

# Geological architecture and history of the Antigua volcano and carbonate platform: Was there an Oligo–Miocene lull in Lesser Antilles arc magmatism?

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## ABSTRACT

Since the acceptance of plate tectonics, the presence of calc-alkaline magmatic rocks has been recognized as evidence of subduction. But under specific geodynamic circumstances, subduction may occur without generating magmas. Here, we investigate the Cenozoic northern Lesser Antilles arc where, from sparsely exposed magmatic records, Eocene–Oligocene and Pliocene magmatic flare-ups and a Miocene lull were postulated. Nevertheless, most of the arc is submarine, so it is challenging to discern lulls and flare-ups from sampling bias. We review the magmatic evidence exposed onshore in the Lesser Antilles and investigate in detail the island of Antigua, which exposes an Eocene to Miocene volcanic sequence and platform carbonate series that coincide with the postulated lull. By combining lithostratigraphic analysis, structural mapping, <sup>40</sup>Ar/<sup>39</sup>Ar geochronology, and biostratigraphy, we refine the magmatic history of the island and date the arrest of extensive arc magmatism at 35 Ma, with minor activity until 27 Ma. No magmatic products are interleaved with the platform sequence until

the latest Oligocene, which confirms a lull in northern Lesser Antilles arc magmatism that may have lasted ca. 20 Ma. Flare-up of magmatic activity contributed to crustal thickening and land emersion, whereas magmatic lulls and related thermal cooling induced subsidence/submersion. Thus, we propose that the paleo-(bio)-geographical evolution of the eastern Caribbean region has been partly controlled by magmatic activity.

Fault kinematic analysis, along with anisotropy of magnetic susceptibility, suggest that, at the island scale, magmatic arrest is not associated with a change in stress field during the Oligocene. We speculate that slab flattening triggered by progressive curvature played a role in the temporal shutdown of the northern Lesser Antilles arc.

## INTRODUCTION

Active volcanic arcs and associated calc-alkaline magmatic rocks are recognized as closely associated with, and the result of, modern subduction zones (e.g., Tatsumi et al., 1983; Davies and Stevenson, 1992; Pearce and Peate, 1995). Based on that observation, geological records of volcanic arcs have become widely used as arguments for dating and reconstructing contemporaneous paleo-subduction zones (e.g., Li et al., 2008; Merdith et al., 2021; Cawood et al., 2009). But is the absence of arc magmatism evidence against subduction?

Worldwide, spatial and temporal lulls in magmatism are identified using the absence of calc-alkaline magmas in the rock record. Such gaps

are observed in the trans-Mexican belt (Ferrari et al., 2001, 2012), the Andes (Kay and Coira, 2009; Gutscher et al., 2000; Rosenbaum et al., 2021), the archetypal Cascade arc of the western USA (Glazner et al., 2004), the Nankai trough near Japan (Cross and Pilger, 1982; McGeary et al., 1985), Taiwan (Yang et al., 1996), Sumatra (Zhang et al., 2019), and Tibet (Ma et al., 2022). In the Western Alps, subduction of the Tethys Ocean did not produce long-lasting magmatic activity (e.g., McCarthy et al., 2020). Such paucity of magmatic evidence is typically explained by specific variations in geodynamical parameters that may affect the downgoing plate and the upper plate, such as flat-slab subduction (McGeary et al., 1985; van Hunen et al., 2002; Martinod et al., 2013; Zhang et al., 2019; Ma et al., 2022); subduction of hyperextended, dry mantle (McCarthy et al., 2020); or changes in convergence rates (Cross and Pilger, 1982), but not as an absence of subduction.

In the eastern Caribbean, the Lesser Antilles region hosts an arc for which magmatic flare-ups and lulls have been postulated (Briden et al., 1979; Bouysse and Westercamp, 1990). The Lesser Antilles arc could appear to be a textbook example of ocean–ocean subduction (e.g., Macdonald et al., 2000). Indeed, the N–S-trending Lesser Antilles trench accommodates 2 cm/yr of westward motion of North and South America relative to the overriding Caribbean plate (Pindell and Dewey, 1982; Boschman et al., 2014; Symithe et al., 2015; Fig. 1). Marine magnetic anomalies in the Cayman trough, which forms a pull-apart basin on the Caribbean–North American plate boundary, documents constant

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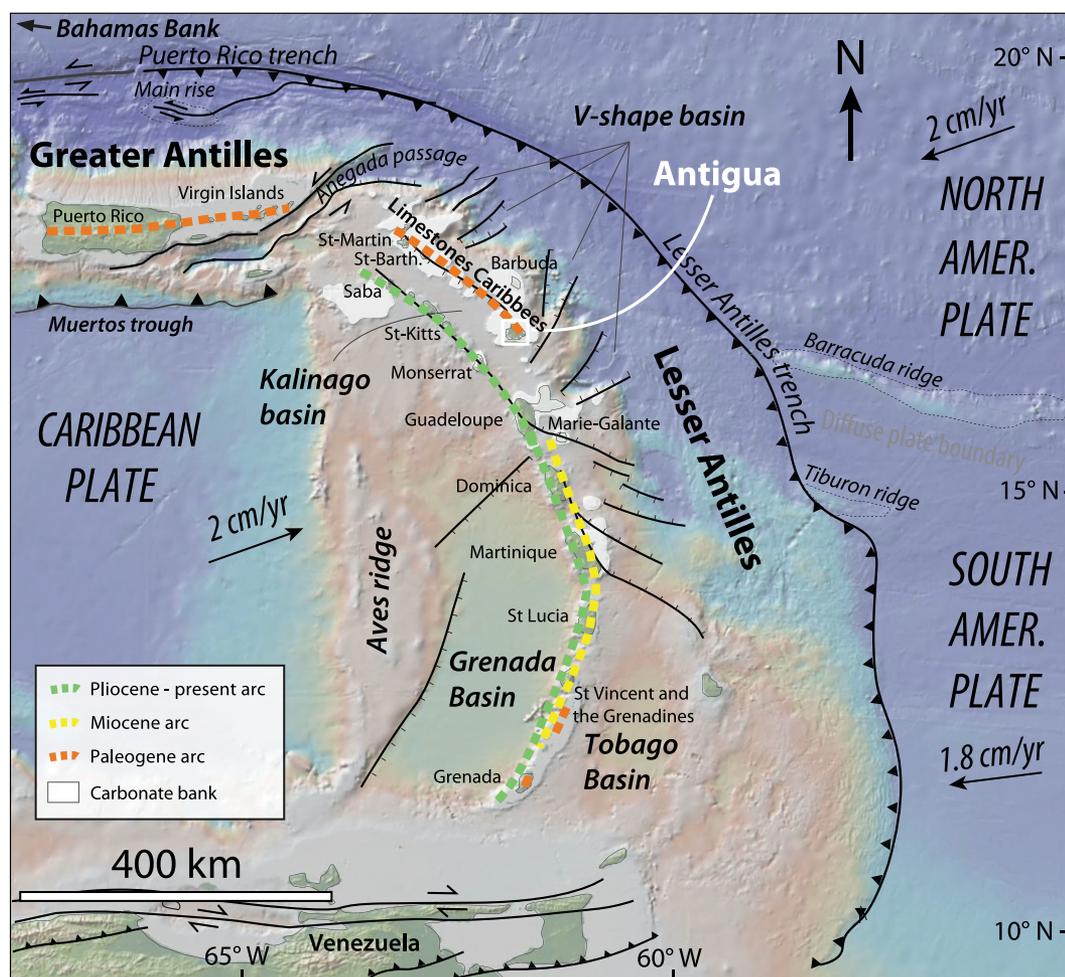


Figure 1. Bathymetric map shows tectonic and volcanic settings of the Lesser Antilles subduction zone. The main faults are drawn from Cornée et al. (2021). The study area (Antigua) is outlined by the white square. Velocity vectors of the Caribbean and North and South American plates are from Symithe et al. (2015).

and stable westward plate motion since 50 Ma (Leroy et al., 2000). This subduction is associated with a currently active, well-developed modern volcanic arc that consists of 17 active volcanoes that formed islands spread over ~800 km. Nevertheless, the arc rocks exposed on these islands are mainly recent (Pliocene to present day), and the magmatic record shows significant breaks in volcanism along sections of the arc during both the late Paleogene and Miocene, especially in the northern part of the arc (Brown et al., 1977; Briden et al., 1979; Bouysse and Westercamp, 1990; Macdonald et al., 2000; Legendre et al., 2018; Noury et al., 2021). In the northern Lesser Antilles, older arc rocks are confined to the islands of the “Limestone Caribbees,” which are located in the forearc of the modern arc (Figs. 1–2, Table 1). The Limestone Caribbees contain well-developed Eocene (ca. 45 Ma) to early Oligocene (ca. 28 Ma) volcanic records that are overlain by platform carbonates. From the period between 28 Ma to 20 Ma, only sparse records of magmatic activity have been reported from the northeastern branch of the Lesser Antilles arc (Briden et al., 1979; Leg-

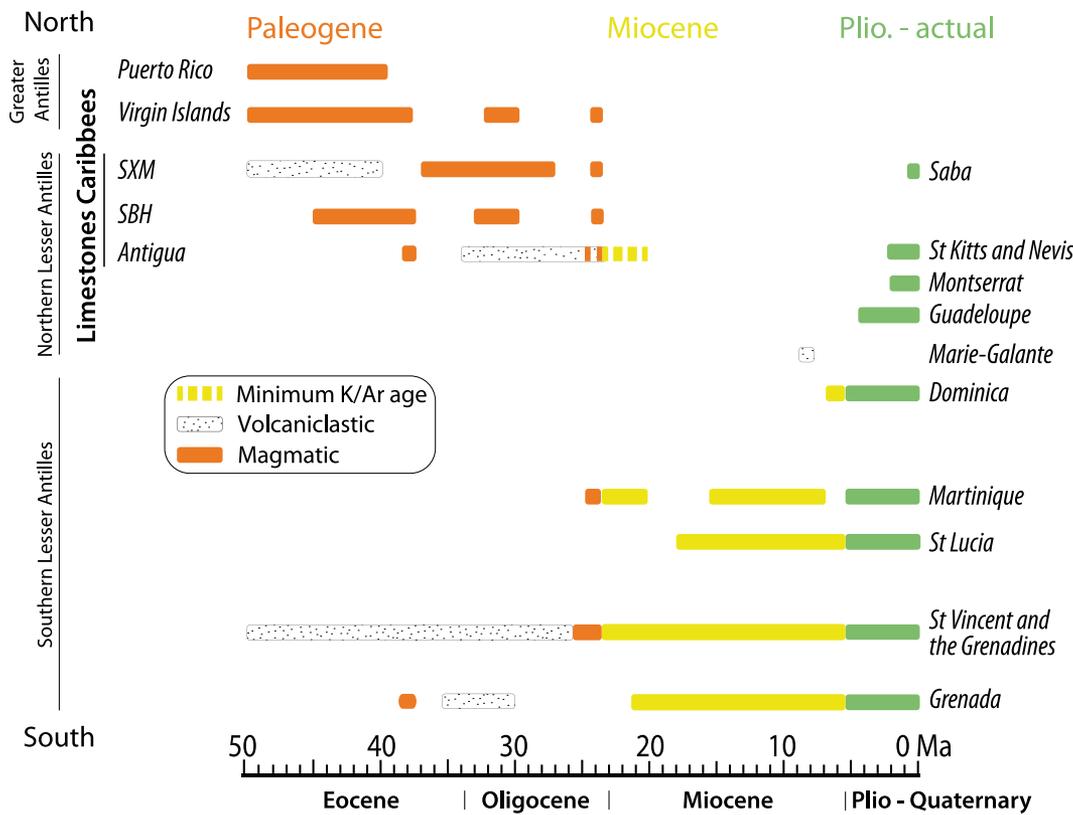
endre et al., 2018; Cornée et al., 2021; Noury et al., 2021; Fig. 2, Table 1). From 20 Ma to 4 Ma, no record of magmatic activity is reported in the northern Lesser Antilles (Fig. 2, Table 1). This interval could represent a lull in magmatism or simply be a sampling bias as exposures are sparsely dispersed on islands. Miocene volcanic record could also be buried below younger volcanoes or located offshore.

An answer to these questions may come from Antigua, which displays a magmatic unit whose age was previously estimated to be from Oligocene to Miocene based on K-Ar whole rock analysis (Nagle et al., 1976; Briden et al., 1979). In Antigua, the only complete Oligocene to Miocene sedimentary sequence of the Limestone Caribbees is preserved.

Uncertainties about the K-Ar ages, which were partly rejuvenated by a hydrothermal event (Gunn and Roobol, 1976; Briden et al., 1979), and the age and thickness of the sedimentary sequence (Mascle and Westercamp, 1983; Multer et al., 1986; Weiss, 1994), in addition to the absence of structural data and a description of the tectonic architecture of the island, lead to

uncertainties about the timing of the geological events that affected the island and consequently about the age of the arrest of the northern Paleogene Lesser Antilles arc. Questions are even raised about the existence of its arrest.

Given that perspective, we constrain the magmatic and stratigraphic architecture and evolution of the island by developing (1) the detailed lithostratigraphic record of the island, in which we analyzed the temporal and spatial relationships between the previously identified lithological units; (2) a first mapping and analysis of the tectonic structures deforming the lithological units; and (3) an updated geological map based on our lithostratigraphic record and structural analysis. We aim to refine the duration of magmatism and the duration and depositional setting of sedimentary rocks (which may or may not overlap) by establishing (4) age estimates of key volcanic and intrusive units using  $^{40}\text{Ar}/^{39}\text{Ar}$  geochronology and (5) an updated facies analysis and foraminiferal biostratigraphy of the marine units. To estimate whether changes in magmatism correlate with tectonic changes, we perform on the scale of Antigua (6) a kinematic analysis



**Figure 2.** Schematic representation shows the dated magmatic and volcanoclastic onshore evidence of the Lesser Antilles arc, from Eocene to present. The references used to construct this sketch are noted in Table 1. SXM—Saint-Martin; SBH—St-Barthelemy.

using small-scale syn- and post-depositional faults and (7) an anisotropy of magnetic susceptibility (AMS) study to complement the structural analysis and estimate the stress field, and changes therein, that existed during sedimentation. Finally, we incorporate our results in the most recent structural, stratigraphic, and paleogeographic frameworks for the region (Philippon et al., 2020a; Marivaux et al., 2020; Cornée et al., 2021) by providing land-sea correlations using seismic lines to the west and south of Antigua acquired during the Antilles IV oceanographic cruise (Bouysse and Mascle, 1994). We use our results to evaluate the presence and duration of a lull in Lesser Antilles magmatism and discuss our results in terms of possible geodynamic triggers for anomalous magmatism.

**GEOLOGICAL SETTINGS**

**Regional Setting**

The Lesser Antilles arc developed above a west-dipping subduction zone that accommodates the convergence between the Caribbean upper plate and the downgoing North and South American plates. The plate boundary between the North and South American plates is diffuse, as relative motion between the two is minor (e.g., Pindell and Kennan, 2009; Fig. 1). To the north and south, the Caribbean plate boundary changes

to E–W-trending transforms that are sinistral in the north and dextral in the south (Pindell and Kennan, 2009; Fig. 1). The transition is abrupt in the south and accommodated by a slab tear that is gradual in the north along a curved plate boundary overlying a curved, amphitheater-shaped slab with a north(west)ward increase in subduction obliquity (Govers and Wortel, 2005; van Benthem et al., 2013, 2014; Fig. 1). This plate configuration developed ~60–40 m.y. ago when the Caribbean plate motion relative to the Americas changed from northeastward to eastward (Boschman et al., 2014; Montes et al., 2019). Before that, the eastern Caribbean plate boundary consisted of an oblique subduction at the origin of the Great Caribbean arc that extended from Cuba to the south of the Aves Ridge (Fox et al., 1971; Bouysse et al., 1985; Christeson et al., 2008; Neill et al., 2011; Allen et al., 2019; Padron et al., 2021; Fig. 1). After the kink, the subduction became orthogonal, and the arc jumped eastward toward the Limestone Caribbees.

At regional scale, a north–south morphostructural dichotomy is observed along the curved Lesser Antilles trench: the southern back-arc region hosts the Eocene Grenada basin with a flat bathymetry and a mean depth of 2800 m and thin oceanic crust (Garroq et al., 2021; Padron et al., 2021). To the north, the seafloor is rough, mean bathymetry is 1400 m, and the arc crust is thicker (Schlaphorst et al., 2018). The

northern Lesser Antilles region hosts a middle to late Eocene fold and thrust belt (Philippon et al., 2020a) in which subsequently, the trench-parallel Kalinago extensional basin opened in the late Eocene to early Oligocene (Bouysse and Westercamp, 1990; Legendre et al., 2018; Philippon et al., 2020a; Cornée et al., 2021; Fig. 1).

From the late Oligocene to middle Miocene, the northern Lesser Antilles forearc was deformed by V-shaped extensional, perpendicular-to-the-trench basins that opened from late Oligocene to middle Miocene time (Fig. 1). That may suggest intensification of the trench curvature (Philippon et al., 2020b; Boucard et al., 2021). These basins have been crosscut by trench-parallel normal faults since the middle Miocene that accommodate fore-arc subsidence (Boucard et al., 2021).

Offshore investigations in the southern Kalinago basin and its extension south of Antigua island (Willoughby sub-basin) showed that the sedimentary infilling was composed of three megasequences (MS) separated by regional unconformities (sequence boundaries [SB]): a late Oligocene to early Miocene megasequence (MS4) resting above a deformed basement and topped by the early–middle Miocene major erosional surface (SB4) indicative of a regional uplift; middle to present-day megasequences (MS5 to MS 6–7) representative of subsidence of the Kalinago and Willoughby basins (Cornée

TABLE 1. GEOCHRONOLOGIC AGES OF MAGMATIC OCCURRENCES OF THE LESSER ANTILLES ARC

Rock type	Mineral	Method	Age range (Ma)	Number of ages	References
<u>Porto Rico</u> Tonalite, granodiorite, basalt, andesite, volcanic breccia	Whole rock, plagioclase, biotite; hornblende	K-Ar	47.3–40	57	Cox et al. (1977); Barabas (1982); Jolly et al. (1998)
<u>US Virgin islands</u> Basalt, andesite, rhyolite	Whole rock	K-Ar	42.1–37.7	8	Vila et al. (1986); Alminas et al. (1994)
<u>BI Virgin Islands</u> Granodiorite–diorite Granodiorite–diorite	Zircon Whole rock, biotite; hornblende	U-Pb K-Ar	43.6–30.6 38.8–24.0	25 6	Schrecengost (2010) Cox et al. (1977); Kesler and Sutter (1979); Vila et al. (1986); Smith et al. (1998)
<u>St-Martin</u> Granodiorite, volcanic breccia, andesite Granodiorite, basalt, andesite	Biotite; hornblende, Plagioclase, groundmass Whole rock	Ar-Ar K-Ar	29.5–24.4 37.2–26.1	8 10	Davidson et al. (1993); Cornée et al. (2021); Noury et al. (2021) Nagle et al. (1976); Briden et al. (1979)
<u>St-Barthelemy</u> Granodiorite, basalt, andesite Granodiorite, basalt, andesite	Plagioclase, groundmass Whole rock	Ar-Ar K-Ar	44.9–24.5 36.3–24.0	7 5	Legendre et al. (2018) Nagle et al. (1976); Briden et al. (1979)
<u>Antigua</u> Andesite, dacite Basalt, andesite, dacite	Whole rock Plagioclase, groundmass	K-Ar Ar-Ar	39.7–20 35.5–27.7	5 5	Nagle et al. (1976); Briden et al. (1979) This study
<u>Saba</u> Andesite	Whole rock	K-Ar	0.4–present	3	Defant et al. (2001)
<u>St-Kitts and Nevis</u> Andesite	Whole rock	K-Ar	2.3–present	4	Baker (1984)
<u>Montserrat</u> Andesite	Plagioclase, groundmass	Ar-Ar	2.1 -present	25	Hatter et al. (2018)
<u>Guadeloupe</u> Andesite	Plagioclase, groundmass	K-Ar, Ar-Ar	4.3–present	47	Samper et al. (2007); Zami et al. (2014); Favier et al. (2019)
<u>Marie Galante</u> Volcaniclastic	Plagioclase	Ar-Ar	8.6–8.1	2	Münch et al. (2013)
<u>Dominica</u> Basalt, andesite, dacite	Groundmass	K-Ar	6.8–present	32	Smith et al. (2013)
<u>Martinique</u> Basalt, andesite Basalt, andesite	Plagioclase, groundmass Whole rock	Ar-Ar K-Ar	24.8–7.1 19.2–present	20 40	Germa et al. (2011) Bellon et al. (1974); Andrieff (1976); Briden et al. (1979)
<u>St-Lucia</u> Basalt, andesite Basalt, andesite	Whole rock Whole rock	K-Ar K-Ar	15.0–present 18.3–present	29 10	Le Guen de Kerneizon et al. (1983) Briden et al. (1979)
<u>St-Vincent and the Grenadines</u> Volcaniclastic Basalt, andesite Basalt, andesite Basalt	Zircon Whole rock, plagioclase Whole rock Whole rock, groundmass	U-Pb K-Ar K-Ar Ar-Ar	34.3–26.3 / 15.8–13.2 25.2–12.4 18.1–present 6.0–present	19 4 40 35	Rojas-Agramonte et al. (2017) Speed et al. (1993) Briden et al. (1979); Westercamp (1985) White et al. (2017)
<u>Grenada</u> Basalt Volcaniclastic Basalt, andesite Basalt, andesite Andesite	Groundmass Zircon Whole rock Zircon Whole rock	Ar-Ar U-Pb K-Ar U-Pb Ar-Ar	38.3 35.9–30.9 21.4–present 5.4–4.2 2–present	1 19 15 11 4	White et al. (2017) Rojas-Agramonte et al. (2017) Briden et al. (1979) Rojas-Agramonte et al. (2017) Speed et al. (1993)

et al., 2021; Supplemental Material 3<sup>1</sup>). No conic structures or chimneys, which could have been representative of fossil volcanic activity, were found on seismic lines.

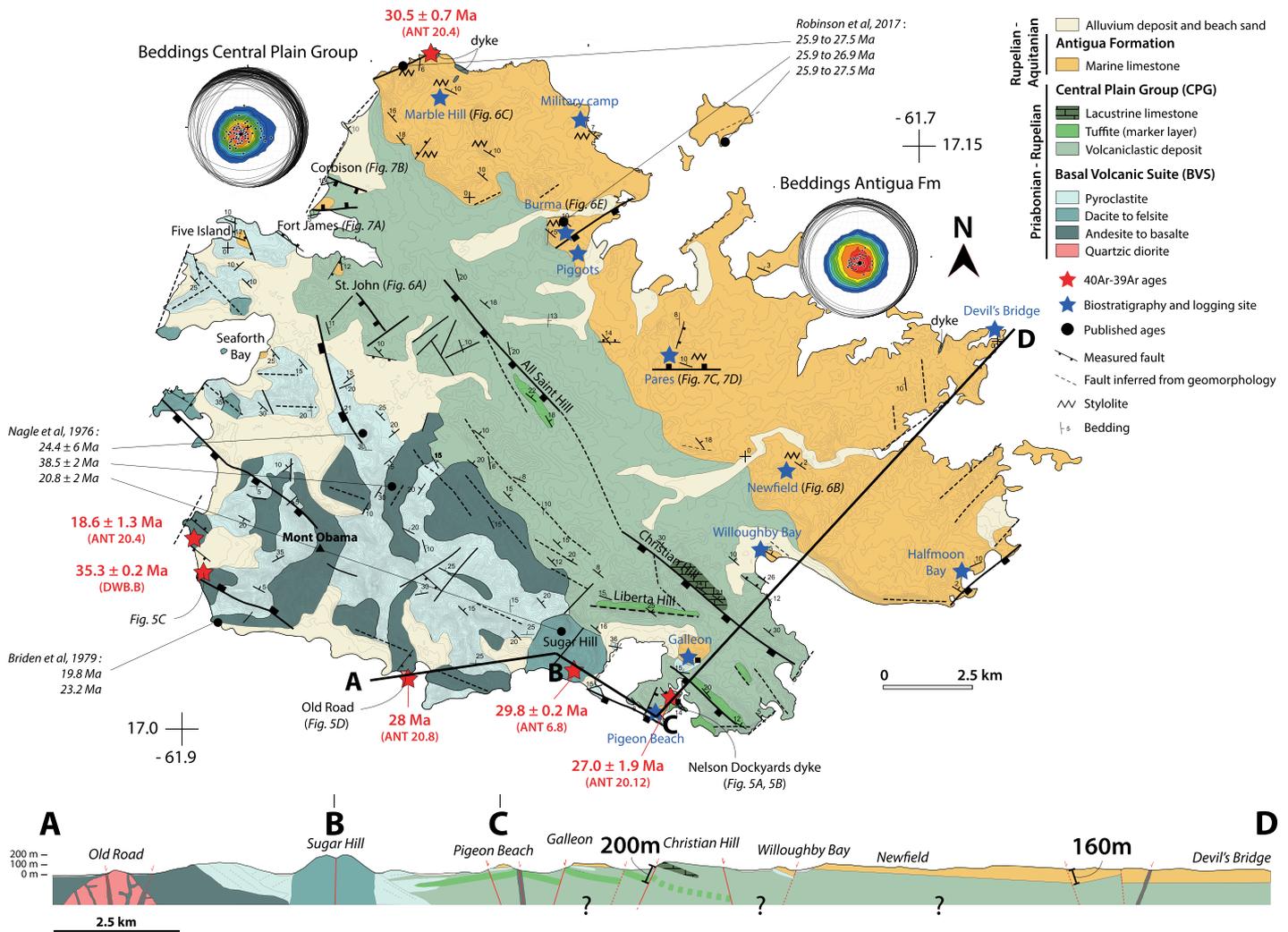
<sup>1</sup>Supplemental Material. Supplemental Material 1: Description of the sections, ages and depositional settings. Supplemental Material 2: Microfacies and larger benthic foraminifera content. Supplemental Material 3: Description of the seismic profiles 508 and 510. Supplemental Material 4: Structural measurements collected in the field. Supplemental Material 5: Detail of Ar/Ar step-heating analyses. Please visit <https://doi.org/10.1130/GSAB.S.20431707> to access the supplemental material, and contact editing@geosociety.org with any questions.

The present-day Lesser Antilles magmatic arc consists primarily of calc-alkaline volcanoes that formed islands from Grenada in the south to Saba in the north (Macdonald et al., 2000; Fig. 1). These islands span a trench-parallel arc with a northwestward kink at the latitude of Martinique (Fig. 1).

To the south, the Pliocene to present-day arc is superimposed on older arc rocks. In Grenada and the Grenadines, these rocks are upper Eocene to Miocene calc-alkaline dikes, volcanic arc rocks, and volcaniclastic deposits (Briden et al., 1979; Westercamp, 1985; Andrieff et al., 1987; Speed et al., 1993; Donovan et al., 2003; Rojas-Agramonte et al., 2017; White et al., 2017; Fig. 2,

Table 1). In Dominica, Martinique, and Sainte-Lucia, older arc rocks consist of well-developed latest Oligocene to Miocene volcanics (Le Guen de Kerneizon et al., 1983; Westercamp, 1985; Briden et al., 1979; Germa et al., 2011; Smith et al., 2013; Fig. 2, Table 1).

In the north, the onset of recent arc volcanoes occurred during the Pliocene to present day on Saba, St-Eustatius, St-Kitts and Nevis, Montserrat, and Guadeloupe (Baker, 1984; Defant et al., 2001; Samper et al., 2007; Hatter et al., 2018; Favier et al., 2019). These eruptive centers are located on the western shoulder of the Kalinago basin, 50–20 km westward from the older eruptive centers preserved in the NW-trending



**Figure 3.** Revised geological map and synthetic cross-section of Antigua are shown. The map was constructed based on the maps of Martin-Kaye (1959, 1969) and Christman (1973) and completed and updated with fieldwork observations, lithostratigraphy, and structural measurements. The cross-section is located by a gray line crosscutting the island from NE to SW, with tipping point marked by the letters A, B, C, and D. The tuffite layer (light green) was used as a marker layer in the construction of the cross-section. Beddings are plotted on the lower hemisphere. The K-Ar ages (Nagle et al., 1976; Briden et al., 1979) and the Sr-Sr ages (Robinson et al., 2017) are represented in italic.

“Limestones Caribbees” islands (St-Martin, St-Barthélemy, and Antigua) on the eastern shoulder of the Kalinago basin (Fig. 1).

The Limestone Caribbees host platform carbonates overlying extinct volcanoes that are preserved in a flat bathymetric high (Anguilla and Antigua Banks). These banks are bounded to the east by V-shaped basins (Fig. 1). On St-Martin, arc volcanism developed an Eocene volcaniclastic series (Dagain et al., 1989) intruded by a 29 Ma granodioritic pluton and associated dikes and sills (Noury et al., 2021), and the last magmatic activity is represented by a submarine volcanic breccia of 24.4 Ma (Cornée et al., 2021; Fig. 2, Table 1). In St-Barthélemy, magmatic rocks consist of submarine tholeiitic and calc-alkaline lavas (45–39 Ma), granodioritic

plutons (33.5–30 Ma), and local tholeiitic lavas (24.5 Ma; Legendre et al., 2018; Philippon et al., 2020a; Fig. 1, Table 1). On Antigua, current dating (K-Ar whole rock analysis; Nagle et al., 1976; Briden et al., 1979) supports arc magmatism from the late Eocene to Miocene and comprises plutonic and volcanic calc-alkaline rocks (Basal Volcanic Suite) and an Oligocene volcaniclastic series (Central Plain Group; Gunn and Roobol, 1976; Briden et al., 1979; Mascle and Westercamp, 1983; Fig. 2, Table 1).

**Geology of Antigua**

The first geological study of the lithologies of Antigua by Vaughan (1919) described coral associations and larger benthic foraminifera.

Martin-Kaye (1959) subdivided the island into three NW–SE-trending geological units: the Basal Volcanic Suite, the Central Plain Group, and the Antigua Formation (Fig. 3).

The Basal Volcanic Suite is a mountainous region exposed on the west side of the island, which includes the island’s highest point (Mont Obama, 402 m). The suite consists of lava flows and pyroclastic rocks, dikes, and domes of predominantly basaltic, andesitic, and dacitic compositions and with island-arc type geochemical signatures (Christman, 1973; Gunn and Roobol, 1976; Fig. 3). Basaltic dikes are also found crosscutting the topmost stratigraphic units (Central Plain Group and Antigua Formation) and are thus considered to be the last magmatic pulses exposed on the island (Martin-Kaye,

1959, 1969; Fig. 2). Whole-rock K-Ar analyses of volcanic rocks of the Basal Volcanic Suite gave ages ranging between  $38.5 \pm 2$  Ma (middle Eocene) and 20 Ma (minimum age, early Miocene) (Nagle et al., 1976; Briden et al., 1979).

The Central Plain Group is described as ~1500 m of terrestrial volcanoclastic rocks (conglomerates, sandstones, mudrocks, and tuffites) derived from volcanism and the erosion of the Basal Volcanic Suite (Masclé and Westercamp, 1983; Multer et al., 1986; Weiss, 1994). These volcanoclastic rocks are interbedded with lacustrine limestones, which contain petrified wood, charophytes, ostracods, and freshwater gastropods (Brown and Pilsbry, 1914; Trechmann, 1941; Martin-Kaye, 1959, 1969; Masclé and Westercamp, 1983; Weiss, 1994; Donovan et al., 2014; Fig. 3). Thomas (1942) and Martin-Kaye (1959, 1969) also suggested that there may be patches of marine carbonates interbedded with the Central Plain Group (e.g., the Seaforth limestone; Fig. 3) that they interpreted as episodic marine invasions.

The Antigua Formation is composed of carbonate deposits overlying the Central Plain Group, in which volcanic deposits have not been recognized. It comprises numerous coral patches and associated bioclastic carbonate deposits that change into, upward and laterally, deeper marine carbonate deposits in the east (e.g., Martin-Kaye, 1959, 1969; Persad, 1969; Masclé and Westercamp, 1983; Multer et al., 1986; Weiss, 1994). The estimated thickness of the Antigua Formation is 350–550 m (Masclé and Westercamp, 1983). Based on the study of larger benthic foraminifera, planktonic foraminifera, ostracods, and calcareous nannofossils, the age of the Antigua Formation has been estimated as late Oligocene (Chattian) (Vaughan, 1919; van den Bold, 1966; Martin-Kaye, 1959, 1969; Persad, 1969; Masclé and Westercamp, 1983), consistent with five strontium isotope ages ranging from 26 Ma to 28 Ma (Robinson et al., 2017; Fig. 3) and with recent preliminary age determinations of Cornée et al. (2021). Previous geological studies did not investigate the tectonic architecture or structural history of Antigua.

The three units comprising the island are affected by sub-greenschist hydrothermal alteration that could be responsible for the rejuvenation of K-Ar ages (Briden et al., 1979). Gunn and Roobol (1976) observed albitized plagioclases and chloritized orthopyroxenes and groundmass in the Basal Volcanic Suite. In the Central Plain Group, bright green, chloritized tuffites were described by Martin-Kaye (1959, 1969) and Jackson (2013). In the Antigua Formation, this alteration phase is found in the form of silicified limestones (Strang et al., 2018).

## METHODS

### Field Investigation

Three field campaigns (2018, 2019, and 2020) were conducted to investigate the tectono-magmatic history of Antigua. Exposure within the interior of the island is poor due to thick alluvial, colluvial, or vegetation cover. Exposure is most abundant along the coast, even though numerous private properties hampered a full remapping of some coastal portions. Nevertheless, we updated the geological map by mapping the first-order lithostratigraphic units and regional structures (unconformities, faults, and dikes) and investigated the relations between units (Fig. 3).

Because the Antigua formation outcrops are isolated, we coupled microfacies and biostratigraphy with lithostratigraphy to reconstruct the general sedimentary organization (Fig. 4, Supplemental Material 1; see footnote 1) and paleo-environments. We logged eight sedimentary sections in the Central Plain Group and 10 in the Antigua Formation (Fig. 4, Supplemental Material 1) and used dated, isolated exposures from road cuts or along the seashore (e.g., Galleon, Military Camp, Marble Hill, Fort James, Nelson Dockyard, Piggots, and Willoughby Bay), continuous cross-sections along the shoreline (e.g., Half Moon Bay, Devil's Bridge, and Pigeon Beach), and quarries (Burma, Newfield, Pares, and Parnham) (Figs. 3–4). We collected some 400 structural measurements (bedding orientations, faults, and striaes) and reported these on the map (Fig. 3, Supplemental Material 4).

### $^{40}\text{Ar}/^{39}\text{Ar}$ Geochronology

For  $^{40}\text{Ar}/^{39}\text{Ar}$  geochronology, we collected samples from volcanic and igneous rocks of the Basal Volcanic Suite (Fig. 3). To understand the magmatic evolution of the island, we collected all types of magmatic material that was recognized on the island while targeting the least altered outcrops. Samples consist of an andesitic lava (ANT 20.4, Fig. 5C) crosscut by a rhyolitic dike (DWB-B, Fig. 5C), a dacitic dome (ANT 6.8, Sugar Hill), a quartz-dioritic pluton (ANT 20.8, Old Road), and two mafic dikes, one crosscutting the Central Plain Group (ANT 20.12, Figs. 3 and 5B) and the other piercing the base of the Antigua Formation (ANT 20.6, Fig. 3).  $^{40}\text{Ar}/^{39}\text{Ar}$  step-heating analyses were performed at Géosciences Montpellier, Géosciences Montpellier, Montpellier, France (detailed methodology in Supplemental Material 5). Raw data for each step and blank were processed and ages were calculated using the ArArCALC software

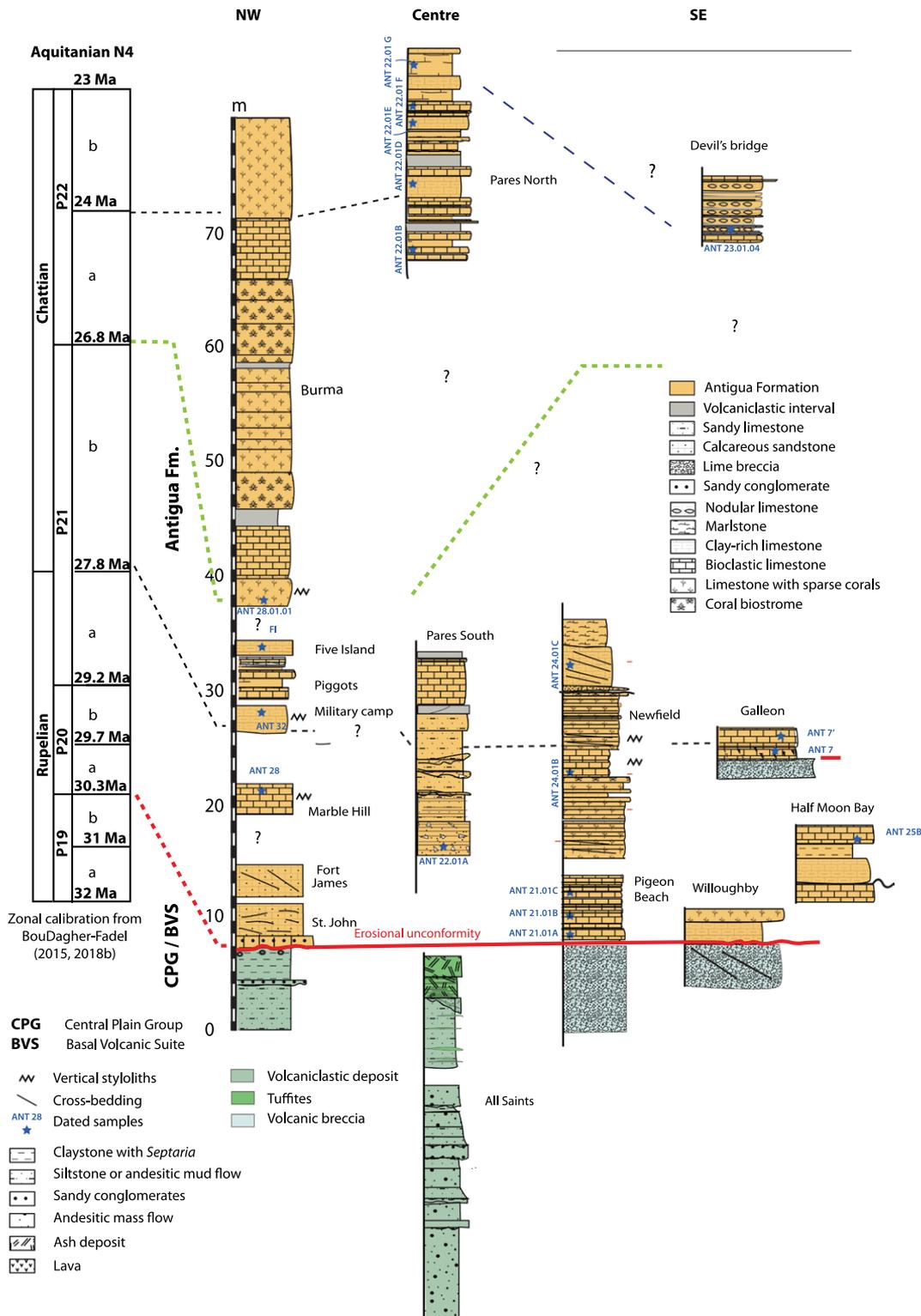
(Koppers, 2002). Ages were statistically analyzed in two ways using  $^{39}\text{Ar}$  released spectra and inverse isochrones. Plateau ages were calculated from at least three consecutive  $^{39}\text{Ar}$  release steps comprising up to 50% of the total  $^{39}\text{Ar}_K$  released and overlapping at the  $2\sigma$  confidence level (Fleck et al., 1977). Isochrone ages were accepted when mean square weighted deviation (MSWD) was close to 1 and the  $^{40}\text{Ar}/^{36}\text{Ar}$  intercept was close to the  $(^{40}\text{Ar}/^{36}\text{Ar})_{\text{atm}}$  value. All results are given at the  $2\sigma$  uncertainty.

### Microfacies and Biostratigraphy

For microfacies and fossil content determination, 35 thin sections were analyzed from the carbonate rocks of the Antigua Formation and 10 from volcanoclastic rocks and lacustrine limestones of the Central Plain Group, which were collected from the 18 logged sections (Fig. 4, Supplemental Material 1). The microfacies were associated with a depositional environment following the classification of Wright and Burchette (1996) and Flügel and Munnecke (2010) and supplemented by the larger benthic foraminiferal content (BouDagher-Fadel, 2018a; Supplemental Material 2). The biostratigraphic analyses were only carried out for the marine Antigua Formation (the Central Plain Group only contains undatable freshwater gastropods) and were based on a complete inventory of larger benthic and planktonic foraminiferal taxa identified in the thin sections. We used the zonal calibrations of BouDagher-Fadel (2015, 2018b) for the foraminifera, calibrated against the timescale of Gradstein et al. (2012).

### Fault Analysis

We employed the right dihedral methods to inverse fault kinematics and interpret the paleo-stress tensor using WinTensor software (Angelier, 1979; Delvaux and Sperner, 2003). WinTensor allows the rotational optimization of the obtained tensor by performing iterative tests of tensors to minimize a misfit function. We performed these inversions with sets of data consisting of fault strike, dip, plunge of the striations, and bedding offsets (Supplemental Material 4). The quality of inversion is given by world stress map quality criteria (QRw) ranging from A (good) to E (poor), depending on the amount of data used for the inversion ( $n$ ), the ratio between the amount of data used for the inversion over the amount of input data ( $n/nt$ ), deviation between observed and theoretical slip direction ( $\alpha_w$ ), the sense of slip confidence level ( $Cl_w$ ), and the type of structure (fault, fracture, or shear zone).



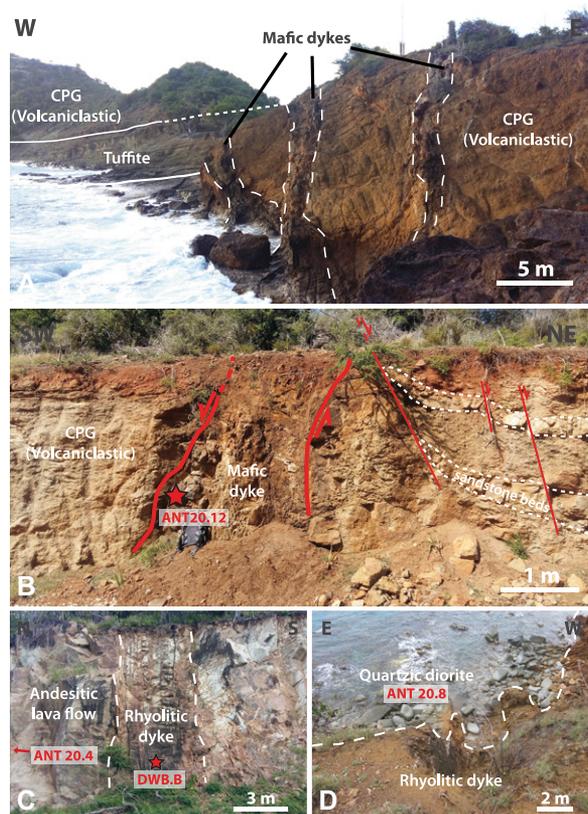
**Figure 4. Stratigraphic synthesis and lithological successions of the Antigua units are shown. The locations of the sections are presented in Figure 3. See Supplemental Material 1 for a detailed description of each section (see text footnote 1). CPG/BVS—Central Plain Group/Basal Volcanic Suite.**

**Anisotropy of Magnetic Susceptibility (AMS)**

To supplement fault kinematics analysis, we used anisotropy of magnetic susceptibility (AMS) to determine the petrofabric of samples

of the Central Plain Group and the Antigua Formation and determine the direction of tectonic paleo-strain. We drilled 71 paleomagnetic cores of fine-grained rocks over eight sites. Samples were collected with a petrol-powered, water-cooled drill and oriented with

a magnetic compass. The local magnetic declination (14°W in January 2020) was corrected on all samples. Non-demagnetized standard specimens were measured using an AGICO KLY-3S Kappabridge at the Geosciences Montpellier Palaeomagnetic Laboratory. AMS



**Figure 5.** Field photography of the Basal Volcanic Suite and the Central Plain Group (CPG). (A) Basaltic andesite dikes crosscut the volcanoclastic deposits of the Central Plain Group; (B) basaltic andesite dike intrudes a NW–SE-trending normal fault and crosscuts the Central Plain Group; (C) 35-m.y.-old rhyolitic dike (sample DWB-B) crosscuts an andesitic lava flow; (D) quartzitic diorite stock is crosscut by a rhyolitic dike.

is defined by an ellipsoid with the principal axes  $k_{\max} > k_{\text{int}} > k_{\min}$  used as the strain indicator (Jelínek and Kropáček, 1978; Parés et al., 1999; Maffione et al., 2015). Undeformed sediments display a so-called sedimentary fabric where  $k_{\max}$  and  $k_{\text{int}}$  are dispersed within the plane of magnetic foliation (F), which is parallel to the sedimentary stratification, and  $k_{\min}$  is the pole of the bedding plane. A tectonic fabric can overprint the sedimentary one and produce a magnetic lineation (L) by aligning the  $k_{\max}$  axes parallel to the maximum axis of stretching ( $\sigma_3$ ). In this study we used the magnetic lineation as a paleo-strain proxy to determine the nature and direction of the tectonic strain. The AMS parameters were evaluated using ANISOFT 5.1 software according to Jelínek and Kropáček, 1978; Chadima et al., 2020).

## RESULTS

### $^{40}\text{Ar}/^{39}\text{Ar}$ Geochronology of the Basal Volcanic Suite

The Basal Volcanic Suite crops out in the southwest of the island and is mainly composed of andesitic lavas and pyroclastic deposits. These are crosscut by rhyolitic dikes (Fig. 5C) and intruded by a dacitic dome at Sugar Hill

and Seaforth Bay (Fig. 3). At the Old Road site, a small stock of quartzitic diorite, previously described by Christman (1973), is intruded by ENE–WSW rhyolitic dikes similar in appearance to the one crosscutting the andesitic lavas (Figs. 5C–5D). To the southeast, at Nelson Dockyard, mafic dikes intrude the Central Plain Group. To the northeast and southeast (Devil’s Bridge), similar dikes are found intruding the Antigua Formation (Figs. 3 and 5A–5B). No consistent bedding dip is apparent in the Basal Volcanic Suite; the pyroclastic rocks have bedding that varies between  $0^\circ$  and  $30^\circ$  around local eruptive centers (Fig. 3).

The andesitic lava (sample ANT20.4) yields a plateau age of  $18.57 \pm 1.26$  Ma with 82.07% of Ar released (Fig. 6, Table 2). Individual steps show large errors, especially for low-temperature degassing steps, resulting from high atmospheric contamination. Consequently, all data cluster along the ordinate axis of the inverse isochron diagram. The inverse isochron calculations for the steps corresponding to the plateau yielded a concordant age of  $19.74 \pm 2.46$  Ma with  $\text{MSWD} = 1.52$  for an initial  $^{40}\text{Ar}/^{36}\text{Ar}$  ratio of  $297.8 \pm 1.5$ , which indicates that the trapped  $^{40}\text{Ar}/^{36}\text{Ar}$  is indistinguishable from atmospheric  $^{40}\text{Ar}/^{36}\text{Ar}$ .

The rhyolitic dike (sample DWB-B) that crosscuts the andesitic lava flow (sample

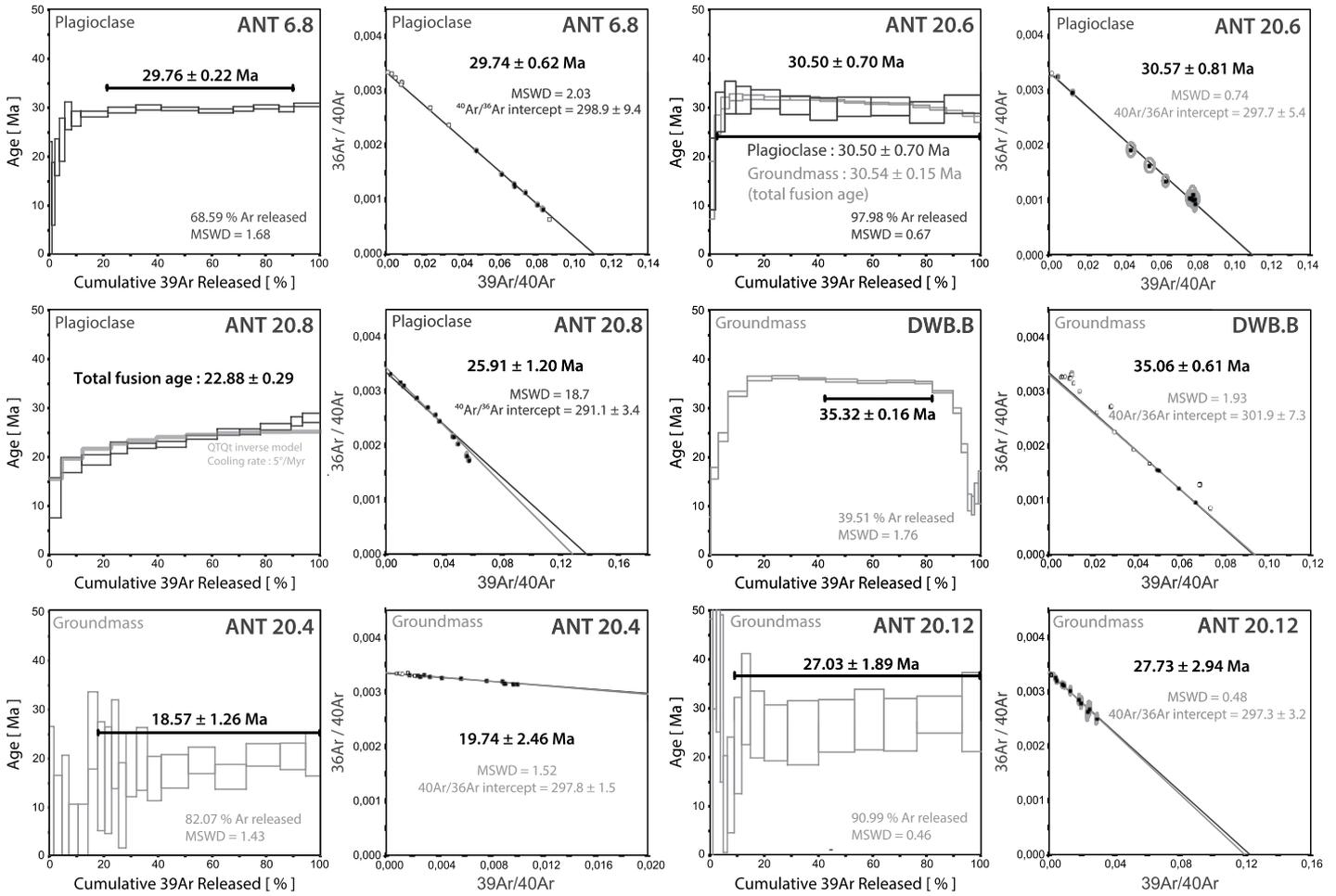
ANT20.4) yields a hump-shaped spectrum with a mini-plateau age ( $\sim 40\%$  cumulative  $^{39}\text{Ar}$  is released) of  $35.32 \pm 0.16$  Ma. The inverse isochron calculations for the steps corresponding to the mini-plateaus yielded concordant ages of  $35.06 \pm 0.61$  Ma ( $\text{MSWD} = 1.93$ , initial  $^{40}\text{Ar}/^{36}\text{Ar}$  ratio of  $301.9 \pm 7.3$ ) with the trapped  $^{40}\text{Ar}/^{36}\text{Ar}$  indistinguishable from atmospheric  $^{40}\text{Ar}/^{36}\text{Ar}$ . The low- and high-temperature degassing steps yield younger ages between 10 Ma and 15 Ma and exhibit compositions poorer in  $^{40}\text{Ar}$  and richer in  $^{38}\text{Ar}$  than the central part, which indicates that they correspond to a secondary phase (Fig. 6, Table 2).

Plagioclase populations from the dacitic dome of Sugar Hill, previously dated with K-Ar geochronology at 20.8 Ma (Nagle et al., 1976; Fig. 3), yield a plateau age of  $29.76 \pm 0.22$  Ma for the central part of the spectra, corresponding to 68.59% of the  $^{39}\text{Ar}$  released. The inverse isochron age of  $29.74 \pm 0.62$  Ma is concordant with an  $\text{MSWD}$  of 2.03 and an initial  $^{40}\text{Ar}/^{36}\text{Ar}$  ratio of  $298.9 \pm 9.4$ , which indicates that the trapped  $^{40}\text{Ar}/^{36}\text{Ar}$  is indistinguishable from atmospheric  $^{40}\text{Ar}/^{36}\text{Ar}$  (Fig. 6, Table 2).

Analysis of the plagioclase population of the quartz diorite (ANT 20.8) yields a staircase spectrum without a plateau ranging from 10 Ma for low-temperature degassing steps to 28 Ma at high temperature. The total fusion age is  $22.88 \pm 0.29$  Ma. To aid interpretation of the measured spectra and the thermal history, we performed inverse modeling of the  $^{40}\text{Ar}/^{39}\text{Ar}$  spectra of the plagioclase population from the dioritic pluton with QTQt software (Gallagher, 2012). The inverse QTQt model for the plagioclase  $^{40}\text{Ar}/^{39}\text{Ar}$  spectrum suggests that the diorite sample underwent a steady cooling at a rate of  $\sim 5^\circ/\text{Ma}$  since ca. 25 Ma (Fig. 6).

Despite a large error on individual steps related to atmospheric contamination, analysis of the groundmass of the basaltic andesite dike crosscutting the Central Plain Group (ANT20.12) yields a plateau age of  $27.03 \pm 1.89$  Ma, which corresponds to 90.99% of the  $^{39}\text{Ar}$  released (Fig. 6, Table 2). The inverse isochron yields a concordant age of  $27.73 \pm 2.94$  Ma with an  $\text{MSWD}$  of 0.48 and an initial  $^{40}\text{Ar}/^{36}\text{Ar}$  ratio of  $297.3 \pm 3.2$ , which is indistinguishable from the atmospheric ratio. Low-temperature individual degassing steps show larger errors and yield younger ages (10–20 Ma).

For the basaltic andesite dike intruding the Antigua Formation (ANT 20.6), we analyzed both plagioclase and groundmass populations. The plagioclase analysis yields a plateau age of  $30.50 \pm 0.70$  Ma with a large error on degassing steps corresponding to 97.98% of the  $^{39}\text{Ar}$  released (Fig. 6, Table 2). Calculated inverse isochrons show indistinguishable



**Figure 6.**  $^{40}\text{Ar}/^{39}\text{Ar}$  spectra and inverse isochrons are shown for the samples from Antigua. MSWD—mean square weighted deviation. The thick gray line in the spectra of ANT 20.8 corresponds to the spectrum modeled using QTQt software (Gallagher, 2012).

trapped and atmospheric  $^{40}\text{Ar}/^{36}\text{Ar}$  and an age of  $30.57 \pm 0.81$  Ma with  $\text{MSWD} = 0.74$  for an initial  $^{40}\text{Ar}/^{36}\text{Ar}$  ratio of  $297.7 \pm 5.4$ . Analysis of the groundmass population did not yield a plateau and a hump-shaped spectrum with younger low- and high-temperature individual steps (ca. 15 Ma) with a poorer  $^{40}\text{Ar}$  and richer  $^{38}\text{Ar}$  composition. However, the total fusion age obtained on the groundmass ( $30.54 \pm 0.27$  Ma) is concordant with the plateau age obtained from plagioclase population analysis.

**Lithostratigraphy, Microfacies, and Biostratigraphy**

**Central Plain Group**

The Central Plain Group comprises various terrestrial clastic rocks, lacustrine limestones, volcanoclastic mud and debris flows and conglomerates interpreted as alluvial fan deposits, volcanoclastic clay-rich deposits, tuffites, and lapilli-rich beds. These are well exposed at St-John, Corbison to the NW, and at the All Saints

site in the central part of the island (Figs. 3–4, Supplemental Material 1). To the SE, the volcanoclastic deposits of the Central Plain Group are topped by the lacustrine limestones of Christian Hill that contain feldspar and quartzitic debris (Figs. 3–4). We interpret the volcanoclastic series as having been deposited in an alluvial plain system with lakes (where the limestones formed) and ephemeral braided alluvial fans (cf. Nemeč and Steel, 1984) that was located at the foot of a volcanic edifice located to the SW and

TABLE 2.  $^{40}\text{Ar}$ - $^{39}\text{Ar}$  ANALYSIS RESULTS

Sample	Long. (°W)	Lat. (°N)	Lithology	Material	Age (Ma)	$\pm 2\sigma$	$^{39}\text{Ar}$ released (%)	MSWD	Inverse isochron age (Ma)	$\pm 2\sigma$	$^{40}\text{Ar}/^{36}\text{Ar}$ intercept	$\pm 2\sigma$	MSWD	Total fusion age (Ma)	$\pm 2\sigma$
ANT 6.8	-61.79	17.01	Dacitic dome	plg	<b>29.76</b>	0.22	68.59	1.68	29.74	0.62	298.92	9.4	2.03	28.96	0.26
ANT 20.4	-61.89	17.04	Andesitic lava flow	grdm	18.57	1.26	82.07	1.43	19.74	2.46	297.75	15.2	1.52	16.29	1.29
ANT 20.6	-61.83	17.17	Mafic dyke	plg	<b>30.50</b>	0.70	97.98	0.67	30.57	0.82	297.74	5.42	0.74	30.14	0.73
				grdm	no plateau	N.A.	N.A.	N.A.	30.88	0.49	294.78	5.18	45.89	30.54	0.15
ANT 20.8	-61.83	17.01	Quartzic diorite	plg	no plateau	N.A.	N.A.	N.A.	25.91	1.20	291.12	3.44	18.70	22.88	0.29
ANT 20.12	-61.76	17.01	Mafic dyke	grdm	<b>27.03</b>	1.89	96.13	0.46	27.79	2.71	297.0	1.56	1.47	26.75	1.87
DWB-B	-61.89	17.04	Rhyolitic dyke	grdm	<b>35.32*</b>	0.16	39.51	1.76	35.06	0.61	301.9	7.3	1.93	32.55	0.12

Note: Boldface indicates the plateau age, selected as characteristic of the sample. plg—plagioclase; grdm—groundmass; MSWD—mean square of weighted deviates. \*mini-plateau.



**Figure 7.** Field photographs show the sedimentary units of Antigua. (A) Unconformity between the Central Plain Group and the lower part of the Antigua Formation observed at Saint-John. Cross-bedded conglomeratic limestones rest above tilted and eroded silty claystones of the Central Plain Group. (B) Abrupt change from reef platform to foreereef slope at Newfield Quarry. Note the presence of blocks that slid toward the NE. (C) Wavy, subhorizontal stylolitic surfaces in foraminiferal wackestones at Marble Hill. (D) Vertical stylolites at Burma Quarry. (E) Coral beds with diverse massive and platy coral colonies (upper part of the Burma Quarry). For location, see Figure 3.

that corresponds nowadays to the Basal Volcanic Suite. Volcanic activity during deposition of the Central Plain Group was ongoing, as recorded

by tuffites and lapilli-rich deposits. We mapped discontinuous exposures of a tuffite member that is bright green in color and easy to recognize in

the field, which we use as a marker in our cross-section (Figs. 3–4, Supplemental Material 1). The Central Plain Group is overall a monoclinical succession striking ~NW–SE and dipping 10° toward the NE (Fig. 3). In the vicinity of normal faults, dips vary and may be up to 30° (Fig. 3).

Previous maps (Martin-Kaye, 1969; Christman, 1973) suggested that the Central Plain Group overlies the Basal Volcanic Suite, but we carefully studied the contacts and did not find this relationship; rather, the Central Plain Group interfingers the Basal Volcanic Suite as a syn-magmatic sedimentary sequence adjacent to the main volcanic edifice. This is supported by the presence of volcanoclastic deposits within the Central Plain Group at Willoughby, Pigeon Beach, and Galleon (Figs. 3–4). At Nelson Dockyard, the volcanoclastic and green tuffite layers of the Central Plain Group are intruded by mafic dikes (Fig. 5A) that follow a NW–SE-trending normal fault (Fig. 5B). From the NE–SW cross-section (Fig. 3), we estimate a minimum thickness of 200 m for the Central Plain Group, but the true thickness may be much thicker as the base of the sequence is not exposed.

#### *Antigua Formation*

At Galleon, Pigeon Beach, Five Island, Fort James, and Seaforth, patches of reefal limestones a few meters thick are observed resting either above the Central Plain Group or the Basal Volcanic Suite (Figs. 3–4). Previous authors (e.g., Thomas, 1942; Martin-Kaye, 1969; Mascle and Westercamp, 1983) interpreted these as interbedded in the Central Plain Group, but our observations show that these patches always unconformably overlie the Central Plain Group and are preserved in the hanging walls of normal faults. From microfacies analyses, we dated these marine limestones from Zones P20 (Rupelian, 30.3–29.2 Ma) to P21b–P22 (Chattian, 28.1–23 Ma).

The Antigua Formation is exposed on the eastern side of the island, upon the Central Plain Group. Local observations (St-John, Fig. 7A) and the regional shallower monoclinical orientation striking N140° and dipping 5° toward the NE indicate an angular unconformity between the Central Plain Group and the Antigua Formation (Fig. 3). The minimum thickness of the Antigua Formation is 80 m and is estimated from section logging in Burma Quarry (Fig. 4, Supplemental Material 1). From the cross-section (Fig. 3), we re-evaluated its thickness and found it to be 160 m. Centimeter-sized, sub-horizontal stylolite are observed within the top beds of the Antigua Formation (Figs. 3, 7C, and 7D), which shows that the thickness of the Antigua Formation must have been perhaps some several hundred meters thicker, but the overburden was later removed.

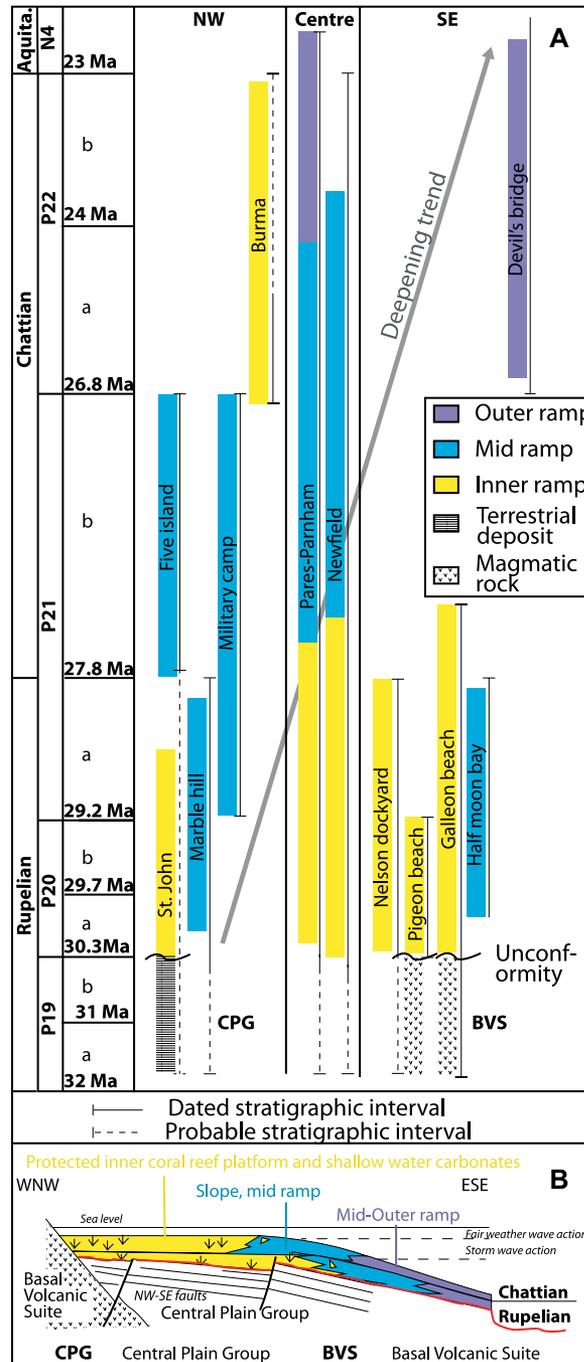
In the northwestern part of the island, the formation comprises ~10-m-thick, coarse-grained, cross-bedded siliciclastic carbonates; 18-m-thick foraminiferal packstones dated from Zones P18–P21a (Rupelian, 33.9–28.1 Ma) to P21b (Chattian, 28.1–26.8 Ma); 45-m-thick coral biostromes (Fig. 7E); and foraminiferal wackestones to packstones, dated to Zones P21–P22 (upper Rupelian–Chattian) in their lowermost part (Figs. 4 and 8, Supplemental Material 1–2). In the central part of the island, the Antigua Formation comprises, from bottom to top: ~10-m-thick, coarse-grained, mixed carbonate and volcanoclastic grainstone with corals, dating from Zones P18 to P21a (Rupelian, 33.9–28.1 Ma); 8-m-thick foraminiferal wackestones and coral beds dating from Zones P21b to P22 (upper Rupelian–Chattian, 29.2–23 Ma); and 5.5-m-thick planktonic foraminiferal wackestones dating from Zones P21b to P22 (Chattian) to P22b–N4 (upper Chattian–Aquitanian, 24–21 Ma) in the uppermost part (Figs. 4 and 8, Supplemental Material 2–3). The southeastern part of the Antigua Formation comprises foraminiferal wackestones of a few meters thick dated to Zones P18–P21a (Rupelian, 33.9–29.2 Ma) (Half Moon Bay) or Zones P22–N4 (upper Chattian–Aquitanian, 26.8–21 Ma; Devil’s Bridge; Figs. 3–4 and 8, Supplemental Material 1–2). The presence of several interbedded, mixed volcanoclastic-carbonate rocks indicates occasional mass flow deposition.

The Antigua Formation was deposited in a low-angle reefal carbonate ramp setting above the Basal Volcanic Suite and the Central Plain Group (Fig. 8). The sequence is composed of inner ramp coral biostromal beds (Fig. 7E) that change southeastwards into planktonic, foraminifera-rich wackestones. An upward deepening trend is also recorded from reefal deposits in the lower part of the formation to muddy, mid- to outer-ramp deposits in its uppermost part. Our biostratigraphic constraints show that the Antigua Formation was deposited from the late Rupelian (Zone P20, 30.3–29.2 Ma) to the late Chattian–Aquitanian (Zones P22–N4, 26.8–21 Ma; Fig. 8, Supplemental Material 2).

The Antigua Formation includes reworked and weathered volcanoclastic beds that we interpreted as turbidites or tempestites (Fig. 4, Supplemental Material 1), but nowhere did we find volcanic ash or pumice deposits interbedded in the Antigua Formation.

**Structure and Kinematics of Antigua**

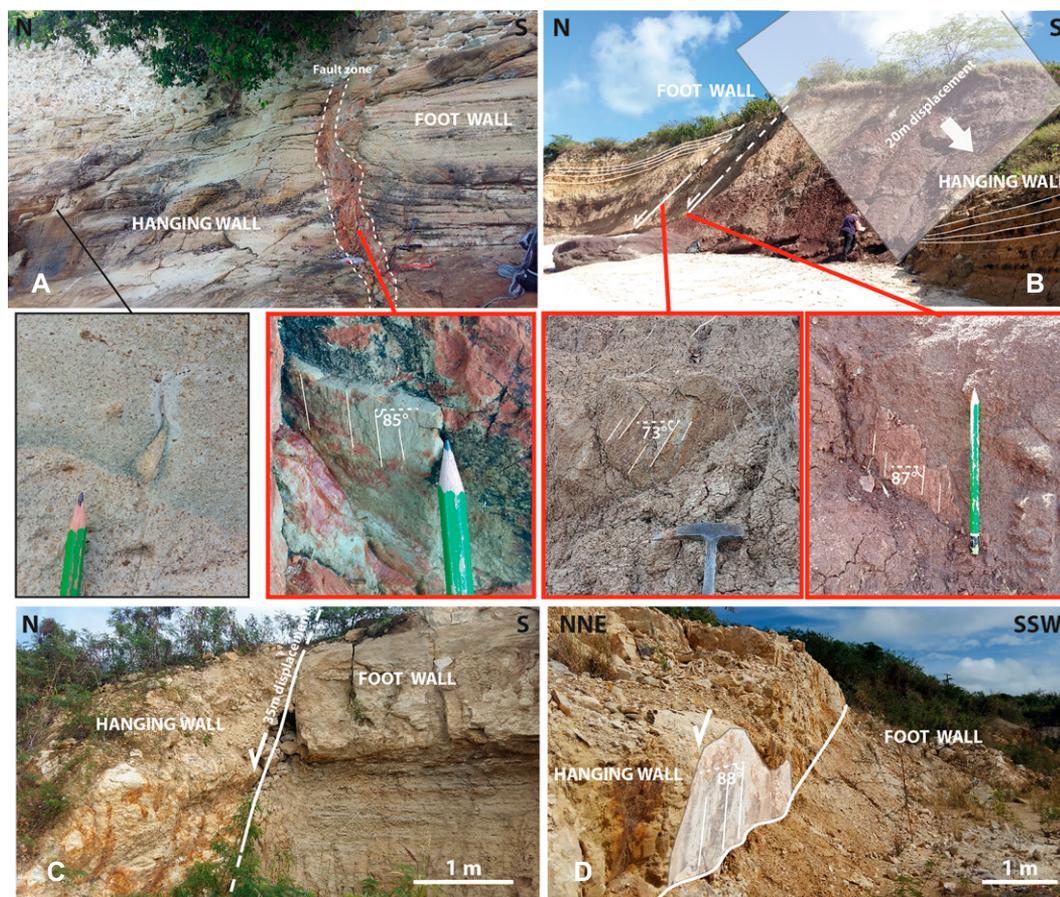
Mapped and outcrop-scale faults are normal faults that may be roughly subdivided into three clusters: ~N020° (±020°), N090° and ~N140° (±020°). On average, the fault planes dip ~60° and show dip-slip slickensides and displace-



**Figure 8. (A) Depositional settings of the Antigua Formation are shown with ages from the biostratigraphic investigations (Supplemental Material 1–2; see text footnote 1). (B) Synthetic sedimentary organization during the late Rupelian and the Chattian. The Antigua Formation was deposited on a northeastward, low-angle dipping ramp. The underlying Central Plain Group was previously faulted and tilted by NW–SE normal faults. For site locations, see Figure 3.**

ments of up to tens of meters (Fig. 9). The N020° and N090° faults are observed in the three units of Antigua, but the N140° faults are absent in the Antigua Formation. The Central Plain Group and Antigua Formation strike parallel to the N140° faults, which suggests that these faults are responsible for the main orientation of the island’s stratigraphy. The trend of the Antigua coastline is controlled by each of these three fault populations: the northwestern and southeastern shores trend NE–SW, the southern shore trends E–W, and the northeastern and southwestern

shore display a series of NW–SE-trending deep and narrow bays and capes. Although no direct crosscutting evidence has been found in the field, at map scale the N20° faults seem to intersect the N90° and N140° faults, especially to the north of Liberta Hill and Christian Hill and on the southeastern and northwestern coasts (Fig. 3). The N090° and N140° faults are well visible in the Central Plain Group, where they shape the landscape with the footwall, which consists of 100–200-m-high hills that trend NW–SE (Christian Hill and All Saint Hill, Fig. 3).



**Figure 9.** Field photographs show normal faults and associated high-angle striation at (A) Fort James, (B) Corbison Beach, and (C–D) Pares Quarry. Note the presence of injectites (caused by sediment injection during earthquakes), which are representative of syn-sedimentary deformation at Fort James.

At site scale, paleo-stress tensors indicate pure to radial extension with the direction of stretching varying from NNW–SSE to E–W. Strike-slip faults were observed only at one site (Old Road, Fig. 10A). Some sites display slightly oblique-slip normal faulting (Fig. 10A). Evidence for syn-sedimentary activity was observed for N090° faults at Fort James (injectites, Fig. 9A) and at Newfield (abrupt facies changes and reworked reefal blocks across the fault, Fig. 7B). At Nelson Dockyard, a N140°-trending fault was intruded by a mafic dike that we dated to  $27.03 \pm 1.89$  Ma (ANT20.12), which gives a minimum fault age.

As the N020° and N090° faults are observed in the three units of Antigua but the N140° faults are absent in the Antigua Formation, we assume that the N20° and N90° faults postdate the N140° faults. Moreover, there are no clear cross-cutting relationships between the N020° and N090° faults, which suggests that they may be synchronous. Therefore, while performing the kinematics inversion, we split our data set, separating N140°-trending faults from the N020°–N090° data set, according to our field constraints and observations. For the N140° fault set, we obtained a paleo-tensor with  $\sigma_1 = N255^\circ/80^\circ$ ,  $\sigma_2 = N120^\circ/0^\circ$ ,  $\sigma_3 = N020^\circ/10^\circ$ , which indi-

cates that these faults accommodated NNE–SSW stretching (Fig. 10C). The paleo-tensors obtained for the E–W- and N–S-trending faults show a subvertical  $\sigma_1$ , a sub-horizontal N130°-trending  $\sigma_2$ , and a sub-horizontal, N030°-trending  $\sigma_3$ , which indicates NNE–SSW stretching (Fig. 10C). Due to the limited number of observed faults and lack of kinematic indicators on fault planes, the quality of the tensors is poor (i.e., class D). Nonetheless, the obtained tensors provide no evidence for major changes in the stress field during deposition of the Antigua Formation; strain is minor and characterized by a NNE–SSW stretching direction that did not vary much during the Oligocene (Fig. 10C).

#### Anisotropy of Magnetic Susceptibility (AMS)

Site mean susceptibility ( $k_m$ ) of sites in the Central Plain Group ( $31.31$ – $76.55 \times 10^{-4}$  SI) and in the Antigua Formation ( $0.052$ – $25.59 \times 10^{-4}$ ) are indicative of a significantly high concentration of ferromagnetic minerals. Corrected anisotropy degrees are low ( $P' < 1.06$ ) with values ranging from 1.014 to 1.058, which suggests a relatively weak tectonic overprint (Table 3). The shape of AMS ellipsoid var-

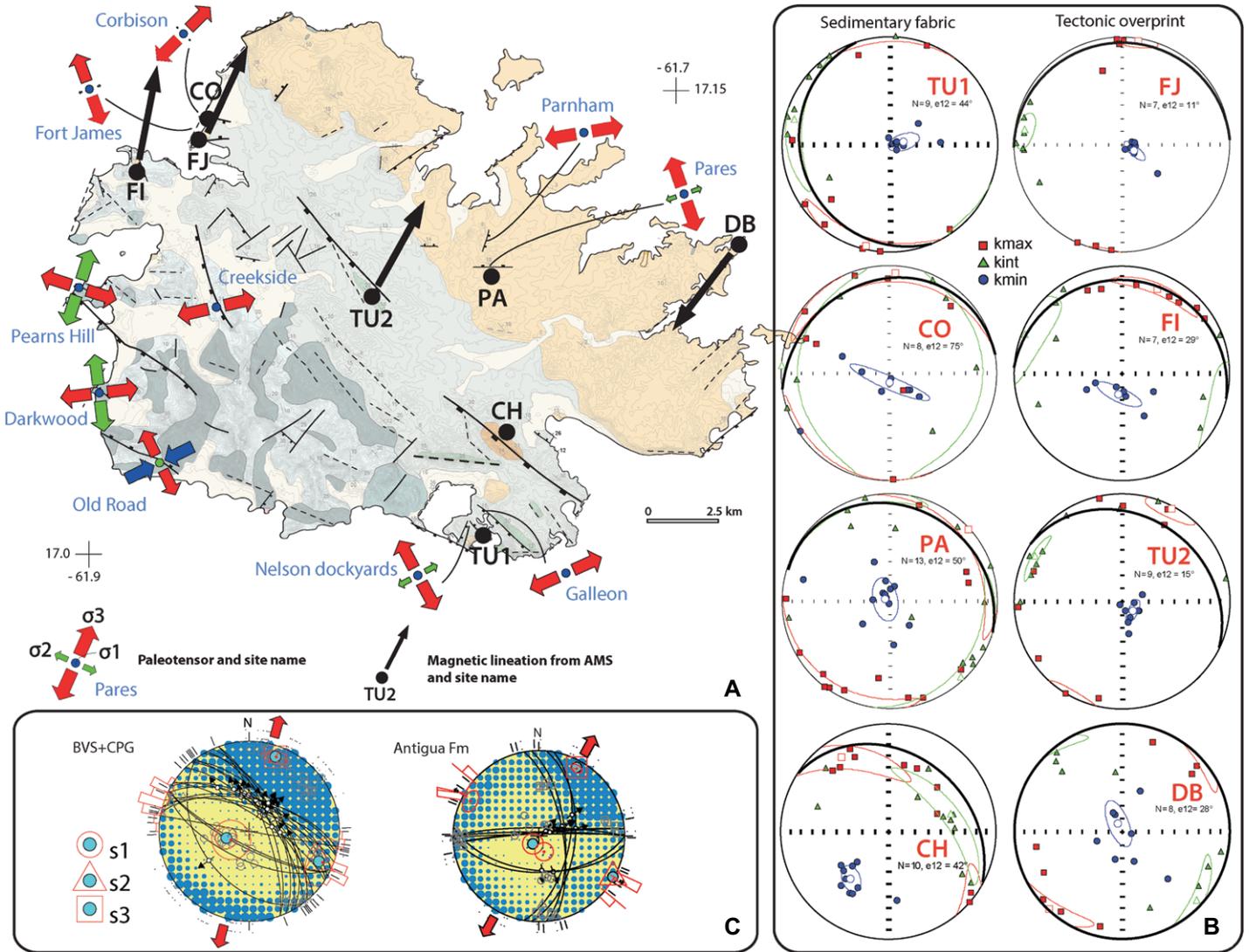
ies from oblate ( $T > 0$ ,  $F > L$ ) to triaxial ( $T \approx 0$ ,  $F \approx L$ ) with a dominance of oblate ellipsoids, which indicates that the tectonic fabric has only partially overprinted the original sedimentary fabric (Borradaile and Jackson, 2004; Table 3). Pure sedimentary fabrics are recognized at sites CH, TU1, CO, and PA, where  $k_{int}$  and  $k_{max}$  are randomly distributed within a plane sub-parallel to the stratification (Fig. 10B). Tectonic fabrics are recognized at sites TU2, FJ, FI, and DB, where a well-defined (i.e.,  $e_{12} < 30^\circ$ ) magnetic lineation is observed (Fig. 10B, Table 3). This magnetic lineation is sub-horizontal to shallowly plunging toward the NNE to NE (Fig. 10B). This direction is sub-parallel to the regional NNE–SSW stretching direction estimated from the fault kinematic analysis from the Basal Volcanic Suite, the Central Plain Group, and the Antigua Formation (Fig. 10C).

## DISCUSSION

### Geological Evolution of Antigua

#### Late Eocene–Early Oligocene Magmatic Activity

With a combination of field observations, lithostratigraphy, biostratigraphy, and  $^{40}\text{Ar}/^{39}\text{Ar}$



**Figure 10.** (A) Kinematic map shows Antigua with paleotensors (colored arrows) estimated per fault and magnetic lineation (black arrows) estimated from anisotropy of magnetic susceptibility (AMS) analysis. (B) Schmidt equal area projections of the AMS ellipsoid principal axis for all sampled sites in geographic coordinates; 95% confidence ellipses and associated mean values for each axis are represented by colored lines and empty symbols, respectively. Bold black arc represents the mean bedding plane of the site. (C) Paleostress tensor obtained from the field data (faults with kinematic criteria and fractures) for the NW–SE-trending faults within the Basal Volcanic Suite (BVS) and the Central Plain Group (CPG; to the left) and for the E–W- and N–S-trending faults of the Antigua Formation (to the right).

dating, we established the duration of magmatic activity in Antigua and subsequent sedimentation. The lithostratigraphy of Antigua shows that magmatism formed the Basal Volcanic Suite and was active during the deposition of the volcanoclastic Central Plain Group but not during the deposition of the Antigua Formation (Fig. 3). The stratigraphically lowest (tuffites) and highest (lacustrine limestones with volcanic debris) units of the Central Plain Group that we observed in the field contain evidence of explosive volcanism (Supplemental Material 1). In the Antigua Formation, all volcanic elements are reworked as clastic sediment, but no evidence of explosive (ash deposits or pumice) or effusive volca-

nism (intercalated lavas) was observed (Fig. 4, Supplemental Material 1). Our field observations corroborate seismic interpretations of lines offshore Antigua, where no magmatic features (conic shaped bumps) were identified in megasequences interpreted to reflect the Oligocene and Neogene deposits (Cornée et al., 2021; Fig. 11, Supplemental Material 3).

From our  $^{40}\text{Ar}/^{39}\text{Ar}$  ages of a rhyolitic dike (sample DWB-B) and a dacitic dome (sample ANT 6.8) crosscutting the andesite-pyroclastic suite ( $35.32 \pm 0.16$  Ma and  $29.76 \pm 0.22$  Ma, respectively; Fig. 6, Table 2), and the K/Ar age of Nagle et al. (1976) of an andesitic lava ( $38.5 \pm 2$  Ma, Fig. 1, Table 1), we infer that the

bulk andesitic and explosive volcanism of the Basal Volcanic Suite occurred in the late Eocene.

We discarded our age of  $18.57 \pm 1.26$  Ma for the andesite from the Basal Volcanic Suite (sample ANT 20.4) because it is inconsistent with the age of the rhyolitic dike that crosscuts it. We assume that this sample was fully reset during a late (post-magmatic) hydrothermal event as exemplified by the very high atmospheric contamination within. This is supported by the occurrence of various degrees of subgreenschist hydrothermal alteration within the Basal Volcanic Suite (albitized plagioclases, chloritized orthopyroxenes, and groundmass; Gunn and Roobol, 1976), the Central Plain

TABLE 3. ANISOTROPY OF THE MAGNETIC SUSCEPTIBILITY (AMS) RESULTS

Site	Formation	Rock type	Age	Latitude (°N)	Longitude (°W)	Beddings	N	kmean	L	F	P'	T	D (°)	I (°)	e12	AMS origin
CH	CPG	Lacustrine Limestone	Rupelian	17°01'22.0"	61°44'29.2"	310/21	10	31.31	1.007	1.04	1.052	0.694	327.9	19.1	53°	S
TU1	CPG	Tuffite	Rupelian	17°00'19.6"	61°45'53.4"	155/14	9	58.80	1.003	1.043	1.052	0.887	193.5	4.4	44°	S
TU2	CPG	Tuffite	Rupelian	17°05'01.1"	61°48'25.8"	316/22	9	32.09	1.006	1.036	1.047	0.696	27.8	2.8	15°	T
CO	CPG	Volcaniclastic	Rupelian	17°08'26.1"	61°51'33.3"	260/16	8	76.55	1.005	1.018	1.025	0.388	3.5	7.2	75°	S
FI	Antigua Fm.	Limy sandstones	L. Rupelian	17°07'51.0"	61°51'49.2"	267/5	7	25.59	1.006	1.047	1.058	0.757	8.9	3.1	11°	T
FA	Antigua Fm.	Limestone	E. Chattian	17°07'15.1"	61°52'48.4"	283/12	7	1.379	1.001	1.012	1.014	0.815	22.5	15.6	29°	T
PA	Antigua Fm.	Clayish limestone	Chattian	17°05'31.3"	61°45'49.7"	290/10	13	0.363	1.005	1.014	1.021	0.377	47.4	1.8	50°	S
DB	Antigua Fm.	Limestone	L. Chattian	17°05'59.1"	61°40'46.2"	0/0	8	0.052	1.181	1.073	1.305	0.026	223.5	2.5	28°	T

Notes: Bedding is expressed as strike/dip. CPG—Central Plain Group; Fm.—formation; N—number of samples studied at a site; kmean—site mean susceptibility in 10–4 SI; Magnetic lineation (L), magnetic foliation (F), corrected anisotropy degree (P'), and shape factor (T) are according to Jelínek and Kropáček (1978). D—Declination; I—Inclination of the maximum susceptibility axis; e12—semi-angle of the 95% confidence ellipse around the declination. "AMS origin" indicates the interpreted nature of the anisotropy of magnetic susceptibility (T—tectonic; S—sedimentary).

Group (chloritized tuffites; Jackson, 2013); and the Antigua Formation (silicified limestones; Strang et al., 2018). Such pervasive subgreenschist hydrothermal metamorphism has been described within the recent arc in Guadeloupe and may take place at 1 km to 3–4 km depth (Favier et al., 2019).

Late volcanic activity occurred in Antigua at the transition between early and late Oligocene and is exemplified by the emplacement of a dacitic dome through the Basal Volcanic Suite and basaltic andesite dikes through the Central Plain Group and the Antigua Formation. The dacitic dome yields a late Rupelian age of 29.76 ± 0.22 Ma (ANT 6.8, Sugar Hill). The mafic dikes intruding the base of the Antigua Formation give an age of 30.50 ± 0.70 Ma (ANT 20.6) that is consistent with the biostratigraphic dating to the late Rupelian (Zone P20, 30.3–29.2 Ma). As the Central Plain Group is synchronous with magmatic activity of the Basal Volcanic Suite and capped by the Antigua Formation, its age must range from at least 38–30 Ma (late Bartonian to late Rupelian). The 27.03 ± 1.89 Ma age of a dike crosscutting the Central Plain Group is the youngest evidence for magmatic activity on Antigua (Figs. 5A–5B).

The quartz diorite stock, now exposed at Old Road, gives only a minimum age of 28 Ma, and could represent a late-stage intrusion (ANT 20.8, Fig. 9, Table 2).

**Late Oligocene–Early Miocene Carbonate Platform**

Following the end of magmatic activity, we observe a change from terrestrial (Central Plain Group) to marine environment (Antigua Formation) during the Rupelian–Chattian transition. The depositional environment of the Antigua Formation varies both geographically from inner ramp coral biostromal beds (Fig. 7E) to the northwest to planktonic, foraminifera-rich wackestones to the southeast and temporally with reefal deposits in the lower Rupelian part of the sequence to muddy, mid- to outer-ramp deposits in its uppermost part (late Chattian to Aquitanian). We reinterpreted the marine carbonate deposits, previously described as interbedded within the Central Plain Group (Masle and Westercamp, 1983), as residual exposures of the Rupelian lower part of the Antigua Formation that are preserved in grabens. This indicates that the Antigua Formation was entirely covering the Central Plain Group and, consequently, that a transgressive event occurred in Antigua during the formation of the P20 Zone (30.3–29.2 Ma, Rupelian). The late Rupelian–Chattian global sea-level variations (tens of meters; Miller et al., 2020) are smaller than those of the depositional environment of the Antigua Formation (tens of

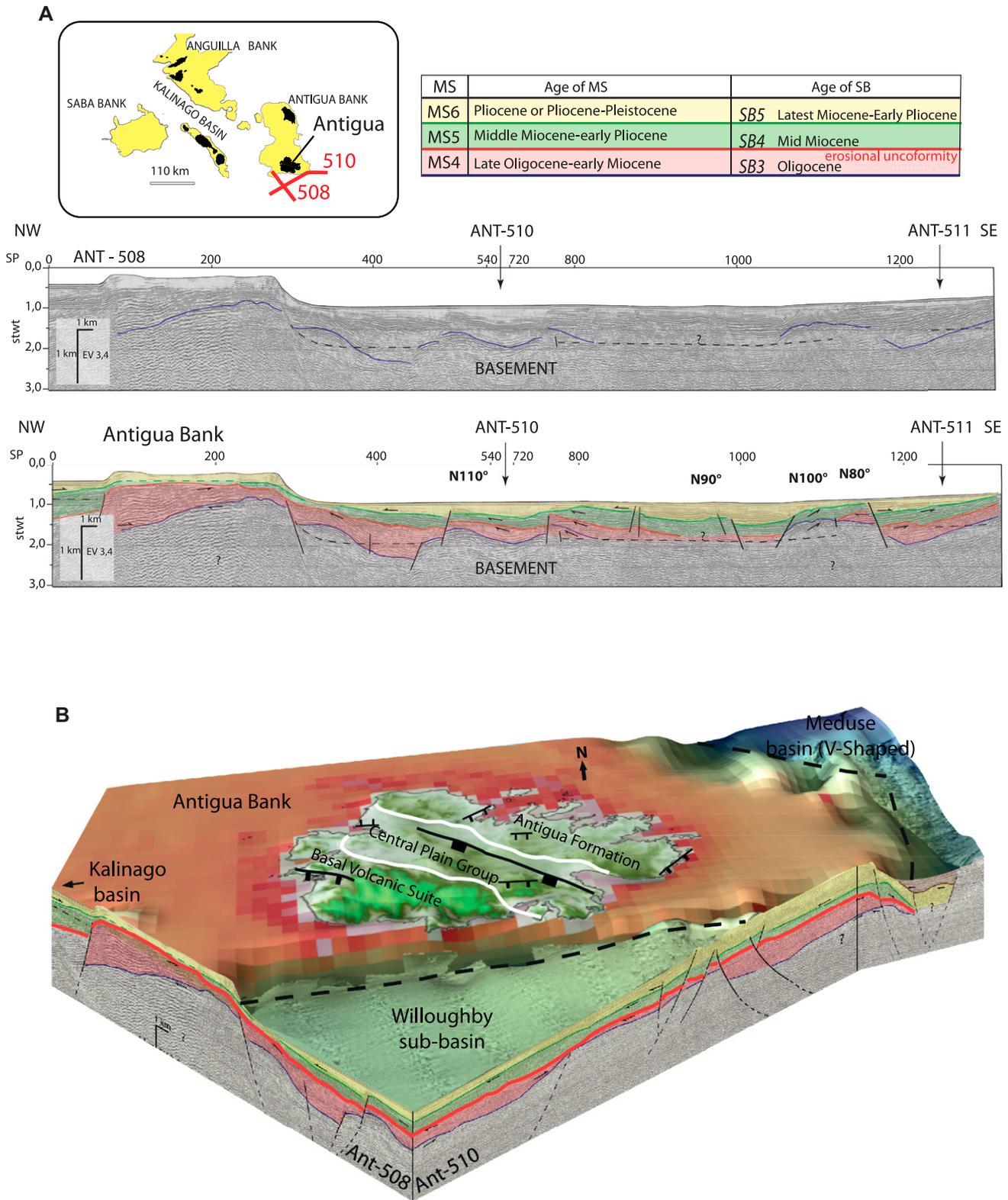
hundreds of meters). Consequently, we infer that the island subsided during the late Rupelian–Chattian interval. Thermal subsidence upon cooling after magmatic activity could have played a role (Moore, 1970; Tallarico et al., 2003).

To account for the presence of centimeter-long, subvertical pressure-dissolution stylolitic peaks (Figs. 7C–7D) the Antigua Formation must have undergone vertical compaction related to a post-Aquitania sedimentary pile that was later eroded. In the neighboring volcanic area, this erosional event and associated exhumation is supported by the outcropping of sub-greenschist facies units that are thought to have originated from 1 km to 3–4 km depth (Favier et al., 2019). In the offshore domain, megasequence 4 (MS4, Oligocene to early–middle Miocene in Cornée et al., 2021), which can be correlated to the Central Plain Group and Antigua Formation, is topped by a major erosive sequence boundary (SB5) that is dated as early–middle Miocene (Fig. 11, Supplemental Material 3) and recognized regionally (Cornée et al., 2021; Boucard et al., 2021). We thus suggest that erosion occurred during early–middle Miocene time. After this period, the island likely remained above sea level as no post-Aquitania sediments (aside from Quaternary alluvium) can be observed. Meanwhile, the Willoughby basin subsided along the N090° faults bounding the Antigua bank to the South (Fig. 11). This stable position appears to be consistent with the slow cooling estimated using QTQt modeling of the <sup>40</sup>Ar/<sup>39</sup>Ar spectrum of plagioclase from the quartz diorite (~5° per Ma; sample ANT 20.8, Fig. 6) since at least 25 Ma; a similar cooling rate was recently reported from the granodioritic plutons of St-Martin island (Noury et al., 2021).

**A 20-m.y.-Long Magmatic Lull**

The geological history we unraveled from Antigua thus includes (1) a main magmatic pulse occurring in Antigua during the late Eocene (continuing to at least 35 Ma), (2) magmatic activity continuing mainly during the early Oligocene in the form of late-stage dikes and plugs (30.5–27 Ma), (3) magmatic activity ceasing during the late Oligocene synchronously with a regional transgressive event and a subsidence of the island that led to the development of a carbonate platform completely devoid of evidence of volcanism; and (4) an early–middle Miocene exhumation and uplift event responsible for the removal of a part of the volcanic and sedimentary series after which Antigua remained emerged and stable.

From our new age model, we find that in the northern Lesser Antilles, the main activity of the



**Figure 11. (A)** Time-calibrated seismic stratigraphic analysis of the seismic profiles CPEM 508 and 510 (in red; ARCANTE 3 Cruise, Bouysse et al., 1985; Bouysse and Mascle, 1994) from Cornée et al., 2021. Top: uninterpreted; below: interpreted. The definition and time calibration follow the seismic stratigraphic chart defined by Cornée al. (2021) in the northeastern Lesser Antilles. MS—megasequence; SB—sequence boundary. **(B)** Three-dimensional block with seismic profiles 508 and 510 shows the onshore and offshore structures in Antigua, Willoughby basin, and the V-shaped Méduse basin. Note the erosional character of the SB4 unconformity that seals prior faults affecting MS4.

Paleogene arc occurred between the late Eocene and the early Oligocene, and we show that magmatism—which was thought to have possibly continued into the Miocene based on K-Ar ages (Nagle et al., 1976; Briden et al., 1979)—waned and came to an arrest by ca. 27 Ma in Antigua. Small-volume, late-stage magmatism on nearby islands St-Martin (24.4 Ma; Cornée et al., 2021) and St-Barthélemy (24.5 Ma; Legendre et al., 2018) continued slightly longer. The recent arc, located further west in the islands of Saba, St-Kitts, Monserrat, and Guadeloupe, has been active since the Pliocene. Our analysis thus supports a long magmatic lull in the northern Lesser Antilles arc despite ongoing subduction at more or less constant rates (Leroy et al., 2000; Boschman et al., 2014). With our new age model, which includes  $^{40}\text{Ar}/^{39}\text{Ar}$  dates and lithostratigraphy, we more accurately constrain it to a ca. 20 Ma period lasting from the early late Oligocene to the Pliocene. As magmatic flareup contributes to crustal thickening (Ma et al., 2022), the Paleogene activity of the Lesser Antilles arc should have contributed to the building of the GrANoLA land (Philippon et al., 2020a), an emerged area that was spread out through the northern Lesser Antilles to Puerto Rico and favored terrestrial fauna dispersal from South America to the Greater Antilles (e.g., Marivaux et al., 2020). The end of magmatic activity and subsequent thermal and tectonic subsidence could have contributed to the removal of this pathway during late Oligocene. This shows the importance of understanding the magmatic history of the Lesser Antilles arc in the frame of the paleo-(bio-)geographical history of the eastern Caribbean region.

### A Correlation Between the End of Magmatism and Tectonic Changes?

In the northern Lesser Antilles, a correlation between tectonics and magmatism is suggested by the NW–SE orientation of the Paleogene arc that aligns with the eastern flank of the NW–SE-oriented Kalinago basin. This basin opened from the Eocene to the late Oligocene, synchronously with magmatic activity (Cornée et al., 2021).

At the scale of Antigua, we found evidence for syn-magmatic and syn-sedimentary tectonic activity. Within the Central Plain Group, we estimate a pre- or Rupelian age for the activity of a normal fault at Nelson Dockyard as it is intruded by a dike dated at 27 Ma (this study). Furthermore, the presence of injectites at Fort James (Fig. 9A), abrupt facies changes and reworked reefal blocks in Newfield (Fig. 7B), and the NE orientation of the magnetic lineation obtained from Central Plain Group and Antigua Formation sites, are consistent with the NE–SW

stretching orientation obtained from paleotensors, which strongly suggests syn-sedimentary tectonic activity.

Synchronously with the arrest of magmatism, we observe a marine transgression and a tilt of the Central Plain Group that resulted in a 5° difference in the mean dip beddings between the Central Plain Group and the Antigua Formation. This strongly suggests that the late Rupelian–Chatian transgression reflects tectonic subsidence, although thermal subsidence could also have played a role. The gentle northeastward 5° dip of the Antigua Formation and a general deepening of the paleo-environment toward the uppermost part of the sequence indicate that the subsidence was active until at least the early Miocene.

Our structural observations and AMS measurements do not show a clear change in deformation behavior or direction during the Oligocene. Both paleostress tensors indicate a NNE–SSW directed stretching, which remained consistent over time. In the Basal Volcanic Suite and Central Plain Group, this stress field is accommodated by dominant N140° faults and N090° and N020° normal faults. In the Antigua Formation, N140° faults are not recognized, and the stress is accommodated by N090° and N020° faults. The absence of N140° faults could be indicative of a lowering of the intensity of the deformation. AMS results show that magnetic lineations acquired during syn-tectonic deposition are consistent with the NNE–SSW stretching direction estimated by paleotensors. This confirms that this stretching direction was dominant in Antigua from the early Oligocene to the early Miocene even though it was accommodated by different families of faults.

With our field and kinematic investigations, we did not see (1) significant deformation in the Antigua Formation and (2) a change in stress field during the Oligocene. The decrease or change in tectonic activity on the island is subtle and only outlined by the absence of N140° faults in the Antigua Formation. Thus, it is unlikely that the end of magmatism can be correlated to a change in tectonic activity on Antigua.

On the islands of St-Martin and St-Barthélemy, a switch from pure orthogonal-to-the-trench to radial extension is recognized during the early Miocene (Legendre et al., 2018; Noury et al., 2021). In the northern Lesser Antilles region, a change in upper-plate deformation style from back arc rifting (Kalinago basin; Cornée et al., 2021) to fore-arc V-shaped basin opening (Boucard et al., 2021) as well as drastic vertical motion (Cornée et al., 2021; Boucard et al., 2021) is observed during the Oligocene. Along with the arrest of magmatism, these changes in upper-plate deformation and vertical motion are

certainly controlled by subduction-related processes and slab parameter variations.

### What Controls Lesser Antilles Magmatic Flare-Ups and Lulls?

The oldest volcanic rocks of the Lesser Antilles arc are ca. 42 Ma (Philippon et al., 2020a), some 10 m.y. younger than the NE–SW to E–W change in Caribbean–North American plate motion that led to the formation of the modern day N–S-trending Lesser Antilles subduction zone (Briden et al., 1979; Boschman et al., 2014).

Such a delay is often seen between the onset of subduction and the development of a mature arc (e.g., Stern et al., 2012). During ca. 15 Ma (late Eocene to late Oligocene), volcanic edifices developed on the eastern flank of the Kalinago basin, and in this period, the arc–trench distance, and by inference, the slab angle, remained more or less constant. But by the end of the Oligocene, magmatism ceased in the northern Lesser Antilles.

Several hypotheses have been proposed to explain this arrest. Bouysse and Westercamp (1990) speculated that slab break-off occurred following the subduction of a postulated buoyant, thick oceanic crust. But seismic tomographic images reveal a continuous, slab reaching the mantle transition zone without evidence of breaks (van Benthem et al., 2013; Braszus et al., 2021), and the subduction rate has not changed (Boschman et al., 2014). If a slab breakoff had occurred, a new slab would have had to quickly form to re-instate the arc as observed.

Boucard et al. (2021) suggested that subduction erosion occurred during middle Miocene time. Subduction erosion could push an arc toward the hinterland, but the timing of this event (post-middle Miocene) cannot explain the late Oligocene onset of the magmatic lull, and subduction erosion would cause an arc to migrate rather than stop its activity.

Slab flattening is then a third option. Subduction of buoyant features (oceanic aseismic ridges, oceanic plateaus, or seamount chains) is commonly thought to possibly trigger flat-slab subduction at mantle-stationary trenches (Yang et al., 1996; Gutscher et al., 2000; Gerya et al., 2009; Martinod et al., 2013; Ma et al., 2022). In the Lesser Antilles region, McCann and Sykes (1984) were the first to propose that a former anomalously thick aseismic ridge connecting the Barracuda Ridge and the Main Rise (Fig. 1) and/or fragments of the Bahamas platform interacted with subduction and could have triggered slab flattening and westward migration of the volcanic center. To the north, the thin continental margin below the Bahamas bank diachronously

subducted and entered the Greater Antilles subduction zone in the late Eocene. At that point, subduction of Cuba ceased (Erikson et al., 1990; Mann et al., 2002; Iturralde-Vinent et al., 2008; Granja-Bruña et al., 2010; Cruz-Orosa et al., 2012; Laó-Dávila, 2014). The eastern, oceanward end of the Bahamas bank had already interacted with the NE Caribbean margin during the Caribbean plate motion change of the mid-Eocene, and there is no clear reason to infer that this interaction allowed for an arc in the late Eocene and early Oligocene but not in the 20 m.y. that followed.

So, while slab flattening may be a viable explanation for the magmatic lull, the obvious solutions to explain the arrival of buoyant features do not seem to be straightforward solutions.

As shown by recent reconstructions of the Caribbean region, a change in the age of the subducted oceanic lithosphere, from a young Proto-Caribbean oceanic crust to an old American oceanic crust, is observed in the eastern Caribbean

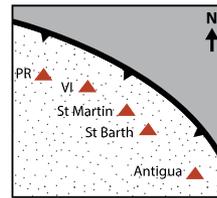
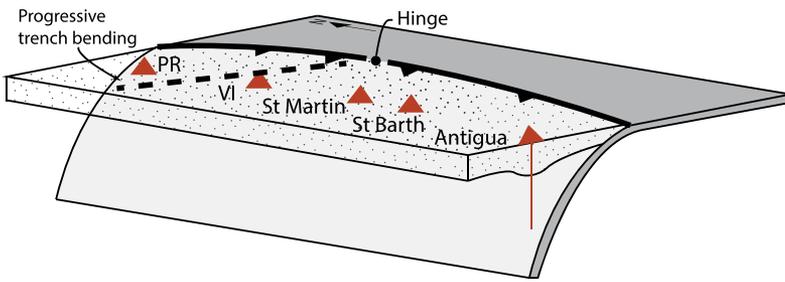
subduction zone during the end of Oligocene (Brasus et al., 2021). Nevertheless, variation in the age of the subducted oceanic lithosphere may not be, by itself, sufficient to change the slab dip (Cruciani et al., 2005).

Schepers et al. (2017) showed, at the Andean subduction zone, that flat slabs may also arise when roll-back of a subducted slab is locally impeded and sub-slab mantle begins to support a flat slab. But the Lesser Antilles trench can be considered to have been more or less stable relative to the mantle since 50 Ma following the kinematic reconstructions of Boschman et al. (2014), and roll-back was short-lived and restricted to the southern portion of the trench, where the Grenada-Tobago basin opened in the Eocene (Aitken et al., 2011) prior to the magmatic lull. So the scenario for the Andes does not apply to the Lesser Antilles arc.

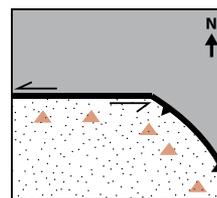
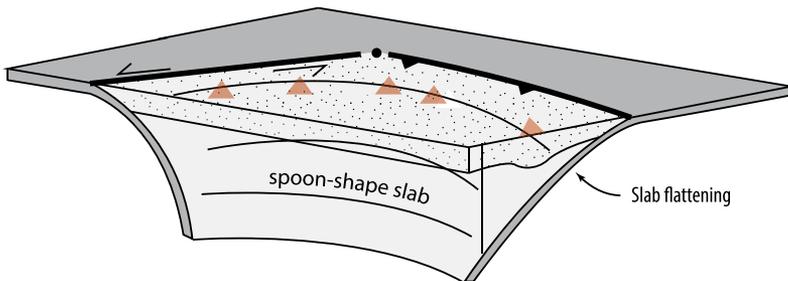
We suggest that slab flattening in the northeastern corner of the Caribbean plate may have originated from the progressive formation of

the spoon-shaped, curved slab (Laurencin et al., 2017; van Benthem et al., 2013; Brasus et al., 2021). The westward motion of North America relative to the nearly mantle-stationary Caribbean plate (Boschman et al., 2014) and Lesser Antilles trench must be associated with westward dragging of the south-dipping Puerto Rican part of the slab (Spakman et al., 2018; van de Lagemaat et al., 2018; Parsons et al., 2021). With ongoing absolute westward motion of North America, the E-W segment of the Puerto Rican segment becomes longer (Boschman et al., 2014), and in the curved area, the slab may flatten (Fig. 12). This curvature and flattening could have been minimal in the Eocene, allowing for a regular volcanic arc, but it became progressively more intense, which may explain the long magmatic lull (Cerpa et al., 2021). We speculate that the recent reinstatement of arc magmatism in the northern Lesser Antilles may relate to a recently observed seismically imaged gap in the slab in the bend area

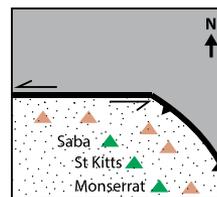
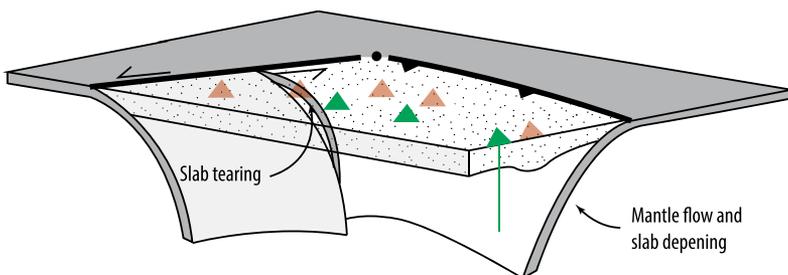
**Paleogene magmatic flare-up : Ante 50 - 25 Ma**



**Oligo-Miocene magmatic lull : 25 - 5 Ma**



**Pliocene magmatic flare-up : 5 Ma - present**



**Figure 12.** Schematic 3-D sketch shows the progressive curvature of the Lesser Antilles slab from Eocene to present and its impact on slab dip variation and the location of magmatism.

(Meighan et al., 2013) that could have allowed the slab to steepen (Fig. 12).

Evaluating whether this may explain the Pliocene magmatic flareup requires further dynamic and magmatic analysis and is beyond the scope of this paper. We note, however, that if the magmatic lull is caused by slab flattening, the Pliocene flareup may provide a proxy for reorganization in slab dynamics in the northeastern Caribbean region. A detailed kinematic reconstruction of the northeastern Caribbean plate boundary zone back to the onset of magmatism would be helpful for establishing the temporal and spatial relationships between subduction, the length of the E–W trench segment east of Puerto Rico, and magmatism.

## CONCLUSION

In this study, we revised the geological architecture and history of Antigua and integrated our findings within the regional framework of both the evolution of the Lesser Antilles arc and the Caribbean upper-plate strain pattern.

Using lithostratigraphy, structural mapping,  $^{40}\text{Ar}/^{39}\text{Ar}$  dating, and biostratigraphy, we revised the geological map of the island and provided a new age model for magmatism and sedimentation of the island. The climax of magmatic activity and concomitant volcanoclastic deposition emplaced during the late Eocene and continued until early Oligocene in the form of late-stage domes and dikes. From the early Oligocene to, at least, the early Miocene, the island subsided along NW–SE faults, resulting in subsequent transgression and deposition of a carbonate platform. The island was then uplifted and partially eroded during the early–middle Miocene, and since that time Antigua has remained emerged and stable. With a kinematic analysis coupling paleotensors and anisotropy of magnetic susceptibility (AMS), we show that minor normal faulting accommodates NE–SW stretching and that, at the island scale, there is no clear change in deformation style concurrent with the end of magmatism during the Oligocene.

This integrative geological study of Antigua restricts the period of magmatic activity on this island and definitively confirms the existence of a 20-m.y.-long magmatic lull in the northern Lesser Antilles. The Paleogene arc, which is also recognized in St-Martin and St-Barthélemy, was emplaced along the eastern flank of the Kalinago basin synchronously with its rifting. The late Eocene to early Oligocene age of the Lesser Antilles arc flare-up is contemporaneous with the emersion of a large area in the northern Lesser Antilles (GrANoLA land). This suggests that the magmatic activity, through crustal thickening, may have influenced the paleo-(bio)

geography of the region and played a role in the rise of emerging lands used as pathway by terrestrial fauna. The lull in magmatic activity and associated thermal subsidence may have, in turn, contributed to the demise of such lands. This subsidence will trigger the apparition of non-connected islands leading to isolation of fauna and flora and distinct evolution of the populations. This could, in part, explain the present-day high endemic rate that characterizes the biodiversity of the Antilles.

During the Oligocene, magmatism ceased contemporaneously with a change in upper-plate deformation from back-arc and perpendicular to the trench stretching (Kalinago basin) to fore-arc and parallel to the trench stretching (V-shaped basins). This episode was followed by regional uplift during the early to middle Miocene. We postulate that the end of magmatic activity and the change in the style of upper-plate deformation and vertical motion may have been triggered by slab flattening caused by the progressive formation of the spoon-shaped, curved slab. Nevertheless, this is still hypothetical. Further study of the kinematic evolution of the northeastern Caribbean plate boundary back to the Eocene is needed to better understand the relationship between magmatism, upper-plate deformation, and subduction.

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