceous to mid-Paleocene: GAC and Aves Ridge

Records of magmatism associated with the Great Arc of the Caribbean (GAC) can be found across the Aruba-Blanquilla archipelago, extending to the south-eastern part of the Aves Ridge. The volcaniclastic and intrusive rocks of Bonaire (c. 98 - 88 Ma) have long been recognised as an island arc suite related to the progressive obduction of the GAC across the South American continent during the Cretaceous (Thompson et al., 2004; Wright and Wyld, 2011).

During the late Cretaceous and Paleocene, widespread, subduction-related crustal melting resulted in the intrusion of adakitic granitoids throughout the Dutch and Venezuelan Antilles (Neill et al., 2011). On Aruba and Curacao, mafic and volcaniclastic rocks associated with the Caribbean Large Igneous Province (CLIP) are intruded by late Cretaceous (c. 89 – 86 Ma) diorites and quartz-diorite dykes and plutons (Sinton et al., 1998; White et al., 1999). Paleocene quartz-diorites, pegmatites and aplites also intrude CLIP basement on Grand Roque, (c. 68 – 64 Ma) (Thompson et al., 2004). Additional tonalites, trondhjemites and aplites of late Cretaceous to Paleocene age (c. 74 – 59 Ma) are found on the island of La Blanquilla and were also recovered from dredge hauls along the Aves Ridge (Neill et al., 2011; Wright and Wyld, 2011). Similar granitoid complexes intrude the metamorphic basement on the island of La Orchila (~60km east of Gran Roque), but are undated. Further eastward; tonalites, trondhjemites, pegmatites and sodic granites are also found on Margarita (c. 76 Ma) and the neighbouring islands of Los Hermanos and Los Testigos. A small, anomalous exposure of undated tholeiitic basalts and gabbros is found on the island of Los Frailes, 5 km to the northeast of Margarita (Santamaria and Schubert, 1974). The
emplacement of the igneous rocks on the islands of Gran Roque to Margarita are likely contemporaneous with back-arc spreading in the eastern VB, the result of the eastward migration of the GAC/Aves Ridge system.

Currently, there is no record of subduction-related magmatism younger than c. 59 Ma in the Dutch or Venezuelan Antilles. The exact timing of abandonment of the Aves Ridge system is difficult to constrain due to the lack of exposure and more detailed geochronological evidence but is likely represented by the youngest granitoid rocks of Gran Roque, La Blanquilla, Margarita and Los Testigo. It may be no coincidence that this is also the estimated date of the collision of the GAC with the Bahamas Bank. It is clear, however, that a drastic change in the composition of igneous rocks in the mid-Paleogene, from mostly adakitic TTG (tonalite-trondhjemite-granodiorite) suites to dominantly sub-marine basaltic and mafic volcaniclastic rocks (see section 3b), must reflect a fundamental change in the tectono-magmatic environment during this time period (Fig. S8). In our new model we use these lines of evidence, together with the magnetic anomaly results, to infer a >300 km eastward shift of the subduction axis during the Paleogene.

**b) Eocene to early Miocene: Outer Arc**

Subduction magmatism in the north-eastern Caribbean was fairly continuous during the Paleogene. The locus of activity migrated steadily from west to east, from Puerto Rico though to the Virgin Islands and then the Limestone Caribbees, as the obliquity of subduction increased following the collision of the GAC with the Bahamas bank. In our model, the gap in records of subduction magmatism in the southern LAA from c. 59 Ma is attributed to the onset of back-arc spreading related to the Outer Arc.
There is direct evidence of renewed subduction-related magmatism in the Lesser Antilles from islands of the Limestone Caribbees, situated to the northeast of the modern LAA (Davidson et al., 1993). On St. Barthelemy, St. Martin and Antigua, upper-Eocene to Oligocene (c. 38 – 22 Ma) sub-marine and sub-aerial volcanic rocks with associated littoral carbonates and hypabyssal plutons are exposed (Andreieff et al., 1988; Weiss, 1994; Westercamp et al., 1985). An older series of early to mid-Eocene (c. 57-40 Ma) intercalated intermediate to felsic volcanogenic strata and pelagic sediments is exposed beneath the upper Eocene series on St. Martin (Andreieff et al., 1988). Given its nearly continuous stratigraphic relationship with the over-riding mid-Eocene arc-related sequence and mineralogical composition, it is also likely to be arc related. To the west, Paleocene igneous rocks (c. 66 – 64 Ma) capped by Eocene limestones were recovered from a drill core off the Saba Bank and Anegada Passage (Warner, 1990), and various early Paleogene to Eocene subduction-related igneous rocks are found throughout the US and British Virgin Islands in the Greater Antilles (Jolly et al., 1998; Lidiak and Jolly, 1998). To the south, there are no ubiquitous exposures of Paleogene igneous rocks along our newly recognised Outer Arc. However, a continuous sequence of regularly interbedded feldspar-quartz-mica bearing volcanioclastic tuffs and hemipelagic sediments of Upper Eocene to Oligocene age are found within the “Oceanic Formation” of Barbados (Saunders et al., 1984; Senn, 1940). The origin of these volcanic rocks is ambiguous but does provide evidence of magmatic activity in the region of the outer arc during the Paleogene.

Jurassic to Cretaceous (c. 145 Ma) rocks believed to have formed as part of the GAC arc and back-arc can also be found much further east, in the Lesser Antilles arc. These exhumed metavolcanics rocks form the basement of La Desirade (Ld) in the central arc segment and on Tobago (Tb) in the south (Corsini et al., 2011; Neill et al., 2010).
tectonic model presented here, with a previously unrecognised Outer Arc, provides a mechanism for the emplacement of these rocks east of the currently active LAA.

Indicators of Paleogene magmatic activity can also be found throughout the southern Lesser Antilles islands however, it is contested whether any of these are indicative of arc magmatism at this time, with most samples demonstrating a back-arc affinity. In the Grenadines, Paleocene to mid-Eocene xenocrystic and detrital zircons (c. 60 – 40 Ma) have been recovered from young igneous rocks of the Tufton Hall Formation, Belvedere Formation and beach sand samples from Grenada and Carriacou (Rojas-Agramonte et al., 2017). On Mayreau, sub-marine pillow basalts and volcanogenic sediments of the Mayreau Basalt are capped by mid-Eocene pelagic limestones, mafic volcanogenic sediments and minor pillow basalts of the Anse Bandeau Formation (c. 50 – 46 Ma, P10 – P11 zonal range). Similar stratigraphy is observed on the neighbouring island of Carriacou, where mid-Eocene sub-marine pillow basalts, intercalated with minor volcanogenic sediment and pelagic limestones of the Cherry Hill Basalt (c. 46 – 45 Ma, P12 zonal range), are capped by foraminiferal and radiolarian carbonates of the Bogles Limestone. Pillow basalts found on Mayreau and Carriacou have similar morphologies, mineralogical compositions, and stratigraphic relationships, and are suggested to be correlative (Speed et al., 1993). Off the east coast of Mayreau, mid-Eocene volcanogenic strata (c. 43 – 46 Ma, NP16 zonal range) are exposed on the islands of Jamesby and Baradel (Westercamp et al., 1985). Around 180 km to the west of Grenada, an off-shore drill core near Los Testigos intersected mid-Eocene mafic volcanogenic sediments (c. 39 – 35 Ma), that are unconformably overlain by lower Miocene limestones.

There is hence a near continual record of sub-marine basaltic magmatism and contemporaneous pelagic basin sedimentation from the Lower Eocene to the Oligocene in the southern Lesser Antilles, beginning roughly 10 Ma after the cessation of the Aves Ridge
system. According to our new tectonic model, these volcanics would have been erupted on the floor of the Grenada-Tobago back-arc basin during the formation of the Outer Arc. Later the westward step of this system into its current position in the Oligocene and growth of the LAA through the southern spreading centres bought them to the surface. The restriction of known samples of this phase of volcanism to the Grenadines is likely to be the result of a lack of exposure on the larger volcanic islands (e.g. St. Lucia and St. Vincent) due to almost continuous arc magmatism from the Miocene to the present-day, combined with the inaccessibility of large areas of these islands.

The Mayreau Basalt and basaltic lavas within overlying Anse de Bandeau Formation (Mayreau) were demonstrated to have elevated whole-rock TiO\textsubscript{2} and Zr concentrations, and Ce/Zr ratios slightly higher than average MORB values (Speed and Walker, 1991). Furthermore, these lavas were also shown to have a clear MORB Nd isotopic compositions, a depletion in LREE elements, concomitant Nb and Tb depletions and elevated radiogenic Sr and Pb isotopic compositions (White et al., 2017). They are also compositionally distinct from Paleogene igneous rocks in the Limestone Caribbees, and recent island arc tholeiites throughout the modern LAA. Such features are characteristic of a depleted, back-arc source region for these magmas e.g. Pearce and Parkinson (1993).

The Tufton Hall Formation (Grenada) and the Belvedere Formation (Carriacou) contain xenocrystic and detrital zircons of mid-Eocene age and are of a similar age to igneous rocks exposed on the proximal islands of Mayreau, Jamesby and Baradel. Furthermore, the rare-earth element (REE) composition of the mid-Eocene zircon population suggests that they formed from a magma that was significantly less oxidised than those found in the recently active southern Lesser Antilles arc (Rojas-Agramonte et al., 2017). This corroborates with a back-arc origin for the mid-Eocene magmas in the southern LAA, as opposed to a highly oxidised arc source. Moreover, because the constrained biostratigraphic depositional ages of
the Tufton Hall Formation and Belvedere Formation agree with detrital U-Pb zircon ages recovered from those same formations, sub-marine igneous activity and pelagic basin sedimentation appear to have occurred coevally during the mid-Eocene. The clear spatial and temporal correlation between sub-marine basalts and mafic volcaniclastic rocks throughout the Grenadines ultimately suggests they were derived from the same tectono-magmatic environment. There are no other igneous rocks of Eocene age found in the Caribbean region that were derived from a depleted, relatively reduced mantle source, and it is difficult to reconcile the geochronological, geochemical and stratigraphic constraints outlined above without invoking the presence of a Paleogene back-arc spreading centre where the currently active LAA resides, as has also been suggested by other authors (e.g. Speed and Walker, 1991; White et al., 2017). In our model, we suggest that this back-arc spreading centre is associated with the Outer Arc, and formed following the migration of the magmatic axis to the east of the Grenada and Tobago basins during the Paleogene

c) Neogene: Lesser Antilles Arc

According to our model, during the Upper Oligocene, the arc migrates westward into the back-arc region of the Eo-Oligocene Outer Arc. By c. 22 Ma, subduction-related magmatism is underway within the currently active LAA. As a result of activity on the more easterly Outer Arc (~40 - 25 Ma) the mantle wedge beneath the current southern LAA is likely to have experienced more prolonged metasomatism than would be predicted from its ~22 Ma volcanic history. This precursory flushing of fluids and melt through the mantle wedge has the capacity to significantly alter the composition of the ambient mantle below the modern LAA, and impart compositional heterogeneities in both the sub-arc mantle and crust here. This process could be responsible for some of the along arc magmatic compositional
variation observed in addition to the crustal contamination of arc magmas arising from the assimilation with Grenada/Tobago Basin sediments and back-arc crust in our model.

Studies rarely invoke the possibility of pre-existing compositional heterogeneities in the sub-arc mantle in the LAA, although many of them suggest that some of the isotopic variability must have been inherited from the mantle wedge (Bezard et al., 2015; Thirlwall and Graham, 1984). A new consideration of a previously unpredicted, pre-existing ambient mantle composition may prove essential to fully reconcile the observed trace element and isotopic variability in the LAA (e.g. Turner et al., 2017).

4. FIGURE CAPTIONS

Figures S1. Map of the ship track data used in construction of the final magnetic grid across the region. The locations of the three magnetic observatories used in the study are shown as green dots.

Figure S2. Diurnal variation curves for two of the magnetic observatories used in the study (SJG and KOU). Fine black lines show measurements made on the 1st January (UTC time) for each year for which data was available. The mean curve is shown in red. The large variation in the magnetic field about this mean, along with the absence of data in the early decades of the study explains the need for levelling of the full dataset.

Figure S3. A) Absolute external Xover errors (difference in magnetic field value) for intersecting ship-tracks prior to levelling. Only errors greater than 20 nT are shown. B) Data shown in A plotted as a histogram. C-D) As A-B following levelling.

Figure S4. Comparison of three magnetic anomaly grids. All grids are shown with the same illumination and contouring. A) The EMAG2 global magnetic grid (Maus et al., 2009) B) Magnetic anomaly grid from this study prior to application of reduction-to-pole. C) Final
magnetic anomaly grid from this study. Figure demonstrates the significantly lower
resolution of the dominantly satellite based grid compared to our result based wholly on
marine data.

Figure S5. A) Locations of VoiLA seismic lines (green) and previous seismic studies
(orange) in the study area. VoiLA OBS network is shown by red dots. Dashed green line
shows the section of profile 58 which is not constrained by reflection data. Seafloor depth
over this section was taken directly from regional bathymetry, whilst basement depth was just
extended linearly from the end of line 5 to the start of line 8 prior to forward modelling. The
key to the historical (orange) profiles is as follows. I = region of oceanic crust identified by
Diebold et al. (1981) in the south-eastern Venezuelan Basin, II - IV = wide angle profiles
from the BOLIVAR project described in Bezada et al. (2010), Clark et al. (2008) and
Christeson et al. (2008) respectively, V = profile gar1 from the GARANTI 2017 experiment
(Lebrun, pers. comm.). Shots from this profile were recorded on the VoiLA OBS network
and used to supplement wide angle modelling along profile 58. VI = wide angle profile
described in Kopp et al. (2011). VII = SISMANTILLES experiment (Evain et al., 2013;
Laigle et al., 2013). VIII = profile from Boynton et al. (1979). IX = Commercial multi-
channel seismic reflection data shown in Gomez et al. (2018). B) P-wave velocity model for
offshore Guadeloupe (historical profile VI) from Kopp et al. (2011). Star highlights fore-arc
high coincident with outer arc identified in magnetics. C) P-wave velocity model for southern
arc transect (historical profile IV) from Christeson et al. (2008).

Figure S6. Examples of six receiver gathers from broadband OBS used in the study. All
images show the vertical component (BHZ) of the instrument and have been reduced to 8
kms\(^{-1}\) such that mantle arrivals will appear flat. A) OBS26, B) OBS19, C) OBS23, receiver
gathers from OBS stations in the northern Grenada Basin. Only sediment (Ps) and crustal
arrivals (Pg) are apparent up to offsets of \(-100\) km. Thick crust in the region is indicated by
the very long offset of the mantle Pn arrivals (>100 km) D) OBS3, E) OBS2, F) OBS1, receiver gathers from stations in the southern Grenada Basin. Clear arrivals are shown from the sediment column, crust and mantle, as are top basement reflections. The first arrivals from the mantle are observed at a relatively short offset (50 km), owing to the thin crust in this region. Secondary sediment arrivals are visible to very long travel times due to the thick sediments of the southern Grenada Basin.

Figure S7. Final p-wave velocity models. Profile locations are given in figure S5 and figure 3 in the main text. All profiles are plotted to the same scale.

Figure S8. Age distribution of igneous activity in the eastern Caribbean from the Late Cretaceous to the present. Ages presented in the diagram have been derived from in situ rock samples, dredge hauls, drill cores and xenocrystic and detrital igneous zircons, using radiometric and biostratigraphic constraints. Samples with an arc origin are highlighted in green, those with a back-arc origin in grey and disputed tectonic affinity are highlighted in pink. Errors bars for radiometric ages are shown when uncertainties are ≥ 1 Ma. Dashed grey lines connect the lower and upper ages of roughly continuous magmatic activity. Individual data points are coloured yellow, blue or red according to the tectonic stage of our new model (see bar at bottom which is shown together with a magnetic polarity timescale for reference). It should be noted that the variable lower greenschist facies metamorphism of rocks found in the Dutch and Venezuelan Antilles, and common post-magmatic alteration of rocks in the Lesser Antilles means that K-Ar ages may not be an accurate representation of crystallisation age and must be interpreted with caution. For locations see Figure S9. Data referencing details are given in Table A1.

Figure S9. Map showing location of all islands discussed in supplementary section 3 and featuring in age distribution Figure S8.

5. TABLE CAPTIONS

Table A1. Referencing information by island for data shown in Figure S8.
6. REFERENCES


Neill, I., Gibbs, J. A., Hastie, A. R., and Kerr, A. C., 2010, Origin of the volcanic complexes of La Desirade, Lesser Antilles Implications for tectonic reconstruction of the Late Jurassic to


Robert Allen
Figure S1
Robert Allen
Figure S2
Robert Allen
Figure S3
Figure S4
VoILA Profiles 5 and 8 are merged to form a single 2D velocity model (S8). Dashed region contains no constraining reflection data.

In summary, the considerable volume of the lower crustal unit (if beneath what could have been originally an oceanic crust thickened by deep magmatic processes and underplating) may comprise segments of oceanic plateau material of thickened crust subductions. In addition, it is not focused underneath the volcanic centers but is imaged from the forearc to the backarc regions. This layer is interpreted as gabbroic plutons (6.8–7.2 km/s) overlying erupted production along the Lesser Antilles arc (Macdonald et al., 2000). The Izu-Bonin arc has a 12- to 18-km-thick oceanic crust thickened by deep magmatic processes and underplating, typical oceanic crust beneath the western Mariana trench system) display a distinct variability along the Mariana arc system. Indeed, the best-studied intra-oceanic subduction zones are the Izu-Bonin (26–32 km [Christensen and Mooney, 2004]) and Izu-Bonin (26–32 km [McLennan, 2002]) arcs.

The model checkerboard and a recovered inverted using the final velocity model as a starting model. The horizontal resolution at each node was obtained by calculating the minimum cell size at that point that could be focused under the island arc backstop and its role in a large accretionary system. J. Geophys. Res. 108, 2358, doi:10.1029/2002JB002040.

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Figure S5
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Figure S6
Northern Basin Profiles

A) Profile 58

C) Profile 4

Southern Basin Profile

B) Profile 6

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Figure S7
Figure S9

Aves Ridge

Dutch Antilles

Venezuelan Antilles

Virgin Islands

Limestone Caribbees

Northern Lesser Antilles

Southern Lesser Antilles

Outer Arc

5

2

4

3

1

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