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Earth-Science Reviews

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	need for future crises to be effectively managed so as to limit the socio-economic impacts and maintain the relationship between scientists, authorities and the population.
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Diverse Magma Storage and Major and Volatile Magma Composition: What are the implications on the Eruptive Style Across a Volcanic Arc? An Example of the Lesser Antilles Arc.

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Key Words: Lesser Antilles; Geochemistry; Magma Storage; Eruption Style; Arc Volcanism; Melt Inclusions; Magma Mush, Volcanic Unrest, Volcanic Eruption

Abstract

Understanding variations in eruptive style and defining the processes and controls on different eruption styles is particularly important at arc volcanoes. This is due to the range of eruption styles which vary from phreatic behaviour to effusive magma extrusion and large explosive Plinian eruptions. Such eruptions can have wide-ranging and long-term impacts on vulnerable communities living near volcanoes. Magma storage and chemistry are key to understanding the origins of such variations along-arc. The Lesser Antilles Arc provides a perfect setting to investigate magma storage, magma composition and eruptive style variations associated with arc volcanism. This arc is comprised of eleven volcanically active islands. Since 1900, there have been 24 eruptions. Many eruptions have resulted in thousands of deaths, large scale and long-term displacements and severe economic impacts, which have had profound impacts on the people living in this region. Through this review we highlight the variations observed in magma storage, which is described as a vertically extensive trans crustal mush system. We hypothesise that across the arc there are generally three levels of the mush storage system. First, a deep system at 25 km (ca. >400 MPa). This system is poorly constrained and contains the most primitive magmas which fractionate to produce magmas which ascend into the second mid-crustal system at 10 – 18 km (ca. 3200 - 400 MPa). The mid-crust stores basalt to basaltic

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Figure 1

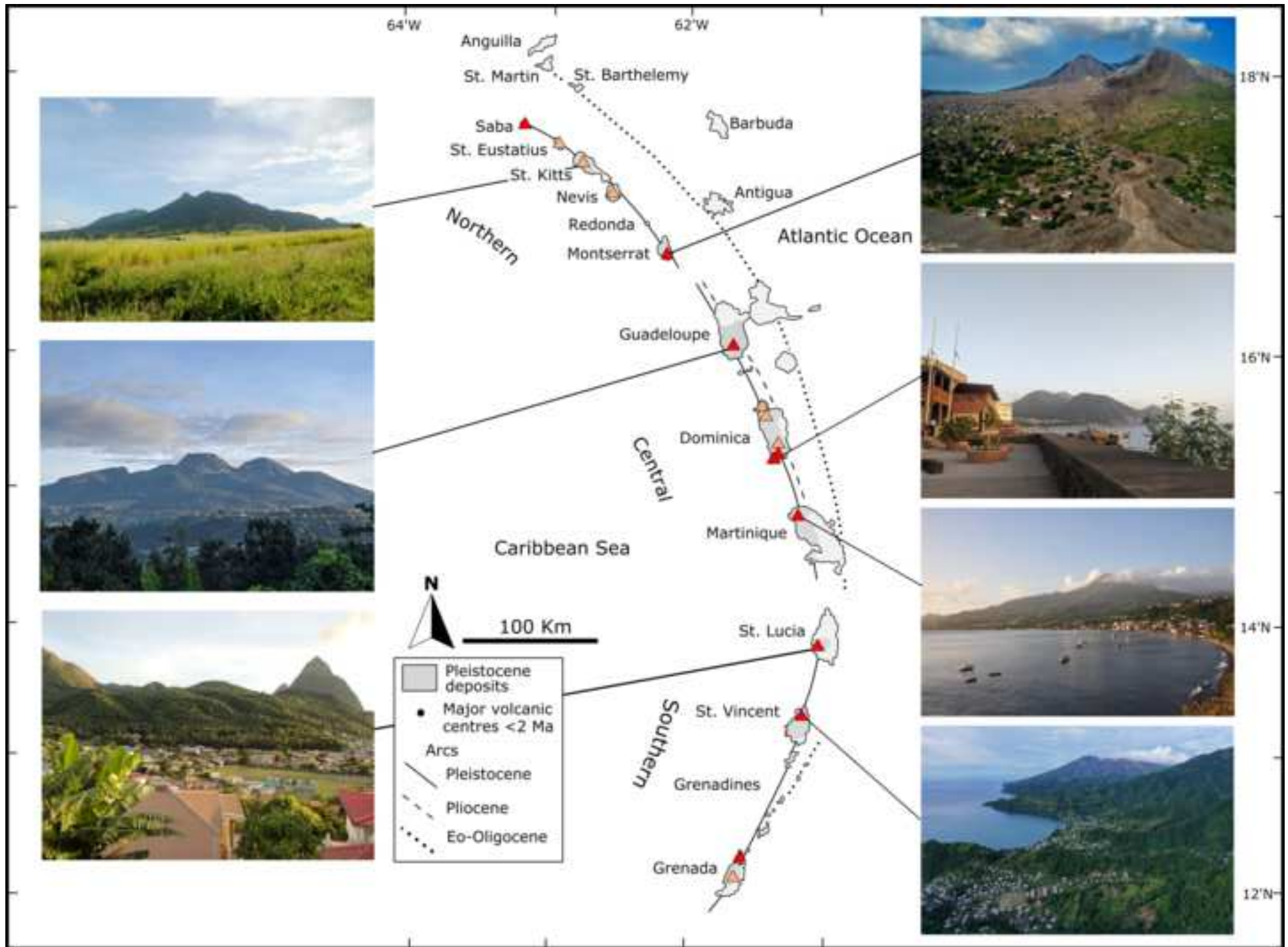


Figure 2

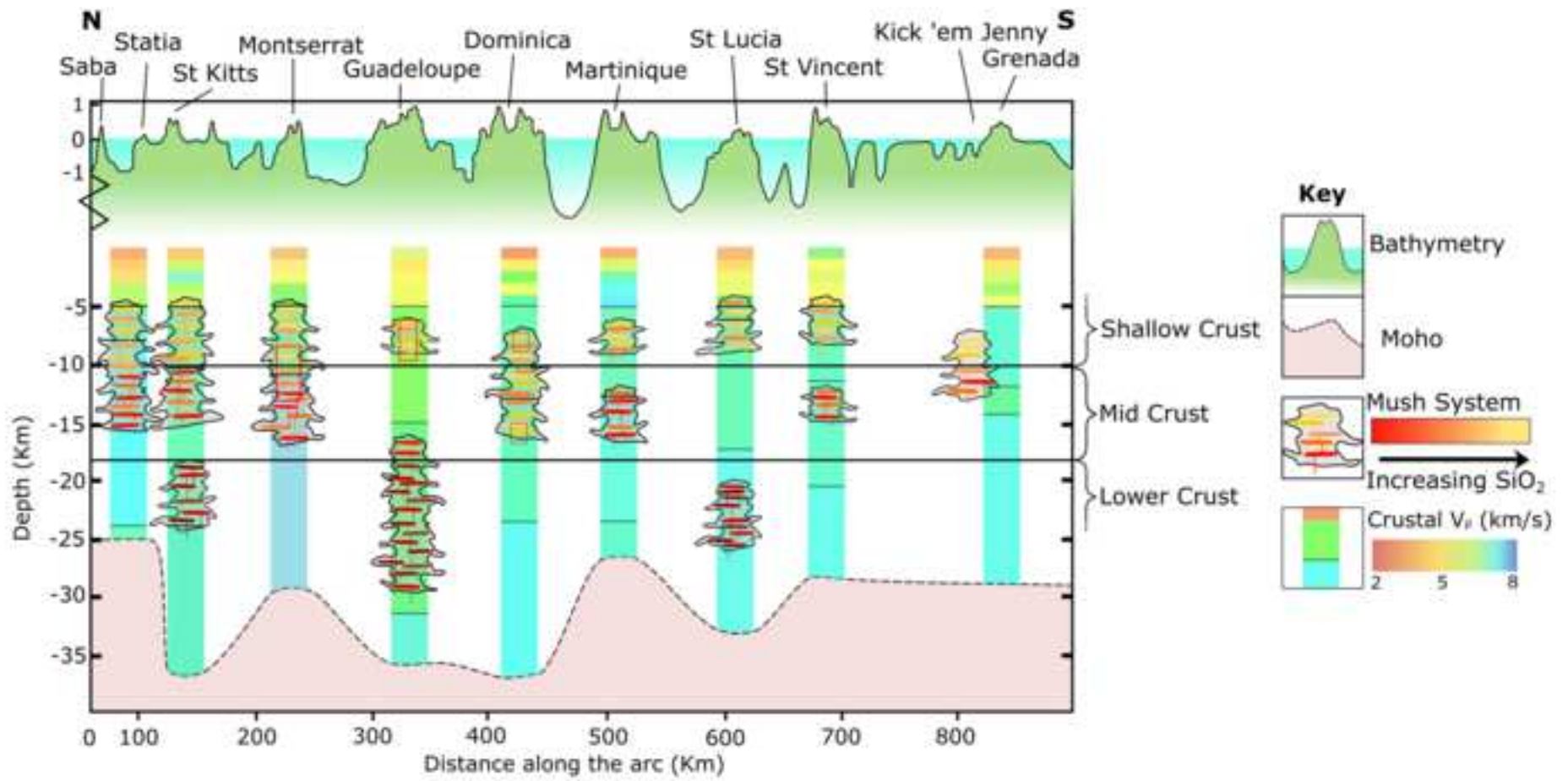


Figure 3

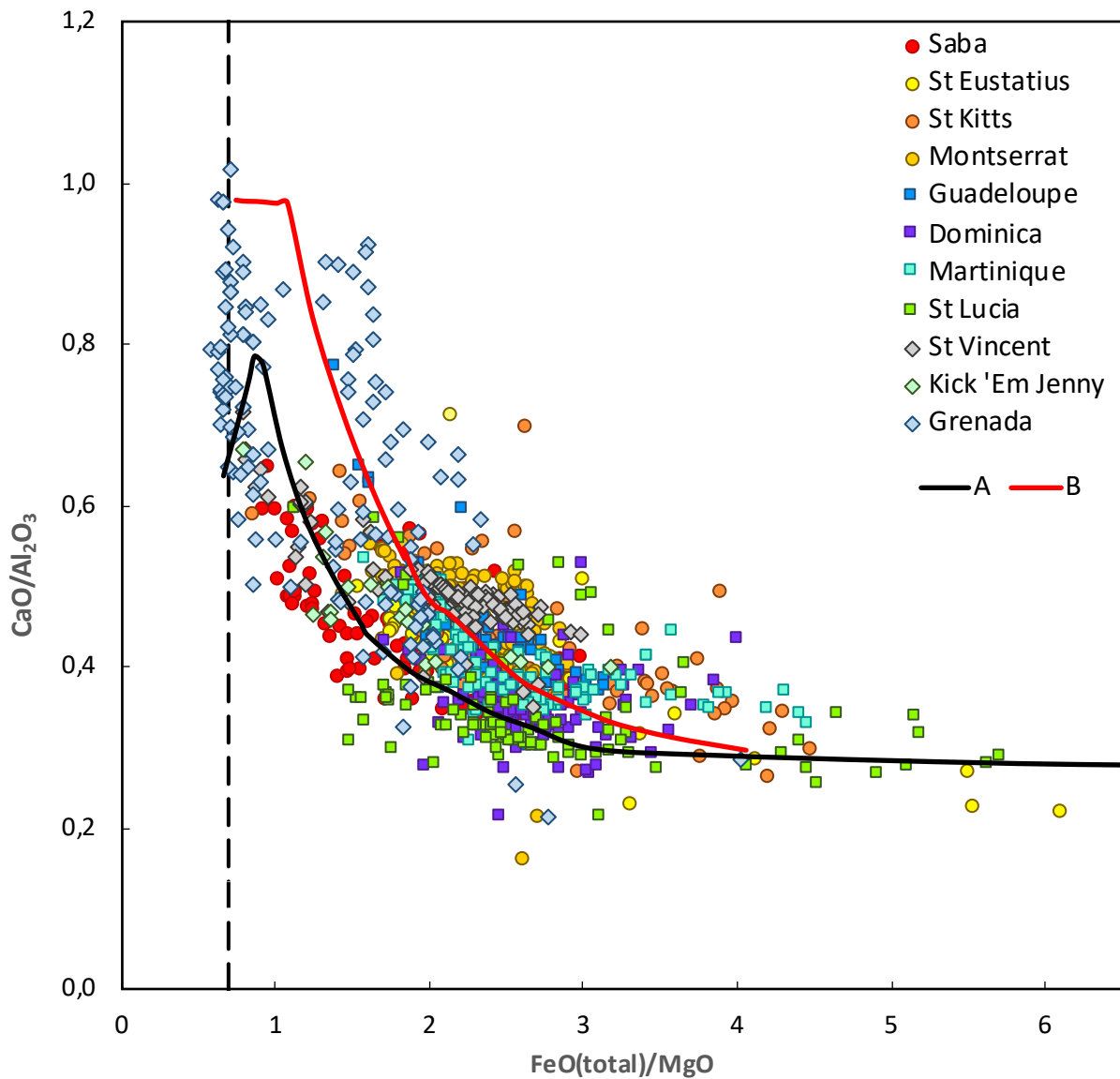


Figure 4

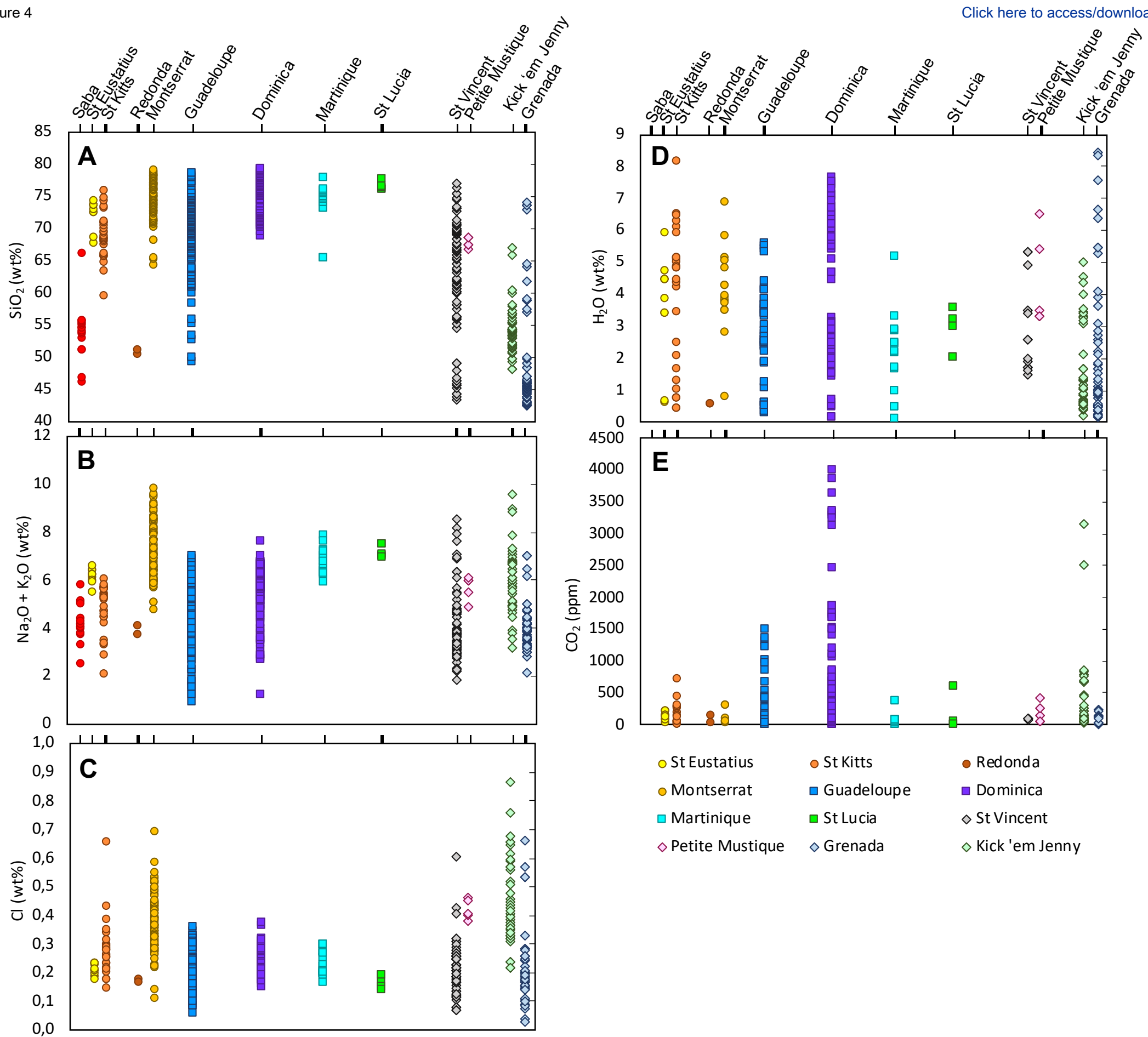
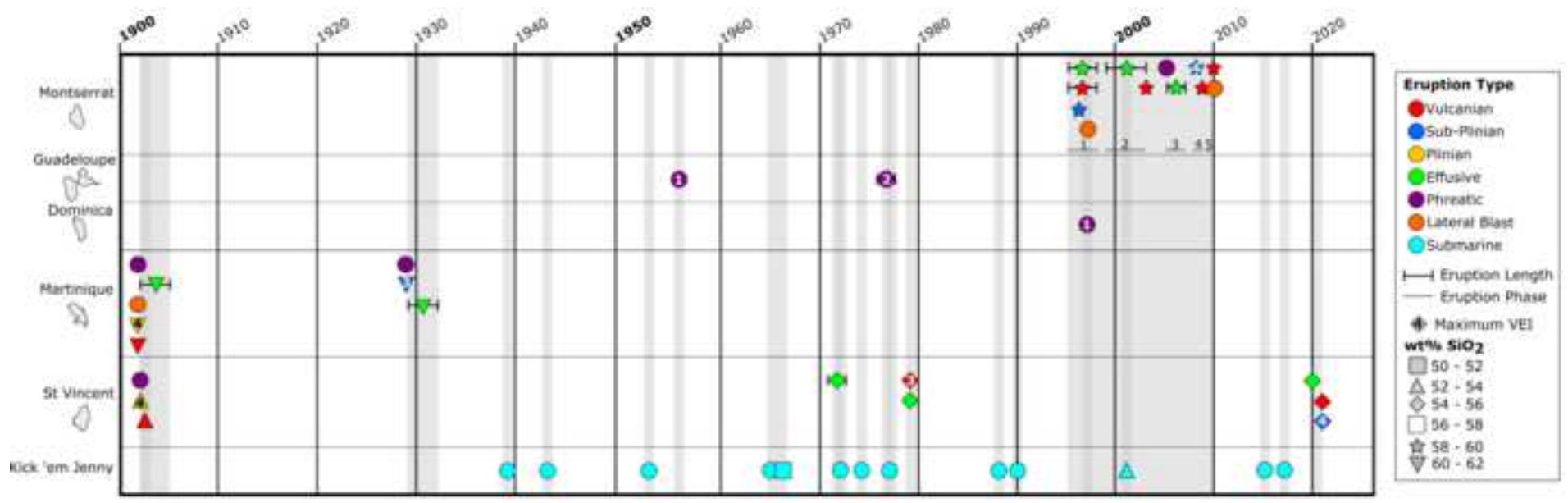
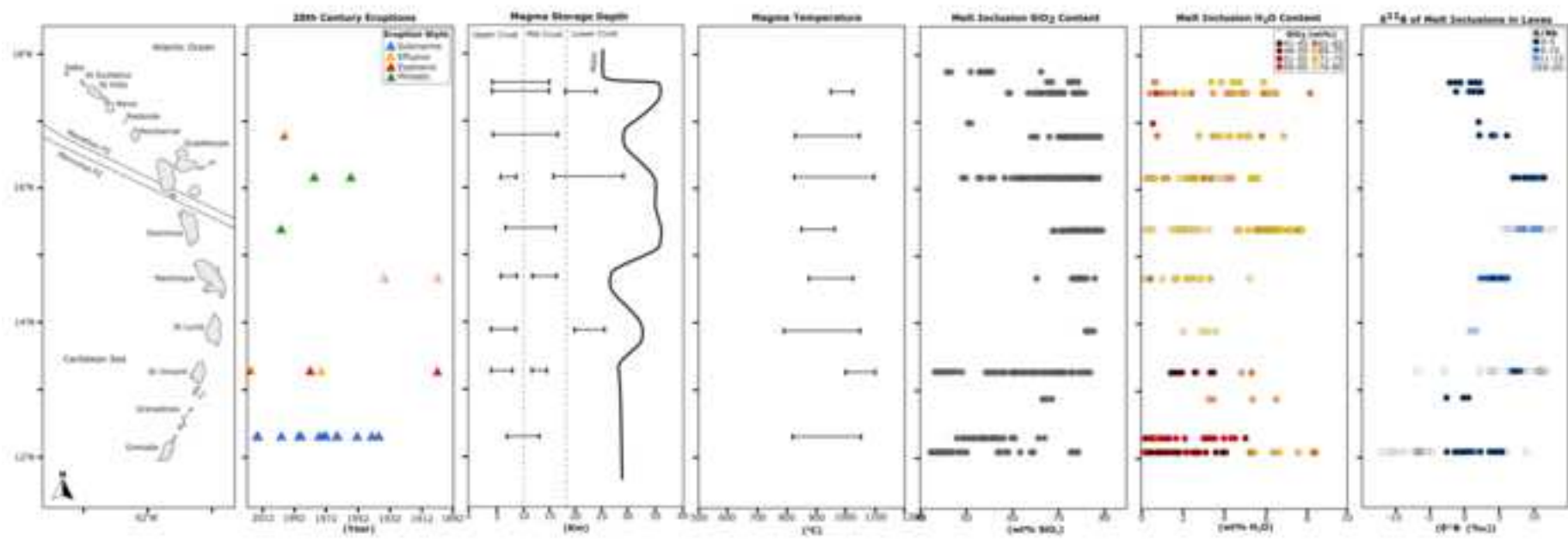
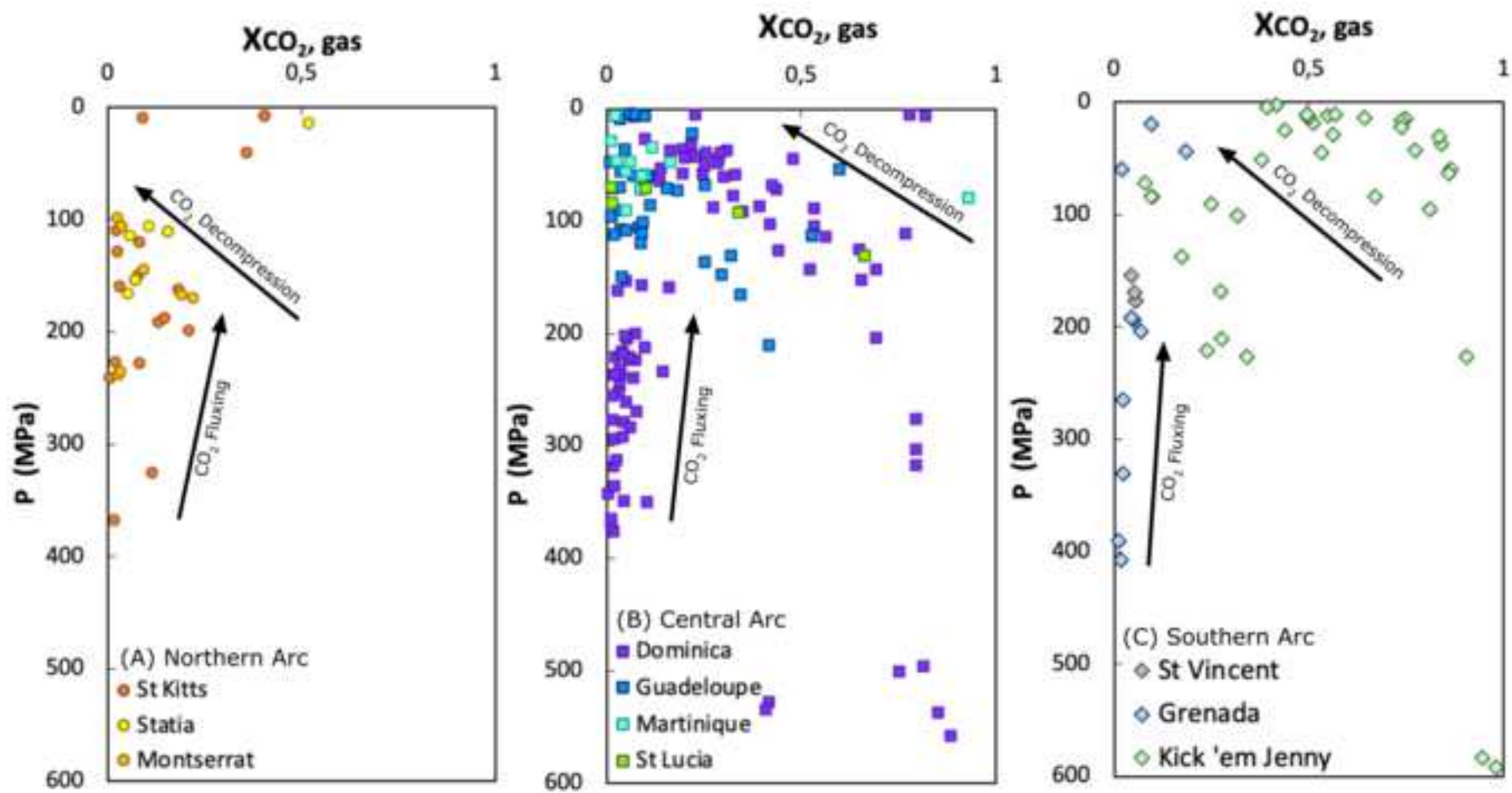


Figure 5







Declaration of interests

The authors declare that they have no known competing financial interests or personal relationships that could have appeared to influence the work reported in this paper.

The authors declare the following financial interests/personal relationships which may be considered as potential competing interests:

1 **Diverse Magma Storage and Major and Volatile Magma Composition: What are the implications**
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16 **Key Words:** Lesser Antilles; Geochemistry; Magma Storage; Eruption Style; Arc Volcanism; Melt
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18 **Abstract**

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47 term displacement and results in major socio-economic impacts. This highlights the need for future
48 crises to be effectively managed so as to limit the socio-economic impacts and maintain the
49 relationship between scientists, authorities and the population.

50 **1 Introduction**

51 Arc volcanism is often considered as one of the most hazardous types of volcanism due to the range
52 of eruption styles observed (e.g., Roobol and Smith, 1989; Brown et al., 2017). This variability makes
53 it difficult to define the processes, timescales and controls on eruptions and resulting hazards.
54 Understanding the drivers of changes in eruptive style is particularly important, as eruption style
55 dictates many of the hazards presented by a volcano (e.g., Komoroswki et al., 2005; Lindsay et al.,
56 2005; Cassidy et al., 2018). Eruptions at arc volcanoes range from phreatic behaviour (generated in
57 the hydrothermal system and not resulting in magma extrusion, e.g., Kilgour et al., 2019) to effusive
58 eruptions in the form of lava domes and spines (e.g., Sheldrake et al., 2016), to explosive behaviour
59 ranging from small vulcanian eruptions (e.g., Druitt et al., 2002) to large Plinian eruptions (e.g., Vidal
60 et al., 2015). There are multiple ways to generate an eruption and processes are unlikely to occur in
61 isolation (Villemant and Boudon, 1998; Di Genova et al., 2017; Cassidy et al., 2018; Popa et al., 2021).
62 For magmatic eruptions, many mechanisms involve recharge of the system by magma from depth and
63 for explosive eruptions the supersaturation of volatiles in the magma and an increase in overpressure,
64 which exert a control on the flow rate into the conduit (e.g., Burnham, 1975).

65 Volcanoes are not limited to one eruption style and there are transitions in eruption style that are
66 observed at arc volcanoes within the same eruption (e.g., effusive to explosive and vice-versa; e.g.,
67 Komorowski et al., 2008; Legendre, 2012; Boudon and Balcone-Boissard, 2021; Joseph et al., 2022)
68 and across several eruptions with an observed repose period (Cassidy et al., 2018). Magma
69 composition is often considered a first order control on eruption style (e.g., Sides et al., 2014; Firth et
70 al., 2016). An explosive eruption is the result of brittle failure of a melt phase when the volume fraction
71 of exsolved volatiles is sufficiently large (Sparks, 1978; Woods and Koyaguchi, 1994; Carr et al., 2020),
72 when the strain rate is sufficiently high (Papale, 1999; Gonnermann and Manga, 2003), or when
73 bubble overpressure exceeds a critical value (Zhang, 1999). Variations in eruption style therefore
74 relate to the ability of that system to dissipate increasing pressure through degassing (Woods and
75 Koyaguchi, 1994; Melnik and Sparks, 2002; Ruprecht and Bachmann, 2010; Kozono and Koyaguchi,
76 2012). Such variations are often linked to processes occurring in the magma storage system including
77 magma mixing and magma recharge (Sparks et al., 1977; Murphy et al., 2000; Plail et al., 2018).
78 However, recent studies also stress the importance of processes occurring external to the magmatic
79 system (e.g., Hlinka et al., 2021; Metcalfe et al., 2022).

80 In recent years the effects of bottom-up vs top-down vs external forces has been shown to play an
81 important role in transitions in the eruptive style, both during a single eruption and across multiple
82 eruptions (e.g., Jaupart and Allègre, 1991; Sparks, 1997; Melnik and Sparks, 1999; Cassidy et al., 2018;
83 Bernard and Bouvet de Maisonneuve, 2020; Hlinka et al., 2021; Metcalfe et al., 2022). Bottom-up
84 controls drive an eruption through changes in the magma storage system. In comparison, top-down
85 controls result from changes in the uppermost plumbing system affecting the late-stage
86 decompression path of the magma that proceed downward. Another important control on eruption
87 style and eruption style transition is caused by external effects such as: flank or dome collapse
88 (Alidibirov and Dingwell, 1996); assimilation of external materials (Deegan et al., 2010; Troll et al.,
89 2012); and external water interacting with magma (Shepherd & Sigurdsson, 1982; Zimanowski et al.,
90 2015). A change in morphology of the volcanic edifice can also impact crater morphology and result
91 in changes to the volcanic hazards produced (e.g., the generation of pyroclastic density currents at
92 Soufriere St Vincent; e.g., Cole et al., 2019).

93 Another important consideration is how the magma is stored at depth, which can vary depending on,
94 for example, the melt proportion, geometry, temperature and compositions (e.g., Cashman et al.,
95 2017). A now widely accepted interpretation of the magma storage zones in arc volcanoes is that
96 magma is stored in a mush system. Mush systems are formed from repeated intrusions of melt from
97 depth and are composed of an interlocking framework of crystals with the presence of melt,
98 containing dissolved volatiles, distributed in varying volumetric proportions within the network of

99 crystals (e.g., Humphreys et al., 2006; Cashman and Blundy, 2013; Cashman and Sparks, 2013; Cooper
100 and Kent, 2014; Christopher et al., 2015; Edmonds et al., 2016). The crystalline parts of the mush
101 system are deformable, meaning melt and fluid can be segregated within the mush system (Mckenzie
102 1984; Solano et al 2012; Connolly et al., 2015; Spiegleman et al., 2001), with the strength of the
103 framework increasing as melt fraction decreases (Marsh, 1981; Rosenberg and Handy, 2005).

104 Geophysical methods identify laterally and vertically elongated low-velocity zones with <10% melt
105 more consistent with a crystal mush system (Koulakov et al., 2013; Ward et al., 2013; Sinton, 1992;
106 Lees, 2007; Farrell et al., 2014; Huang et al., 2015; Kiser et al., 2016; Pritchard and Gregg, 2016;
107 Pritchard et al., 2018; Cordell et al., 2018). A magma mush can exist at a lower temperature due to
108 the close crystal contacts and the high viscosity which prevents convection and thereby limits most
109 heat lost through conduction (Balmforth and Rust, 2009; Dufek et al 2010; Blundy and Annen, 2016).
110 The heat loss prevents large volumes of eruptible magma from forming but allows the mush system
111 to be maintained over long periods of time (Blundy and Annen, 2016; Cashman et al., 2017; Karakas
112 et al., 2017; Glazner, 2021). Periodic injections of recharge magma have also been shown by numerical
113 models to be sufficiently high or long enough to maintain some eruptible melt within a crystal mush
114 (Huber et al., 2012; Annen et al., 2015; Cooper, 2017). Long-lived, trans-crustal mush systems are
115 consistent with many of the features of arc volcanoes and the long-lived volcanic activity observed in
116 the arc (e.g., Balcone-Boissard et al., 2018; Metcalfe et al., 2022; Higgins et al., 2022).

117 The transition from crystal mush to an eruptible magma occurs over a narrow crystal content (Shaw,
118 1969; Caricchi et al., 2007; Costa et al 2009; Burgisser and Bergantz, 2011; Sparks et al., 2019). The
119 energy required to remobilize a crystal mush depends on the initial melt content; mushes with a higher
120 initial melt content require less energy to remobilise (Bachmann and Bergantz, 2006; Bergantz et al.,
121 2017; Sparks et al., 2019). Magma recharge and addition of volatiles are considered the main drivers
122 of remobilization (Scandone et al., 2007; Costa et al., 2009; Mader et al., 2013; Rubin et al., 2017;
123 Cooper and Kent, 2014; Sparks and Cashman, 2017; Jackson et al 2018). There are several hypotheses
124 on how a mush is remobilised. Many models involve a reduction in crystallinity through, for example,
125 reactive melt flow (Jackson et al., 2018) and the growth of a buoyant mobile layer (Burgisser and
126 Bergantz, 2011). This increases the melt fraction and leads to remobilization of the cold mush system
127 through overturn or unzipping of the mush system (Burgisser and Bergantz, 2011; Bergantz et al.,
128 2015, 2017; Schleicher et al., 2016; Schleicher and Bergantz 2017; Carrara et al., 2019) The different
129 proposed mechanisms of remobilisation indicate that this is a complex process and may involve a
130 combination of mechanisms.

131 The Lesser Antilles Arc (LAA) is a good example of an active island arc, with a range of eruption styles
132 recorded and a wealth of geochemical data available, making it an excellent arc to investigate links

133 between magma storage, magma composition and the eruptive style. Few petrologic studies
134 investigate the arc as a whole (e.g., Cooper et al, 2020; Melekoheva et al., 2019), instead choosing to
135 focus on individual islands and magmatic systems. This has resulted in a significant amount of data
136 available for the LAA, which has expanded greatly in the last few years (e.g., VOILA project; Goes et
137 al., 2019). Here, we provide a review of the available LAA data, focusing on the geochemistry, magma
138 storage and eruptive style, adding to the reviews already available from Macdonald et al. (2000)
139 (geochemistry), Lindsay et al. (2005) (volcanic hazard atlas) and Smith et al. (2013) (geological and
140 tectonic setting).

141 **2 The Lesser Antilles Arc**

142 The Lesser Antilles Arc is an active volcanic arc in the Caribbean, which spans 850 km from Saba in the
143 north to Grenada in the south (Fig 1). The active arc is made up of eleven main volcanic islands, an
144 archipelago of nineteen smaller islands (the Grenadines) with a total of twenty-one volcanoes active
145 in the Holocene, including a submarine volcano. The LAA is formed due to subduction of the American
146 plate beneath the Caribbean plate at a rate of approximately 1.0 – 2.4 cm/yr (Macdonald and
147 Holcombe, 1978; McCann and Sykes, 1984; Rosencrantz and Sclater, 1986; Jarrard, 1986; Deng and
148 Sykes, 1995; Dixon et al., 1998; DeMets et al., 2000; Jarrad, 2003; Feuillet et al., 2011; Symithe et al.,
149 2015). This slow subduction rate results in low magma production rates in comparison to other
150 subduction zones. Wadge (1984) reports a magma production rate of $5 \text{ km}^3 \text{ Ma}^{-1} \text{ km}^{-1}$ for the past 300
151 years in the LAA. In comparison, the Central American arc has a magma production rate of $62 \text{ km}^3 \text{ Ma}^{-1}$
152 km^{-1} for the past 300 years (Wadge, 1984).

153 To the north of Martinique, the arc bifurcates into an inactive eastern limb and an active western limb
154 (Schlaphorst et al., 2018). The older eastern arc is composed of several islands which are now covered
155 in a carbonate platform – the Limestone Caribbees, which experienced active volcanism from the
156 Eocene to the mid-Oligocene. Activity on the arc has migrated westward forming the western arc,
157 which has been active since 20 Ma (Early Miocene to present (Fig 1); Westercamp and Tazieff, 1980).
158 Pleistocene to recent volcanism has occurred in narrow 10 km wide zones which are split into three
159 segments: the Northern Arc (Saba to Montserrat), the Central Arc (Guadeloupe to Martinique) and
160 the Southern Arc (St Lucia to Grenada (Sigurdsson and Carey, 1981; Wadge and Shepherd, 1984; Fig
161 1).

162 Each island has at least one active volcanic complex except for Dominica, which has nine potentially
163 active volcanoes (e.g., Brown et al., 2015; Leone et al., 2019; Fig 1) and Guadeloupe which has two
164 active volcanoes (Komorowski et al., 2005; Komorowski et al., 2022). Eruptions range from small
165 phreatic to effusive dome forming (ca. $9 - 47 \times 10^6 \text{ m}^3$; e.g., Shepard and Sigurdsson, 1982; Loughlin

166 et al., 2010; Martel, 2012) to large explosive eruptions (VEI 2 – 6; Legendre, 2012; Howe et al., 2014;
167 Palmer et al., 2016; Carazzo et al., 2020; Joseph et al., 2022). The twenty-three eruptions that have
168 occurred in the LAA since 1900 provide examples of how devastating an eruption in the LAA can be
169 (Table 1). Recent eruptions have resulted in thousands of deaths (e.g., 29,800 at Montagne Pelée in
170 1902 (Lacroix, 1904); 1350 at La Soufrière St Vincent in 1902 (Brown et al., 2017; Pyle et al., 2018;
171 Table 1), the short and long-term displacement of thousands more (e.g., La Soufrière de Guadeloupe,
172 1976; Soufrière Hills Volcano, Montserrat, 1995; Soufrière St. Vincent, 2021 (Lepointe, 1999;
173 Komorowski et al., 2005; Feuillard, 2011; Joseph et al., 2022); Table 1) and long-lasting economic
174 impacts (e.g., La Soufrière de Guadeloupe, 1976; Soufrière Hills, Montserrat, 1995-present; Lepointe,
175 1999; Blérald, 1986; Kokelaar, 2002), all of which have profound impacts on the lives and livelihoods
176 of people in the LAA. Approximately one million people are threatened by the direct effect of volcanic
177 activity in the Lesser Antilles, and related phenomena threaten approximately two and a half million
178 people (Lindsay et al., 2005a; Leone et al., 2019).

179 There are notable variations in geochemistry, volcanic activity, crustal structure and seismicity
180 observed along the LAA, including variations (Boynton et al., 1979; Macdonald et al., 2000, Melekhova
181 et al., 2019; Hayes et al., 2014; Schlaphorst et al., 2018; 2021; Cooper et al., 2020). An example of this
182 is the variation in Wadati-Benioff zone dip, which diverges from 50° – 60° in the Northern Arc to sub-
183 vertical in the Southern Arc (e.g., Wadge and Shepard, 1984; Schlaphorst et al., 2018). In the Central
184 Arc the slab dips at ca. 20° until 50 km depth where it increases to 45° – 55° (Paulatto et al., 2017).
185 Systematic variations in crustal thickness are observed along the Lesser Antilles Arc, with a greater
186 crustal thickness observed in the Northern and Central arcs. North of Martinique crustal thickness is
187 estimated between 30 – 37 km (Arnaiz-Rodríguez et al., 2016; Melekhova et al., 2019; Schlaphorst et
188 al., 2021). South of Martinique the crust thins with crustal thickness estimated between 20 – 30 km
189 (Arnaiz-Rodríguez et al., 2016; Melekhova et al., 2019; Schlaphorst et al., 2021). The range of crustal
190 thickness (20 – 37 km) is considered to relate to long-term variations in magma productivity (Melekhova
191 et al., 2019; Cooper et al., 2020). Another contributing factor to the variation in magma production is
192 the shift of the Marathon–Mercurius fracture zone from St Kitts (Northern Arc) to Dominica (Central
193 Arc), which is reflected in the greater crustal thickness in the Northern Arc (Paulatto et al., 2017; Bie
194 et al 2019; Cooper et al., 2020; Bie et al 2022).

195 Two crustal layers are observed using seismic discontinuities (Boynton et al., 1979). The first
196 discontinuity is the upper crust, which extends to a variable depth of 2 -30 km along the arc and is a
197 plutonic igneous layer (Wadge, 1986; Schlaphorst et al., 2018). Below this is a layer of dense mafic
198 igneous rocks including cumulates directly overlying the mantle (the Moho), estimated between 30 -
199 60 km depth (e.g., Boynton et al., 1979; Christeson et al., 2008; Kopp et al., 2011; Laigle et al., 2013;

200 Arnaiz-Rodríguez et al., 2016; Schlaphorst et al., 2021). Both discontinuities are found at greater
201 depths in the northern part of the arc, with the along arc variation in the discontinuities being
202 unusually strong (Schlaphorst et al., 2018). The strong variation in the nature of the discontinuities
203 results from variation in melt generation and differentiation (Schlaphorst et al., 2018).

204 Variations in upper plate mantle seismic velocities are also observed, seismicity is shown to be closely
205 related to hydration and dehydration processes and are controlled by thermal structure, composition
206 and state of stress (Paulatto et al., 2017; Ezenwaka et al., 2022). Three main phases of dehydration
207 occur in the Atlantic slab: (1) 5 – 20 km - compaction and draining of pore fluids from the sediments;
208 (2) 40 – 100 km - chemical dehydration of the crust; (3) 120 – 160 km lithospheric mantle
209 deserpentinization (van Keken et al., 2011; Paulatto et al., 2017). Notably slower seismic velocities are
210 observed below Guadeloupe, Dominica and Martinique where thicker crust occurs and is bound by
211 two major fracture zones to the north (Paulatto et al., 2017; Cooper et al., 2020; Harmon et al., 2019;
212 Schlaphorst et al., 2021). The anomalies are thought to be related to either the upwelling of hot mantle
213 material or partial melt and fluids. A high $\delta^{11}\text{B}$ anomaly is also observed in the Guadeloupe and
214 Dominica melt inclusions thought to be related to dewatering of serpentinised oceanic crust, occurring
215 close to the fracture zone to the north of the Central Arc (Cooper et al., 2020). Alternatively, the
216 curvature of the arc (e.g., Harris et al., 2018; Schlaphorst et al., 2017), ponding of melt and resulting
217 redistribution of fluids (Ha et al., 2020) or along-strike variability in the subducted plate morphology
218 could also account for the velocity anomalies observed.

219 **3 Magma Composition and Magma Storage Across the Lesser Antilles Arc**

220 A crystal mush is often described to be vertically extensive and consist of melt-rich lenses and solid
221 crystals (Cashman and Blundy, 2013; Cashman and Sparks, 2013; Cashman et al., 2017; Blundy et al.,
222 2019). The mush exists under sub-solidus conditions for long timescales before remobilisation (e.g.,
223 Bachmann and Bergantz, 2006; Bergantz et al., 2015; Burgisser and Bergantz, 2011; Parmigiani et al.,
224 2014). The chemical and dynamic features of magmatic systems of the LAA volcanoes can be explained
225 by the model of a transcrustal, vertically (and often laterally) extensive mush system as described by
226 Cashman et al. (2017).

227 Conditions of the mush system are calculated through: [1] Melt inclusion volatile contents which are
228 converted into minimum entrapment pressures (e.g., Balcone-Boissard et al., 2018; Pichavant et al.,
229 2018; Camejo-Harry et al., 2019; Metcalfe et al., 2022) using $\text{H}_2\text{O}-\text{CO}_2$ solubility models (e.g., Papale
230 et al., 2006); [2] Thermobarometry (e.g., Bezard et al., 2017; Higgins et al., 2022) which relates mineral
231 phase chemistry to a known pressure, temperature and melt composition (e.g., Nimis & Taylor, 2000;
232 Blundy and Cashman, 2008; Putirka, 2008; Ridolfi et al., 2010); [3] Equilibrium experiments with the

233 pre-eruptive physical parameters defined by the experimental phase equilibria and chemistry (e.g.,
234 Pichavant et al., 2002, 2018; Solaro et al., 2019, 2020).

235 Pressures determined by melt inclusions are considered minimum estimates of storage pressures due
236 to degassing processes that can occur prior to entrapment or if the melt was not volatile saturated at
237 the point of entrapment (e.g., Steele-Macinnis et al., 2011; Cannatelli et al., 2016). Additionally, post-
238 entrapment processes can result in volatiles leaking (e.g., Steele-Macinnis et al., 2011; Wallace et al.,
239 2015; Cannatelli et al., 2016). In particular, H₂O loss can occur through H⁺ diffusion through the host
240 mineral and CO₂ is lost to vapour bubbles forming during post-entrapment cooling (e.g., Wallace et
241 al., 2015). However, when melt inclusions are used to determine pressure careful checks are made to
242 ensure the melt inclusion has not experienced post entrapment processes (e.g., using deuterium
243 values for instance; e.g., Metcalfe et al., 2022) and to ensure the melts were volatile-saturated at the
244 time of entrapment (e.g., using CO₂ vs Th; Balcone-Boissard et al., 2018).

245 Additionally, crustal density affects depth-pressure conversions, depth-pressure conversions range
246 from 2450 – 2800 kg/m³ across the arc (e.g., Kopp et al., 2011; Christeson et al., 2008; Barnoud et al.,
247 2016, Gómez-García et al., 2019). The mean crustal density of 2660 kg m⁻³ is also widely used for
248 conversions (Christeson et al., 2008).

249 Studies of the LAA magmatic systems indicate the deepest storage areas occur at >25 km, but will
250 likely depend on the slab dip and crustal thickness which varies across the arc (Schlaphorst et al., 2018;
251 Melekhova et al., 2019; Schlaphorst et al., 2021). A product of the differentiation and fractionation
252 processes are cumulates with the mineral proportions dependent on the magmatic conditions and
253 depths. Cumulates, along with xenoliths, are brought to the surface and can provide further insights
254 into the plumbing system (e.g., Baker 1968; Parkinson et al., 2003; Stamper et al., 2014; Cooper et al.,
255 2016; Melekhova et al., 2019).

256 Many studies indicate that differentiation to produce evolved melts is occurring deeper in the system,
257 generally in the range 13 to 16 km (ca. 300 - 400 MPa; Edmonds et al., 2016; Pichavant et al., 2002;
258 Boudon and Balcone, 2021). Though this is often less well resolved by geophysical and geochemical
259 data in comparison to the shallow regions that are better sampled by eruptions and imaged through
260 geophysics. This is consistent with the transc crustal magma mush model (e.g., Cashman et al., 2017).

261 The existence of this deeper region has various implications for the system, with recharge of the
262 shallow storage system with less evolved magma often cited as the principal cause of unrest and
263 eruption (e.g., Humphreys et al., 2010; Howe et al., 2015; Pichavant et al., 2018). From the deeper
264 region fluids, heat and magma inputs into shallow depths may promote the formation of eruptible
265 magma. However, recent studies also suggest melts migrating within the shallow zone and volatile

266 oversaturation can result in an eruption without the requirement for a deep mafic recharge event
267 (e.g., Moretti et al., 2019; Boudon and Balcone-Boissard, 2021). Despite this, the deeper system is
268 important for the production of more-evolved magmas through fractionation processes, the triggering
269 of eruption through recharge, and may also be evacuated in the more voluminous eruptions.

270 The active part of the systems supplying eruptions are located in the range of 5 - 9 km (ca. 100 - 200
271 ± 50 MPa), with some evidence of melt at < 5 km (e.g., Martel et al., 1998; Pichavant et al., 2002;
272 Christopher et al. 2015; Edmonds et al., 2016; Pichavant et al., 2018; Balcone-Boissard et al., 2018 and
273 Solaro et al., 2019; d'Augustin, et al., 2020; Metcalfe et al., 2021; Fig 2). These shallow storage regions
274 are typically formed of evolved magmas and are often the main zone from which magma is erupted.

275 Magma production rate is relatively low compared to the global range of subduction zones, with the
276 rate varying along the arc, the Central Arc produces the largest volumes of volcanic deposits,
277 particularly in the past 0.1 Ma (40 km^3 for Dominica and 8 km^3 for Guadeloupe and Martinique; Wadge
278 1984; 1986; Macdonald et al., 2000). In comparison, the Northern and Southern parts of the arc have
279 produced smaller volumes (ca. $0 - 5 \text{ km}^3$; Macdonald et al., 2000).

280 A wide range of magma types are observed across the arc, with variation in sediment input into the
281 arc playing a role in the changing chemistry of primitive parental magmas (Carpentier et al., 2008). In
282 the Southern Arc, the incoming plate is rich in clastic detritus, in comparison to the Northern Arc,
283 which has limited sediment supply and is richer in pelagic marine sediments (Faugères et al., 1993;
284 Carpentier et al., 2008). The LAA magmas have fractionated from a parental magma hypothesised by
285 MacDonald et al. (2000) as a picritic, silica-undersaturated magma, though due to the variation in
286 sediment input the trace element compositions is likely to vary across the arc.

287 The LAA magmas vary from tholeiitic suites in the Northern Arc, to calc-alkaline suites in the Central
288 Arc and silica-undersaturated, highly magnesian suites in the Southern Arc (Macdonald et al., 2000).
289 The trends observed in the geochemical data show the various magmas are the result of fractional
290 crystallisation (Fig 3; MacDonald et al., 2000). However, crustal contamination, and minor magma
291 mixing are also occurring, particularly in the Central Arc (MacDonald et al., 2000), minor assimilation
292 of crust or sediments is also observed in St Lucia and Martinique magmas (Bezard et al., 2015).

293 Here, we provide a summary of the magma storage systems and a summary of the wealth of
294 geochemistry data available for the LAA.. This is important to understand any common depths of
295 magma storage, along arc variations and to eventually provide a coherent model of magma storage
296 along the arc. We note that, the geochemical dataset is biased towards both what is erupted and what
297 is analysed out of large amounts of samples, providing an incomplete view of the system. Our dataset
298 is available in Supplementary dataset 1 (built on compilation by DIGIS Team, 2022).

299 **3.1 The Northern Arc**

1
2 300 In the Northern Arc (Saba to Montserrat) the dominant rock type ranges from basalt to andesite (e.g.,
3
4 301 Rea and Baker, 1980; Fig 3). However, on several islands the compositional range can vary greatly,
5
6 302 from basalt to rhyolite.

8 303 **3.1.1 Saba**

9
10 304 Eruption deposit compositions from Saba range from basalts to andesites (49 – 66 wt% SiO₂), with the
11
12 305 basalts representing some of the least evolved compositions observed in the Northern Arc (Fig 3;
13
14 306 Defant et al., 2001; Roobol and Smith, 2004). Evidence of magma mixing in banded pumices and
15
16 307 rounded enclaves is observed (Defant et al., 2001). However, no compositional differences between
17
18 308 the enclaves and other magmas are discussed (Defant et al., 2001).

19
20 309 Melt inclusions (MIs) are also unevolved, particularly in comparison to other MIs in the Northern Arc,
21
22 310 and generally fall in the range 46 - 56 wt% SiO₂ (Fig 4A; Roobol and Smith, 2004). However, Na₂O +
23
24 311 K₂O contents (1.7 – 4.6 wt% Na₂O, 0.36 – 0.45 wt% K₂O; Fig 4B) fall into the range observed in the.
25
26 312 Northern Arc (Roobol and Smith, 2004). No volatile data or pressure data is reported for Saba.

27 313 **3.1.2 St Eustatius (Statia)**

28
29 314 Magmas erupted from The Quill on St Eustatius (Statia) are dominantly andesitic in composition,
30
31 315 however, a broader range of compositions are observed ranging from basalt to rhyolite (50 - 72 wt%
32
33 316 SiO₂; Roobol and Smith 2004). The compositions are generally relatively evolved, particularly in
34
35 317 comparison to magmas erupted from Saba and St Kitts (Fig 3; Roobol and Smith 2004). MIs hosted in
36
37 318 clinopyroxene from Statia shows a similar range of SiO₂ contents (68 – 74 wt% SiO₂; Fig 4A) to those
38
39 319 from other systems in the Northern Arc. The Na₂O + K₂O contents also fall into a range typical of the
40
41 320 Northern arc (0.9 – 1.2 wt% K₂O, 4.4 – 5.3 wt% Na₂O; Fig 4B). Volatiles recorded in MIs from Statia
42
43 321 show Cl content is relatively low in comparison to other systems in the Northern Arc (0.18 – 0.23 wt%,
44
45 322 Fig 4C). A wide range of H₂O is recorded (0.61 - 5.93 wt%; Figure 4D), and CO₂ is comparable to other
46
47 323 systems in the Northern arc (26 – 216 ppm) (Cooper et al., 2019). Bubbles are reported by Cooper et
48
49 324 al. (2019), and the pressure estimated calculated are considered minima.

50 325 Cumulates from Statia have mineral assemblages and textures that are similar to those observed in
51
52 326 xenoliths from Martinique (Cooper et al., 2019) with plagioclase, olivine, clinopyroxene,
53
54 327 orthopyroxene, amphibole and spinel interstitial melt also present (Cooper et al., 2016, 2019). Cooper
55
56 328 et al. (2019) calculate minimum volatile saturation pressures ranging from 0 – 426 MPa, which
57
58 329 indicates storage at <15 km.

59 330 **3.1.3 St Kitts**

331 St Kitts eruption products show a wide range of compositions ranging from basalt to rhyolite (e.g.,
332 Toothill et al., 2007; Melekhova et al., 2017; Higgins et al., 2021). The most recent active volcano on
333 St Kitts is Mt Liamuiga where basaltic-andesites and andesites are most common (42.1 – 65.4 wt%
334 SiO₂; Baker, 1984; Toothill et al., 2007; Melekhova et al., 2017; Higgins et al., 2021). Many Mt Liamuiga
335 eruption products are considered to be hybrid magmas from mixing of basalt, basaltic andesite and
336 andesite magmas, and are the product of open-system processes (Toothill et al., 2007). These magmas
337 cover nearly the whole range of compositions observed in the LAA (Fig 3).

338 MIs hosted in plagioclase, clinopyroxene, orthopyroxene, ilmenite and amphibole from St Kitts range
339 from 59 – 76 wt% SiO₂, 1.3 – 4.8 wt% Na₂O and 0.6 – 2.5 wt% K₂O, which is within the range observed
340 in the Northern Arc (Fig 4A, B; Toothill et al., 2007; Melekhova et al., 2017; Cooper et al., 2020).
341 Volatiles in MIs from St Kitts record Cl contents in the range 0.15 – 0.43 wt% (Fig 4C; Toothill et al.,
342 2007; Melekhova et al., 2017; Cooper et al., 2020) with one MI recording 0.66 wt% Cl. H₂O recorded
343 in the MI is variable ranging from 0.4 - 8.2 wt%; Fig 4D; Toothill et al., 2007; Melekhova et al., 2017;
344 Cooper et al., 2020), this is the highest H₂O content recorded in the Northern Arc. CO₂ ranges from 15
345 – 721 ppm (Fig 4E; Toothill et al., 2007; Melekhova et al., 2017; Cooper et al., 2020), though this is the
346 highest CO₂ recorded in the northern Arc, this is still relatively low in comparison to the rest of the arc.
347 Bubbles are reported in the St Kitts MIs (Toothill et al., 2007; Melekhova et al., 2017), however, MIs
348 with lower H₂O or CO₂ are not associated with larger bubbles (Melekhova et al., 2017). Xenoliths from
349 St Kitts are reported to be rich in plagioclase and amphibole.

350 An extensive, transcrustal magmatic mush system has been proposed by Melekhova et al. (2017) and
351 Higgins et al. (2022) for the production of magmas at St Kitts. Depths are estimated from
352 thermobarometry of xenoliths and seismic modelling and range from 3 – >25 km. Though the deepest
353 part of the system is more difficult to distinguish as the cumulates are rarely observed, the deep
354 system is hypothesised to generate basalts -basaltic andesites through fractionation which ascend to
355 18 – 25 km where they experience differentiation (Melekhova et al., 2017) (Fig 2). These basaltic-
356 andesites are considered the parental magmas for more evolved compositions which migrate upwards
357 and equilibrate with existing crystal mush (Melekhova et al., 2017; Fig 2). This means many eruption
358 products will have an equilibration pressure corresponding to shallow depths despite differentiation
359 beginning much deeper in the system. The majority of lavas erupted are supplied from the upper
360 portion of the transcrustal system, this is consistent with experimental data which indicates many
361 eruption products from St Kitts can be generated at 200 – 550 MPa (ca. 3 km – 15 km; Melekhova et
362 al., 2017; Higgins et al., 2022). Though many magmas correspond to shallow depths, some melts are
363 shown to bypass equilibration with the shallower mush system and so preserve evidence of their
364 origin in the deeper system (Melekhova et al., 2017).

365 3.1.4 Montserrat

1
2 366 The ongoing, multiphase eruption of Soufrière Hills Volcano on Montserrat has allowed many in depth
3
4 367 studies to be conducted on the eruption products. Soufrière Hills Volcano on Montserrat, has
5
6 368 produced a restricted bulk composition of andesite (56 - 62 wt% SiO₂; e.g., Murphy et al., 2000; Wadge
7
8 369 et al., 2014). No systematic changes in bulk composition have been observed since the beginning of
9
10 370 the most recent eruption in 1995. Mixing and mingling textures are reported on Montserrat with rare,
11
12 371 banded lava and pumice observed, however no significant difference in composition is noted between
13
14 372 layers (Murphy et al., 2000).

15 373 Mafic enclaves are hosted throughout the andesite, with the proportion of mafic enclaves increasing
16
17 374 from the Phase 1 (1995) deposits to the Phase 5 (2010) deposits (Barclay et al. 2010). The mafic
18
19 375 enclaves are generally less evolved than the andesite they are hosted in (49 – 56 wt% SiO₂ (Murphy et
20
21 376 al., 2000; Christopher et al., 2014; Plail et al., 2018)), and some overlap in composition is observed
22
23 377 with the South Soufrière Hills magmas found in the south of the island (e.g., Baker, 1984; Murphy et
24
25 378 al., 2000; Wadge et al., 2014). The eruption products are some of the most evolved in the Northern
26
27 379 Arc, and do not extend to the same unevolved compositions observed on Saba and St Kitts (Fig 3).

28 380 Montserrat has a large amount of MI data available due to the large amount of work done on the 1995
29
30 381 – present eruption deposits. MI hosted in plagioclase, quartz and clinopyroxene from Soufrière Hills
31
32 382 volcano are the most evolved in the Northern Arc (64 – 79 wt% SiO₂; Fig 4A; Devine et al., 1998;
33
34 383 Humphreys et al., 2020; Mann et al., 2013; Cooper et al., 2020). The highest Na₂O + K₂O values
35
36 384 observed across the whole LAA arc are also from Soufrière Hills volcano MIs (2.4 - 6.3 wt% Na₂O, 0.9
37
38 385 – 6 wt% K₂O; Fig 4B; Devine et al., 1998; Humphreys et al., 2020; Mann et al., 2013; Cooper et al.,
39
40 386 2020). Xenoliths from Montserrat are reported to be comparable to those from Guadeloupe and St
41
42 387 Lucia and are felsic and olivine-free (Kiddle et al., 2010; Melekhova et al., 2019).

43 388 MIs from Montserrat record the highest Cl contents in the Northern Arc, generally ranging from 0.22
44
45 389 - 0.59 wt% (Fig 4C; Devine et al., 1998; Humphreys et al., 2010; Mann et al., 2013; Cooper et al., 2020).
46
47 390 H₂O contents range from 0.78 – 6.9 wt%, which falls in the range recorded across the Northern arc
48
49 391 (Fig 4D; Mann et al., 2013; Cooper et al., 2020). CO₂ contents are low, as observed in the other systems
50
51 392 in the Northern Arc, ranging from 23 – 304 ppm (Fig 4E; Cooper et al., 2020). Vapour bubbles are not
52
53 393 reported by Cooper et al. (2020).

54 394 This provides a snapshot of the magma system as it was directly prior to the eruption. Our
55
56 395 understanding of this system is based on the key hypothesis that an andesitic magma body is
57
58 396 recharged by an intruding basaltic magma, which releases heat and volatiles and drives eruption (e.g.,
59
60 397 Murphy et al., 2000; Wadge et al., 2014). Studies of South Soufrière Hills Volcano reveal a distinct

398 geochemical signatures in comparison to the active Soufrière Hills Volcano. This indicates that isolated
399 melt bodies within the crystal mush result in a heterogenous system, which is vertically and laterally
400 extensive (Devine et al., 2003; Edmonds et al., 2016).

401 As well as being chemically heterogenous, the Soufrière Hills system is thought to be composed of
402 layers of mush (solid and crystalline), liquid (sub-solidus, eruptible magma) and exsolved fluid layers
403 (volatiles exsolved from the magma on saturation; Christopher et al., 2015; Edmonds et al., 2016). The
404 presence of fluid layers with exsolved fluids, resulting from magma recharge, may lead to instabilities
405 in the mush system and may result in overpressures that could fuel explosive eruptions (Christopher
406 et al., 2015; Edmonds et al., 2016). The cyclicity observed in the vulcanian eruptions was thought to
407 be related to the ascent of different magma batches. The ascent of magma to the surface was
408 calculated to take 5 – 13 hours, consistent with the 12-hour cyclicity of the Vulcanian eruptions.
409 Geochemistry, petrology (e.g., Christopher et al., 2015; Edmonds et al., 2016) and geophysical studies
410 (e.g., Paulatto et al., 2012) show the mush system extends from <4 to 16 km, with the main active
411 reservoir between 8 – 12 km (Paulatto et al., 2012; Christopher et al., 2015; Edmonds et al., 2016; Fig
412 2).

413 **3.2 The Central Arc (including St Lucia)**

414 In the Central Arc (Guadeloupe to Dominica) andesites are dominant, with some basalts and dacites
415 (Fig. 3; 4). St Lucia is more commonly included in the Southern Arc based on seismic data (Wadge and
416 Shepherd, 1984), but fits into the Central Arc based on geochemistry and volcanology, so is also
417 discussed in this section.

418 **3.2.1 Guadeloupe**

419 The whole rock composition of La Soufrière de Guadeloupe eruption products have also remained
420 consistent, though they do span a wide range of compositions (51 – 61 wt% SiO₂; Touboul et al., 2007;
421 Boudon et al., 2008; Samper et al., 2009; Ricci et al., 2017; Pichavant et al., 2018; Metcalfe et al.,
422 2022). The 1530 CE sub-Plinian eruption (Komorowski et al., 2008; Boudon et al., 2008; Esposti Ongaro
423 et al., 2020) shows black-and-white banded pumices with characteristic mixing and mingling textures
424 between andesitic (ca. 62 wt% SiO₂) and basaltic-andesite (55 wt% SiO₂) compositions (Semet et al.,
425 1981; Poussineau, 2005; Boudon et al., 2008; Pichavant et al., 2018). Deposits from the Echelle scoria
426 cone also show a bimodal composition, with mixing textures between basaltic andesite and dacitic
427 compositions observed. To the south of La Soufrière de Guadeloupe, the Mont Caraïbes complex
428 erupted a basaltic composition (47 – 50 wt% SiO₂) and is often interpreted to represent the parental
429 magma for La Soufrière de Guadeloupe (Bissainte, 1995; Metcalfe, 2022). Eruption products Mont

1 430 Caraïbes are the least differentiated in the Central Arc, while eruption products from La Soufrière de
2 431 Guadeloupe are more typical of those observed (Fig 3).

3
4 432 Analysis of MIs found in clinopyroxene, orthopyroxene and plagioclase from La Soufrière de
5 433 Guadeloupe, and the neighbouring monogenetic cone of Echelle range from 58 – 79 wt% SiO₂ (Fig 4
6 434 A; Poussineau, 2005; Boudon et al., 2008; Pichavant et al., 2018; Cooper et al., 2020; Metcalfe et al.,
7 435 2022). Major elements show no systematic changes across the Holocene eruptions (Poussineau, 2005;
8 436 Boudon et al., 2008; Pichavant et al., 2018; Metcalfe et al., 2022; Fig 4). MIs hosted in clinopyroxene,
9 437 orthopyroxene, plagioclase and olivine from the basaltic parental magma from Mont Caraïbes have
10 438 also been measured for Guadeloupe. These MIs are notably less evolved in comparison to the MIs
11 439 from La Soufrière de Guadeloupe (49 – 56 wt% SiO₂) and are the least evolved inclusions observed in
12 440 the Central Arc (Metcalfe, 2022; Fig 4). Guadeloupe xenoliths are reported as plagioclase-rich,
13 441 isotropic and homogeneous, but texturally diverse. Bissainte (1999) describe gabbroic cumulates and
14 442 granodioritic enclaves which are interpreted as fragments of a granodioritic basement.

15
16 443 Guadeloupe MIs record variable Cl contents ranging from 0.06 - 0.36 (Fig 4C; Cooper et al., 2020;
17 444 Metcalfe et al., 2022; Metcalfe, 2022). H₂O generally falls in the range 1.85 - 4.42 wt%, with a
18 445 maximum of 5.6 wt% H₂O (Fig 4D; Pichavant et al., 2018; Cooper et al., 2020; Metcalfe et al., 2022).
19 446 The CO₂ contents from Guadeloupe are much higher than those recorded in the Northern arc ranging
20 447 from <50 – 1500 ppm (Fig 4E; Cooper et al., 2020; Metcalfe et al., 2022). MIs from from La Soufrière
21 448 de Guadeloupe are reported to be bubble-free (Metcalfe et al., 2022).

22
23 449 Most models of the current La Soufrière de Guadeloupe magma storage system have been based on
24 450 data collected for the 1530 CE eruption (Semet et al., 1981; Touboul et al., 2007; Boudon et al., 2008;
25 451 Pichavant et al., 2018). The observation of the compositionally banded 1530 CE eruption products
26 452 imply that the La Soufrière de Guadeloupe reservoir is chemically and physically heterogenous.
27 453 However, the relative compositional homogeneity of mineral phases observed in the 1530 CE eruption
28 454 products is difficult to reconcile with this scenario.

29
30 455 In the absence of any geophysical evidence for a deep long-lived magma reservoir, early
31 456 interpretations hypothesise the presence of an andesite magma body formed from fractional
32 457 crystallisation of a basaltic-andesite parental melt (Touboul et al., 2007; Pichavant et al., 2018). The
33 458 shallow andesite body must be remobilised by a magma, as the andesite body is considered to be
34 459 homogenous until remobilisation caused by the arrival of basaltic magma (Semet et al., 1981; Touboul
35 460 et al., 2007; Pichavant et al., 2018). With the arrival of a mafic recharge magma chemical and physical
36 461 changes occur in the magma storage zone. A hybrid magma may also be formed through incomplete
37 462 mixing (Pichavant et al., 2018). The magma mixing and development of instabilities in the reservoir

463 can lead to eruption (e.g., Gourgaud et al., 1989; Murphy et al., 1998, 2000; Pichavant et al., 2002).
464 Pichavant et al. (2018) suggest, based on melt inclusion volatile concentrations and experimental data,
465 a minimum depth of the La Soufrière Guadeloupe magma reservoir between 5.6 and 7.1 km (135 –
466 170 MPa, based on a density of 2450 kg m⁻³; Barnoud et al., 2016), and a maximum depth of 8.5 km
467 given the absence of amphibole phenocrysts in the juvenile eruptive products .

468 Moretti et al. (2020a) proposed a deep gas supply from a crystal mush conceptual model as used at
469 many similar systems to explain observations and activity (e.g., Cashman et al., 2017). This was refined
470 by Metcalfe et al. (2021, 2022) to a vertically and laterally extensive mush system extending 6 – 9 km
471 in depth (190 – 220 MPa, based on a density of 2450 kg m⁻³; Barnoud et al., 2016), which hosts evolved
472 magmas (Fig 2). Remobilization of this mush system to produce eruptible magma may occur through
473 the injection of mafic magma from depth into the shallow system, resulting in magma mixing.
474 Alternatively, in eruptions where no mixing textures are observed (e.g., 1657 CE), exsolved fluids and
475 heat from magma at depth may allow remobilization with no interaction with a mafic magma.
476 Investigation of major and volatile elements of melt inclusions across several Holocene eruptions
477 shows the composition of the storage zone has remained stable during this time period (Metcalfe et
478 al., 2022). There is some evidence for a deeper storage zone, hosting less -evolved compositions based
479 on the presence of a basaltic parental magma (Monts Caraïbes) and the basalt produced by the Echelle
480 and Citerne scoria cones (Bissante, 1999; Legendre, 2012; Metcalfe et al., 2022). Bissante (1999)
481 calculated crystallization pressures of 400 – 900 MPa for Mont Caraïbes cumulates, which is equivalent
482 to ca. 16 to 30 km depth (Fig 2).

483 Moretti et al. (2020) show the mush system is the source of heat and deep magmatic fluids that rise
484 and contribute to the hydrothermal system. The heat and fluids supplied by intrusions into the mush
485 system below 5 km are considered to be one possible cause of seismic activity (White and McCausland,
486 2016) and unrest episodes, as well as the potential cause of some of the more violent and long-lasting
487 historical non-magmatic phreatic activity produced from the long-lived hydrothermal system
488 (Feuillard et al., 1983; Komorowski et al., 2005; Villemant et al., 2005; Rosas-Carbajal et al., 2015,
489 Moretti et al., 2020a).

490 **3.2.2 Dominica**

491 The whole rock composition of eruption products from Dominica ranges from 50 – 67 wt% SiO₂
492 (Lindsay et al., 2005; Howe et al., 2015; Boudon et al., 2017; d’Augustin et al., 2020), with no large-
493 scale magma mixing or mingling textures reported (e.g., Howe et al., 2015; Boudon et al., 2017; Solaro
494 et al., 2020). The eruption products are relatively evolved within the LAA whole rock trend (Fig 3),
495 which is typical of the eruption products in the central arc. In comparison, the composition of MIs

496 hosted in orthopyroxene, clinopyroxene and plagioclase from Dominica shows a narrower range (69
1 497 – 79 wt% SiO₂, 0.2 – 4.5 wt% K₂O, 1.1 - 3.9 wt%; Fig 4A, B) and is homogenous, regardless of eruption
2
3 498 (Balcone-Boissard et al., 2018; d’Augustin et al., 2020; Cooper et al., 2020; Ostorero et al., 2021).
4
5 499 Xenoliths from Dominica are dominated by olivine with well-equilibrated textures (Ziberna et al.,
6
7 500 2017; Melekhova et al., 2019).

8
9 501 MIs from Dominica record a similar range of Cl contents (0.15 - 0.38 wt% Cl) as observed in
10
11 502 Guadeloupe and across the Central arc in general (Fig 4C; Balcone-Boissard et al., 2018; d’Augustin et
12
13 503 al., 2020; Cooper et al., 2020). H₂O contents from Dominica are the highest observed in the Central
14
15 504 Arc, with two distinct data clusters, one ranging from 4.5 – 7.3 wt% H₂O and the second from 1.4 - 3.3
16
17 505 wt% H₂O (Fig 4D; Balcone-Boissard et al., 2018; d’Augustin et al., 2020; Cooper et al., 2020). The
18
19 506 highest CO₂ contents in the LAA are recorded in Dominica, CO₂ is variable but again generally falls into
20
21 507 two clusters (3137 – 4012 ppm and <1856 ppm; Fig 4E; Balcone-Boissard et al., 2018; d’Augustin et
22
23 508 al., 2020; Cooper et al., 2020). Bubble-free MIs were analysed by Balcone-Boissard et al. (2018) and
24
25 509 d’Augustin et al. (2020).

26 510 A vertically and laterally extensive crystal mush system is also proposed for Dominica, with three
27
28 511 components: an underplating basalt, a basaltic– andesitic crystalline mush with a liquid hybrid layer,
29
30 512 and an interstitial rhyolitic melt which may accumulate at the top of the reservoir in a silicic cap (e.g.,
31
32 513 Howe et al., 2015). This system is hypothesised to have formed through continuous fractionation,
33
34 514 accumulation, and disaggregation of the crystalline material of a crystal-mush along with interstitial
35
36 515 melt. This is highlighted by crystals with disequilibrium features indicating they have experienced
37
38 516 different histories (Howe et al., 2015). Volatile release from the underplating basalt is thought to
39
40 517 cause vigorous convection prior to eruption. Mafic phases observed in the Dominica eruption products
41
42 518 are remnants of incomplete mixing of basaltic material (Halama et al., 2006; Howe et al., 2015).
43
44 519 Following the eruption and the cessation of recharge of the underplating basalt, the liquid remaining
45
46 520 in the reservoir will begin to crystallise to ultimately remain in a crystalline state between eruptive
47
48 521 periods (Howe et al., 2015).

49 522 However, more recently the model of mafic recharge has been discarded for some eruptions (e.g.,
50
51 523 Layou and Roseau) due to the lack of evidence for mixing in groundmass glasses and melt inclusion
52
53 524 compositions (Balcone-Boissard et al., 2018; Solaro et al., 2020). Instead, mixing of different parts of
54
55 525 the mush, without the need of a mafic magma, can explain many features observed for the Layou and
56
57 526 Roseau eruptions. The system prior to the eruption is considered as a cooling system which is
58
59 527 disturbed by a slightly hotter magma of a similar composition a decade prior to eruption (Solaro et al.,
60
61 528 2020). This suggests the system is heterogeneous in temperature (Solaro et al., 2019, 2020).
62
63
64
65

529 The large ignimbrite eruptions observed on Dominica (e.g., 60 and 24 ka BP) were fed from magmas
530 stored at ca. 16 km, ca. 400 MPa (based on a crustal density of 2450 g/m³; Balcone-Boissard et al.,
531 2018 and Solaro et al., 2019; Fig 2). In comparison, more recent Plinian eruptions (e.g., 8–18 ka) were
532 fed from magma that was sourced in shallower storage regions ≤ 7.3 km, ca. <200 MPa (based on a
533 mean crustal density of 2800 kg/m³; d’Augustin et al., 2020), this is consistent with the system
534 described for Montserrat (Christopher et al., 2015; Edmonds et al., 2016; Fig 2). The similarities in
535 mineralogical and chemical composition, including comparable volatile element ratios for magma that
536 fed the ignimbritic eruption and the more recent Plinian eruptions indicate a common source. This
537 supports the existence of a transcrustal magmatic system extending from <7 – 16 km (d’Augustin et
538 al., 2020; Fig 2).

539 3.2.3 Martinique

540 Most studies on Martinique focus on the homogeneity of the composition of eruption products in the
541 last 13,500 years of activity, though a wide range of compositions are recorded (51 - 66 wt% SiO₂;
542 Gourgand et al., 1989; Smith and Roobol, 1990; Pichavant et al., 2002; Annen et al., 2008). The
543 majority of eruption products appear to be evolved in comparison to the Northern and Southern Arc
544 eruption products (Fig 3). Mafic lavas, mafic enclaves, and mingled/mixed products are also reported
545 in the eruptive history of the volcano (Bourdier et al., 1985; Fichaut et al., 1989a, 1989b; Gourgand et
546 al., 1989). MIs hosted in orthopyroxene and plagioclase from Montagne Pelée, Martinique are evolved
547 in composition and fall into a similar range to MIs from Guadeloupe and Dominica (65 – 78 wt% SiO₂;
548 Fig 4A; Cooper et al., 2020; Martel et al., 1998). Na₂O + K₂O falls in the upper range observed in the
549 central arc (3.1 - 5.0 wt% Na₂O, 1.7 - 4.1 wt% K₂O; Fig 4 B; Cooper et al., 2020; Martel et al., 1998).
550 Martinique xenoliths are comparable to those from St Kitts and are plagioclase-bearing (Cooper et al.,
551 2016).

552 Volatiles from Martinique MI are reported by Cooper et al. (2020). Cl contents (0.27 - 0.30 wt% Cl; Fig
553 4C) fall into the range observed in the Central arc. H₂O contents, however, are generally lower than
554 the others systems in the Central arc (<3.3 wt% H₂O; Fig 4D) with one MI recording 5.2 wt% H₂O. CO₂
555 in Martinique MIs (<381 ppm; Fig 4E) is lower than observed in Guadeloupe and Dominica and is most
556 comparable to St Lucia (Cooper et al., 2020). Vapour bubbles are not reported by Cooper et al. (2020).

557 The main andesitic magma reservoir beneath Montagne Pelée is inferred using the petrology of
558 erupted rocks as well as results from experimental petrology to be at pressures of 6 – 9 km, ca. 200 ±
559 50 MPa (based on a density of 2650 kg m³; Martel et al., 1998; Fig 2). Mafic magmas have been
560 experimentally determined to be stored at > 16 km (400 to 500 MPa), suggesting that the magmatic
561 transcrustal system below Montagne Pelée also has a similar deeper component, like magmatic

562 feeding systems identified in Dominica and Montserrat (Pichavant et al., 2002; Edmonds et al., 2016).
563 The model proposes that magma rising from a depth > 16 km stalls and forms a deeper storage zone
564 located between 12 - 16 km depth (Pichavant et al., 2018; Boudon and Balcone, 2021).

565 The presence of mafic magmas with an andesite bulk composition leads to the interpretation of the
566 existence of a zoned magma storage system with a mafic zone below an andesitic zone (Pichavant et
567 al., 2002), similar to that presented for Guadeloupe by Pichavant et al. (2018). Annen et al. (2008)
568 proposed a mush system for Montagne Pelée based on experimental determination of the thermal
569 evolution of the system. The formation of a long-lived persistent magma chamber was determined to
570 be unlikely based on realistic magma injection and cooling rates. Instead, the magma system is built
571 up by accreting sills. This injected magma then completely solidifies, existing as a highly crystalline
572 mush with only a small part (10–20 vol%) remaining as eruptible melt to feed eruptions (Annen et al.,
573 2008). Magmas are calculated to ascend in < 1 day from a shallow reservoir of 6 km (ca. 200 ± 50 MPa)
574 (Martel, 2012).

575 **3.2.4 St Lucia**

576 In the Soufrière volcanic complex of St Lucia no large-scale mixing or mingling textures are reported
577 in the silica-rich deposits (62 – 68 wt % SiO₂; Bezard et al., 2014, 2015, 2017). A wider range of
578 compositions (basalt to rhyolite) are observed in older deposits extending this range to 50 - 68 wt%
579 SiO₂ (18 and 1.1 Ma; Briden et al., 1979; Aquater, 1982; Le Guen De Kerneizon et al., 1983; Samper et
580 al., 2008; Bezard et al., 2017). The most differentiated eruption products in the central arc are found
581 on St Lucia, though generally the data fits with the trends observed in the Central Arc (Fig 3).

582 Fewer MIs have been studied for St Lucia; the clinopyroxene hosted MIs fall at the more evolved end
583 of the Central Arc compositional range (76 – 78 wt% SiO₂, 3.3 – 3.9 wt% Na₂O, 3.2 – 3.8 wt% K₂O; Fig
584 4A, B; Cooper et al., 2020). The xenoliths from St Lucia contain abundant amphibole and plagioclase
585 and only scarce olivine and are more comparable in composition to those from the central arc in
586 comparison to xenoliths from the Southern Arc (Melekhova et al., 2019).

587 Cooper et al. (2020) report volatiles from St Lucia MIs, Cl is relatively low compared to other systems
588 in the Central Arc (0.14 - 0.20 wt% Cl; Fig 4C). MI H₂O content falls into the range observed in the
589 central arc, ranging from 2.0 – 3.6 wt%. CO₂ content (<600 ppm; Fig 4E) is comparable to that observed
590 in Martinique MIs. Vapour bubbles are not reported by Cooper et al. (2020).

591 Bezard et al. (2017) present a schematic model for the St Lucia magma system. This model
592 hypothesises three storage zones with fractionation occurring at the base of the crust (>25 km).
593 Storage of basaltic-andesite to andesite magma is hypothesised to occur in the mid-crust, in a chamber
594 at 580 – 720 MPa (ca. 20–26 km depth, based on a crustal density of 2660 kg/m (Christeson et al.,

595 2008). A shallow chamber is also hypothesised at, 110 – 210 MPa (4–8 km depth based on a crustal
596 density of 2660 kg/m (Christeson et al., 2008). Magma from the mid-crustal chamber recharges the
597 shallow chamber with mixing hypothesised between the two magmas, shown by resorption textures
598 observed in plagioclase.

599 **3.3 The Southern Arc**

600 **3.3.1 St Vincent**

601 Magmas from St Vincent have a narrow range of compositions, with Soufrière St Vincent producing
602 basalt to andesite compositions (47 - 62 wt% SiO₂; Graham and Thirlwall, 1981; Heath et al., 1998;
603 Cole et al., 2019; Fedele et al., 2021; Joseph et al., 2022). The majority of eruption deposits are basalt
604 to basaltic andesite, with andesite mainly found in mixed magma deposits as the more evolved
605 component. For example, banded and mingled textures are observed in the 1902–03 and 1979
606 eruption deposits (Carey and Sigurdsson, 1978; Graham and Thirlwall, 1981; Heath et al., 1998a). The
607 greatest compositional variation is observed in samples from the 1902-03 eruption (50 - 55 wt% SiO₂;
608 Cole et al., 2019) with magmas from older eruptions generally being more homogenous and more
609 evolved (e.g., 1440 CE and 1580 CE; Cole et al., 2019). Despite the range of compositions, most of the
610 whole rock compositions appear relatively evolved within the LAA whole rock trend (Fig 3).

611 MIs hosted in plagioclase, orthopyroxene, clinopyroxene and titanomagnetite from Soufrière St
612 Vincent (55 – 77 wt% SiO₂; Heath et al., 1998; Cooper et al., 2020), are more comparable to those
613 from the Central Arc, spanning a similar range to the MIs from La Soufrière de Guadeloupe (Fig 4A).
614 As with the systems in the Central Arc (e.g., La Soufrière de Guadeloupe and Dominica) the MIs from
615 Soufrière St Vincent generally, indicate the system is fed by homogenous magmas with episodic
616 rejuvenation events (Fedele et al., 2021). Devine and Sigurdsson (1982) and Bouvier et al. (2010b)
617 measured MIs in olivine from basaltic-andesite eruption products (43 – 60 wt% SiO₂). St Vincent
618 xenoliths lack orthopyroxene and are rich in olivine (Melekhova et al., 2019).

619 The majority of MIs from St Vincent record Cl contents ranging from 0.08 – 0.29 wt%, with a few MIs
620 extending the range to 0.61 wt% Cl (Fig 4C; Heath et al., 1998; Bouvier et al., 2010b; Cooper et al.,
621 2020;). H₂O content is variable ranging from 1.50 – 5.3 wt% (Fig 4D; Bouvier et al., 2010b; Cooper et
622 al., 2020), and CO₂ contents are low (76 – 103 ppm), however, this may be due to the small amount
623 of CO₂ data available (Fig 4E; Cooper et al., 2020;). Vapour bubbles are reported by Heath et al. (1998)
624 and Bouvier et al. (2010b), therefore pressures must be considered minimum estimates.

625 Soufrière St Vincent experiences relatively steady-state behaviour, fed by homogenous magmas
626 initially stored at 7 km, ca. 200 MPa (based on a 2660 kg/m (Christeson et al., 2008)). Petrological and
627 geochemistry studies indicate magmas are stored in a shallow crustal reservoir at ca. 7 km (and

628 perhaps as shallow as 3.5 km. ca. 100 MPa (based on a 2660 kg/m (Christeson et al., 2008)). The
629 magmas evolve and undergo homogenisation, leading to the eruption of chemically homogeneous
630 magmas (Fedele et al., 2021) (Fig 2). The stored magmas exsolve volatiles which contribute to the
631 build-up of pressure in the shallow system which may trigger eruptions (Cole et al., 2019; Fedele et
632 al., 2021). The 2020 -2021 eruption is inferred to have begun with the remaining 1979 high-viscosity,
633 degassed magma in the conduit being pushed up by the injection of new gas-rich magma (Joseph et
634 al., 2022). The injection of gas rich magma is thought to have occurred at ca. 6 km based on the depth
635 of seismic events (Joseph et al., 2022). The displacement of the high-viscosity magma cap by the new
636 lower viscosity gas-rich magma resulted in an acceleration in extrusion rate and on reaching the
637 surface experienced explosive fragmentation (Joseph et al., 2022).

638 However, eruption products from the 1902 -03 eruption and the presence of cumulates indicates the
639 shallower reservoir experiences episodic perturbations by magma rejuvenation events from a less-
640 evolved basaltic, parental magma at a deeper level calculated at ca. 13 km, ca. 400 MPa (Fedele et al.,
641 2021) (Fig 2).

642 3.3.2 Kick 'em-Jenny

643 Kick 'em-Jenny has produced the most eruptions in the Lesser Antilles in the 20th and 21st centuries.
644 Whole rock compositions range from 48 – 59 wt% SiO₂, with an evolution in magma composition
645 observed (Devine and Sigurdsson, 1995; Camejo-Harry et al., 2019). The 1972 eruption produced
646 predominantly basalts while the more recent eruptions (e.g., 2013/2014) produced basaltic andesites
647 (Camejo-Harry et al., 2019). Kick 'em Jenny eruption products are some of the least differentiated
648 across the arc (Fig 3).

649 MIs hosted in olivine, clinopyroxene and hornblende from Kick 'em-Jenny also extend to unevolved
650 compositions (48 – 67 wt% SiO₂; Fig 4A; Devine and Sigurdsson, 1995; Camejo-Harry et al., 2019)
651 particularly when compared to the Central arc. Kick 'em-Jenny MIs also have a higher Na₂O + K₂O
652 content in comparison to other systems in the Southern arc (2.2 - 7.2 wt% Na₂O, 0.7 – 4.3 wt% K₂O;
653 Fig 4B; Devine and Sigurdsson, 1995; Camejo-Harry et al., 2019). MIs record the highest Cl contents
654 across the LAA, with the majority of Cl contents generally in the range 0.31 - 0.66 wt%, with some MIs
655 recording up to 0.87 wt% Cl (Fig 4C; Camejo-Harry et al., 2019). H₂O content ranges from 0.2 - 5.0
656 wt%, which is comparable to H₂O contents recorded in the Southern Arc (Fig 4D; Camejo-Harry et al.,
657 2019). Kick 'em Jenny MIs record the highest CO₂ contents in the southern Arc, with the majority of
658 MIs falling in the range <50 – 784 ppm, with two MIs at 2513 ppm and 3157 ppm (Fig 4E; Camejo-
659 Harry et al., 2019). Bubble-free MIs were analysed by Camejo-Harry et al. (2019).

660 Camejo-Harry et al. (2019) present data on the magma storage system for the submarine volcano Kick
661 'em Jenny, suggesting the presence of a main storage region at ca. 7 – 12 km, ca. 170 - 325 MPa (based
662 on densities of 2500 – 2750 kg/m³; Camejo-Harry et al., 2019), with some shallower networks of melt
663 and mush (Fig 2). Combining petrology, geochemistry and seismic data they show that the magmatic
664 system extends through the crust with fractionation occurring deep in the system. This feeds the
665 shallower system, which becomes increasingly evolved in composition. This vertically extensive
666 system produces eruptions through the build-up of volatiles, generation of overpressure and
667 destabilisation of the system. Periods of dormancy are ended by basaltic recharge at depth, which
668 leads to fractionation and the migration of more evolved melts and the liberation of volatiles and
669 renewed eruptive activity of the system.

670 3.3.3 Grenada

671 Eruptions on Grenada produce basalt to basaltic-andesite from scoria cones and explosion craters
672 (Arculus, 1976; Van Soest, 2000; Stamper et al., 2014), with andesite and dacite much less common in
673 comparison to other islands (Arculus, 1976). The most primitive compositions are observed on
674 Grenada (42 – 67 wt% SiO₂; Cawthorn et al., 1973; Arculus et al., 1976; Thirwall and Graham, 1984;
675 Thirwall et al., 1996; Woodland et al., 2002; Rojas-Agramonte et al., 2017), the least evolved
676 compositions can be considered the primitive magmas of the LAA (Fig 3). The basalts in Grenada can
677 be split into two groups: the 'low-Sr' series which are Mg-rich olivine basalts, and the 'high-Sr' series
678 which are clinopyroxene-phyric series (Stamper et al., 2014).

679 The most primitive MI host in olivine are recorded in eruption deposits from Grenada however, more
680 evolved MIs are also present (42 - 74 wt% SiO₂; Fig 4A; Bouvier et al., 2010a; Cooper et al., 2020). Na₂O
681 +K₂O fall into a range typical of the Southern Arc (1.5 - 4.5 wt% Na₂O, 0.3 – 3.1 wt% K₂O; Fig 4B; Bouvier
682 et al., 2010a; Cooper et al., 2020). Grenada xenoliths are dominated by mafic minerals but have few
683 orthopyroxenes and plagioclase-free samples (Melekhova et al., 2019).

684 MIs from Grenada record Cl contents in the range 0.04 - 0.33 wt%, with three points in the range 0.54
685 – 0.67 wt% Cl and fall into the range observed in the Northern arc (Fig 4C Bouvier et al., 2010a; Cooper
686 et al., 2020). The highest H₂O contents in the LAA are recorded in MIs which fall in the range 0.2 - 8.4
687 wt% (Fig 4D; Bouvier et al., 2010a; Cooper et al., 2020). CO₂ contents are low <50 – 111 ppm and are
688 similar to the other systems in the Northern Arc (Fig 4E; Bouvier et al., 2010a; Cooper et al., 2020).
689 Vapour bubbles are reported by Bouvier et al. (2010a), however, no relation between bubble size and
690 water content is observed.

691 4 Volcanism, Eruptive Style and Volcanic Unrest Across the Arc

692 The LAA has a long history of eruptions with 24 eruptions of varying types occurring since 1900,
693 including 9 from volcanoes on Guadeloupe, Martinique, St Vincent, Montserrat and Dominica and 14
694 from the submarine volcano Kick ‘em Jenny (Lindsay et al., 2005a; Table 1; Fig. 5). Most islands have
695 a single volcano from which Holocene eruptions have occurred, and which are likely to erupt again in
696 the future (e.g., Saba, Statia, Nevis, Montserrat, Martinique, St Vincent). However, more complicated
697 systems are present with two active volcanoes in Guadeloupe (La Soufrière de Guadeloupe volcano,
698 and the Trois-Rivières-Madeleine volcanic complex; Komorowski et al., 2005; Samper et al., 2009;
699 Komorowski et al., 2013; 2022) and nine potentially active volcanoes in Dominica (Lindsay et al.,
700 2003, 2005c). Since 1900, eruptions have varied in size and style ranging from subaerial phreatic (e.g.,
701 Guadeloupe 1956 and 1976-1977 (Le Guern et al., 1980; Feuillard et al., 1983; Zlotnicki et al., 1992;
702 Komorowski et al., 2005; 2018; Hincks et al., 2014); Dominica 1997 (Mayer et al., 2017; ; Global
703 Volcanism Program, 2023) to magmatic (e.g., Montserrat 1995 – present (Wadge et al., 2014) and St
704 Vincent in 2020-2021 (Joseph et al., 2022). Submarine eruptions have also occurred at Kick ‘em Jenny
705 (Devine and Sigurdsson, 1995; Allen et al., 2018); Table 1; Fig 5).

706 Most volcanoes in the LAA are stratovolcanoes which have often experienced lava domes eruptions,
707 particularly in the Northern and Central arcs (including St Lucia). The Northern and Central volcanoes
708 can produce large explosive eruptions generating pyroclastic flows as well as lava domes and
709 associated block and ash-flows (e.g., Macdonald et al., 2000; Lindsay et al., 2013; Smith et al.,
710 2013; Bezard et al., 2014; Howe et al., 2015). Volcanoes of the Central Arc (Soufrière of Guadeloupe
711 and Montagne Pelée) also have produced numerous explosive Plinian, VEI 4 eruptions (e.g., the P3 79
712 CE Plinian eruption of Montagne Pelée (Boudon and Balcone, 2021; Carazzo et al., 2020; Michaud-
713 Dubuy et al., 2021), the 1010 CE GDS 3 eruption of La Soufrière (Legendre, 2012; Komorowski et al.,
714 2012; Metcalfe et al., 2021a; 201b) as well as Subplinian VE3 eruptions such as the 1530 CE eruption
715 of Soufrière of Guadeloupe (Boudon et al., 2008; Komorowski et al. 2008; Esposti Ongaro et al.,
716 2020). Larger magnitude VEI 5 Plinian eruptions have also been recorded, for example in Guadeloupe
717 (e.g., the Pintade eruption ca. 42 ka from the Grande Découverte-La Soufrière volcanic complex,
718 Komorowski et al., 2005), Dominica (e.g., the Roseau Tuff; Carey and Sigurdsson, 1980; Solaro et al.,
719 2020) and St Lucia (e.g., the Belfond unit; Lindsay et al., 2013). However, such activity occurred 20,000
720 to 45,000 years ago, and Plinian eruptions of this scale are not recorded in recent history. As well as
721 stratovolcanoes, the Southern Arc activity has produced a diversity of eruptive phenomena including
722 small explosion craters and lava flows, whereas large dome-forming eruptions are rare due to the
723 magma chemistry (Lindsay and Robertson, 2018).

724 Some of the most recent sub-aerial, magmatic activity has occurred on the islands of Martinique (e.g.,
725 1902 – 1905; 1929 - 1932), St Vincent (e.g., 1902; 1971-1972; 1979 and 2020 - 2021) and Montserrat

726 (e.g., 1995 – present). These eruptions exemplify the multi-phase, transitional activity style that occurs
727 at volcanoes in the Lesser Antilles (Table 1; Fig 5). Here, we discuss some of the most recent and
728 important eruptions in the Lesser Antilles. Therefore, we demonstrate the hazards the volcanoes
729 pose, and the potential impact for the people living on the islands. We also discuss recent unrest at
730 these systems and the implications for future eruptions.

731 **4.1 Soufrière Hills Volcano, Montserrat**

732 Soufrière Hills volcano in Montserrat has experienced a long period of eruptive activity, beginning in
733 1995 and continuing to the present day. This eruption has provided insights into a range of
734 phenomena regarding andesite dome building eruptions including: [1] cyclicity of magma transport
735 linked with ground deformation, [2] shallow seismicity and eruptive activity, [3] growth and collapse
736 of andesite lava domes, [4] sector collapse and formation of debris avalanches, [5] dynamics of lateral
737 blasts and vulcanian explosions, [6] eruption transitions from effusive to explosive activity triggered
738 by dome collapse, and [7] sub-surface processes and structure of the system, amongst many others.

739 This eruption is split into five phases with intervals of minimal activity which escalated to the next
740 phase (Cole et al., 2014a). Phases 1 - 3 were long-lived in comparison to Phases 4 - 5 which were
741 shorter but more vigorous (Wadge et al., 2014a). The onset of the eruption was preceded by three
742 years of elevated local seismicity (Robertson et al., 2000; Young et al., 2002; Shepherd et al., 2002;
743 Odbert et al., 2014b) which was followed by an explosive phreatic phase that began on 18 July 1995
744 and lasted until 15 November 1995. At this point the onset of dome growth was observed marking
745 the transition into the magmatic phases of the eruption.

746 Phase 1 (15 November 1995 – 10 March 1998) began with a cycle of dome extrusion and Vulcanian
747 explosions leading to dome collapse and pyroclastic density current surges (Voight et al., 1998; Sparks
748 et al., 1998; Cole et al., 1998, 2002; Robertson et al., 1998; Calder et al., 2002; Kokelaar, 2002; Wadge
749 et al., 2014). The explosion on 17 September 1996 produced a column that reached up to 15 km and
750 may have been of Sub-Plinian magnitude and intensity (Robertson et al., 1998). A noticeable pumice
751 deposit was produced with ashfall in the Caribbean region, which was particularly heavy in
752 Guadeloupe. In 1997 activity changed and two series of Vulcanian explosions occurred with a sub-
753 daily cyclic pattern that produced a total of 88 explosions (Druitt et al., 2002b). On 26 December 1997,
754 this phase of the eruption ended with partial edifice collapse of the old pre-eruption crater wall and
755 part of the growing dome. This led to the emplacement of a debris avalanche deposit that reached
756 the South-West coast and triggered a small tsunami with a maximum wave run-up height of 1 m on
757 the western shores of Montserrat (Sparks et al., 2002). The partial edifice collapse also resulted in a
758 sudden decompression at the base of the growing dome, triggering a violent laterally-directed

1 759 explosion. This generated a high-energy, high-velocity, turbulent, dilute pyroclastic density current,
2 760 and resulted in the most widespread destruction recorded during the eruption (Young et al., 2002;
3 761 Voight et al., 2002). This first paroxysmal event in Montserrat was followed by an 18 month pause in
4
5 762 lava extrusion (Kokelaar, 2002).
6

7 763 Phase 2 (27 November 1999 – 28 July 2003) was preceded by seismic swarms, ash venting and small
8
9 764 explosions (Norton et al., 2002). This culminated in pumice-bearing explosions and later initiated the
10
11 765 longest period of continuous lava extrusion (Wadge et al., 2014). This phase was characterised by
12
13 766 major dome collapse events, with cyclic processes of magma flux and gas flux observed. Dome collapse
14
15 767 events in this phase resulted in ash columns up to 9 km high (Carn et al., 2004), which also caused
16
17 768 repeated minor ashfalls in the Caribbean region, with much of the collapse material deposited by
18
19 769 block-and-ash flows into the sea (e.g., Le Friant et al., 2010). Phase 2 ended with a series of Vulcanian
20
21 770 explosions and a total dome collapse of ca. $2 \times 10^6 \text{ m}^3$ of lava (Edmonds et al., 2006; Voight et al.,
22 771 2010b).
23

24 772 Prior to Phase 3 (1 August 2005 – 20 April 2007) a 3.5 month-long series of phreatic episodes was
25
26 773 observed, these eruptions resulted in additional fracturing of the shallow regions of the volcanic
27
28 774 edifice which facilitated magma transport (Wadge et al., 2014). Dome growth began in August 2005,
29
30 775 with the growth rate increasing four-fold by December 2005, allowing rapid growth (Ryan et al., 2010;
31
32 776 Wadge et al., 2014). Through the beginning of 2006 dome growth rate and direction varied and by
33
34 777 May 2006 a lava spine with an approximate diameter of 50 m had grown. Complete dome collapse,
35
36 778 which included the remnants of the Phase 2 dome, occurred in May 2006, producing PDCs. Following
37
38 779 this further dome and growth collapses were observed, with magma extrusion varying, some ash
39
40 780 venting was observed at the end of 2006 on the rapidly expanding dome. This phase was different to
41
42 781 the previous two phases with little explosive activity observed in comparison to Phase 1 - 2 (Wadge et
43 782 al., 2014).
44

45 783 Phase 4 is split into two periods of activity, 4A (Aug 8th to Oct 8th) and 4B (Dec 2nd to Jan 3rd), with some
46
47 784 possible explosive activity preceding this in March 2008. Phase 4A began with escalating seismicity
48
49 785 over 48 hours, 12 hours after this seismicity ended, a large explosion occurred producing a 12 km high
50
51 786 eruption column and dome collapses (Chardot et al., 2010; Komorowski et al., 2010). Additional
52
53 787 Vulcanian explosions that produced pumice fallout deposits and small pumice-rich pyroclastic density
54
55 788 currents occurred on 3 December 2008, and 3 January 2009 (Komorowski et al., 2010). Phase 4 was
56
57 789 short but was characterised by a high degree of explosivity leading to pyroclastic density currents.
58
59 790 Lava extrusion rates were also high and there was a different cyclicity than observed in the previous
60
61 791 phases (Wadge et al., 2014).
62
63
64
65

792 Phase 5 (9th October 2009 – 11th February 2010) began with ash venting followed by lava extrusion
1 793 and pyroclastic density currents. This was followed by large Vulcanian explosions and the termination
2 794 of phase 5 with a major dome collapse on 11 February 2010 (Stinton et al., 2014b). This generated the
3 795 second laterally directed explosion of the eruption, which directed high-energy, high-velocity,
4 796 turbulent, dilute pyroclastic density currents northward. This resulted in significant damage on areas
5 797 previously affected by pyroclastic density currents in 1997. This paroxysmal event, the last of the
6 798 ongoing eruption so-far, also generated widespread ashfalls in the Caribbean region from Guadeloupe
7 799 to Saint-Lucia (Komorowski et al., 2011; Cadelis et al., 2013). Since 11th February 2010, a pause in the
8 800 dome growth began, which has continued up to the time of writing, making it the longest pause so far
9 801 in the eruption.

10 802 Overall, the eruption at Soufrière Hills has been characterised by periods of dome-growth, dome
11 803 collapse and a diversity of Vulcanian to Sub-Plinian explosions, with the eruption becoming
12 804 increasingly vigorous (Druitt and Kokelaar, 2002; Fig 5). In particular, the cyclicities of this eruption
13 805 have provided important observations on the nature of eruptions and have allowed a better
14 806 understanding of the drivers and behaviour of this volcano (Odbert et al., 2014). The cyclicity in
15 807 eruptive styles and the unique longevity of the eruption, highlight the importance of the continuous
16 808 monitoring of this system, which remains active and hazardous even during quiescence periods.

17 809 Following the end of lava extrusion and the last collapse event which occurred on 11th February 2010,
18 810 the Hazard Level was lowered from Level 4 to Level 3 and a pause in the eruption began (MVO
19 811 Scientific Updates, 2010-21). This followed the previous pattern in the eruption where activity was
20 812 minimal between eruption phases. However, it is unclear how long this pause in activity will last.
21 813 Through 2010 seismicity and SO₂ emissions continued, as well as minor incandescence on the dome
22 814 related to hot gasses from a deep source passing through fractures. The observed deformation pattern
23 815 showed slow inflation of the volcano, with ash venting and degradation of the lava dome observed
24 816 along with occasional, small pyroclastic flows. This low level of activity continued into 2011, with the
25 817 Hazard Level reduced again to Level 2 in November 2011 (MVO Scientific Updates, 2010-21).

26 818 Activity remained at a low level in 2012 and was characterized by spots of incandescence on the lava
27 819 dome and continued fumarolic activity and seismic swarms. Peaks in activity were often associated
28 820 with ash venting, with significant ash venting observed in March 2012, resulting in the formation of a
29 821 new crater. Large VT events associated with the venting are interpreted to be related to increasing
30 822 pressure below the volcano from uprising magma, while the ash venting was interpreted to be
31 823 phreatic in origin (MVO Scientific Updates, 2010-21).

824 This low level of activity continued through 2013 and 2014 with periods of elevated seismicity and gas
825 emissions. However, no changes to fumarolic activity or the dome were observed and in August 2014
826 the Hazard Level was lowered again to Level 1 (MVO Scientific Updates, 2010-21). Since 2017, activity
827 has remained low and consistent with short-lived seismic swarms and SO₂ output generally at 200 -
828 400 tons per day, though some short-lived peaks in seismic activity and SO₂ output do occur (MVO
829 Scientific Updates, 2010-21). In recent years, lahars, pyroclastic flows and rockfalls pose the most
830 significant threat, particularly following heavy rainfall. Some larger earthquakes continue to be felt by
831 the population, however, most felt earthquakes are now unrelated to Soufriere Hills Volcano (MVO
832 Scientific Updates, 2010-21). The monitoring data is consistent with the previous pauses in activity
833 observed during this eruption, with escalation in activity possible as the eruption has not been
834 declared to be officially over. For the eruption to be considered over, the seismic activity, gas
835 emissions and deformation must all return to substantially lower levels than have so far been
836 observed.

837 **4.2 La Soufrière de Guadeloupe**

838 Though La Soufrière de Guadeloupe has not experienced a magmatic eruption since 1657 CE, the 1976
839 – 1977 phreatic eruption demonstrates the impact a limited VEI 2, non-magmatic, explosive eruption
840 can have (Komorowski et al., 2005; Legendre, 2012; Komorowski et al., 2012; Rosas-Carbajal et al.,
841 2016; Metcalfe et al., 2021a, b; Komorowski et al., 2022). This eruption also highlights the importance
842 of clear, effective and efficient crisis response management, and risk reduction plans. The eruption
843 posed a significant risk to the population of Guadeloupe and resulted in severe socio-economic
844 difficulties associated with the evacuation of >70,000 people from Basse-Terre, and had a lasting
845 influence on the population of Guadeloupe (Le Guern et al., 1980; Feuillard et al., 1983; Komorowski
846 et al., 2005; Hincks et al., 2014; Chenet et al., 2014; Komorowski et al., 2015). One year prior to the
847 1976 - 77 event, volcanic seismicity began to increase with three distinct earthquake swarms (Feuillard
848 et al., 1983; Komorowski et al., 2005; Hincks et al., 2014). Despite this increase in seismicity, no large
849 changes in fumarolic activity were observed (Dorel and Feuillard, 1980; Feuillard et al., 1983). One
850 month prior to the eruption a minor landslide occurred and new fractures formed (Feuillard, 2011).
851 Between July 1976 and March 1977, 26 phreatic explosions occurred, which changed the mechanical
852 state of the dome through the opening of new major fractures and widening and deepening of existing
853 fractures and craters.

854 Some controversy has surrounded the scientific management of this eruption and the resulting crisis
855 response decisions made by authorities (Feuillard, 2011; Komorowski et al., 2005; Beauducel, 2006;
856 Hincks et al., 2014; Chenet et al., 2014; Komorowski et al., 2015; Komorowski et al., 2018). Particularly
857 concerning the report of fresh, unaltered, glassy andesite which was taken as evidence for the

858 presence of new magma close to the surface and that the eruption was escalating rapidly (Marinelli,
1 859 1976; Brousse et al., 1977; Heiken et al., 1980). This observation had a major influence on the scientific
2
3 860 management of the crisis, though ultimately the material was identified as non-juvenile clasts
4
5 861 removed from a pumice fallout deposited during the 1530 CE eruption (Boudon et al., 2008;
6
7 862 Komorowski et al., 2008).

8
9 863 There is also some debate on the cause of this eruption with two models proposed. The more widely
10
11 864 accepted model relates the phreatic eruption to a 'still-born' magmatic event (Feuillard et al., 1983).
12
13 865 Evidence for this model is provided by the depth of seismicity at ca. 6 km, consistent with the depth
14
15 866 of the magma reservoir and SO₂ and Cl signatures which were magmatic in origin. The migration of
16
17 867 earthquake hypocentres from depth towards the shallower system over a few hours was also
18
19 868 observed and considered to support this model (Feuillard et al., 1983; Villemant et al., 2005; Boichu
20
21 869 et al., 2011; Li et al., 2016). Ruzié et al. (2012) suggest new magma is being continuously injected into
22
23 870 a shallow magma reservoir at ca. 6 km based on modelling of the flux of noble gas isotopes (⁴He/³He;
24
25 871 R/Ra) in hydrothermal fluids and on petrological analysis of the 1530 CE eruption deposits (Semet et
26
27 872 al., 1981; Pichavant et al., 2018). Magmatic gases are buffered by the hydrothermal system but an
28
29 873 increase in heat flux and migration of chemically reactive fluids in the hydrothermal system are
30
31 874 responsible for the sealing of the host-rock and resulting pressurization of the system leading to the
32
33 875 phreatic activity observed. The injection of magma into the reservoir is interpreted as the triggering
34
35 876 source of several unrest episodes in La Soufrière de Guadeloupe's history including new pulses in
36
37 877 1992, 1997-1998 (Villemant et al., 2014) and 2018 (Moretti et al., 2020).

38
39 878 However, Zlotnicki et al. (1992) proposed that new magma arrival is not required for a phreatic
40
41 879 eruption to occur. Therefore, hydrothermal activity alone, with deposition of impermeable clay
42
43 880 minerals that seal of fractures combined with high fluid circulation can result in pressurization of
44
45 881 aquifers and phreatic eruption. Moune et al. (2022) also discuss this model for generating phreatic
46
47 882 eruptions in the context of the more recent unrest. Independent of the exact mechanism required to
48
49 883 generate the phreatic event of 1976-77, all gas data and the hydrothermal fluid C, He and Cl isotopic
50
51 884 ratios indicate a persistent magma reservoir degassing and supplying fluids and heat to the shallow
52
53 885 hydrothermal system (Brombach et al., 2000; Ruzié et al., 2012; 2013; Allard et al., 2014; Jean-Baptiste
54
55 886 et al., 2014; Villemant et al., 2014; Tamburello et al., 2019; Inostroza et al., 2022a, b; Moretti et al.,
56
57 887 2020a, b; Moune et al., 2022).

58
59 888 Following the 1976-77 eruption, monitoring of this volcano was increased to include continuous
60
61 889 seismic and geodetic monitoring (using global navigation satellite system; GNSS), gas plume
62
63 890 monitoring, meteorological networks, periodic extensometric surveys of lava dome fractures,
64
65 891 fumarole gas and thermal spring sampling and thermal monitoring (OVSG-IPGP, 1999–2023;

892 Komorowski et al., 2005; Jacob et al., 2005; Beauducel, 2006; Villemant et al., 2005; 2014; Moretti et
1 893 al., 2020a,b; Jessop et al., 2021; Moune et al., 2022; Komorowski et al., 2022). This allowed decreasing
2
3 894 activity to be observed from 1984, with minimal activity observed through the 1980s until May 1992
4
5 895 when seismic swarms and a new degassing phase began with summit fumarole reactivation (Zlotnicki
6
7 896 et al, 1994; Komorowski et al., 2001; Komorowski et al., 2005; OVSG-IPGP, 1999-2023). Over the next
8
9 897 five years fumarolic activity extended at the summit, and then more recently along its eastern flanks,
10
11 898 and to the top of the lava dome with the reactivation of three fumaroles (OVSG-IPGP, 1999-2021;
12
13 899 Komorowski et al., 2005; Tamburello et al., 2019; Jessop et al., 2021). Late 1997 and early 1998 marked
14
15 900 the onset of highly acidic chlorine-rich fluids degassing, which has continued till present day
16
17 901 (Komorowski et al., 2005; OVSG-IPGP, 1999-2023; Villemant et al., 2014; Moretti et al., 2020a,b).

18 902 Seismic activity is characterised by numerous, low magnitude ($M < 1$), high-frequency volcano-tectonic
19
20 903 earthquakes which have been observed to increase in frequency and rate with twenty felt volcanic
21
22 904 earthquakes since 1992, including ten since 2013 (Komorowski et al., 2005; Moretti et al., 2020a).
23
24 905 Between April 2017-December 2018 unrest reached the highest level since 1976-77 (Moretti et al.,
25
26 906 2020a), with a peak in seismicity observed on 27 April 2018 when a M 4.1 earthquake was felt by the
27
28 907 Basse-Terre population, the largest earthquake in 42 years (OVSG-IPGP, 1999-2023; Moretti et al.,
29
30 908 2020a).

31 909 Changes in gas chemistry were also detected before and after the April 2018 earthquake. Low $\text{CO}_2/\text{H}_2\text{S}$
32
33 910 and $\text{SO}_2/\text{H}_2\text{S}$ ratios were observed until November 2017 when a strong chemical perturbation
34
35 911 occurred. Perturbations at the time of the earthquake resulted in high $\text{SO}_2/\text{H}_2\text{S}$, CO_2/CH_4 , $\text{H}_2/\text{H}_2\text{O}$ and
36
37 912 CO/CO_2 and He/CH_4 recorded by MultiGas and Giggenbach bottles (Moretti et al., 2020a,b; Moune et
38
39 913 al., 2022). From June to December 2018 a decrease in $\text{SO}_2/\text{H}_2\text{S}$ and $\text{CO}_2/\text{H}_2\text{S}$ ratios was then observed.
40
41 914 These ratios indicate a pulse of fluid with deep hydrothermal/magmatic origin heated and pressurised
42
43 915 the hydrothermal system (Moretti et al., 2020a).

44 916 This episode period of accelerating unrest is interpreted as a failed phreatic eruption triggered by the
45
46 917 ascent of heat and fluids from the deep hydrothermal and magmatic system to shallow depths
47
48 918 (Moretti et al. 2020). Two hypotheses are formulated by Moretti et al. (2020a): [1] The deep magmatic
49
50 919 fluids, stored at mid- to low-crustal depths in the mush system, generated an upstreaming of
51
52 920 magmatic gases through the mush system and the crystallizing magma reservoir. This process leads
53
54 921 to cyclic rejuvenation of the magma and exsolution of dissolved volatiles that are released into the
55
56 922 hydrothermal system and cause heating and pressurisation. [2] Episodic fresh injections of basaltic
57
58 923 magma from mid-crustal depths into the shallow magma reservoir of La Soufrière de Guadeloupe,
59
60 924 located between a depth of 5.1 and 8 km triggers the pulsatory ascent of high-temperature and high-

925 pressure magmatic fluids into the hydrothermal system. This causes heating and pressurisation that
1 926 may exceed the tensile strength of the rocks and lead to a phreatic eruption.

2
3
4 927 However, for both hypotheses, Moretti et al. (2020a) show that the multiparameter unrest data
5
6 928 recorded prior to and after the M 4.1 earthquake, indicate that the high degree of fracturing and
7
8 929 permeability of the dome allowed fluid pressure to be released laterally, on pre-existing deep tectonic
9
10 930 discontinuities of the upper regions of the Grande-Découverte-La Soufrière volcanic complex (Moretti
11
12 931 et al., 2020a). Moreover, Moretti et al. (2020a) concluded, regardless of which of these hypotheses is
13
14 932 more realistic, the reoccurrence of unrest sequences could escalate to a magmatic eruption. This
15
16 933 scenario would result from the presence of either; a rejuvenated magma at shallow depth (ca. 3 km
17
18 934 below the summit) resulting from stalled small intrusions (Villemant et al., 2014), or injections of fresh
19
20 935 undegassed magma from depth resupplying the shallow magma storage zone of the upper mush
21
22 936 system.

23
24 937 Measurements of temperature and mass flux at fumaroles and thermal springs and ground
25
26 938 temperature measurements shows that between 2010 and 2020 widespread ground heating of the
27
28 939 summit occurred with an increase of the heat output from 0.2 ± 0.1 MW to 5.7 ± 0.9 MW (Jessop et
29
30 940 al., 2021). The total heat flux density has also increased over the 2010 – 2020 period from 326 ± 69
31
32 941 W/m^2 - $403 \pm 24 \text{ W/m}^2$, indicating an increase in thermal intensity at the summit. These changes are
33
34 942 related to the spread of fumarolic sites over the dome and also the propagation of thermal anomalies
35
36 943 on the summit. The observed increases in temperature, mass and heat flux correlate well with the
37
38 944 observed accelerating seismic and degassing unrest of La Soufrière de Guadeloupe over the last
39
40 945 decade. On the basis of this measured heat loss per unit (heat flux density, W/m^2), La Soufrière de
41
42 946 Guadeloupe is one of the most intense heat emitters globally among volcanic domes with an active
43
44 947 hydrothermal system (Jessop et al., 2021).

45
46 948 Looking more closely at the long term unrest at La Soufrière de Guadeloupe, particularly at the rate
47
48 949 of fracture opening and evolution of the permeability of the system through degassing and gas ratios,
49
50 950 could be interpreted as cycles of sealing and reopening of the hydrothermal system (Moune et al.,
51
52 951 2022) For example, in March 2017 during a phase of increased permeability, liberation of native S and
53
54 952 sulphate and resulting increase of H_2S degassing and a decrease in the $\text{CO}_2/\text{H}_2\text{S}$ ratio could indicate
55
56 953 the release of the seal which prevented escalation to a phreatic eruption (Zlotnicki et al., 1992; Moune
57
58 954 et al., 2022). Despite this, the cycle of sealing and reopening of fractures and transfer of heat from a
59
60 955 lower aquifer, to an upper one (with high fluid circulation) raises concerns for a future phreatic
61
62 956 eruption (Zlotnicki et al., 1992; Moune et al., 2022). As the shallow hydrothermal system is hosted in
63
64 957 the hydrothermally altered core of the dome (Heap et al., 2021a; 2021b; 2022a; 2022b), any
65
66 958 pressurisation may reduce friction and increase pore pressure along mechanically weak pre-existing

1 959 structural discontinuities. This may lead to a partial edifice collapse with potentially devastating
2 960 consequences (e.g., Peruzzetto et al., 2019; Heap et al., 2021b). Such edifice collapses could occur
3 961 without the involvement of magma, as observed in the La Soufrière de Guadeloupe stratigraphic
4 962 record (Komorowski et al., 2005; Rosas-Carbajal et al., 2016). These collapses are associated with
5 963 moderate laterally directed explosions (up to VEI2) that can generate turbulent pyroclastic density
6 964 currents of limited extension (< 5 km) as described for the 1797-1798, 1836-1837, and 1976-1977
7 965 phreatic eruptions (Komorowski et al., 2005; Rosas-Carbajal et al., 2016).

12 966 **4.3 Dominica**

15 967 Dominica has nine active Holocene volcanic centres spread across the island which makes it one of
16 968 the highest concentrations of active volcanoes in the world (Lindsay et al. 2005; Mayer et al., 2017).
18 969 These volcanoes have produced the highest erupted magma volume in the LAA (40 km³; Wadge 1984;
20 970 1986; Macdonald et al., 2000; Di Napoli et al. 2014; Mayer et al., 2017). Only one recent eruption is
22 971 recorded on Dominica, but unrest in the form of seismicity and hydrothermal activity has been
24 972 recorded for the nine volcanic centres. The Valley of Desolation-Boiling Lake area is one of the more
26 973 active centres and experiences vigorous hydrothermal activity (Sapper 1903; Robson and Wilmore
28 974 1955; Lindsay et al., 2005; Mayer et al., 2017) and is monitored regularly by UWI-SRC.

30 975 The two most recent eruptions on Dominica have occurred from the Valley of Desolation-Boiling Lake
31 976 area in 1880 and 1997, with neither eruption witnessed (Lindsay et al., 2005; Fig 1). The 1880 eruption
33 977 was a large phreatic eruption with ash covering 55 km², resulting in a slight ash fall 19 km out to sea
35 978 and the generation of lahars (Nicholls et al., 1880b). In comparison, the phreatic eruption occurring
37 979 on 8th – 9th July 1997 was very small magnitude, however, as this eruption was unobserved little is
39 980 known about the dynamics. Subsequent investigation reported non-altered lava, hydrothermally
41 981 altered igneous minerals and hydrothermal minerals present (Komorowski and Hammouya, 1998).
43 982 This eruption may have been triggered by the occurrence of several landslides. Fumarolic vents were
45 983 observed to be covered by landslide deposits, which has been hypothesised to have resulted in
47 984 pressurization of the hydrothermal system and produced a phreatic eruption (James, 1997;
49 985 Komorowski and Hammouya, 1998; Lindsay et al. 2005; Mayer 2017). The Boiling Lake itself
51 986 experiences considerable water level and temperature variations and experienced a small explosion
53 987 of CO₂ and other gases in 1901, killing two people (Ober 1880; James 1988).

55 988 Activity at the other complexes varies greatly, with the majority of seismic and hydrothermal activity
57 989 occurring on the south of the island (Lindsay et al., 2005; Fig 1). The Plat Pays volcanic complex, which
59 990 is mainly comprised of lava domes, experiences a high number of seismic swarms and fumarolic
61 991 activity, this complex was active as recently as 450 years ago. Also, in the south of the island are the

992 Morne Anglais and Morne Watt complexes which have a history of explosive eruptions and experience
1 993 regular seismic swarms (Lindsay et al., 2005; Fig 1). One of the potentially most dangerous complexes
2
3 994 is the Wotten Waven complex, which has a history of large explosive eruptions and produced the
4
5 995 Roseau Tuff. This complex experiences hydrothermal and fumarolic activity and a future explosive
6
7 996 eruption has the potential to affect most of Dominica (Lindsay et al., 2005).
8
9 997 In the north of the island the Morne Diablotins and Morne Aux Diables complexes also experience
10
11 998 considerable periods of unrest (Fig 1). Though no historical eruptions are recorded from these
12
13 999 complexes, and the majority of seismicity occurs in the south of the island, several periods of increased
14
15 1000 seismic activity have been recorded in 1841, 1893, 2000 and 2003 (Lindsay et al., 2005).
16
17 1001 Finally, the Trois Pitons and Grand Soufriere Hills complexes are the least active complexes and have
18
19 1002 no evidence for recent eruptions (Lindsay et al., 2005; Fig 1). However, they are poorly studied and do
20
21 1003 still experience periods of increased seismic activity. The eruptive record is poorly understood,
22
23 1004 particularly at Grand Soufriere Hills, and until new data is available on the eruptive history, the
24
25 1005 complex is considered active (Lindsay et al., 2005).
26
27 1006 All of Dominica's volcanoes are at the green volcanic alert level and continue to be monitored (UWI
28
29 1007 SRC Scientific Updates, 2001-22). A background level of seismicity is recorded with occasional swarms
30
31 1008 of elevated seismicity which, though significant, are considered part of the normal pattern of activity
32
33 1009 (UWI SRC Scientific Updates, 2001-22). Volcanic unrest has in the past been observed in northern
34
35 1010 (Morne aux Diables) and southern Dominica (Plat Pays Volcanic Centre). Swarms occurring in the
36
37 1011 southern complexes, particularly Morne Plat Pays and Morne Anglais, which are the youngest
38
39 1012 volcanoes on the island, are particularly significant as this is considered the most likely site of the next
40
41 1013 eruption (UWI SRC Scientific Updates, 2001-22).

41 1014 **4.4 Montagne Pelée, Martinique**

42
43 1015 The eruptive record of Montagne Pelée shows a history of sub-Plinian to Plinian and dome-forming
44
45 1016 eruptions (Westercamp and Traineau, 1983; Carazzo et al., 2012; Michaud-Dubuy, 2019; Michaud-
46
47 1017 Dubuy et al., 2019; Boudon et al., 2015; Boudon and Balcone-Boissard, 2021). Many of the 8 Plinian
48
49 1018 eruptions of the last 9000 years at Montagne Pelée (e.g., P1 eruption at ca. 1350 CE; P2 eruption at
50
51 1019 ca. 330 CE, P3 eruption at ca. 79 CE) are thought to be associated with 20 – 30 km high columns which
52
53 1020 underwent collapse and produced pyroclastic density currents (Carazzo et al., 2012, 2019, 2020;
54
55 1021 Michaud-Dubuy et al., 2019). At least 5 of the numerous phases of eruptions in the last 2000 years are
56
57 1022 associated with lava dome growth and collapse (Carazzo et al., 2019, 2020; Michaud-Dubuy, 2019;
58
59 1023 Boudon and Balcone-Boissard, 2021) along with violent laterally directed explosions at the base of the
60
61 1024 dome. These explosions produced devastating, highly energetic and mobile pyroclastic density
62
63
64
65

1025 currents with a laterally spreading, dilute flow, as well as valley-channelled denser flows (Lacroix,
1 1026 1904; Perret, 1935; Boudon et al., 2005; Boudon and Balcone-Boissard, 2021) as exemplified by one
2
3 1027 of the world's most devastating eruptions that occurred on 8th May 1902.
4
5 1028 Prior to this eruption, a repose period of three centuries was observed. This was then interrupted by
6
7 1029 two short-lived phreatic eruptions in 1792 and 1851-52 (Tanguy, 1994; Boudon et al., 2005). Following
8
9 1030 these minor eruptions activity, a new period of unrest began in 1889 in the form of renewed fumarolic
10
11 1031 activity reaching a peak in 1902 (Chrétien, 1983; Chrétien and Brousse, 1988; Boudon et al., 2005).
12
13 1032 Seismic activity also increased, inferred from the breaking of submarine cables due to submarine
14
15 1033 landslides and mudflows and the increasing number of earthquakes felt by the population of St Pierre.
16
17 1034 The onset of eruptive activity began with phreatic eruptions on 23rd – 25th April 1902 which increased
18
19 1035 in intensity and transitioned to phreatomagmatic activity, perhaps as early as the 2nd or 3rd May.
20
21 1036 Activity transitioned to magmatic activity on the night of 5th May, and a large glowing dome was seen
22
23 1037 on the night of 6th – 7th May 1902, and continued to grow rapidly (Lacroix, 1904; Tanguy, 1994; Tanguy,
24
25 1038 2004; Boudon et al., 2005; Boudon and Balcone-Boissard, 2021).
26
27 1039 The rapid rise of pressurized hot hydrothermal fluids and groundwater led to the sudden drainage of
28
29 1040 the summit crater lake, causing devastating lahars that killed 23 people on 5th May 1902 in St Pierre.
30
31 1041 Continuing explosive phreatomagmatic activity triggered new lahars during the night of 7 – 8th May
32
33 1042 that destroyed the town of Le Prêcheur killing another 400 people (Tanguy, 1994). Sudden
34
35 1043 decompression of the over-pressurised and rapidly growing lava dome resulted in a small vertical
36
37 1044 column but a large laterally-directed explosion on 8th May 1902 (Lacroix, 1904; Tanguy, 1994; Boudon
38
39 1045 and Lajoie, 1989; Boudon et al., 1990; Boudon et al., 2005; Boudon and Balcone-Boissard, 2021. This
40
41 1046 produced a very mobile pyroclastic density current which expanded rapidly and moved across
42
43 1047 topographic ridges as a high-energy, low-density pyroclastic surge (Boudon and Lajoie, 1989; Boudon
44
45 1048 et al., 1990; Bourdier et al., 1989). More concentrated block-and-ash pyroclastic flows rich in solid
46
47 1049 material from the dome were channelled in numerous valleys down to the coast. These high-velocity
48
49 1050 and highly turbulent surges caused the complete destruction of St Pierre and the death of 28,000
50
51 1051 people on 8th May 1902 (Lacroix, 1904; Tanguy, 1994; Tanguy, 2004; Bourdier et al., 1989; Boudon and
52
53 1052 Lajoie, 1989; Boudon et al., 1990; Boudon et al., 2005; Brown et al., 2017; Gueugneau et al., 2020;
54
55 1053 Boudon and Balcone-Boissard, 2021).
56
57 1054 Following this deadly phase, activity remained high but intermittent with strong vertical eruptions and
58
59 1055 further concentrated, high-energy, turbulent pyroclastic density currents which again affected St
60
61 1056 Pierre on 20th and 26th May, 6th June, 9th July and 30th August 1902. The 30th August flow destroyed
62
63 1057 the towns of Morne Rouge and Ajoupa Bouillon which caused an additional 1500 - 2000 deaths (Brown
64
65 1058 et al., 2017). During this explosive phase, a vertical Vulcanian explosion on 9th July occurred on the

1059 lava dome, producing a small volume pumiceous fallout and an eruption column that collapsed to
1
2 1060 generate a concentrated, channelised pyroclastic density current (Boudon and Lajoie, 1989; Bourdier
3
4 1061 et al., 1989; Boudon et al., 1990; 2005; Boudon and Balcone-Boissard, 2021). Following this initial
5
6 1062 explosive phase, the eruption proceeded until 1905 with effusive dome and lava spine growth
7
8 1063 (Tanguy, 1994; Boudon et al., 2005; Boudon and Balcone-Boissard, 2021). This continued to produce
9
10 1064 numerous block-and-ash pyroclastic density currents from gravitational dome collapse. The currents
11
12 1065 were channelled in valleys to the south and south-west of the volcano without overflowing.

13 1066 This 1902-1905 eruption highlights the long periods of unrest which can be observed preceding an
14
15 1067 eruption, in this case unrest began in 1889 and continued to build to a magmatic eruption in 1902, 13
16
17 1068 years after the onset of phreatic unrest. In comparison, the less destructive eruption of 1929 had only
18
19 1069 a couple of months of felt unrest, yet followed a similar sequence of events as the 1902 eruption. The
20
21 1070 1929 eruption began with phreatic eruptions followed by the main explosive magmatic phase, which
22
23 1071 was followed by dome building activity and then a decrease in activity until the eruption ended in
24
25 1072 1932 (Bourdier et al., 1989). This eruption was less powerful compared to the 1902 eruption, and as
26
27 1073 a result the impact on Martinique was much less severe (Bourdier et al., 1989).

28 1074 As with the current Soufrière Hills Volcano, Montserrat eruption (1995 - ongoing), the 20th century
29
30 1075 eruptions of Montagne Pelée eruption also provides an account of transition in styles with phreatic
31
32 1076 and explosive and effusive magmatic activity observed during one eruption. Finally, the 1902 eruption
33
34 1077 in particular, reminds us of the continuing hazards posed by active LAA volcanoes.

35 1078 Montagne Pelée is currently experiencing a new phase of volcanic unrest which began in late
36
37 1079 December 2018. This was observed by the network which recorded a swarm of 17 deep, distal,
38
39 1080 volcano-tectonic (VT) earthquakes (the hypocentral distance is ≤ 25 km, depth between 10-20 km
40
41 1081 below sea level) as well as a swarm of shallow-depth (down to 4 km depth) volcano-tectonic
42
43 1082 earthquakes located on the volcano (Fontaine et al., 2021; 2022). At the time the distal VT seismicity
44
45 1083 had not been entirely recognized as being of volcanic origin. Therefore the shallow-depth VT seismicity
46
47 1084 was interpreted as evidence for a reactivation of the hydrothermal systems, which is not necessarily
48
49 1085 associated with the arrival at depth of magmatic fluids.

50 1086 In November 2020, the seismic unrest accelerated with respect to the rate and energy of shallow-
51
52 1087 depth, proximal VT seismicity but also, for the first time, of low frequency, volcanic seismicity.
53
54 1088 Together this indicated the onset of an increased circulation of pressurized hydrothermal fluids. This
55
56 1089 unprecedented volcanic unrest prompted the Observatoire Volcanologique et Sismologique de
57
58 1090 Martinique (OVSM-IPGP) to recommend an increase to the yellow volcanic alert level to the
59
60 1091 authorities in charge of civil protection. This was implemented by the authorities on 4th December

1092 2020 (OVSM-IPGP, 2020-2022). In addition, 41 hybrid and 25 long-period seismic signals have been
1 recorded up to the time of writing on 1st September 2022 (OVSM, 2020-2022; Fontaine et al., 2021;
2 1093 2022).
3 1094

4
5 1095 Signs of the current unrest have been detailed in OVSM, 2020-2022 and Fontaine et al., (2021; 2022),
6 1096 these signals are as follows: [1] an ongoing perturbation of the hydrothermal system, perhaps related
7 1097 to the circulation of hot and pressurized fluids of hydrothermal and/or magmatic origin, observed via
8 1098 the seismicity with high and low frequencies content. No felt earthquakes have been reported since
9 1099 the onset of unrest, though the number of VTs is above the base level recorded between 1st January
10 1100 1999 and 1st January 2019; [2] strongly degraded, browned and dead vegetation zones observed in
11 1101 2021 on the southwest flank of the volcano which show a pulsatory increase and decrease in size; [3]
12 1102 a minor deformation from June 2021 to June 2022 is revealed by continuous GNSS data, best modelled
13 1103 as a source of inflation located about 1 km below and slightly SW of the summit, above the
14 1104 hydrothermal system; [4] the detection of an area of underwater gas emission at shallow depths since
15 1105 June 2021.

16 1106 The ongoing unrest features most likely reflect the ascent of a limited volume of deep magmatic fluids
17 1107 which reinvigorated the shallow hydrothermal circulation. Though no surface fumarolic activity is
18 1108 observed, extensive monitoring of this system is still required given the fact that there is no pre-
19 1109 eruptive unrest monitored by a multiparameter network. Although an extended period of 13 years
20 1110 preceded the major eruption of Montagne Pelée in 1902, we cannot expect that this is a characteristic
21 1111 of magmatic eruptions at Montagne Pelée after extensive times of quiescence.

22 1112 **4.5 Soufrière St Vincent**

23 1113 Soufrière St Vincent is the most active subaerial volcano in the Lesser Antilles due to numerous recent
24 1114 eruptions (Table 1). The major historic eruptions of 1440, 1580, 1718 and 1812, and recent eruptions
25 1115 of 1902-03, 1971-72, 1979 and 2020 - 21 are well documented, however, the prehistoric eruptive
26 1116 history is less well known. The recent eruptive history reveals a range of activity from major explosive
27 1117 phases to periods of completely effusive dome growth (e.g., Anderson and Flett, 1903; Aspinall et al.,
28 1118 1973; Shepard et al., 1979; Robertson, 1992; Robertson, 1995; Cole et al, 2019; Fedele et al., 2021;
29 1119 Joseph et al., 2022). Two additional prehistoric eruptions (1580 and 1440 CE) suggest a cyclicity with
30 1120 repose period ranging between 77 and 140 years that appears to be systematically decreasing (Cole
31 1121 et al 2019).

32 1122 The 1902-3 eruption was the deadliest eruption recorded for St Vincent with 1500 deaths (Pyle et al.,
33 1123 2018; Table 1). As with the 1902 Montagne Pelée eruption, unrest prior to the 1902 Soufrière St
34 1124 Vincent began with an increase in seismic activity. However, unlike Montagne Pelée's extended period

1125 of unrest, increases in seismic activity are only reported from 1901, one year prior to the eruption
1126 (though a longer period of unrest, with earthquakes going unfelt by the population, cannot be
1127 discounted (Pyle et al., 2018)). Seismic activity culminated in phreatic eruptions which transitioned
1128 into a violent magmatic eruption producing pyroclastic density currents and lahars (Anderson and
1129 Flett in 1903; Roobol and Smith, 1975; Pyle et al., 2018). The eruption continued with intermittent
1130 vulcanian explosions until March 1903 (Roobol and Smith, 1975; Pyle et al., 2018).

1131 In comparison, a completely effusive eruption was observed in 1971-72. This eruption saw the
1132 aseismic emplacement of a basaltic-andesite dome, a direct contrast to the major explosive eruption
1133 observed in 1902 (Aspinall et al., 1973; Shepard et al., 1979; Robertson, 1992). Explosive activity began
1134 13th April 1979 following elevated seismicity and lasted for 13 days until 26 April. The explosive phase
1135 of the eruption consisted of 11 vulcanian explosions that produced eruptions columns up to 18 km.
1136 Pyroclastic flows affected several valleys surrounding the volcano, but only reached the sea at Larikai
1137 valley. The explosive phase of the eruption was followed by six months of effusive dome growth
1138 (Shepard et al., 1979; Robertson, 1995).

1139 The most recent eruption of Soufrière St Vincent began in December 2020, making this the most
1140 recent period of volcanic activity in the Lesser Antilles (Global Volcanism Program, 2021; Joseph et al.,
1141 2022). This eruption began with an effusive phase on 27th December 2020 with dome growth in the
1142 south-west sector of the crater floor, adjacent to the 1979 dome. Increasing seismicity culminated in
1143 April 2021 with a transition from effusive to explosive behaviour, following three months of dome
1144 growth. On recommendation from the scientists from the Seismic Research Center (UWI Trinidad)
1145 the volcanic alert level was raised by the authorities from orange to red on the afternoon of 8th April
1146 2021. This is the highest level on a four-level scale (UWI-SRC Scientific Updates, 2020-21). The increase
1147 in alert level was followed that day by the onset of rapid evacuations from the red zone (Joseph et al.,
1148 2022). The first explosion occurred on the morning of 9th April and was vulcanian in nature. This was
1149 then followed by a pause of roughly 5 hours after which sub-Plinian activity began, lasting until the
1150 11th of April, when the first pyroclastic density currents appeared. This was followed by a period of
1151 pulsing, explosive activity that continued for the next 20 hours. Between 9th April and 22nd April 2021,
1152 32 identifiable explosions (using seismicity, satellite imagery and visual observations) occurred with
1153 eruption columns reaching up to 16 km. After the initial intense phase of activity, discrete vulcanian
1154 explosions continued with decreasing frequency and energy (UWI SRC scientific updates 2020-21;
1155 Global Volcanism Program, 2021; Joseph et al 2022).

1156 The explosions were associated with ashfall and pyroclastic density currents with lahars occurring later
1157 after heavy rainfall. The explosions produced significant damage to the northern parts of the island
1158 with tephra accumulation leading to collapse of buildings and destruction of crops. In May 2021, the

1159 alert level was reduced to orange, though the volcano continued in a state of unrest with escalation
1160 in activity possible with little warning (UWI SRC scientific updates 2020-21). No casualties occurred
1161 from the eruption but with >16,000 people evacuated from the North of the island, this eruption will
1162 no doubt have socio-economic impacts on the population of St Vincent.

1163 Through June and July 2021 activity continued to decline, thermal anomalies continued to be detected
1164 on the crater floor and were interpreted to be from a small body of magma remaining close to the
1165 floor of the Summit Crater (UWI SRC scientific updates 2020-21). Along with thermal anomalies, daily
1166 seismic activity and substantial degassing continued to occur with persistent steam emissions as the
1167 dominant observable feature, indicating the system was still in unrest. During August 2021 heavy
1168 rainfall resulted in lahars in most of the valleys on the volcano, showing the volcano still posed
1169 significant hazards. In September 2021, the alert level was reduced to yellow given that visual
1170 observations confirmed that no major changes had occurred in the crater since the end of explosive
1171 activity in April. At the end of 2021 and beginning of 2022 seismic activity continued at a low level
1172 with no significant deformation recorded since May 2021. In February 2022 plume chemistry showed
1173 decreasing SO₂ and CO₂ concentrations and by March the levels had become lower than the detection
1174 threshold, as observed in 2020 pre-eruption.

1175 The latest update on the volcano was in March 2022 when seismic activity was reported to have
1176 reached a minimum, no significant change in deformation signal had been observed and no change in
1177 gas composition has been observed in the plume. This activity is considered to be at levels similar to
1178 background activity recorded prior to the 2020 – 2021 eruption and the alert level was lowered to
1179 green.

4.6 Kick ‘em Jenny

1181 The submarine volcano Kick ‘em Jenny has erupted 14 times in the 20th and 21st centuries, making this
1182 the most active volcano in the LAA (Figure 5). The most recent eruption occurred in 2017, following
1183 this eruption the alert level was reduced to yellow (UWI SRC scientific updates 2020). This alert level
1184 has remained at yellow due to periods of seismic unrest being recorded (UWI SRC scientific updates
1185 2020).

1186 Prior to the 2001 eruption, Kick ‘em Jenny eruptions were not associated with precursory seismicity,
1187 as far as is known (UWI SRC scientific updates 2020). The 2001 eruption marked a change in behaviour,
1188 with this and the following 2015 and 2017 eruptions associated with precursory seismicity. Following
1189 the 2017 eruption several periods of high level seismic unrest have been recorded. The most recent
1190 period of unrest occurred from April to June 2020, which included >1000 volcanic earthquakes with a
1191 maximum magnitude 3.1 earthquake recorded (UWI SRC scientific updates 2020).

1192 This period of unrest is interpreted to relate to magma intrusion or pressure changes (due to
1193 crystallisation or degassing) (UWI SRC scientific updates 2020). It is also hypothesised the system may
1194 have experienced sufficient magma recharge to generate eruptions without further high level
1195 seismicity, as observed for the pre-2001 eruptions. However, further periods of seismic unrest without
1196 eruption are also considered likely (UWI SRC scientific updates 2020).

5 Discussion

5.1 How Reliable is the Melt Inclusion Data?

1199 MIs can undergo significant post entrapment modifications (PEM) which can result in changes to the
1200 composition of the MI (e.g., Steele-Macinnis et al., 2011; Wallace et al., 2015; Cannatelli et al., 2016).
1201 In order to ensure the MIs are representative of the original melt care must be taken when choosing
1202 the MIs and also when processing the data.

1203 The majority of studies have avoided or discarded MIs with obvious signs of post entrapment
1204 crystallization (PEC), cracks or bubbles (e.g., Mann et al., 2013; Cooper et al., 2020; Metcalfe et al.,
1205 2022; Supplementary datasheet 2). This ensures the MIs have experienced minimal PEMs and the
1206 composition has not been altered. For studies which report PEC, MIs with >10% PEC have been
1207 discarded (e.g., Camejo-Harry et al., 2019; d'Augustin et al., 2020). This leaves some MIs in the dataset
1208 presented here which have experienced <10% PEC, however, PEC has been shown to have negligible
1209 effects on both major and volatile elements for up to 11% PEC (Moretti et al., 2018). (Supplementary
1210 datasheet 2). Bubbles can also have a major impact on the volatile content, particularly CO₂ with up
1211 to 90% CO₂ sequestered in the bubble (e.g., Moore et al., 2015; Venugopal et al., 2020). In most studies,
1212 MIs with bubbles were not analysed (e.g., Balcone-Boissard et al., 2018; d'Augustin et al., 2020). In
1213 cases where they were analysed, care has been taken to ensure the bubble size and volatile
1214 concentration did not correlate and the volatile contents were comparable to MIs without bubbles
1215 present (Supplementary datasheet 2).

1216 Many studies have also included additional checks on the MIs to ensure the MI has experienced
1217 minimal PEM. For example, studies which report bubbles in the MIs ensure lower H₂O and CO₂ values
1218 are not associated with larger bubbles (e.g., Melekhova et al., 2017). Balcone-Boissard et al. (2018)
1219 ensure their MIs were volatile saturated by comparing CO₂ content of melt inclusions with increasing
1220 crystallization, shown through increasing contents of incompatible trace elements (Wallace et al.,
1221 1999; Blundy and Cashman, 2008). d'Augustin et al. (2020) also show minimum H⁺ diffusion by
1222 comparing H₂O content with Li content. The effect of H⁺ diffusion has also been ensured to be minimal
1223 through analysis of deuterium (e.g., Metcalfe et al., 2022; Bouvier et al., 2010b), with MIs outside the

1224 accepted range of δD values discarded. Finally, ensuring the MIs fall in the same liquid line of descent
1225 can provide assurance that there has been minimal PEM affecting the inclusions.

1226 Corrections and checks to the MIs have been applied by: Toothill et al., 2007, Bouvier et al., 2010 a,b,
1227 Melekhova et al., 2017, Balcone-Boissard et al., 2018, Camejo-Harry et al., 2019, Cooper et al., 2019
1228 d'Augustin et al., 2020, and Metcalfe et al., 2022 (Supplementary dataset 2). However, for the Martel
1229 et al., 1998, Mann et al., 2013 and Cooper et al., 2020 may have a higher error on pressure estimates
1230 (Supplementary dataset 2). Based on the checks performed by the studies included here, and the fact
1231 that MIs have minimal PEC and are not associated with cracks or Fe-Ti oxides we can be confident that
1232 the MI dataset presented here have experienced minimal PEM. This indicates the MIs provide good
1233 preservation of the original major and volatile composition.

1234 **5.2 Controls on Along Arc Variations**

1235 Slab structure controls are also hypothesised to influence the central portion of the LAA. Slab age does
1236 not vary greatly along strike (Cooper et al., 2020; Braszus et al., 2021) and instead: (1) sediment input
1237 (Plank and Langmuir, 1998; Carpentier et al., 2008), (2) subduction of the Tiburon and Baracuda
1238 bathymetric ridges (Bouysse and Westercamp, 1990) and, (3) subduction of strongly hydrated fracture
1239 zones (Schlaphorst et al., 2019; Cooper et al., 2020; Bie et al., 2022) have been hypothesised to control
1240 along arc variations (e.g., Allen et al, 2019). The increased presence of water in the subduction zone
1241 lowers the melting point of the rocks promoting melt genesis. Fracture zones have recently been
1242 proposed to have a primary control on along arc variations in various subduction zones (e.g.,
1243 Venugopal et al. 2020), but also in the LAA, in particular the Marathon and Mercurius fracture zones
1244 which underlay the Northern and Central LAA (Müller et al., 2019; Cooper et al., 2020; Bie et al., 2020;
1245 Fig. 6). Cooper et al. (2020) demonstrated this through $\delta^{11}B$ studies of melt inclusions across the arc
1246 with higher $\delta^{11}B$ values observed in the Central Arc (Fig 6) related to fluids from serpentine
1247 dehydration. In comparison, $\delta^{11}B$ values in the Northern and Southern arc are related to fluids from
1248 altered oceanic crust and sediment dehydration. The higher $\delta^{11}B$ values in the Central Arc coincide
1249 with the location of the Marathon and Mercurius fracture zones subducted beneath the Central Arc.
1250 These fracture zones most likely host the serpentine and supply substantial volumes of fluid to the
1251 Central Arc (Fig 6). This suggests that the along arc compositional variation and magmatic productivity
1252 is controlled by the distribution of subducted serpentine (Cooper et al. 2020). In turn, these variations,
1253 will result in variations in magma storage due to changes in density contrasts and influence the ascent
1254 of the magma, volatiles exsolving will reduce the density further until the neutral buoyancy point is
1255 reached. This is particularly important when considering the Central Arc which has the highest magma

1256 productivity, shallowest storage depths and the largest number of volcanoes in a state of unrest (Fig
1 1257 6).

1258 The distribution of serpentine in the slab is hypothesised to control magma productivity and crustal
1259 thickness (Fig 6). This could also relate to the variation in composition observed along the arc with the
1260 Northern and Central (including St Lucia) arcs erupting generally more evolved compositions and the
1261 Southern Arc producing less evolved compositions and dominated by basalts and basaltic andesites
1262 (Fig 6). The correlation between the eruption of evolved magmas and the thicker crust is hypothesised
1263 to occur due to basaltic magmas becoming more likely to pond as the crust thickens, allowing further
1264 fractionation to more evolved compositions (Baker, 1982; Devine, 1995; Macdonald et al., 2000; Fig
1265 6). This suggests that the homogenous composition produced from a narrow range of shallow storage
1266 depths in the central arc, is due to the thicker crust which is allowing for the ponding and fractionation
1267 of basaltic magmas (Fig 6). Subsequently, magma composition, which is at least partly controlled by
1268 density contrasts, crustal thickness, magma productivity and serpentine distribution, is a first order
1269 control on eruption style. This is observed in the Central arc which produce evolved compositions and
1270 large volume, explosive, caldera eruptions, dome building eruptions and associated voluminous
1271 pyroclastic density current deposits from stratovolcanoes (Macdonald et al., 2000; Lindsay et al.,
1272 2013; Smith et al., 2013; Bezard et al., 2014; Howe et al., 2015a; Lindsay and Robertson, 2018; Fig 6).
1273 In contrast, the thinner crust in the Southern Arc which allows the eruption of more basaltic magmas
1274 erupting in explosion craters and mafic lava flows, with large dome-forming eruptions being less
1275 common (e.g., Arculus, 1976; Brown et al., 1977; Thirlwall and Graham, 1984; Robertson, 2003;
1276 Lindsay and Robertson, 2018; Fedele et al., 2021).

1277 **5.3 Magma Composition, Storage and Controls on Eruption Style**

1278 **5.3.1 Magma Chemistry**

1279 The Grenada magmas are the most primitive and are particularly silica-poor and could represent a
1280 potential parental magma composition which would fractionate to produce the suites of magma
1281 observed across the Arc. The whole rock compositions in equilibrium with a primitive magma are
1282 shown on Figure 3, and plot below the FeO/MgO ratio corresponding to olivine Fo_{90} , using $Kd^{ol-liq} =$
1283 0.3 (e.g., Melekhova et al., 2019). Using the compositions of the primitive Grenada whole rock we
1284 provide two models of fractional crystallisation starting at pressures of 300 MPa (White et al., 2017)
1285 using RhyoliteMELTS v1.2.0 (Fig 3; Ghiorso and Sack, 1995; Gualda et al., 2012).

1286 In the first model (A), the starting composition is the average of the most primitive whole rock
1287 compositions (FeO/MgO <0.7; Fig 3). In the second model (B), we used the average of the most
1288 primitive whole rock compositions with the highest CaO/Al₂O₃ ratio. The same input parameters were

1289 used for both starting compositions which included a starting pressure of 300 MPa, a starting
1290 temperature of 1200°C (liquidus calculated by MELTS) and an oxygen fugacity buffered at NNO
1291 (stamper et al., 2014). The melt was then cooled and decompressed in three iterations due to failed
1292 convergence. The iterations were completed at increasing dP/dT (details in supplementary datasheet
1293 3), determined by the decompression and cooling intervals. This simulated fractional crystallisation
1294 involving phases including olivine, plagioclase, orthopyroxene, clinopyroxene, amphibole, hornblende
1295 and spinel (supplementary datasheet 3). This modelled assemblage is coherent with the observed
1296 assemblage in whole rock samples across the LAA. The fractional crystallisation models predict non-
1297 linear liquid lines of descent which both reproduce the general trend of the LAA whole rock data. This
1298 indicates the LAA magmas are related to one another by fractional crystallisation, and the
1299 differentiation trend observed can be explained by fractionation of a primitive melt with a
1300 composition similar to the most primitive Grenada eruption products (Fig 3). However, there is a large
1301 spread of data, indicating magma mixing and crustal contamination are also occurring to produce the
1302 trends within the individual systems. For example, minor assimilation of crust or sediments is
1303 observed in St Lucia and Martinique magmas (e.g., Bezard et al., 2015) and mixing textures observed
1304 in Guadeloupe (e.g., Boudon et al., 2008).

1305 The magmas from individual volcanoes are often compositionally homogenous across long lasting
1306 eruptions (e.g., Montserrat) and across several eruptions (e.g., Guadeloupe). For the volcanoes which
1307 have been most recently active (i.e., those in the central arc, Montserrat and St Vincent) the eruption
1308 products are described as having restricted bulk compositions and few systematic compositional
1309 changes. Together this indicates the 'active region' of the magma mush system being tapped and
1310 supplying eruptions has remained consistent in composition and also stable. Indeed, in their study of
1311 CO₂-SO₂-H₂S gas emissions from Soufrière Hills in Montserrat, Edmonds et al. (2010) demonstrated
1312 that released gases must result from the mixing of shallow volatile-depleted rhyolitic gases with deep
1313 volatiles supplied by unerupted underplating mafic magma (basaltic andesite). Water content across
1314 the LAA from melt inclusions varies greatly, with large ranges observed at individual systems (Fig 6;
1315 Cooper et al., 2020). Lower water contents are generally observed in melt inclusions from the central
1316 arc, with higher water contents observed in the northern and southern arcs (Fig 6). However, the fact
1317 that melt inclusion water contents can often be affected by processes including: magma
1318 differentiation, degassing and post-entrapment crystallization or diffusion, suggests that the observed
1319 trend may be the result of additional processes following magma differentiation. Therefore, the
1320 highest water content recorded at a given volcano is considered the best estimate of the water
1321 content (Rasmussen et al., 2022).

1322 **5.3.2 Magma Storage**

1323 At many systems along the arc the mush model is now the accepted storage model, with a transcrustal
1324 mush system often described. Looking closer at the individual systems, we can hypothesise that across
1325 the arc there are generally three levels of the mush storage system which varies in depth across the
1326 northern, central and southern arcs. Storage depths are often considered to be the depths at which
1327 the magma is at neutral buoyancy (Sparks et al., 2019), and due to the rheological properties of the
1328 magma (e.g., composition, crystal content, volatile abundance, gas/liquid ratios and temperature)
1329 which drive magma emplacement at different depths (e.g., Burov et al., 2003). However, more
1330 recently magma H₂O concentration has been hypothesised as a major control on the storage depth
1331 (Rasmussen et al., 2022). Storage depths have in some cases been shown to be consistent with the
1332 predicted depths of water degassing and related processes (Rasmussen et al., 2022). The variations in
1333 storage depths in the LAA are not reflected in magma storage temperature variations (Fig 6), with
1334 temperature remaining consistent across the arc. This may be due to the different methods and
1335 techniques used across the arc to calculate temperature. Bias towards measurement of one phase
1336 (e.g., cpx+liquid) and crystal rims + matrix glass, may exacerbate this and contribute to the
1337 homogenous temperature profile (Higgins et al., 2022). Melekhova et al. (2019) use an inversion
1338 approach which combines xenolith petrology and seismic receiver functions to present crustal
1339 structure along the LAA. This model shows four crustal layers which vary laterally (Figure 2). The
1340 magma system depths presented in the literature generally fit within layers 2 (6 - 15 km) and 3 (12 –
1341 32 km), except for St Lucia where the lower crust system fits into layer 4 (26 – 38 km).

1342 **5.3.2.a Lower Crust (ca. >18 km, ca. >400 MPa)**

1343 Melekhova et al. (2017) infer a depth of 25 km from primitive cumulate composition of ol+cpx+sp
1344 from St Kitts which are in equilibrium with the most primitive magmas. This is in agreement with the
1345 estimates for cumulates from other islands (>15 - >30 km; e.g., Bissante, 1999; Parkinson and Arculus,
1346 1999; Stamper et al., 2014; Fig 6). Fractionation occurs to produce magmas which ascend into the
1347 middle crustal storage zone. However, as cumulates from this lower storage level are rarely observed
1348 it is unclear whether this is applicable across the arc. Despite this, this does provide a good estimation
1349 for the deepest part of the LAA system.

1350 **5.3.2.b Mid Crust (ca. 10 – 18 km, ca. 200 - 400 MPa)**

1351 Above the lower crustal zone is a zone of mafic-intermediate magma which, depending on the system,
1352 is generally described as basalt to basaltic-andesite in composition and associated cumulates (Tollan
1353 et al., 2012; Stamper et al., 2014; Melekhova et al., 2019). The magmas stored here are often
1354 described as the parental magmas for the more evolved magmas stored in the shallow system. Again,
1355 this mid crustal part of the system is less well studied than shallower parts but is commonly reported

1356 between 10 – 18 km (Pichavant et al., 2002; Edmonds et al., 2016; Boudon and Balcone-Boissard 2021;
1357 e.g., Melekhova et al., 2017; Balcone-Boissard et al., 2018; Fedele et al., 2021; Fig 6). Within the
1358 transcrustal mush system, the parental magmas can migrate into the shallow part of the system and
1359 undergo processes including fractional crystallisation and magma mixing. For example, during the
1360 most recent eruption of Soufriere St Vincent seismicity indicates the initial activity during the explosive
1361 phase (23 - 24 March 2021) involved the shallow system (~6 km). This was then followed by deeper
1362 seismicity (~10 km depth; 5-6 April 2021) interpreted as fresh magma migrating from the mid system
1363 up into the shallow system (Joseph et al., 2022).

1364 **5.3.2.c Upper Crust (ca. <10 km, ca. <200 MPa)**

1365 The most well-defined parts of the magma systems across the arc are the shallow zones. The shallow
1366 area of the system contains evolved magmas which are often described as homogenous (e.g.,
1367 Dominica, St Vincent, Guadeloupe; e.g., Fedele et al., 2021; Metcalfe et al., 2022). Magmas from this
1368 system can display homogeneity from one eruption to the next, even with considerable repose
1369 periods (e.g., Martinique, Guadeloupe). Evolved magmas are reported to extend to the mid-crust in
1370 the Northern Arc (down to 15 km), but is consistent in the Central Arc, between 6 – 9 km (e.g.,
1371 Pichavant et al., 2002; d’Augustin et al., 2020; Fedele et al., 2021; Metcalfe et al., 2022; Fig 6). Less
1372 data exists for the Southern Arc, but in St Vincent there is some evidence that the shallow system may
1373 extend up to <3.5 km (Fedele et al., 2021). However, we do acknowledge that more data is required
1374 in order to conclusively observe this trend. The upper crust system (<10 km) presented here fits into
1375 the layer 2 (6 – 15 km) proposed by Melekhova et al. (2019). This is particularly well defined in
1376 Guadeloupe, Martinique and St Vincent (Fig 2). Layer 3 (12 – 32 km) proposed by Melekhova et al.
1377 (2019) spans a wide range of depths and encompasses both the mid-crust (10 – 18 km) and lower
1378 crust (>18 km) systems presented here. This means, for example, in Guadeloupe the lower crust
1379 system fits into the Melekhova et al. (2019) layer 3 (fig 2). This highlights that the magma storage
1380 structure which is defined by magma chemistry and petrology provides a snapshot of the crustal
1381 structure along the arc.

1382 **5.3.3 Processes Occurring in the Storage Zone**

1383 The shallow systems are hypothesised to contain an ‘active region’ with eruptible magma which
1384 supplies recent eruptions (e.g., Montserrat; Christopher et al., 2015; Edmonds et al., 2016). The active
1385 region of mush within the shallow system is generally hypothesised to be remobilised by recharge
1386 from the middle basalt to basaltic andesite region (12 – 18 km). Evidence for this is observed in the
1387 textures exhibited in lavas and pumices (e.g., banded pumices from Guadeloupe and enclaves from
1388 Montserrat; e.g., Boudon et al., 2008; Wadge et al., 2014), which are present in recent eruption

1389 products showing this is an active process in triggering eruptions. However, it is important to note
1
2 1390 mixing textures are generally not consistent through a system's eruptive history (e.g., Guadeloupe;
3 1391 Metcalfe et al., 2022). This suggests that though recharge by the parental magma from the middle
4
5 1392 region may occur it does not always result in the eruption of mixed magma compositions. Instead, this
6
7 1393 may result in the remobilisation of the shallow system and eruption of homogenous magmas and
8
9 1394 dominantly more evolved liquids in the form of melt inclusions. Also suggested as a hypothesis for
10
11 1395 remobilisation of the shallow system, is mixing of magmas with a similar composition (Solaro et al.,
12 1396 2019). Or alternatively, through volatiles and heat released from the parental magma migrating up
13
14 1397 through the system and resulting in remobilisation in the shallower system, with no need for direct
15
16 1398 mixing (Christopher et al., 2010; Metcalfe et al., 2022).

17
18 1399 However, we cannot discount the effect of the well-developed hydrothermal systems present. The
19
20 1400 hydrothermal systems present at the active volcanoes, including Montserrat, Guadeloupe, Dominica,
21
22 1401 Martinique and St Vincent. This often complicates the interpretation of monitoring signals due to the
23
24 1402 scrubbing of magmatic gases, in particular SO₂, which often buffers and suppresses the magmatic
25
26 1403 signal (Moune et al., 2022). Recent research highlights the potential importance of the hydrothermal
27 1404 systems in generating eruptions (Zlotnicki et al., 1992; Mayer et al., 2017; Moune et al., 2022).

28
29 1405 Phreatic eruptions in particular, which can be generated without any change in the magma recharge
30
31 1406 zone at depth, can be generated due to the sealing of fractures in the hydrothermal system which
32
33 1407 increases the pressure in deep aquifers to potentially generate an eruption (e.g., Moune et al., 2022).
34
35 1408 This has been discussed as a potential cause of the 1976-77 La Soufrière phreatic eruption (Zlotnicki
36
37 1409 et al., 1992) and given the build-up of hydrothermal pressure observed in 2018, it may have led to a
38 1410 phreatic eruption if the fractured state, hence the high permeability of the dome had not allowed for
39
40 1411 release of the pressure (Moretti et al., 2020a; Moune et al., 2022). The effect of the upper system has
41
42 1412 also been discussed in the context of magmatic eruptions from La Soufriere Guadeloupe, where the
43
44 1413 eruption style of magma ascending from the active region of the storage zone is affected by the state
45
46 1414 of the upper system. Though ultimately the main driver of an explosive eruption is the supersaturation
47 1415 of volatiles in the magma and brittle failure of a magma. Metcalfe et al. (2022) hypothesise that
48
49 1416 eruption style at La Soufriere de Guadeloupe is controlled by magma ascent rate, with faster magma
50
51 1417 ascents resulting in larger, more explosive eruptions. Along with magma composition and ascent rate,
52
53 1418 the state of the hydrothermal system can also influence eruption style and could be responsible for
54 1419 the range of eruption styles produced by each individual system. A slow ascending magma
55
56 1420 experiencing feedbacks between degassing and crystallization will be expected to erupt with low
57
58 1421 explosivity or effusively, however, sealing of the hydrothermal system and a reduction in permeability

1422 may favour the accumulation of gas, leading to over pressurisation and an explosive eruption (e.g.,
1423 Cassidy et al., 2018; Metcalfe et al., 2022).

1424 In order to provide an appraisal of the prevalent degassing process (decompression degassing vs CO₂-
1425 infiltration), integrated into a fast and comprehensive visualization of the volcano plumbing systems
1426 from their source to the surface, we provide pressure vs CO₂ (gas) diagrams (Fig 7). This has been used
1427 in literature to: [1] identify degassing processes such as decompression degassing and CO₂-fluxing
1428 (Mangiacapra et al., 2008; Barsanti et al., 2009; Arienzo et al., 2010; Moretti et al., 2013, 2019; Caricchi
1429 et al., 2018; Papale et al., 2022), and [2] provide a straightforward petrologic/geochemical image of
1430 volcano plumbing systems which integrates geophysical information and highlights depth intervals of
1431 magma storage. The accuracy of this representation is highly dependent on the performance of the
1432 adopted model for volatile melt-saturation, in particular, the model's sensitivity to compositional
1433 variations from deep to shallow regions (Barsanti et al., 2009; Papale et al., 2022). Among
1434 compositional variations, the redox-induced partitioning of iron between FeO and Fe₂O₃ plays a major
1435 role in shifting the equilibrium pressure of H₂O-CO₂ saturation by a factor comparable, or even
1436 superior to, that of analytical errors on dissolved H₂O and especially CO₂ contents measured in melt
1437 inclusions (Moretti, 2022).

1438 For given fO_2 values we calculated the iron oxidation state of each MI by means of the Ottonello et al.
1439 (2001) model extended to $P > 1$ bar and hydrous conditions, as in Moretti (2005). We first computed
1440 $Fe^{3+}/Fe_{(tot)}$ of MIs, and therefore FeO and Fe₂O₃ contents, by using guess pressure values. Saturation
1441 pressures ($P_{(fluid)}$) from the Papale et al. (2006) model were then used to update $Fe^{3+}/Fe_{(tot)}$ values via
1442 the redox model of Moretti (2005) and the new FeO and Fe₂O₃ values were entered into the Papale et
1443 al. (2006) computation for pressure. Two to three iterations were sufficient to obtain consistent sets
1444 of fO_2 - $Fe^{3+}/Fe_{(tot)}$ - $P_{(fluid)}$ values for the two models combined and solve the mutual effects played by
1445 FeO-Fe₂O₃ partitioning on equilibrium fluid pressures at H₂O-CO₂ saturation and the role of dissolved
1446 water on iron oxidation state (Moretti and Papale, 2004; Papale et al., 2006; Moretti, 2022, Moretti
1447 and Ottonello, 2022; Papale et al., 2022 for detailed discussions and examples on different melt
1448 compositions including basalt and andesite. When we begin to consider data from across the arc, the
1449 P vs CO₂ data (Fig 7) shows the processes of decompression and fluxing are occurring above the Moho,
1450 and between ca. 200 and 100 MPa (ca. 4 - 8 km). The data is spread over the whole H₂O-CO₂ range,
1451 giving us an estimate of the main levels of emplacement in the crust. Further spread in the gas
1452 composition may occur at this crustal level (ca. 200 -100 MPa, ca. 4 - 8 km) due to magma
1453 differentiation and magma mixing. There are two main scenarios: (1) the first scenario is a 'static'
1454 scenario, where the data is spread over a large XCO₂(gas) domain for a pressure interval. This is
1455 indicative of magma ponding at a given depth interval where H₂O and CO₂ fractionate due to fractional

1456 crystallization or CO₂-driven dehydration processes. (2) the second scenario is a ‘dynamic’ scenario
1457 where we identify paths of magma ascent with either XCO₂ increase (CO₂-fluxing) or XCO₂ decrease
1458 (decompression degassing).

1459 The P vs CO₂ data indicates in the Northern Arc, St Kitts and Montserrat MI CO₂ decreases with
1460 decompression of the magma across pressures of ca. 240 – 100 MPa (Fig 7A). However, MIs from
1461 Statia are experiencing CO₂ fluxing, shown by the CO₂ increasing with decompression across the same
1462 pressure range of ca. 240 – 100 MPa (Fig 7A). There are also a small cluster of St Kitts MIs at a higher
1463 pressure (>300 MPa) which are affected by CO₂ fluxing (Fig 7A). In the Central arc decreasing with
1464 decompression is affecting the CO₂ content in the majority of the MIs, this generally occurs across
1465 pressures of ca. <200 MPa (Fig 7B). Martinique and St Lucia MIs experience decompression over a
1466 narrower range (ca. <130 MPa) (Fig 7B). The Dominica MIs form a bimodal distribution, the data from
1467 Balcone-Boissard et al. (2018) (Fig 7B), has a very hydrous source and experiences CO₂ fluxing over
1468 pressures from 375 – 150 MPa. In comparison, the data from d'Augustin et al. (2020) is comparatively
1469 CO₂-rich, with CO₂ decreasing with decompression. The majority of these MIs from d'Augustin et al.
1470 (2020) experience decompression at <200 MPa, however, there are some MIs which are deeper at
1471 >500 MPa. In the Southern Arc, decompression and CO₂ fluxing are affecting MIs, generally these
1472 processes occur at < 200 MPa, however, Grenada has a deeper source down to ca. 400 MPa (Fig 7C).
1473 Though there are only a few MIs data points for St Vincent, MI CO₂ is decreasing with decompression
1474 across a narrow pressure range (150 – 180 MPa; Fig 7C), however, with more data this range may be
1475 extended. The Kick ‘em Jenny MIs show a wide spread of data, which also shows CO₂ decreasing with
1476 decompression. There are also two high pressure (ca.600 MPa), CO₂-rich data points which are outliers
1477 from the distribution (Fig 7C). Grenada has a very hydrous source and experiences CO₂ fluxing over a
1478 wide range of pressures (60 – 400 MPa; Fig 7C).

1479 **5.4 Perspectives and Limitations**

1480 Despite the wealth of knowledge and understanding of the LAA there is still much to be learned about
1481 each magmatic system and about the LAA as a whole. In particular, we need to improve our
1482 understanding of the geometry of the deeper parts of the LAA magma storage systems to provide a
1483 more complete picture of this system. Understanding the variations in pressure, temperature and
1484 depth of the deep magma storage system may help provide links with the variations in mineralogy
1485 and chemical composition observed through the system. This could be done, for example, through
1486 seismic imaging across larger areas and over longer seismic lines which would extend the data
1487 collected by the VOILA project (refs). In addition, deep penetration multichannel seismic (e.g., Laigle
1488 et al., 2013), gravity surveys and magnetotellurics imaging at sea using ocean bottom stations (e.g.,
1489 Darnet et al., 2020) may help to resolve and provide additional information on the structure of the

1490 magmatic reservoir. This should include understanding the lateral extent of the storage systems and
1
2 1491 ‘active regions’ of mush, with particular focus on Dominica which appears to have a much more
3
4 1492 laterally extensive system in comparison to the other islands (Balcone-Boissard et al., 2018; d’Augustin
5
6 1493 et al., 2020).

7
8 1494 To understand the role of the parental magmas in the different systems and how this varies along the
9
10 1495 arc could include refining our understanding of how the parental magma remobilises the shallow
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12 1496 system. Though remobilisation by a parental magma has been proven in many cases, there are many
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14 1497 eruption products, which show no evidence for mixing a less evolved magma demonstrating the
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16 1498 parental magma is not always erupted. This creates a bias in the geochemistry dataset as the evolved
17
18 1499 magmas from the shallow system are more often erupted, with fewer examples of the parental and
19
20 1500 primitive magmas. This along with the homogenous compositions observed in magmas erupted from
21
22 1501 the shallow system requires detailed investigation of eruption deposits to better understand if fine-
23
24 1502 scale mixing textures are present. This will allow us to better understand how the parental magma
25
26 1503 interacts with the shallow system and why it is not consistently erupted.

27
28 1504 Understanding the volatile processes occurring, and in particular how degassing from various depths
29
30 1505 across the vertically extended magma system contributes to the gas ratios measured at the surface
31
32 1506 (e.g., Metcalfe et al., 2023,; Moretti et al., 2023). This requires an integration of melt inclusion volatile
33
34 1507 data and gas monitoring data which is available for many systems which have long-term monitoring.
35
36 1508 In addition, understanding the presence and accumulation of exsolved fluids in the system as the
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38 1509 abundance of exsolved volatiles can affect the evolution and growth of the magma reservoir (e.g.,
39
40 1510 Degruyter et al., 2017; Rasmussen et al., 2022). Understanding the relationship between the presence
41
42 1511 of exsolved volatiles to the depth and growth of the magmatic systems in the LAA, may be key to
43
44 1512 understanding along arc variations in the magmatic systems.

45
46 1513 The majority of eruptive activity in the 20th century has occurred in the Central and Southern arcs,
47
48 1514 with the only eruption in the northern arc being in Montserrat. The majority of eruptions in this period
49
50 1515 have been vulcanian eruptions, dome eruptions and phreatic eruptions (Fig 5) and it would be
51
52 1516 reasonable to assume any future activity will also take this form, with less frequent sub-Plinian to
53
54 1517 Plinian size eruptions interspersed. However, it is also important to consider the large pumiceous
55
56 1518 pyroclastic flow deposits of Guadeloupe, Dominica and St Lucia. The Roseau and Layou tuffs in
57
58 1519 Dominica (Howe et al., 2014), the Belfond Unit on St Lucia (Lindsay et al., 2013) and large Plinian
59
60 1520 eruptions from Grande-Découverte and Carmichaël volcanoes on Guadeloupe (Komorowski et al.,
61
62 1521 2005; Legendre, 2012) and though much less frequent in the eruptive record, these large Plinian
63
64 1522 eruptions are important to consider, as such eruptions would impact the entire region (Lindsay and
65
60 1523 Robertson, 2018)

1524 Volcano monitoring has shown major advancements over the last fifty years, however, the period of
1 1525 monitorable unrest can vary greatly even within the same system. In the case of St Vincent, the two
2 1526 most recent explosive eruptions have had very short periods of monitorable unrest (on the order of
3 1527 days for the 1979 eruption and 2 months for the 2020 eruption (onset in November 2020; Joseph et
4 1528 al., 2022; UWI SRC scientific updates 2020-21). In the case of the St Vincent 2020 eruption, the 2
5 1529 months of unrest was followed by an effusive phase of eruption with evacuation not taking place until
6 1530 the explosive phase of eruption had begun (Joseph et al., 2022). Other systems such as La Soufrière
7 1531 de Guadeloupe could also potentially have short periods of monitorable unrest prior to eruption, as
8 1532 shown by diffusion timescales (e.g., Metcalfe et al., 2021 a,b). Diffusion timescales allow the
9 1533 calculation of the time between the last magmatic process that destabilizes the magma in the storage
10 1534 zone and eruption, for La Soufrière de Guadeloupe, this is calculated to be on the order of weeks for
11 1535 even large explosive eruptions at this system (Metcalfe et al., 2021a,b). The short time frames of
12 1536 escalation from monitorable unrest to eruption and then an explosive phase requires efficient and
13 1537 effective monitoring plans, to allow for timely and successful crisis management and evacuation, as
14 1538 was demonstrated by the remarkable crisis response to the Soufrière of St Vincent eruption in 2020
15 1539 (Joseph et al., 2022). Much can be learnt from the handling of the 2020 Soufrière St Vincent eruption
16 1540 and the support that has been provided for the evacuated people. However, though people begun to
17 1541 return to their homes a few months after the eruption, some people remain displaced and support
18 1542 had to be continued over years after the eruption in order to facilitate social and economic recovery
19 1543 (Joseph et al., 2022). This is highlighted by the 1995 Soufrière Hills eruption on Montserrat, where
20 1544 following the evacuation of >6,000 people and a population reduction from ca. 11,000 pre-eruption
21 1545 to ca. 4,000 post-eruption, only 1000 people have returned (Kokelaar, 2002; Aspinall et al., 2009;
22 1546 Barclay et al., 2019).

41 1547 As previously discussed, a wide variation in eruption styles is observed across the arc as exemplified
42 1548 by the eruptions in the 20th century (Fig 5). The hazards posed by eruptions need to be clearly
43 1549 conveyed to the populations with volcanoes in a state of unrest, particularly on islands where volcanic
44 1550 risk is not necessarily considered high by the population (e.g., Guadeloupe; Chenet et al., 2014).
45 1551 Though the eruptions occurring in the early 20th century (e.g., Montagne Pelée 1902 and Soufrière St
46 1552 Vincent 1902) resulted in a large number of casualties (>30,000 deaths), advancements in science and
47 1553 technology have allowed a better understanding and monitoring of the volcanoes of the LAA. This has
48 1554 undoubtedly prevented large numbers of casualties in eruptions in the later 20th century, however, in
49 1555 most cases, evacuations of people remain unavoidable (Table 1). In some cases, evacuation becomes
50 1556 long term displacement and results in major socio-economic impacts and in the case of the La
51 1557 Soufrière de Guadeloupe 1976 eruption, a loss of trust in scientists and authorities (Hincks et al., 2014;

1558 Chenet et al., 2014; Komorowski et al., 2018). This highlights the need for future crises to be managed
1559 with care and transparency to limit the socio-economic impacts and maintain the relationship
1560 between scientists, authorities and the population.

1561 **6 Conclusions**

1562 In this review of the LAA we have focused on the magma storage system, geochemistry and recent
1563 eruptive history. We have highlighted investigating trends across the arc can be useful to provide more
1564 context for each system. The distribution of subducted serpentine has a first order control on
1565 magmatic productivity (Fig 6), compositional variation and variation in magma storage depth as
1566 serpentine can supply substantial volumes of fluid to the arc (Cooper et al., 2020). This is particularly
1567 important to consider in the Central Arc where evolved magmas are stored at shallow depths and are
1568 likely to erupt explosively in the near future, given the unrest observed at many systems in LAA.

1569 Through this review we show that there appears to be three different levels of magma storage within
1570 a vertical and laterally extensive, transcrustal magma mush. The deepest storage area is the least well
1571 defined but is most likely at >18 km depth. Above this, at ca. 10 – 18 km are the parental magmas
1572 which undergo fractional crystallization to produce more evolved magmas which are stored in the
1573 shallow system at ca. <10 km. Within the shallow mush system there is an ‘active region’ which has
1574 supplied the most recent eruptions, and is most likely remobilised by magma, heat and volatiles from
1575 the parental magmas at 10 – 18 km. Eruption products from recent eruptions are often described as
1576 having restricted bulk compositions with few systematic compositional changes. This indicates the
1577 ‘active region’ of the magma systems supplying eruptions can remain homogenous in composition
1578 and also stable over long periods of time, perhaps by buffering recharge from depth.

1579 Relating the sub-surface processes occurring leading up to an eruption, and the monitorable surface
1580 observables remains a challenge. Adapting current monitoring strategies to detect, in a timely
1581 manner, the entire eruption timeline from the point at which sub-surface processes effectively
1582 generate recordable signals on the surface would greatly enhance monitoring and crisis management
1583 strategies. However, many uncertainties must be addressed within this including understanding the
1584 surface observables in terms of the sub-surface processes, timescales, and dynamics. In addition,
1585 understanding the modulation, abortion and escalation of sub-surface processes and the probability
1586 it can lead to an eruption are also important and complicated factors to consider for unrest thresholds,
1587 alert levels and early-warning systems.

1588 Each island in the LAA has an active volcano with eruptions occurring in the Holocene, and 24
1589 eruptions since 1900. Recent eruptions demonstrate the volcanoes of the LAA can result in thousands
1590 of deaths, large scale and long-term displacements and severe economic impacts. Though

1591 advancements in science, technology and our understanding of LAA volcanoes has undoubtedly
1592 prevented large numbers of casualties in the late 20th and 21st centuries, large evacuations of people
1593 remain unavoidable. Long term displacements of people require continued support to facilitate social
1594 and economic recovery. The majority of recent eruptive activity has occurred in the Central and
1595 Southern arcs, and though the northern arc should not be neglected, research focus should be based
1596 on where the next eruption is most likely to occur. This should include understanding the lateral extent
1597 of the storage systems and ‘active regions’ of mush, and as mentioned previously, with particular focus
1598 on Dominica which appears to have a much more laterally extensive system in comparison to the
1599 other islands. This also provides an area where monitoring efforts should be concentrated, in
1600 particular renewed unrest in Martinique and ongoing unrest in Guadeloupe should continue to be
1601 monitored carefully. This should include further studies of past eruptions to provide further
1602 constraints on the magma storage and behaviour and also careful monitoring of the hydrothermal
1603 systems. Overall, further research on the LAA must continue to be underpinned by the hazardous
1604 nature of the system and the tens of thousands of people that can be affected by even a small
1605 eruption. Continuing our advancement in understanding of the LAA is essential to continue to improve
1606 monitoring techniques, reduce the impact of eruption and provide clear crisis management.

1607 **Data Availability**

1608 Our compiled whole rock and melt inclusion data set are available (supplementary material dataset
1609 1), along with the MELTS models (supplementary dataset 3).

1610 **Author Contribution Statement**

1611 AM collated the data and drafted the manuscript and figures. J-CK and SM supervised AM during the
1612 PhD project and fieldwork this paper stemmed from. RM conceptualized the Pressure-CO₂,gas
1613 diagram, performed calculations and with AM discussed and interpreted the results. All authors
1614 discussed the data, wrote and revised the manuscript and the figures.

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1635 **ClerVolc contribution number XX.**

1636 **Figure Captions**

1637 **Figure 1** Schematic map of the Lesser Antilles (adapted from Wadge 1986 and Leone et al., 2019). The
1638 Lesser Antilles arc bifurcates at Martinique with the outer Eo-Oligocene arc making up the Limestone
1639 Caraïbes and the inner Pliocene and Pleistocene making up the current active arc. The major volcanic
1640 centres which are <2 Ma, are split into three areas of the arc: Northern, Central and Southern (Wadge
1641 and Shepherd, 1984) including the submarine volcano Kick ‘em Jenny. Also shown are the active
1642 volcanoes which have had eruptions since 1270 CE (red triangles) and eruptions in the Holocene
1643 (orange triangles). These twenty-one Holocene active volcanoes have the potential to impact tens of
1644 thousands of people living in the Lesser Antilles. The photographs show the variation in morphology:
1645 St Kitts – the northern flank of Mt Lliamuiga; Montserrat – Soufriere Hills Volcano morphology
1646 following eruption, with collapse features and lahar deposits; Guadeloupe – the La Soufriere de
1647 Guadeloupe lava dome emplaced in 1530 CE (Boudon et al., 2008) and surrounding scoria cones (right
1648 of the image); Dominica – view of the lava domes within the Plat Plays Complex; Martinique – the
1649 edifice of Mont Pelee; St Lucia – the Pitons which make up the Soufriere Volcanic Centre; St Vincent
1650 Soufriere St Vincent prior to the recent eruption, visible is the crater structure in which the 2020 dome
1651 grew. Images from: Guadeloupe and Dominica (Metcalf, 2020); St Kitts, St Lucia and St Vincent (UWI-
1652 SRC, 2015 – 2022); Martinique (Michaud-Dubuy, 2017), Montserrat (Fournier, 2005; Hickson et al.
1653 2013).

1654 **Figure 2** Schematic summary diagram of magma storage across the arc adapted from Melekohva et
1655 al. (2019). Storage models from along the arc are superimposed over the crustal structure (see key).
1656 Laterally and vertically extensive mush systems are now proposed for many systems across the arc,
1657 which can be divided broadly into three levels: (1) >18 km; (2) 10 – 18 km; (3) <10 km. The location of

1658 the mush systems, generally fit with the location of crustal layers (e.g., Guadeloupe, Martinique, St
1659 Vincent). The majority of research has focused on the shallow and middle systems. Many systems are
1660 hypothesised to have an active region of mush which has supplied recent eruptions in the shallow
1661 area of the system (<10 km). The shallow system is remobilised by less evolved magmas from >10 km.
1662 This highlights the need to refine storage models, particularly at depths greater than 10 km. Magmas
1663 may be produced from parental magmas fractionating at >18 km, as hypothesised for St Kitts,
1664 Guadeloupe and St Lucia. (Bissante, 1999 ; Melekhova et al., 2015 ; Melekhova et al., 2017 ; Paulatto
1665 et al., 2012 ; Christopher et al., 2015 ; Edmonds et al., 2016 ; Pichavant et al., 2018 ; Metcalfe et al.,
1666 2022 ; Balcone-Boissard et al., 2018 ; d'Augustin et al., 2020 ; Martel et al., 1998 ; Fedele et al., 2021 ;
1667 Camejo-Harry et al., 2019 ; Boudon and Balcone-Boissard, 2021).

Figure 3 Plot of $\text{CaO}/\text{Al}_2\text{O}_3$ vs $\text{FeO}_{(\text{total})}/\text{MgO}$ using whole rock data compilation from across the arc. As
1668 well as data from the northern (circle symbols), central (square symbols) and southern (diamond
1669 symbol) arcs, we present two MELTS models of fractional crystallisation (see discussion). This shows
1670 that the whole rock data is related through fractional crystallisation, with the variability observed
1671 related to mixing. (Saba (Defant et al., 2001; Roobol and Smith, 2004); Statia (Roobol and Smith, 2004);
1672 St Kitts (Baker, 1968, 1980, 1984; Toothill et al., 2007; Higgins et al., 2021); Montserrat (Murphy et al.,
1673 2000; Zellmer et al., 2003; Christopher et al., 2015; Plail et al., 2018); Guadeloupe (including La
1674 Soufrière and Mont Caraïbes, Toubol et al., 2007; Boudon et al., 2008; Samper et al., 2009; Ricci et al.,
1675 2017; Pichavant et al., 2018; Metcalfe et al., 2021, 2022); Dominica (Lindsay et al., 2005; Howe et al.,
1676 2015; Boudon et al., 2017; Balcone-Boissard et al., 2018; d'Augustin et al., 2020); Martinique
1677 (Gourgand et al., 1989; Pichavant et al., 2002); St Lucia (Wright et al., 1984; Vidal et al., 1991; Lindsay
1678 et al., 2013; Bezaud et al., 2015); St Vincent (Graham and Thirwall, 1981; Heath et al., 1988; Fedele et
1679 al., 2021; Cole et al., 2019; Joseph et al., 2022); Kick 'em Jenny (Devine and Sigurdsson, 1996; Camejo-
1680 Harry et al., 2019); Grenada (Arculus et al., 1976; Arculus, 1978; Cawthorn et al., 1973; Thirwall and
1681 Graham, 1984; Thirwall et al., 1994, 1996; Woodland et al., 2002; Rojas-Agramonte et al., 2017).

Figure 4 Melt inclusion data from across the Lesser Antilles **(A)** SiO_2 ; **(B)** $\text{Na}_2\text{O} + \text{K}_2\text{O}$; **(C)** Cl; **(D)** H_2O ; **(E)**
1683 CO_2 vs volcano latitude. The most SiO_2 rich magmas are observed in the Central Arc, while the least
1684 evolved magmas are found in the Southern Arc. Though the most evolved magmas are in the Central
1685 Arc, the most alkali rich magmas are found on Montserrat in the Northern Arc. Volatile content varies
1686 greatly across the arc, particularly H_2O , however, magmas from the Central Arc notably more CO_2 rich
1687 than the Northern and Southern Arcs. For full melt inclusion dataset see supplementary dataset 1.
1688 (Saba (Roobol and Smith, 2004); Statia (Cooper et al., 2020); St Kitts (Toothill et al., 2007; Melekhova
1689 et al., 2017; Cooper et al., 2020); Redonda (Cooper et al., 2020); Montserrat (Devine et al., 1998;
1690 Humphreys et al., 2010; Mann et al., 2013; Cooper et al., 2020); Guadeloupe (Poussineau, 2005;

1692 Boudon et al., 2008; Pichavant et al., 2018; Cooper et al., 2020; Metcalfe et al., 2022; Metcalfe, 2022);
1693 Dominica (Baalcone-Boissard et al., 2018; Cooper et al., 2020; d'Augustin et al., 2020; Ostorero et al.,
1694 2021); Martinique (Martel et al., 1998; Cooper et al., 2020); St Lucia (Cooper et al., 2020); St Vincent
1695 (Devine and Sigurdsson, 1983; Heath et al., 1998; Bouvier et al., 2010b; Cooper et al., 2020); Petite
1696 Mustique (Cooper et al., 2020); Kick 'em Jenny (Camejo-Harry et al., 2019); Grenada (Bouvier et al.,
1697 2010a; Cooper et al., 2020).

Figure 5 Timeline showing the distribution of LAA 20th and 21st Century eruption styles. Grey bars indicate years an eruption was occurring in the Lesser Antilles. For each eruption we show the main eruptive styles which occurred and indicate the maximum VEI. Where the data is available, we show the average bulk rock SiO₂. Smaller and less explosive eruptions are more common (e.g., phreatic, vulcanian and dome) than larger explosive eruptions. Submarine eruptions are also common due to the high level of activity observed at Kick 'em Jenny.

Figure 6 Panels showing: The Lesser Antilles and the location of the Marathon and Mercurius fault zones, 20th century eruptions, crustal thickness, magma storage depth, magma temperature, MI SiO₂ content, MI H₂O content and MI δ¹¹B content variation. Crustal thickness is highest in the central arc, which coincides with the shallow storage magma zones, the most evolved magma compositions, the highest MI δ¹¹B content and the location of the Marathon and Mercurius fault zones.

Figure 7 XCO₂ in the gas phase v pressure for **(A)** The Northern Arc; **(B)** The Central Arc; **(C)** The Southern Arc. From this we can see the process of decompression and CO₂ fluxing occurring across the arc, the direction of these processes are shown by the arrows. We highlight that in Dominica we see both process of CO₂ fluxing (increasing CO₂ with decreasing pressure) and decompression (decreasing CO₂ with decreasing CO₂). For full melt inclusion dataset see supplementary dataset 1. (Saba (Roobol and Smith, 2004); Statia (Cooper et al., 2020); St Kitts (Toothill et al., 2007; Melekhova et al., 2017; Cooper et al., 2020); Redonda (Cooper et al., 2020); Montserrat (Devine et al., 1998; Humphreys et al., 2010; Mann et al., 2013; Cooper et al., 2020); Guadeloupe (Poussineau, 2005; Boudon et al., 2008; Pichavant et al., 2018; Cooper et al., 2020; Metcalfe et al., 2022; Metcalfe, 2022); Dominica (Baalcone-Boissard et al., 2018; Cooper et al., 2020; d'Augustin et al., 2020; Ostorero et al., 2021); Martinique (Martel et al., 1998; Cooper et al., 2020); St Lucia (Cooper et al., 2020); St Vincent (Devine and Sigurdsson, 1983; Heath et al., 1998; Bouvier et al., 2010b; Cooper et al., 2020); Petite Mustique (Cooper et al., 2020); Kick 'em Jenny (Camejo-Harry et al., 2019); Grenada (Bouvier et al., 2010a; Cooper et al., 2020).

Table 1 Sub-aerial eruptions in the LAA occurring in the 20th and 21st centuries (Lacroix, 1904; Jolivet, 1958; Aspinall et al., 1973; Shepard et al., 1979; Feuillard et al., 1983; Fournier et al., 2009; Wadge et

1725 al., 2014; Komorowski et al., 2005, 2015, 2022; Pyle et al., 2018; Barclay et al., 2019; Joseph et al.,
1726 2022; Global Volcanism Program, 2023). 14 submarine eruptions have also occurred from Kick ‘em
1727 Jenny (Lindsay et al., 2005a). *Current day summit temperatures Lindsay et al., 2005; Moretti et al.,
1728 2020a,b; Basan et al., 2022). **UWI-SRC 2020-22; Global Volcanism Program, 2020. ***some ‘self-
1729 evacuation’ occurred but no numbers are available (Pyle et al., 2018).

1730 **Figures and Tables**

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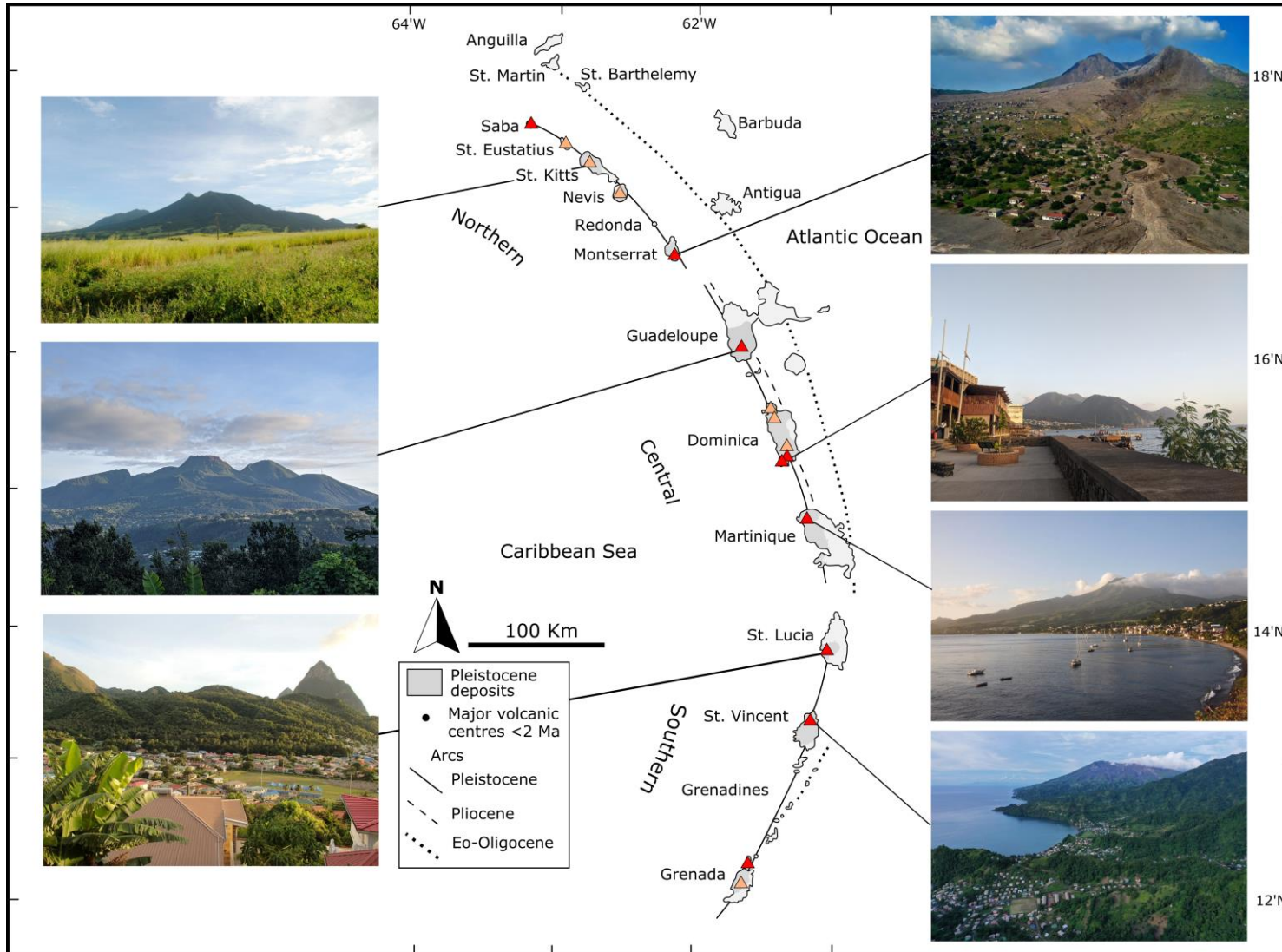
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FIGURE 1

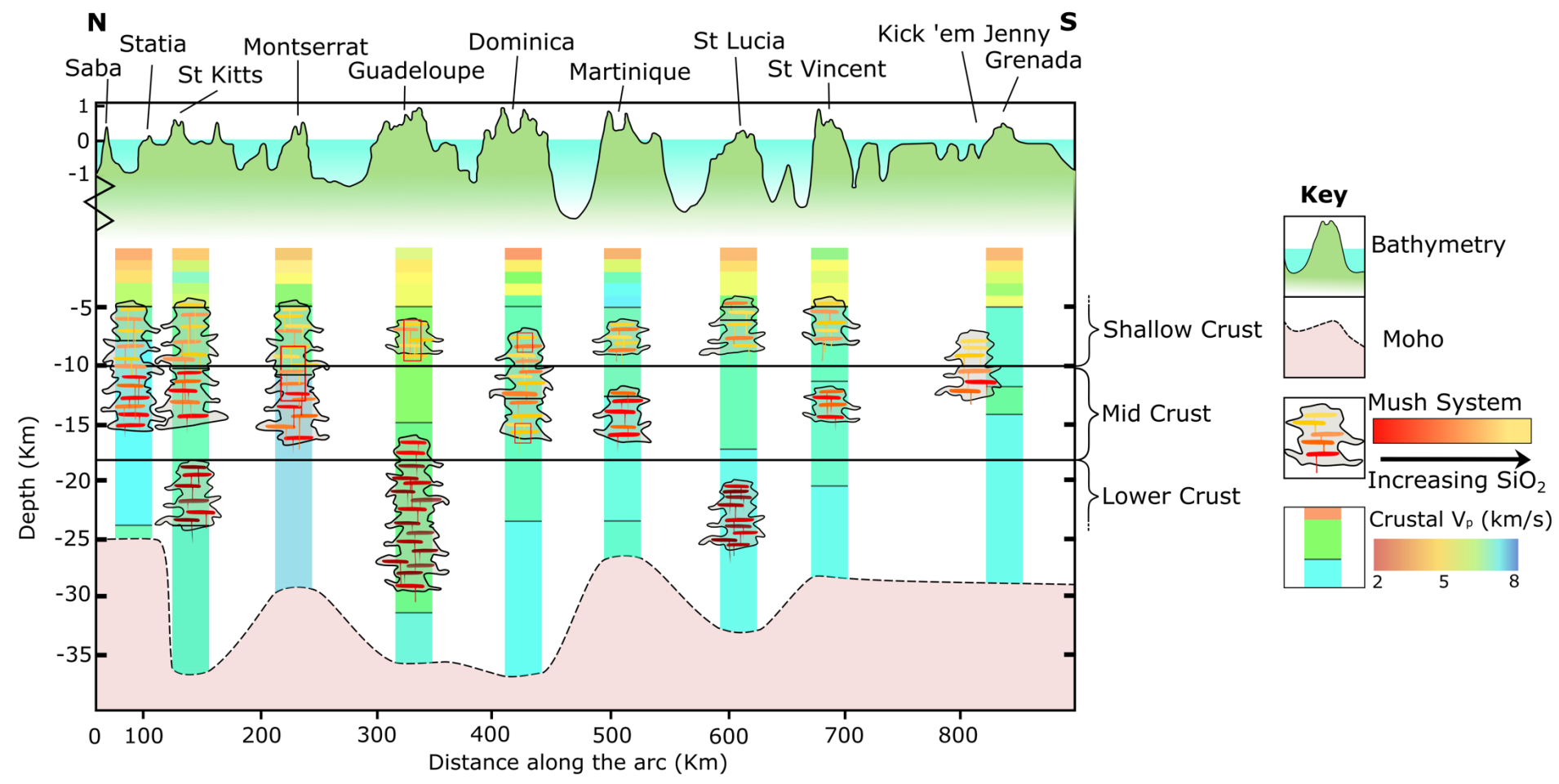


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FIGURE 2



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FIGURE 3

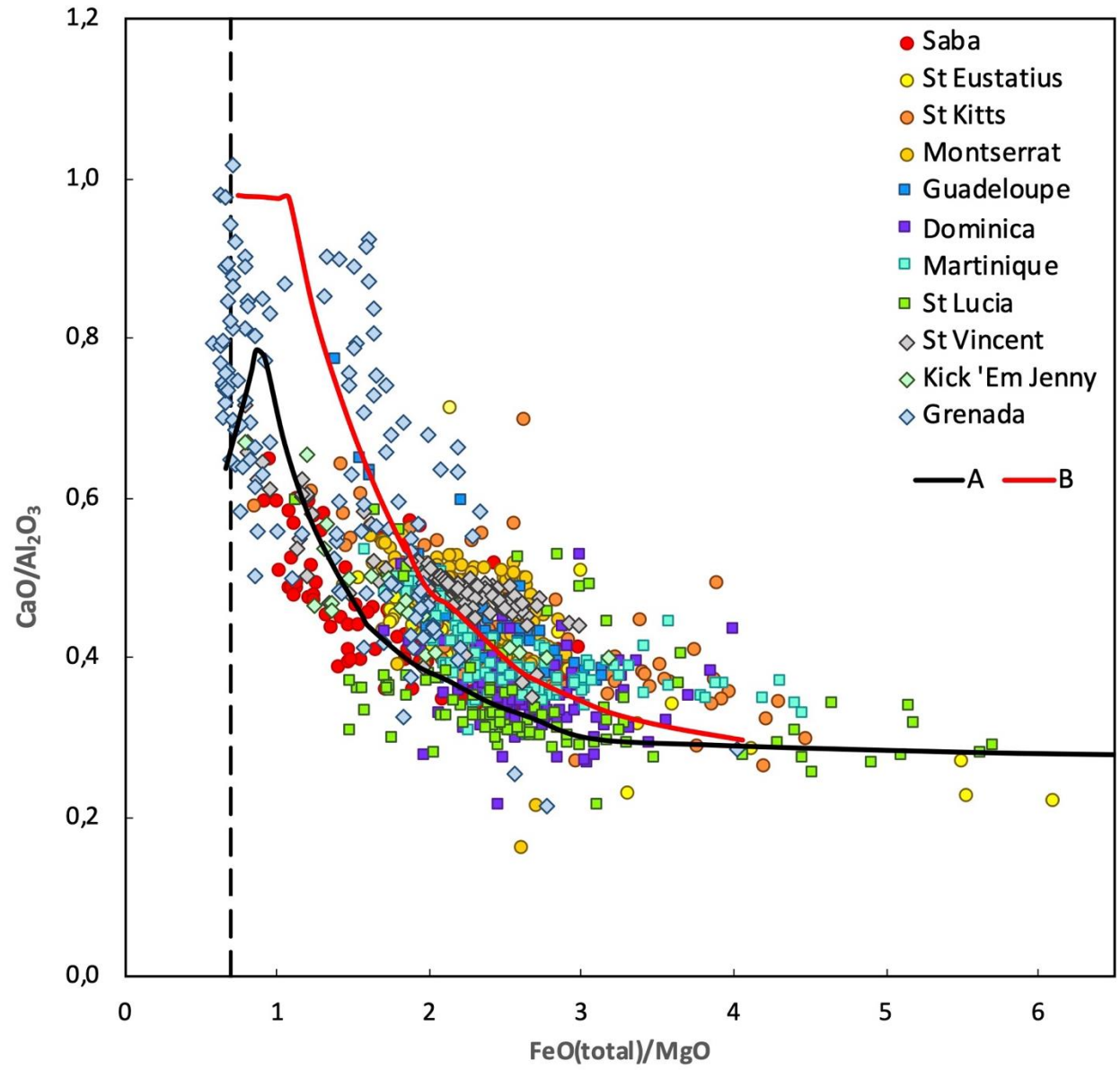
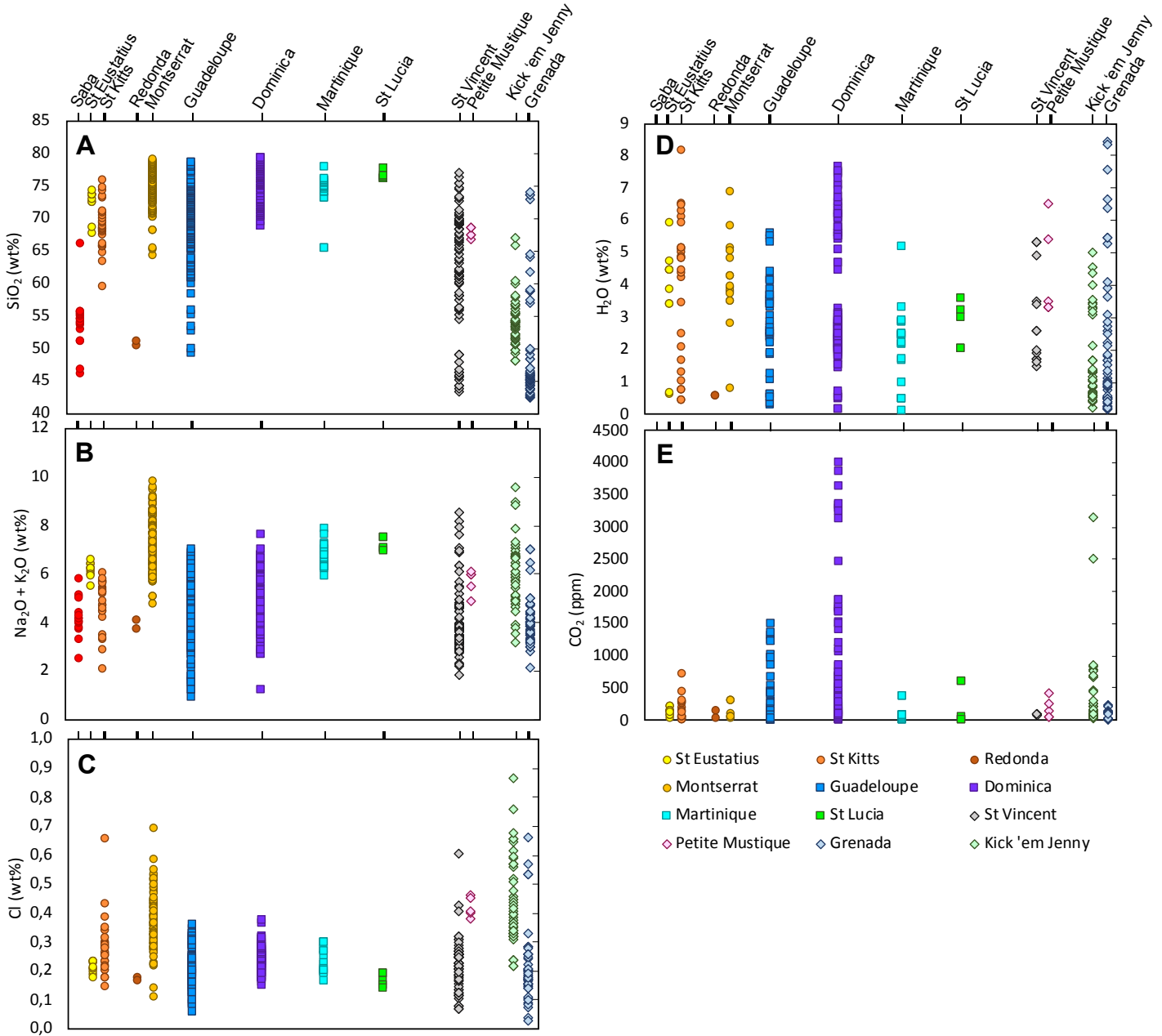
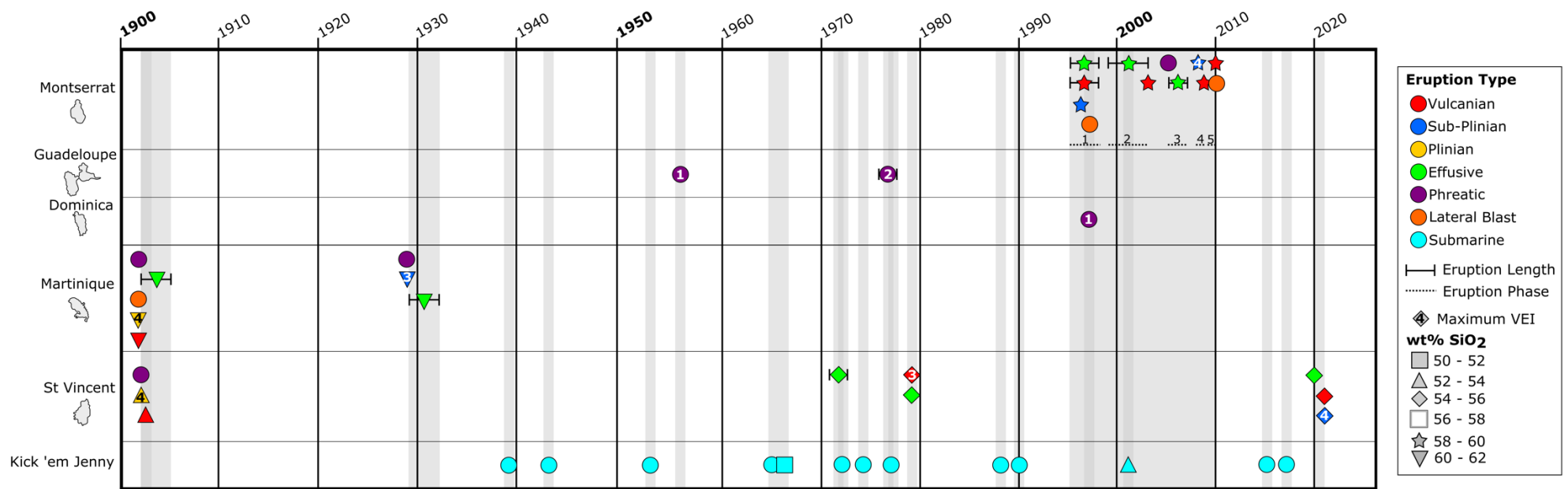


FIGURE 4



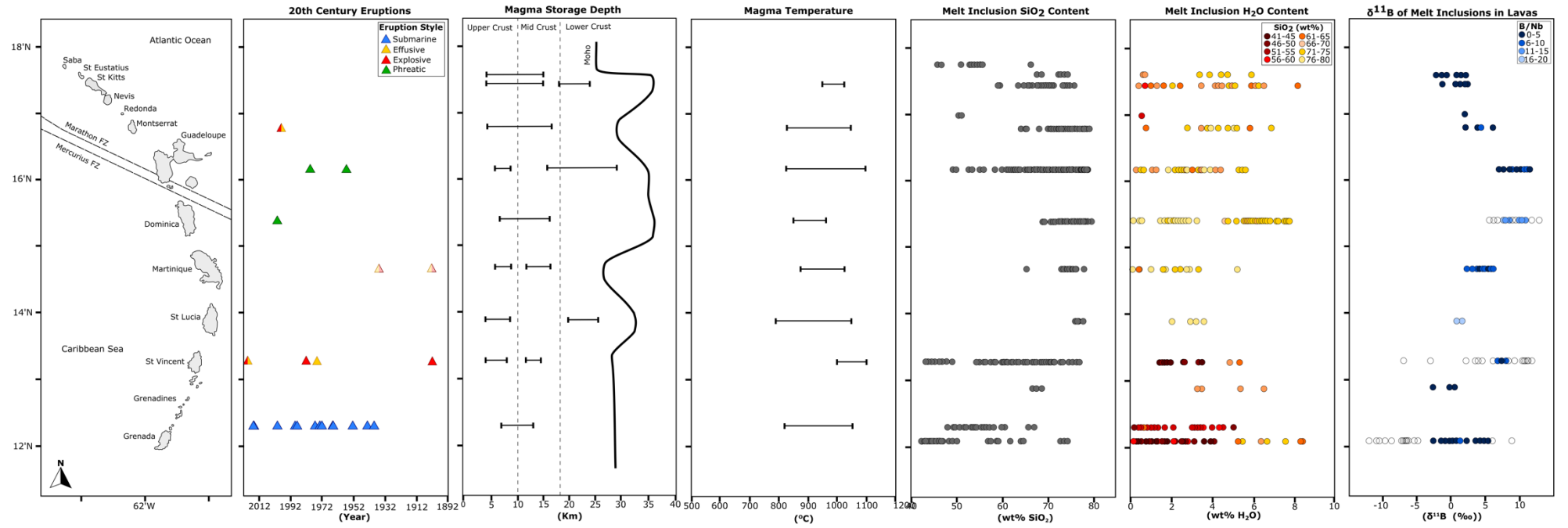
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FIGURE 5



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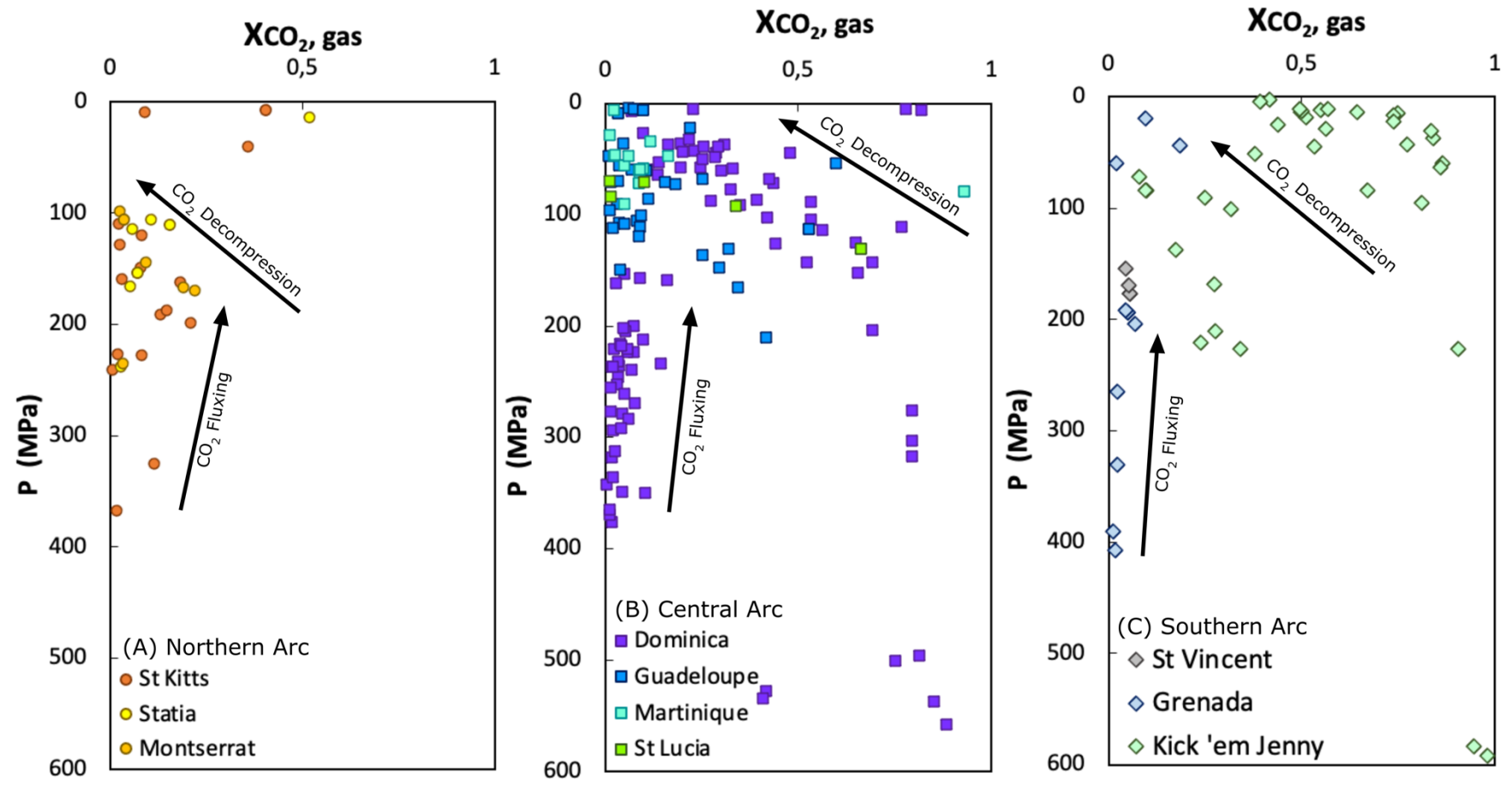
FIGURE 6



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FIGURE 7



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Island	Volcano	Summit Temperature (°C) *	Eruption start	Eruption end	Main Eruptive Style	Maximum VEI	Eruptive Phases	Unrest Period Reported Prior to Eruption	Number of Deaths	People Evacuated
St Vincent	Soufrière St Vincent	>95	2020	2021	Explosive	4	- Dome growth - Cyclic vulcanian explosions - Sub-Plinian activity	~2 months**	0	>16,000
Dominica	Morne Watt	91 - 99	1997	1997	Phreatic	1	- Unobserved phreatic explosion	N/A	0	0
Montserrat	Soufrière Hills	98 - 99	1995	Ongoing	Explosive	3 - 4	- Cyclic dome growth and collapse, - Cyclic vulcanian explosions - Sub-Plinian activity - Lateral blasts	3 years	19	>6,000
St Vincent	Soufrière St Vincent	-	1979	1979	Explosive	3	- Vulcanian explosions - Dome growth	~days	2	20,000
Guadeloupe	La Soufrière de Guadeloupe	~100	1976	1977	Phreatic	2	- 26 phreatic explosions	13 months	0	70,000
St Vincent	Soufrière St Vincent	80-100	1971	1972	Effusive	0	- Aseismic dome growth	N/A	0	0

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Guadeloupe	La Soufrière de Guadeloupe	~100	1956	1956	Phreatic	1	- Phreatic explosion	~months	0	0
Martinique	Montagne Pelée	-	1929	1932	Explosive	3	-Phreatic eruptions - Pelean activity - Dome growth	~months	N/A	N/A
Martinique	Montagne Pelée	-	1902	1907	Explosive	4	- Vulcanian explosions - Sub-Plinian and Pelean activity - Dome growth and collapse	~13 years	~30,000	N/A
St Vincent	Soufrière St Vincent	-	1902	1903	Explosive	4	- Vulcanian explosions - Sub-Plinian activity	1 year	~1500	N/A***

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8 **Diverse Magma Storage and Major and Volatile Magma Composition: What are the implications**
9 **on the Eruptive Style Across a Volcanic Arc? An Example of the Lesser Antilles Arc.**

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32 **Key Words:** Lesser Antilles; Geochemistry; Magma Storage; Eruption Style; Arc Volcanism; Melt
33 Inclusions; Magma Mush, Volcanic Unrest, Volcanic Eruption

34
35 **Abstract**

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37 Understanding variations in eruptive style and defining the processes and controls on different
38 eruption styles is particularly important at arc volcanoes. This is due to the range of eruption styles
39 which vary from phreatic behaviour to effusive magma extrusion and large explosive Plinian eruptions.
40 Such eruptions can have wide-ranging and long-term impacts on vulnerable communities living near
41 volcanoes. Magma storage and chemistry are key to understanding the origins of such variations
42 along-arc. The Lesser Antilles Arc provides a perfect setting to investigate magma storage, magma
43 composition and eruptive style variations associated with arc volcanism. This arc is comprised of
44 eleven volcanically active islands. Since 1900, there have been 24 eruptions. Many eruptions have
45 resulted in thousands of deaths, large scale and long-term displacements and severe economic
46 impacts, which have had profound impacts on the people living in this region. Through this review we
47 highlight the variations observed in magma storage, which is described as a vertically extensive trans
48 crustal mush system. We hypothesise that across the arc there are generally three levels of the mush
49 storage system. First, a deep system at 25 km (ca. >400 MPa). This system is poorly constrained and

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contains the most primitive magmas which fractionate to produce magmas which ascend into the second mid-crustal system at 10 – 18 km (ca. 3200 - 400 MPa). The mid-crust stores basalt to basaltic andesite which supplies and remobilises the third and last shallow system at <10 km (ca. <200 MPa) which hosts evolved dacitic – rhyolitic magmas. This is the most well-defined part of the magma system and is the region supplying eruptions. The remobilisation of the shallow system and the production of eruptible magma can result in eruption, with magma composition considered a first order control on eruption style. This is observed with the northern and central arcs, which produce evolved compositions, large explosive and dome-forming eruptions, with associated hazards. Magmatic productivity compositional variation and variation in magma storage depth is controlled by the distribution of subducted serpentine. This is particularly important to consider in the central arc which has the highest magma productivity and evolved magmas stored at shallow depths. These magmas are likely to erupt explosively in the near future, given the unrest observed at many systems in the central arc. Despite the fact that advancements in science, technology and our understanding of volcanoes in the arc undoubtedly prevented large numbers of casualties in the late 20th and 21st centuries, large evacuations of people remain unavoidable. In some cases, evacuation becomes long term displacement and results in major socio-economic impacts. This highlights the need for future crises to be effectively managed so as to limit the socio-economic impacts and maintain the relationship between scientists, authorities and the population.

1 Introduction

Arc volcanism is often considered as one of the most hazardous types of volcanism due to the range of eruption styles observed (e.g., Roobol and Smith, 1989; Brown et al., 2017). This variability makes it difficult to define the processes, timescales and controls on eruptions and resulting hazards. Understanding the drivers of changes in eruptive style is particularly important, as eruption style dictates many of the hazards presented by a volcano (e.g., Komoroswki et al., 2005; Lindsay et al., 2005; Cassidy et al., 2018). Eruptions at arc volcanoes range from phreatic behaviour (generated in the hydrothermal system and not resulting in magma extrusion, e.g., Kilgour et al., 2019) to effusive eruptions in the form of lava domes and spines (e.g., Sheldrake et al., 2016), to explosive behaviour ranging from small vulcanian eruptions (e.g., Druitt et al., 2002) to large Plinian eruptions (e.g., Vidal et al., 2015). There are multiple ways to generate an eruption and processes are unlikely to occur in isolation (Villemant and Boudon, 1998; Di Genova et al., 2017; Cassidy et al., 2018; Popa et al., 2021). For magmatic eruptions, many mechanisms involve recharge of the system by magma from depth and for explosive eruptions the supersaturation of volatiles in the magma and an increase in overpressure, which exert a control on the flow rate into the conduit (e.g., Burnham, 1975).

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Volcanoes are not limited to one eruption style and there are transitions in eruption style that are observed at arc volcanoes within the same eruption (e.g., effusive to explosive and vice-versa; e.g., Komorowski et al., 2008; Legendre, 2012; Boudon and Balcone-Boissard, 2021; Joseph et al., 2022) and across several eruptions with an observed repose period (Cassidy et al., 2018). Magma composition is often considered a first order control on eruption style (e.g., Sides et al., 2014; Firth et al., 2016). An explosive eruption is the result of brittle failure of a melt phase when the volume fraction of exsolved volatiles is sufficiently large (Sparks, 1978; Woods and Koyaguchi, 1994; Carr et al., 2020), when the strain rate is sufficiently high (Papale, 1999; Gonnermann and Manga, 2003), or when bubble overpressure exceeds a critical value (Zhang, 1999). Variations in eruption style therefore relate to the ability of that system to dissipate increasing pressure through degassing (Woods and Koyaguchi, 1994; Melnik and Sparks, 2002; Ruprecht and Bachmann, 2010; Kozono and Koyaguchi, 2012). Such variations are often linked to processes occurring in the magma storage system including magma mixing and magma recharge (Sparks et al., 1977; Murphy et al., 2000; Plail et al., 2018). However, recent studies also stress the importance of processes occurring external to the magmatic system (e.g., Hlinka et al., 2021; Metcalfe et al., 2022).

In recent years the effects of bottom-up vs top-down vs external forces has been shown to play an important role in transitions in the eruptive style, both during a single eruption and across multiple eruptions (e.g., Jaupart and Allègre, 1991; Sparks, 1997; Melnik and Sparks, 1999; Cassidy et al., 2018; Bernard and Bouvet de Maisonneuve, 2020; Hlinka et al., 2021; Metcalfe et al., 2022). Bottom-up controls drive an eruption through changes in the magma storage system. In comparison, top-down controls result from changes in the uppermost plumbing system affecting the late-stage decompression path of the magma that proceed downward. Another important control on eruption style and eruption style transition is caused by external effects such as: flank or dome collapse (Alidibirov and Dingwell, 1996); assimilation of external materials (Deegan et al., 2010; Troll et al., 2012); and external water interacting with magma (Shepherd & Sigurdsson, 1982; Zimanowski et al., 2015). A change in morphology of the volcanic edifice can also impact crater morphology and result in changes to the volcanic hazards produced (e.g., the generation of pyroclastic density currents at Soufriere St Vincent; e.g., Cole et al., 2019).

Another important consideration is how the magma is stored at depth, which can vary depending on, for example, the melt proportion, geometry, temperature and compositions (e.g., Cashman et al., 2017). A now widely accepted interpretation of the magma storage zones in arc volcanoes is that magma is stored in a mush system. Mush systems are formed from repeated intrusions of melt from depth and are composed of an interlocking framework of crystals with the presence of melt, containing dissolved volatiles, distributed in varying volumetric proportions within the network of

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7 99 crystals (e.g., Humphreys et al., 2006; Cashman and Blundy, 2013; Cashman and Sparks, 2013; Cooper
8 and Kent, 2014; Christopher et al., 2015; Edmonds et al., 2016). The crystalline parts of the mush
9 system are deformable, meaning melt and fluid can be segregated within the mush system (Mckenzie
10 1984; Solano et al 2012; Connolly et al., 2015; Spiegleman et al., 2001), with the strength of the
11 framework increasing as melt fraction decreases (Marsh, 1981; Rosenberg and Handy, 2005).
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14 104 Geophysical methods identify laterally and vertically elongated low-velocity zones with <10% melt
15 more consistent with a crystal mush system (Koulakov et al., 2013; Ward et al., 2013; Sinton, 1992;
16 Lees, 2007; Farrell et al., 2014; Huang et al., 2015; Kiser et al., 2016; Pritchard and Gregg, 2016;
17 Pritchard et al., 2018; Cordell et al., 2018). A magma mush can exist at a lower temperature due to
18 the close crystal contacts and the high viscosity which prevents convection and thereby limits most
19 heat lost through conduction (Balmforth and Rust, 2009; Dufek et al 2010; Blundy and Annen, 2016).
20 The heat loss prevents large volumes of eruptible magma from forming but allows the mush system
21 to be maintained over long periods of time (Blundy and Annen, 2016; Cashman et al., 2017; Karakas
22 et al., 2017; Glazner, 2021). Periodic injections of recharge magma have also been shown by numerical
23 models to be sufficiently high or long enough to maintain some eruptible melt within a crystal mush
24 (Huber et al., 2012; Annen et al., 2015; Cooper, 2017). Long-lived, trans-crustal mush systems are
25 consistent with many of the features of arc volcanoes and the long-lived volcanic activity observed in
26 the arc (e.g., Balcone-Boissard et al., 2018; Metcalfe et al., 2022; Higgins et al., 2022).
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32 117 The transition from crystal mush to an eruptible magma occurs over a narrow crystal content (Shaw,
33 1969; Caricchi et al., 2007; Costa et al 2009; Burgisser and Bergantz, 2011; Sparks et al., 2019). The
34 energy required to remobilize a crystal mush depends on the initial melt content; mushes with a higher
35 initial melt content require less energy to remobilise (Bachmann and Bergantz, 2006; Bergantz et al.,
36 2017; Sparks et al., 2019). Magma recharge and addition of volatiles are considered the main drivers
37 of remobilization (Scandone et al., 2007; Costa et al., 2009; Mader et al., 2013; Rubin et al., 2017;
38 Cooper and Kent, 2014; Sparks and Cashman, 2017; Jackson et al 2018). There are several hypotheses
39 on how a mush is remobilised. Many models involve a reduction in crystallinity through, for example,
40 reactive melt flow (Jackson et al., 2018) and the growth of a buoyant mobile layer (Burgisser and
41 Bergantz, 2011). This increases the melt fraction and leads to remobilization of the cold mush system
42 through overturn or unzipping of the mush system (Burgisser and Bergantz, 2011; Bergantz et al.,
43 2015, 2017; Schleicher et al., 2016; Schleicher and Bergantz 2017; Carrara et al., 2019) The different
44 proposed mechanisms of remobilisation indicate that this is a complex process and may involve a
45 combination of mechanisms.
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51 131 The Lesser Antilles Arc (LAA) is a good example of an active island arc, with a range of eruption styles
52 recorded and a wealth of geochemical data available, making it an excellent arc to investigate links
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between magma storage, magma composition and the eruptive style. Few petrologic studies investigate the arc as a whole (e.g., Cooper et al, 2020; Melekofova et al., 2019), instead choosing to focus on individual islands and magmatic systems. This has resulted in a significant amount of data available for the LAA, which has expanded greatly in the last few years (e.g., VOILA project; Goes et al., 2019). Here, we provide a review of the available LAA data, focusing on the geochemistry, magma storage and eruptive style, adding to the reviews already available from Macdonald et al. (2000) (geochemistry), Lindsay et al. (2005) (volcanic hazard atlas) and Smith et al. (2013) (geological and tectonic setting).

2 The Lesser Antilles Arc

The Lesser Antilles Arc is an active volcanic arc in the Caribbean, which spans 850 km from Saba in the north to Grenada in the south (Fig 1). The active arc is made up of eleven main volcanic islands, an archipelago of nineteen smaller islands (the Grenadines) with a total of twenty-one volcanoes active in the Holocene, including a submarine volcano. The LAA is formed due to subduction of the American plate beneath the Caribbean plate at a rate of approximately 1.0 – 2.4 cm/yr (Macdonald and Holcombe, 1978; McCann and Sykes, 1984; Rosencrantz and Sclater, 1986; Jarrard, 1986; Deng and Sykes, 1995; Dixon et al., 1998; DeMets et al., 2000; Jarrard, 2003; Feuillet et al., 2011; Symithe et al., 2015). This slow subduction rate results in low magma production rates in comparison to other subduction zones. Wadge (1984) reports a magma production rate of $5 \text{ km}^3 \text{ Ma}^{-1} \text{ km}^{-1}$ for the past 300 years in the LAA. In comparison, the Central American arc has a magma production rate of $62 \text{ km}^3 \text{ Ma}^{-1} \text{ km}^{-1}$ for the past 300 years (Wadge, 1984).

To the north of Martinique, the arc bifurcates into an inactive eastern limb and an active western limb (Schlaphorst et al., 2018). The older eastern arc is composed of several islands which are now covered in a carbonate platform – the Limestone Caribbees, which experienced active volcanism from the Eocene to the mid-Oligocene. Activity on the arc has migrated westward forming the western arc, which has been active since 20 Ma (Early Miocene to present (Fig 1); Westercamp and Tazieff, 1980). Pleistocene to recent volcanism has occurred in narrow 10 km wide zones which are split into three segments: the Northern Arc (Saba to Montserrat), the Central Arc (Guadeloupe to Martinique) and the Southern Arc (St Lucia to Grenada (Sigurdsson and Carey, 1981; Wadge and Shepherd, 1984; Fig 1).

Each island has at least one active volcanic complex except for Dominica, which has nine potentially active volcanoes (e.g., Brown et al., 2015; Leone et al., 2019; Fig 1) and Guadeloupe which has two active volcanoes (Komorowski et al., 2005; Komorowski et al., 2022). Eruptions range from small phreatic to effusive dome forming (ca. $9 - 47 \times 10^6 \text{ m}^3$; e.g., Shepard and Sigurdsson, 1982; Loughlin

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et al., 2010; Martel, 2012) to large explosive eruptions (VEI 2 – 6; Legendre, 2012; Howe et al., 2014; Palmer et al., 2016; Carazzo et al., 2020; Joseph et al., 2022). The twenty-three eruptions that have occurred in the LAA since 1900 provide examples of how devastating an eruption in the LAA can be (Table 1). Recent eruptions have resulted in thousands of deaths (e.g., 29,800 at Montagne Pelée in 1902 (Lacroix, 1904); 1350 at La Soufrière St Vincent in 1902 (Brown et al., 2017; Pyle et al., 2018; Table 1), the short and long-term displacement of thousands more (e.g., La Soufrière de Guadeloupe, 1976; Soufrière Hills Volcano, Montserrat, 1995; Soufrière St. Vincent, 2021 (Lepointe, 1999; Komorowski et al., 2005; Feuillard, 2011; Joseph et al., 2022); Table 1) and long-lasting economic impacts (e.g., La Soufrière de Guadeloupe, 1976; Soufrière Hills, Montserrat, 1995-present; Lepointe, 1999; Blérald, 1986; Kokelaar, 2002), all of which have profound impacts on the lives and livelihoods of people in the LAA. Approximately one million people are threatened by the direct effect of volcanic activity in the Lesser Antilles, and related phenomena threaten approximately two and a half million people (Lindsay et al., 2005a; Leone et al., 2019).

There are notable variations in geochemistry, volcanic activity, crustal structure and seismicity observed along the LAA, including variations (Boynton et al., 1979; Macdonald et al., 2000, Melekhova et al., 2019; Hayes et al., 2014; Schlaphorst et al., 2018; 2021; Cooper et al., 2020). An example of this is the variation in Wadati-Benioff zone dip, which diverges from 50° – 60° in the Northern Arc to sub-vertical in the Southern Arc (e.g., Wadge and Shepard, 1984; Schlaphorst et al., 2018). In the Central Arc the slab dips at ca. 20° until 50 km depth where it increases to 45° – 55° (Paulatto et al., 2017). Systematic variations in crustal thickness are observed along the Lesser Antilles Arc, with a greater crustal thickness observed in the Northern and Central arcs. North of Martinique crustal thickness is estimated between 30 – 37 km (Arnaiz-Rodríguez et al., 2016; Melekhova et al., 2019; Schlaphorst et al., 2021). South of Martinique the crust thins with crustal thickness estimated between 20 – 30 km (Arnaiz-Rodríguez et al., 2016; Melekhova et al., 2019; Schlaphorst et al., 2021). The range of crustal thickness (20 – 37 km) is considered to relate to long-term variations in magma productivity (Melekhova et al., 2019; Cooper et al., 2020). Another contributing factor to the variation in magma production is the shift of the Marathon–Mercurius fracture zone from St Kitts (Northern Arc) to Dominica (Central Arc), which is reflected in the greater crustal thickness in the Northern Arc (Paulatto et al., 2017; Bie et al 2019; Cooper et al., 2020; Bie et al 2022).

Two crustal layers are observed using seismic discontinuities (Boynton et al., 1979). The first discontinuity is the upper crust, which extends to a variable depth of 2 -30 km along the arc and is a plutonic igneous layer (Wadge, 1986; Schlaphorst et al., 2018). Below this is a layer of dense mafic igneous rocks including cumulates directly overlying the mantle (the Moho), estimated between 30 - 60 km depth (e.g., Boynton et al., 1979; Christeson et al., 2008; Kopp et al., 2011; Laigle et al., 2013;

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Arnaiz-Rodríguez et al., 2016; Schlaphorst et al., 2021). Both discontinuities are found at greater depths in the northern part of the arc, with the along arc variation in the discontinuities being unusually strong (Schlaphorst et al., 2018). The strong variation in the nature of the discontinuities results from variation in melt generation and differentiation (Schlaphorst et al., 2018).

Variations in upper plate mantle seismic velocities are also observed, seismicity is shown to be closely related to hydration and dehydration processes and are controlled by thermal structure, composition and state of stress (Paulatto et al., 2017; Ezenwaka et al., 2022). Three main phases of dehydration occur in the Atlantic slab: (1) 5 – 20 km - compaction and draining of pore fluids from the sediments; (2) 40 – 100 km - chemical dehydration of the crust; (3) 120 – 160 km lithospheric mantle deserpentinization (van Keken et al., 2011; Paulatto et al., 2017). Notably slower seismic velocities are observed below Guadeloupe, Dominica and Martinique where thicker crust occurs and is bound by two major fracture zones to the north (Paulatto et al., 2017; Cooper et al., 2020; Harmon et al., 2019; Schlaphorst et al., 2021). The anomalies are thought to be related to either the upwelling of hot mantle material or partial melt and fluids. A high $\delta^{11}\text{B}$ anomaly is also observed in the Guadeloupe and Dominica melt inclusions thought to be related to dewatering of serpentinised oceanic crust, occurring close to the fracture zone to the north of the Central Arc (Cooper et al., 2020). Alternatively, the curvature of the arc (e.g., Harris et al., 2018; Schlaphorst et al., 2017), ponding of melt and resulting redistribution of fluids (Ha et al., 2020) or along-strike variability in the subducted plate morphology could also account for the velocity anomalies observed.

3 Magma Composition and Magma Storage Across the Lesser Antilles Arc

A crystal mush is often described to be vertically extensive and consist of melt-rich lenses and solid crystals (Cashman and Blundy, 2013; Cashman and Sparks, 2013; Cashman et al., 2017; Blundy et al., 2019). The mush exists under sub-solidus conditions for long timescales before remobilisation (e.g., Bachmann and Bergantz, 2006; Bergantz et al., 2015; Burgisser and Bergantz, 2011; Parmigiani et al., 2014). The chemical and dynamic features of magmatic systems of the LAA volcanoes can be explained by the model of a transcrustal, vertically (and often laterally) extensive mush system as described by Cashman et al. (2017).

Conditions of the mush system are calculated through: [1] Melt inclusion volatile contents which are converted into minimum entrapment pressures (e.g., Balcone-Boissard et al., 2018; Pichavant et al., 2018; Camejo-Harry et al., 2019; Metcalfe et al., 2022) using H₂O-CO₂ solubility models (e.g., Papale et al., 2006); [2] Thermobarometry (e.g., Bezard et al., 2017; Higgins et al., 2022) which relates mineral phase chemistry to a known pressure, temperature and melt composition (e.g., Nimis & Taylor, 2000; Blundy and Cashman, 2008; Putirka, 2008; Ridolfi et al., 2010); [3] Equilibrium experiments with the

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pre-eruptive physical parameters defined by the experimental phase equilibria and chemistry (e.g., Pichavant et al., 2002, 2018; Solaro et al., 2019, 2020).

Pressures determined by melt inclusions are considered minimum estimates of storage pressures due to degassing processes that can occur prior to entrapment or if the melt was not volatile saturated at the point of entrapment (e.g., Steele-Macinnis et al., 2011; Cannatelli et al., 2016). Additionally, post-entrapment processes can result in volatiles leaking (e.g., Steele-Macinnis et al., 2011; Wallace et al., 2015; Cannatelli et al., 2016). In particular, H₂O loss can occur through H⁺ diffusion through the host mineral and CO₂ is lost to vapour bubbles forming during post-entrapment cooling (e.g., Wallace et al., 2015). However, when melt inclusions are used to determine pressure careful checks are made to ensure the melt inclusion has not experienced post entrapment processes (e.g., using deuterium values for instance; e.g., Metcalfe et al., 2022) and to ensure the melts were volatile-saturated at the time of entrapment (e.g., using CO₂ vs Th; Balcone-Boissard et al., 2018).

Additionally, crustal density affects depth-pressure conversions, depth-pressure conversions range from 2450 – 2800 kg/m³ across the arc (e.g., Kopp et al., 2011; Christeson et al., 2008; Barnoud et al., 2016, Gómez-García et al., 2019). The mean crustal density of 2660 kg m⁻³ is also widely used for conversions (Christeson et al., 2008).

Studies of the LAA magmatic systems indicate the deepest storage areas occur at >25 km, but will likely depend on the slab dip and crustal thickness which varies across the arc (Schlaphorst et al., 2018; Melekhova et al., 2019; Schlaphorst et al., 2021). A product of the differentiation and fractionation processes are cumulates with the mineral proportions dependent on the magmatic conditions and depths. Cumulates, along with xenoliths, are brought to the surface and can provide further insights into the plumbing system (e.g., Baker 1968; Parkinson et al., 2003; Stamper et al., 2014; Cooper et al., 2016; Melekhova et al., 2019).

Many studies indicate that differentiation to produce evolved melts is occurring deeper in the system, generally in the range 13 to 16 km (ca. 300 - 400 MPa; Edmonds et al., 2016; Pichavant et al., 2002; Boudon and Balcone, 2021). Though this is often less well resolved by geophysical and geochemical data in comparison to the shallow regions that are better sampled by eruptions and imaged through geophysics. This is consistent with the transcristal magma mush model (e.g., Cashman et al., 2017).

The existence of this deeper region has various implications for the system, with recharge of the shallow storage system with less evolved magma often cited as the principal cause of unrest and eruption (e.g., Humphreys et al., 2010; Howe et al., 2015; Pichavant et al., 2018). From the deeper region fluids, heat and magma inputs into shallow depths may promote the formation of eruptible magma. However, recent studies also suggest melts migrating within the shallow zone and volatile

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oversaturation can result in an eruption without the requirement for a deep mafic recharge event (e.g., Moretti et al., 2019; Boudon and Balcone-Boissard, 2021). Despite this, the deeper system is important for the production of more-evolved magmas through fractionation processes, the triggering of eruption through recharge, and may also be evacuated in the more voluminous eruptions.

The active part of the systems supplying eruptions are located in the range of 5 - 9 km (ca. 100 - 200 ±50 MPa), with some evidence of melt at <5 km (e.g., Martel et al., 1998; Pichavant et al., 2002; Christopher et al. 2015; Edmonds et al., 2016; Pichavant et al., 2018; Balcone-Boissard et al., 2018 and Solaro et al., 2019; d'Augustin, et al., 2020; Metcalfe et al., 2021; Fig 2). These shallow storage regions are typically formed of evolved magmas and are often the main zone from which magma is erupted.

Magma production rate is relatively low compared to the global range of subduction zones, with the rate varying along the arc, the Central Arc produces the largest volumes of volcanic deposits, particularly in the past 0.1 Ma (40 km³ for Dominica and 8 km³ for Guadeloupe and Martinique; Wadge 1984; 1986; Macdonald et al., 2000). In comparison, the Northern and Southern parts of the arc have produced smaller volumes (ca. 0 – 5 km³; Macdonald et al., 2000).

A wide range of magma types are observed across the arc, with variation in sediment input into the arc playing a role in the changing chemistry of primitive parental magmas (Carpentier et al., 2008). In the Southern Arc, the incoming plate is rich in clastic detritus, in comparison to the Northern Arc, which has limited sediment supply and is richer in pelagic marine sediments (Faugères et al., 1993; Carpentier et al., 2008). The LAA magmas have fractionated from a parental magma hypothesised by MacDonald et al. (2000) as a picritic, silica-undersaturated magma, though due to the variation in sediment input the trace element compositions is likely to vary across the arc.

The LAA magmas vary from tholeiitic suites in the Northern Arc, to calc-alkaline suites in the Central Arc and silica-undersaturated, highly magnesian suites in the Southern Arc (Macdonald et al., 2000). The trends observed in the geochemical data show the various magmas are the result of fractional crystallisation (Fig 3; MacDonald et al., 2000). However, crustal contamination, and minor magma mixing are also occurring, particularly in the Central Arc (MacDonald et al., 2000), minor assimilation of crust or sediments is also observed in St Lucia and Martinique magmas (Bezard et al., 2015).

Here, we provide a summary of the magma storage systems and a summary of the wealth of geochemistry data available for the LAA.. This is important to understand any common depths of magma storage, along arc variations and to eventually provide a coherent model of magma storage along the arc. We note that, the geochemical dataset is biased towards both what is erupted and what is analysed out of large amounts of samples, providing an incomplete view of the system. Our dataset is available in Supplementary dataset 1 (built on compilation by DIGIS Team, 2022).

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3.1 The Northern Arc

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In the Northern Arc (Saba to Montserrat) the dominant rock type ranges from basalt to andesite (e.g., Rea and Baker, 1980; Fig 3). However, on several islands the compositional range can vary greatly, from basalt to rhyolite.

3.1.1 Saba

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Eruption deposit compositions from Saba range from basalts to andesites (49 – 66 wt% SiO₂), with the basalts representing some of the least evolved compositions observed in the Northern Arc (Fig 3; Defant et al., 2001; Roobol and Smith, 2004). Evidence of magma mixing in banded pumices and rounded enclaves is observed (Defant et al., 2001). However, no compositional differences between the enclaves and other magmas are discussed (Defant et al., 2001).

Melt inclusions (MIs) are also unevolved, particularly in comparison to other MIs in the Northern Arc, and generally fall in the range 46 - 56 wt% SiO₂ (Fig 4A; Roobol and Smith, 2004). However, Na₂O + K₂O contents (1.7 – 4.6 wt% Na₂O, 0.36 – 0.45 wt% K₂O; Fig 4B) fall into the range observed in the Northern Arc (Roobol and Smith, 2004). No volatile data or pressure data is reported for Saba.

3.1.2 St Eustatius (Statia)

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Magmas erupted from The Quill on St Eustatius (Statia) are dominantly andesitic in composition, however, a broader range of compositions are observed ranging from basalt to rhyolite (50 - 72 wt% SiO₂; Roobol and Smith 2004). The compositions are generally relatively evolved, particularly in comparison to magmas erupted from Saba and St Kitts (Fig 3; Roobol and Smith 2004). MIs hosted in clinopyroxene from Statia shows a similar range of SiO₂ contents (68 – 74 wt% SiO₂; Fig 4A) to those from other systems in the Northern Arc. The Na₂O + K₂O contents also fall into a range typical of the Northern arc (0.9 – 1.2 wt% K₂O, 4.4 – 5.3 wt% Na₂O; Fig 4B). Volatiles recorded in MIs from Statia show Cl content is relatively low in comparison to other systems in the Northern Arc (0.18 – 0.23 wt%, Fig 4C). A wide range of H₂O is recorded (0.61 - 5.93 wt%; Figure 4D), and CO₂ is comparable to other systems in the Northern arc (26 – 216 ppm) (Cooper et al., 2019). Bubbles are reported by Cooper et al. (2019), and the pressure estimated calculated are considered minima.

Cumulates from Statia have mineral assemblages and textures that are similar to those observed in xenoliths from Martinique (Cooper et al., 2019) with plagioclase, olivine, clinopyroxene, orthopyroxene, amphibole and spinel interstitial melt also present (Cooper et al., 2016, 2019). Cooper et al. (2019) calculate minimum volatile saturation pressures ranging from 0 – 426 MPa, which indicates storage at <15 km.

3.1.3 St Kitts

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St Kitts eruption products show a wide range of compositions ranging from basalt to rhyolite (e.g., Toothill et al., 2007; Melekhova et al., 2017; Higgins et al., 2021). The most recent active volcano on St Kitts is Mt Liamuiga where basaltic-andesites and andesites are most common (42.1 – 65.4 wt% SiO₂; Baker, 1984; Toothill et al., 2007; Melekhova et al., 2017; Higgins et al., 2021). Many Mt Liamuiga eruption products are considered to be hybrid magmas from mixing of basalt, basaltic andesite and andesite magmas, and are the product of open-system processes (Toothill et al., 2007). These magmas cover nearly the whole range of compositions observed in the LAA (Fig 3).

MI's hosted in plagioclase, clinopyroxene, orthopyroxene, ilmenite and amphibole from St Kitts range from 59 – 76 wt% SiO₂, 1.3 – 4.8 wt% Na₂O and 0.6 – 2.5 wt% K₂O, which is within the range observed in the Northern Arc (Fig 4A, B; Toothill et al., 2007; Melekhova et al., 2017; Cooper et al., 2020). Volatiles in MI's from St Kitts record Cl contents in the range 0.15 – 0.43 wt% (Fig 4C; Toothill et al., 2007; Melekhova et al., 2017; Cooper et al., 2020) with one MI recording 0.66 wt% Cl. H₂O recorded in the MI is variable ranging from 0.4 - 8.2 wt%; Fig 4D; Toothill et al., 2007; Melekhova et al., 2017; Cooper et al., 2020), this is the highest H₂O content recorded in the Northern Arc. CO₂ ranges from 15 – 721 ppm (Fig 4E; Toothill et al., 2007; Melekhova et al., 2017; Cooper et al., 2020), though this is the highest CO₂ recorded in the northern Arc, this is still relatively low in comparison to the rest of the arc. Bubbles are reported in the St Kitts MI's (Toothill et al., 2007; Melekhova et al., 2017), however, MI's with lower H₂O or CO₂ are not associated with larger bubbles (Melekhova et al., 2017). Xenoliths from St Kitts are reported to be rich in plagioclase and amphibole.

An extensive, transcrustal magmatic mush system has been proposed by Melekhova et al. (2017) and Higgins et al. (2022) for the production of magmas at St Kitts. Depths are estimated from thermobarometry of xenoliths and seismic modelling and range from 3 – >25 km. Though the deepest part of the system is more difficult to distinguish as the cumulates are rarely observed, the deep system is hypothesised to generate basalts -basaltic andesites through fractionation which ascend to 18 – 25 km where they experience differentiation (Melekhova et al., 2017) (Fig 2). These basaltic-andesites are considered the parental magmas for more evolved compositions which migrate upwards and equilibrate with existing crystal mush (Melekhova et al., 2017; Fig 2). This means many eruption products will have an equilibration pressure corresponding to shallow depths despite differentiation beginning much deeper in the system. The majority of lavas erupted are supplied from the upper portion of the transcrustal system, this is consistent with experimental data which indicates many eruption products from St Kitts can be generated at 200 – 550 MPa (ca. 3 km – 15 km; Melekhova et al., 2017; Higgins et al., 2022). Though many magmas correspond to shallow depths, some melts are shown to bypass equilibration with the shallower mush system and so preserve evidence of their origin in the deeper system (Melekhova et al., 2017).

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3.1.4 Montserrat

The ongoing, multiphase eruption of Soufrière Hills Volcano on Montserrat has allowed many in depth studies to be conducted on the eruption products. Soufrière Hills Volcano on Montserrat, has produced a restricted bulk composition of andesite (56 - 62 wt% SiO₂; e.g., Murphy et al., 2000; Wadge et al., 2014). No systematic changes in bulk composition have been observed since the beginning of the most recent eruption in 1995. Mixing and mingling textures are reported on Montserrat with rare, banded lava and pumice observed, however no significant difference in composition is noted between layers (Murphy et al., 2000).

Mafic enclaves are hosted throughout the andesite, with the proportion of mafic enclaves increasing from the Phase 1 (1995) deposits to the Phase 5 (2010) deposits (Barclay et al. 2010). The mafic enclaves are generally less evolved than the andesite they are hosted in (49 – 56 wt% SiO₂ (Murphy et al., 2000; Christopher et al., 2014; Plail et al., 2018)), and some overlap in composition is observed with the South Soufrière Hills magmas found in the south of the island (e.g., Baker, 1984; Murphy et al., 2000; Wadge et al., 2014). The eruption products are some of the most evolved in the Northern Arc, and do not extend to the same unevolved compositions observed on Saba and St Kitts (Fig 3).

Montserrat has a large amount of MI data available due to the large amount of work done on the 1995 – present eruption deposits. MI hosted in plagioclase, quartz and clinopyroxene from Soufrière Hills volcano are the most evolved in the Northern Arc (64 – 79 wt% SiO₂; Fig 4A; Devine et al., 1998; Humphreys et al., 2020; Mann et al., 2013; Cooper et al., 2020). The highest Na₂O + K₂O values observed across the whole LAA arc are also from Soufrière Hills volcano MIs (2.4 - 6.3 wt% Na₂O, 0.9 – 6 wt% K₂O; Fig 4B; Devine et al., 1998; Humphreys et al., 2020; Mann et al., 2013; Cooper et al., 2020). Xenoliths from Montserrat are reported to be comparable to those from Guadeloupe and St Lucia and are felsic and olivine-free (Kiddle et al., 2010; Melekhova et al., 2019).

MIs from Montserrat record the highest Cl contents in the Northern Arc, generally ranging from 0.22 - 0.59 wt% (Fig 4C; Devine et al., 1998; Humphreys et al., 2010; Mann et al., 2013; Cooper et al., 2020). H₂O contents range from 0.78 – 6.9 wt%, which falls in the range recorded across the Northern arc (Fig 4D; Mann et al., 2013; Cooper et al., 2020). CO₂ contents are low, as observed in the other systems in the Northern Arc, ranging from 23 – 304 ppm (Fig 4E; Cooper et al., 2020). Vapour bubbles are not reported by Cooper et al. (2020).

This provides a snapshot of the magma system as it was directly prior to the eruption. Our understanding of this system is based on the key hypothesis that an andesitic magma body is recharged by an intruding basaltic magma, which releases heat and volatiles and drives eruption (e.g., Murphy et al., 2000; Wadge et al., 2014). Studies of South Soufrière Hills Volcano reveal a distinct

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geochemical signatures in comparison to the active Soufrière Hills Volcano. This indicates that isolated melt bodies within the crystal mush result in a heterogenous system, which is vertically and laterally extensive (Devine et al., 2003; Edmonds et al., 2016).

As well as being chemically heterogenous, the Soufrière Hills system is thought to be composed of layers of mush (solid and crystalline), liquid (sub-solidus, eruptible magma) and exsolved fluid layers (volatiles exsolved from the magma on saturation; Christopher et al., 2015; Edmonds et al., 2016). The presence of fluid layers with exsolved fluids, resulting from magma recharge, may lead to instabilities in the mush system and may result in overpressures that could fuel explosive eruptions (Christopher et al., 2015; Edmonds et al., 2016). The cyclicity observed in the vulcanian eruptions was thought to be related to the ascent of different magma batches. The ascent of magma to the surface was calculated to take 5 – 13 hours, consistent with the 12-hour cyclicity of the Vulcanian eruptions. Geochemistry, petrology (e.g., Christopher et al., 2015; Edmonds et al., 2016) and geophysical studies (e.g., Paulatto et al., 2012) show the mush system extends from <4 to 16 km, with the main active reservoir between 8 – 12 km (Paulatto et al., 2012; Christopher et al., 2015; Edmonds et al., 2016; Fig 2).

3.2 The Central Arc (including St Lucia)

In the Central Arc (Guadeloupe to Dominica) andesites are dominant, with some basalts and dacites (Fig. 3; 4). St Lucia is more commonly included in the Southern Arc based on seismic data (Wadge and Shepherd, 1984), but fits into the Central Arc based on geochemistry and volcanology, so is also discussed in this section.

3.2.1 Guadeloupe

The whole rock composition of La Soufrière de Guadeloupe eruption products have also remained consistent, though they do span a wide range of compositions (51 – 61 wt% SiO₂; Touboul et al., 2007; Boudon et al., 2008; Samper et al., 2009; Ricci et al., 2017; Pichavant et al., 2018; Metcalfe et al., 2022). The 1530 CE sub-Plinian eruption (Komorowski et al., 2008; Boudon et al., 2008; Esposti Ongaro et al., 2020) shows black-and-white banded pumices with characteristic mixing and mingling textures between andesitic (ca. 62 wt% SiO₂) and basaltic-andesite (55 wt% SiO₂) compositions (Semet et al., 1981; Poussineau, 2005; Boudon et al., 2008; Pichavant et al., 2018). Deposits from the Echelle scoria cone also show a bimodal composition, with mixing textures between basaltic andesite and dacitic compositions observed. To the south of La Soufrière de Guadeloupe, the Mont Caraïbes complex erupted a basaltic composition (47 – 50 wt% SiO₂) and is often interpreted to represent the parental magma for La Soufrière de Guadeloupe (Bissainte, 1995; Metcalfe, 2022). Eruption products Mont

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Caraïbes are the least differentiated in the Central Arc, while eruption products from La Soufrière de Guadeloupe are more typical of those observed (Fig 3).

Analysis of MIs found in clinopyroxene, orthopyroxene and plagioclase from La Soufrière de Guadeloupe, and the neighbouring monogenetic cone of Echelle range from 58 – 79 wt% SiO₂ (Fig 4 A; Poussineau, 2005; Boudon et al., 2008; Pichavant et al., 2018; Cooper et al., 2020; Metcalfe et al., 2022). Major elements show no systematic changes across the Holocene eruptions (Poussineau, 2005; Boudon et al., 2008; Pichavant et al., 2018; Metcalfe et al., 2022; Fig 4). MIs hosted in clinopyroxene, orthopyroxene, plagioclase and olivine from the basaltic parental magma from Mont Caraïbes have also been measured for Guadeloupe. These MIs are notably less evolved in comparison to the MIs from La Soufrière de Guadeloupe (49 – 56 wt% SiO₂) and are the least evolved inclusions observed in the Central Arc (Metcalfe, 2022; Fig 4). Guadeloupe xenoliths are reported as plagioclase-rich, isotropic and homogeneous, but texturally diverse. Bissainte (1999) describe gabbroic cumulates and granodioritic enclaves which are interpreted as fragments of a granodioritic basement.

Guadeloupe MIs record variable Cl contents ranging from 0.06 - 0.36 (Fig 4C; Cooper et al., 2020; Metcalfe et al., 2022; Metcalfe, 2022). H₂O generally falls in the range 1.85 - 4.42 wt%, with a maximum of 5.6 wt% H₂O (Fig 4D; Pichavant et al., 2018; Cooper et al., 2020; Metcalfe et al., 2022). The CO₂ contents from Guadeloupe are much higher than those recorded in the Northern arc ranging from <50 – 1500 ppm (Fig 4E; Cooper et al., 2020; Metcalfe et al., 2022). MIs from La Soufrière de Guadeloupe are reported to be bubble-free (Metcalfe et al., 2022).

Most models of the current La Soufrière de Guadeloupe magma storage system have been based on data collected for the 1530 CE eruption (Semet et al., 1981; Touboul et al., 2007; Boudon et al., 2008; Pichavant et al., 2018). The observation of the compositionally banded 1530 CE eruption products imply that the La Soufrière de Guadeloupe reservoir is chemically and physically heterogeneous. However, the relative compositional homogeneity of mineral phases observed in the 1530 CE eruption products is difficult to reconcile with this scenario.

In the absence of any geophysical evidence for a deep long-lived magma reservoir, early interpretations hypothesise the presence of an andesite magma body formed from fractional crystallisation of a basaltic-andesite parental melt (Touboul et al., 2007; Pichavant et al., 2018). The shallow andesite body must be remobilised by a magma, as the andesite body is considered to be homogeneous until remobilisation caused by the arrival of basaltic magma (Semet et al., 1981; Touboul et al., 2007; Pichavant et al., 2018). With the arrival of a mafic recharge magma chemical and physical changes occur in the magma storage zone. A hybrid magma may also be formed through incomplete mixing (Pichavant et al., 2018). The magma mixing and development of instabilities in the reservoir

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can lead to eruption (e.g., Gourgaud et al., 1989; Murphy et al., 1998, 2000; Pichavant et al., 2002). Pichavant et al. (2018) suggest, based on melt inclusion volatile concentrations and experimental data, a minimum depth of the La Soufrière Guadeloupe magma reservoir between 5.6 and 7.1 km (135 – 170 MPa, based on a density of 2450 kg m⁻³; Barnoud et al., 2016), and a maximum depth of 8.5 km given the absence of amphibole phenocrysts in the juvenile eruptive products .

Moretti et al. (2020a) proposed a deep gas supply from a crystal mush conceptual model as used at many similar systems to explain observations and activity (e.g., Cashman et al., 2017). This was refined by Metcalfe et al. (2021, 2022) to a vertically and laterally extensive mush system extending 6 – 9 km in depth (190 – 220 MPa, based on a density of 2450 kg m⁻³; Barnoud et al., 2016), which hosts evolved magmas (Fig 2). Remobilization of this mush system to produce eruptible magma may occur through the injection of mafic magma from depth into the shallow system, resulting in magma mixing. Alternatively, in eruptions where no mixing textures are observed (e.g., 1657 CE), exsolved fluids and heat from magma at depth may allow remobilization with no interaction with a mafic magma. Investigation of major and volatile elements of melt inclusions across several Holocene eruptions shows the composition of the storage zone has remained stable during this time period (Metcalfe et al., 2022). There is some evidence for a deeper storage zone, hosting less -evolved compositions based on the presence of a basaltic parental magma (Monts Caraïbes) and the basalt produced by the Echelle and Citerne scoria cones (Bissante, 1999; Legendre, 2012; Metcalfe et al., 2022). Bissante (1999) calculated crystallization pressures of 400 – 900 MPa for Mont Caraïbes cumulates, which is equivalent to ca. 16 to 30 km depth (Fig 2).

Moretti et al. (2020) show the mush system is the source of heat and deep magmatic fluids that rise and contribute to the hydrothermal system. The heat and fluids supplied by intrusions into the mush system below 5 km are considered to be one possible cause of seismic activity (White and McCausland, 2016) and unrest episodes, as well as the potential cause of some of the more violent and long-lasting historical non-magmatic phreatic activity produced from the long-lived hydrothermal system (Feuillard et al., 1983; Komorowski et al., 2005; Villemant et al., 2005; Rosas-Carbajal et al., 2015, Moretti et al., 2020a).

3.2.2 Dominica

The whole rock composition of eruption products from Dominica ranges from 50 – 67 wt% SiO₂ (Lindsay et al., 2005; Howe et al., 2015; Boudon et al., 2017; d’Augustin et al., 2020), with no large-scale magma mixing or mingling textures reported (e.g., Howe et al., 2015; Boudon et al., 2017; Solaro et al., 2020). The eruption products are relatively evolved within the LAA whole rock trend (Fig 3), which is typical of the eruption products in the central arc. In comparison, the composition of MIs

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hosted in orthopyroxene, clinopyroxene and plagioclase from Dominica shows a narrower range (69 – 79 wt% SiO₂, 0.2 – 4.5 wt% K₂O, 1.1 - 3.9 wt%; Fig 4A, B) and is homogenous, regardless of eruption (Balcone-Boissard et al., 2018; d’Augustin et al., 2020; Cooper et al., 2020; Ostorero et al., 2021). Xenoliths from Dominica are dominated by olivine with well-equilibrated textures (Zibera et al., 2017; Melekhova et al., 2019).

MIs from Dominica record a similar range of Cl contents (0.15 - 0.38 wt% Cl) as observed in Guadeloupe and across the Central arc in general (Fig 4C; Balcone-Boissard et al., 2018; d’Augustin et al., 2020; Cooper et al., 2020). H₂O contents from Dominica are the highest observed in the Central Arc, with two distinct data clusters, one ranging from 4.5 – 7.3 wt% H₂O and the second from 1.4 - 3.3 wt% H₂O (Fig 4D; Balcone-Boissard et al., 2018; d’Augustin et al., 2020; Cooper et al., 2020). The highest CO₂ contents in the LAA are recorded in Dominica, CO₂ is variable but again generally falls into two clusters (3137 – 4012 ppm and <1856 ppm; Fig 4E; Balcone-Boissard et al., 2018; d’Augustin et al., 2020; Cooper et al., 2020). Bubble-free MIs were analysed by Balcone-Boissard et al. (2018) and d’Augustin et al. (2020).

A vertically and laterally extensive crystal mush system is also proposed for Dominica, with three components: an underplating basalt, a basaltic– andesitic crystalline mush with a liquid hybrid layer, and an interstitial rhyolitic melt which may accumulate at the top of the reservoir in a silicic cap (e.g., Howe et al., 2015). This system is hypothesised to have formed through continuous fractionation, accumulation, and disaggregation of the crystalline material of a crystal-mush along with interstitial melt. This is highlighted by crystals with disequilibrium features indicating they have experienced different histories (Howe et al., 2015). Volatile release from the underplating basalt is thought to cause vigorous convection prior to eruption. Mafic phases observed in the Dominica eruption products are remnants of incomplete mixing of basaltic material (Halama et al., 2006; Howe et al., 2015). Following the eruption and the cessation of recharge of the underplating basalt, the liquid remaining in the reservoir will begin to crystallise to ultimately remain in a crystalline state between eruptive periods (Howe et al., 2015).

However, more recently the model of mafic recharge has been discarded for some eruptions (e.g., Layou and Roseau) due to the lack of evidence for mixing in groundmass glasses and melt inclusion compositions (Balcone-Boissard et al., 2018; Solaro et al., 2020). Instead, mixing of different parts of the mush, without the need of a mafic magma, can explain many features observed for the Layou and Roseau eruptions. The system prior to the eruption is considered as a cooling system which is disturbed by a slightly hotter magma of a similar composition a decade prior to eruption (Solaro et al., 2020). This suggests the system is heterogeneous in temperature (Solaro et al., 2019, 2020).

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The large ignimbrite eruptions observed on Dominica (e.g., 60 and 24 ka BP) were fed from magmas stored at ca. 16 km, ca. 400 MPa (based on a crustal density of 2450 g/m³; Balcone-Boissard et al., 2018 and Solaro et al., 2019; Fig 2). In comparison, more recent Plinian eruptions (e.g., 8–18 ka) were fed from magma that was sourced in shallower storage regions ≤ 7.3 km, ca. <200 MPa (based on a mean crustal density of 2800 kg/m³; d’Augustin et al., 2020), this is consistent with the system described for Montserrat (Christopher et al., 2015; Edmonds et al., 2016; Fig 2). The similarities in mineralogical and chemical composition, including comparable volatile element ratios for magma that fed the ignimbritic eruption and the more recent Plinian eruptions indicate a common source. This supports the existence of a transcrustal magmatic system extending from <7 – 16 km (d’Augustin et al., 2020; Fig 2).

3.2.3 Martinique

Most studies on Martinique focus on the homogeneity of the composition of eruption products in the last 13,500 years of activity, though a wide range of compositions are recorded (51 - 66 wt% SiO₂; Gourgard et al., 1989; Smith and Roobol, 1990; Pichavant et al., 2002; Annen et al., 2008). The majority of eruption products appear to be evolved in comparison to the Northern and Southern Arc eruption products (Fig 3). Mafic lavas, mafic enclaves, and mingled/mixed products are also reported in the eruptive history of the volcano (Bourdier et al., 1985; Fichaut et al., 1989a, 1989b; Gourgard et al., 1989). MIs hosted in orthopyroxene and plagioclase from Montagne Pelée, Martinique are evolved in composition and fall into a similar range to MIs from Guadeloupe and Dominica (65 – 78 wt% SiO₂; Fig 4A; Cooper et al., 2020; Martel et al., 1998). Na₂O + K₂O falls in the upper range observed in the central arc (3.1 - 5.0 wt% Na₂O, 1.7 - 4.1 wt% K₂O; Fig 4 B; Cooper et al., 2020; Martel et al., 1998). Martinique xenoliths are comparable to those from St Kitts and are plagioclase-bearing (Cooper et al., 2016).

Volatiles from Martinique MI are reported by Cooper et al. (2020). Cl contents (0.27 - 0.30 wt% Cl; Fig 4C) fall into the range observed in the Central arc. H₂O contents, however, are generally lower than the others systems in the Central arc (<3.3 wt% H₂O; Fig 4D) with one MI recording 5.2 wt% H₂O. CO₂ in Martinique MIs (<381 ppm; Fig 4E) is lower than observed in Guadeloupe and Dominica and is most comparable to St Lucia (Cooper et al., 2020). Vapour bubbles are not reported by Cooper et al. (2020).

The main andesitic magma reservoir beneath Montagne Pelée is inferred using the petrology of erupted rocks as well as results from experimental petrology to be at pressures of 6 – 9 km, ca. 200 ± 50 MPa (based on a density of 2650 kg m³; Martel et al., 1998; Fig 2). Mafic magmas have been experimentally determined to be stored at > 16 km (400 to 500 MPa), suggesting that the magmatic transcrustal system below Montagne Pelée also has a similar deeper component, like magmatic

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feeding systems identified in Dominica and Montserrat (Pichavant et al., 2002; Edmonds et al., 2016). The model proposes that magma rising from a depth > 16 km stalls and forms a deeper storage zone located between 12 - 16 km depth (Pichavant et al., 2018; Boudon and Balcone, 2021).

The presence of mafic magmas with an andesite bulk composition leads to the interpretation of the existence of a zoned magma storage system with a mafic zone below an andesitic zone (Pichavant et al., 2002), similar to that presented for Guadeloupe by Pichavant et al. (2018). Annen et al. (2008) proposed a mush system for Montagne Pelée based on experimental determination of the thermal evolution of the system. The formation of a long-lived persistent magma chamber was determined to be unlikely based on realistic magma injection and cooling rates. Instead, the magma system is built up by accreting sills. This injected magma then completely solidifies, existing as a highly crystalline mush with only a small part (10–20 vol%) remaining as eruptible melt to feed eruptions (Annen et al., 2008). Magmas are calculated to ascend in < 1 day from a shallow reservoir of 6 km (ca. 200 ± 50 MPa) (Martel, 2012).

3.2.4 St Lucia

In the Soufrière volcanic complex of St Lucia no large-scale mixing or mingling textures are reported in the silica-rich deposits (62 – 68 wt % SiO₂; Bezard et al., 2014, 2015, 2017). A wider range of compositions (basalt to rhyolite) are observed in older deposits extending this range to 50 - 68 wt% SiO₂ (18 and 1.1 Ma; Briden et al., 1979; Aquater, 1982; Le Guen De Kerneizon et al., 1983; Samper et al., 2008; Bezard et al., 2017). The most differentiated eruption products in the central arc are found on St Lucia, though generally the data fits with the trends observed in the Central Arc (Fig 3).

Fewer MIs have been studied for St Lucia; the clinopyroxene hosted MIs fall at the more evolved end of the Central Arc compositional range (76 – 78 wt% SiO₂, 3.3 – 3.9 wt% Na₂O, 3.2 – 3.8 wt% K₂O; Fig 4A, B; Cooper et al., 2020). The xenoliths from St Lucia contain abundant amphibole and plagioclase and only scarce olivine and are more comparable in composition to those from the central arc in comparison to xenoliths from the Southern Arc (Melekhova et al., 2019).

Cooper et al. (2020) report volatiles from St Lucia MIs, Cl is relatively low compared to other systems in the Central Arc (0.14 - 0.20 wt% Cl; Fig 4C). MI H₂O content falls into the range observed in the central arc, ranging from 2.0–3.6 wt%. CO₂ content (<600 ppm; Fig 4E) is comparable to that observed in Martinique MIs. Vapour bubbles are not reported by Cooper et al. (2020).

Bezard et al. (2017) present a schematic model for the St Lucia magma system. This model hypothesises three storage zones with fractionation occurring at the base of the crust (>25 km). Storage of basaltic-andesite to andesite magma is hypothesised to occur in the mid-crust, in a chamber at 580 – 720 MPa (ca. 20–26 km depth, based on a crustal density of 2660 kg/m (Christeson et al.,

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2008). A shallow chamber is also hypothesised at, 110 – 210 MPa (4–8 km depth based on a crustal density of 2660 kg/m (Christeson et al., 2008). Magma from the mid-crustal chamber recharges the shallow chamber with mixing hypothesised between the two magmas, shown by resorption textures observed in plagioclase.

3.3 The Southern Arc

3.3.1 St Vincent

Magmas from St Vincent have a narrow range of compositions, with Soufrière St Vincent producing basalt to andesite compositions (47 - 62 wt% SiO₂; Graham and Thirlwall, 1981; Heath et al., 1998; Cole et al., 2019; Fedele et al., 2021; Joseph et al., 2022). The majority of eruption deposits are basalt to basaltic andesite, with andesite mainly found in mixed magma deposits as the more evolved component. For example, banded and mingled textures are observed in the 1902–03 and 1979 eruption deposits (Carey and Sigurdsson, 1978; Graham and Thirlwall, 1981; Heath et al., 1998a). The greatest compositional variation is observed in samples from the 1902-03 eruption (50 - 55 wt% SiO₂; Cole et al., 2019) with magmas from older eruptions generally being more homogenous and more evolved (e.g., 1440 CE and 1580 CE; Cole et al., 2019). Despite the range of compositions, most of the whole rock compositions appear relatively evolved within the LAA whole rock trend (Fig 3).

MIs hosted in plagioclase, orthopyroxene, clinopyroxene and titanomagnetite from Soufrière St Vincent (55 – 77 wt% SiO₂; Heath et al., 1998; Cooper et al., 2020), are more comparable to those from the Central Arc, spanning a similar range to the MIs from La Soufrière de Guadeloupe (Fig 4A). As with the systems in the Central Arc (e.g., La Soufrière de Guadeloupe and Dominica) the MIs from Soufrière St Vincent generally, indicate the system is fed by homogenous magmas with episodic rejuvenation events (Fedele et al., 2021). Devine and Sigurdsson (1982) and Bouvier et al. (2010b) measured MIs in olivine from basaltic-andesite eruption products (43 – 60 wt% SiO₂). St Vincent xenoliths lack orthopyroxene and are rich in olivine (Melekhova et al., 2019).

The majority of MIs from St Vincent record Cl contents ranging from 0.08 – 0.29 wt%, with a few MIs extending the range to 0.61 wt% Cl (Fig 4C; Heath et al., 1998; Bouvier et al., 2010b; Cooper et al., 2020;). H₂O content is variable ranging from 1.50 – 5.3 wt% (Fig 4D; Bouvier et al., 2010b; Cooper et al., 2020), and CO₂ contents are low (76 – 103 ppm), however, this may be due to the small amount of CO₂ data available (Fig 4E; Cooper et al., 2020;). Vapour bubbles are reported by Heath et al. (1998) and Bouvier et al. (2010b), therefore pressures must be considered minimum estimates.

Soufrière St Vincent experiences relatively steady-state behaviour, fed by homogenous magmas initially stored at 7 km, ca. 200 MPa (based on a 2660 kg/m (Christeson et al., 2008)). Petrological and geochemistry studies indicate magmas are stored in a shallow crustal reservoir at ca. 7 km (and

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perhaps as shallow as 3.5 km. ca. 100 MPa (based on a 2660 kg/m (Christeson et al., 2008)). The
magmas evolve and undergo homogenisation, leading to the eruption of chemically homogeneous
magmas (Fedele et al., 2021) (Fig 2). The stored magmas exsolve volatiles which contribute to the
build-up of pressure in the shallow system which may trigger eruptions (Cole et al., 2019; Fedele et
al., 2021). The 2020 -2021 eruption is inferred to have begun with the remaining 1979 high-viscosity,
degassed magma in the conduit being pushed up by the injection of new gas-rich magma (Joseph et
al., 2022). The injection of gas rich magma is thought to have occurred at ca. 6 km based on the depth
of seismic events (Joseph et al., 2022). The displacement of the high-viscosity magma cap by the new
lower viscosity gas-rich magma resulted in an acceleration in extrusion rate and on reaching the
surface experienced explosive fragmentation (Joseph et al., 2022).

However, eruption products from the 1902 -03 eruption and the presence of cumulates indicates the
shallower reservoir experiences episodic perturbations by magma rejuvenation events from a less-
evolved basaltic, parental magma at a deeper level calculated at ca. 13 km, ca. 400 MPa (Fedele et al.,
2021) (Fig 2).

3.3.2 Kick 'em-Jenny

Kick 'em-Jenny has produced the most eruptions in the Lesser Antilles in the 20th and 21st centuries.
Whole rock compositions range from 48 – 59 wt% SiO₂, with an evolution in magma composition
observed (Devine and Sigurdsson, 1995; Camejo-Harry et al., 2019). The 1972 eruption produced
predominantly basalts while the more recent eruptions (e.g., 2013/2014) produced basaltic andesites
(Camejo-Harry et al., 2019). Kick 'em Jenny eruption products are some of the least differentiated
across the arc (Fig 3).

MIs hosted in olivine, clinopyroxene and hornblende from Kick 'em-Jenny also extend to unevolved
compositions (48 – 67 wt% SiO₂; Fig 4A; Devine and Sigurdsson, 1995; Camejo-Harry et al., 2019)
particularly when compared to the Central arc. Kick 'em-Jenny MIs also have a higher Na₂O + K₂O
content in comparison to other systems in the Southern arc (2.2 - 7.2 wt% Na₂O, 0.7 – 4.3 wt% K₂O;
Fig 4B; Devine and Sigurdsson, 1995; Camejo-Harry et al., 2019). MIs record the highest Cl contents
across the LAA, with the majority of Cl contents generally in the range 0.31 - 0.66 wt%, with some MIs
recording up to 0.87 wt% Cl (Fig 4C; Camejo-Harry et al., 2019). H₂O content ranges from 0.2 - 5.0
wt%, which is comparable to H₂O contents recorded in the Southern Arc (Fig 4D; Camejo-Harry et al.,
2019). Kick 'em Jenny MIs record the highest CO₂ contents in the southern Arc, with the majority of
MIs falling in the range <50 – 784 ppm, with two MIs at 2513 ppm and 3157 ppm (Fig 4E; Camejo-
Harry et al., 2019). Bubble-free MIs were analysed by Camejo-Harry et al. (2019).

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Camejo-Harry et al. (2019) present data on the magma storage system for the submarine volcano Kick 'em Jenny, suggesting the presence of a main storage region at ca. 7 – 12 km, ca. 170 - 325 MPa (based on densities of 2500 – 2750 kg/m³; Camejo-Harry et al., 2019), with some shallower networks of melt and mush (Fig 2). Combining petrology, geochemistry and seismic data they show that the magmatic system extends through the crust with fractionation occurring deep in the system. This feeds the shallower system, which becomes increasingly evolved in composition. This vertically extensive system produces eruptions through the build-up of volatiles, generation of overpressure and destabilisation of the system. Periods of dormancy are ended by basaltic recharge at depth, which leads to fractionation and the migration of more evolved melts and the liberation of volatiles and renewed eruptive activity of the system.

3.3.3 Grenada

Eruptions on Grenada produce basalt to basaltic-andesite from scoria cones and explosion craters (Arculus, 1976; Van Soest, 2000; Stamper et al., 2014), with andesite and dacite much less common in comparison to other islands (Arculus, 1976). The most primitive compositions are observed on Grenada (42 – 67 wt% SiO₂; Cawthorn et al., 1973; Arculus et al., 1976; Thirwall and Graham, 1984; Thirwall et al., 1996; Woodland et al., 2002; Rojas-Agramonte et al., 2017), the least evolved compositions can be considered the primitive magmas of the LAA (Fig 3). The basalts in Grenada can be split into two groups: the 'low-Sr' series which are Mg-rich olivine basalts, and the 'high-Sr' series which are clinopyroxene-phyric series (Stamper et al., 2014).

The most primitive MI host in olivine are recorded in eruption deposits from Grenada however, more evolved MIs are also present (42 - 74 wt% SiO₂; Fig 4A; Bouvier et al., 2010a; Cooper et al., 2020). Na₂O +K₂O fall into a range typical of the Southern Arc (1.5 - 4.5 wt% Na₂O, 0.3 – 3.1 wt% K₂O; Fig 4B; Bouvier et al., 2010a; Cooper et al., 2020). Grenada xenoliths are dominated by mafic minerals but have few orthopyroxenes and plagioclase-free samples (Melekhova et al., 2019).

MIs from Grenada record Cl contents in the range 0.04 - 0.33 wt%, with three points in the range 0.54 – 0.67 wt% Cl and fall into the range observed in the Northern arc (Fig 4C Bouvier et al., 2010a; Cooper et al., 2020). The highest H₂O contents in the LAA are recorded in MIs which fall in the range 0.2 - 8.4 wt% (Fig 4D; Bouvier et al., 2010a; Cooper et al., 2020). CO₂ contents are low <50 – 111 ppm and are similar to the other systems in the Northern Arc (Fig 4E; Bouvier et al., 2010a; Cooper et al., 2020). Vapour bubbles are reported by Bouvier et al. (2010a), however, no relation between bubble size and water content is observed.

4 Volcanism, Eruptive Style and Volcanic Unrest Across the Arc

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The LAA has a long history of eruptions with 24 eruptions of varying types occurring since 1900, including 9 from volcanoes on Guadeloupe, Martinique, St Vincent, Montserrat and Dominica and 14 from the submarine volcano Kick ‘em Jenny (Lindsay et al., 2005a; Table 1; Fig. 5). Most islands have a single volcano from which Holocene eruptions have occurred, and which are likely to erupt again in the future (e.g., Saba, Statia, Nevis, Montserrat, Martinique, St Vincent). However, more complicated systems are present with two active volcanoes in Guadeloupe (La Soufrière de Guadeloupe volcano, and the Trois-Rivières-Madeleine volcanic complex; Komorowski et al., 2005; Samper et al., 2009; Komorowski et al., 2013; 2022) and nine potentially active volcanoes in Dominica (Lindsay et al., 2003, 2005c). Since 1900, eruptions have varied in size and style ranging from subaerial phreatic (e.g., Guadeloupe 1956 and 1976-1977 (Le Guern et al., 1980; Feuillard et al., 1983; Zlotnicki et al., 1992; Komorowski et al., 2005; 2018; Hincks et al., 2014); Dominica 1997 (Mayer et al., 2017; ; Global Volcanism Program, 2023) to magmatic (e.g., Montserrat 1995 – present (Wadge et al., 2014) and St Vincent in 2020-2021 (Joseph et al., 2022). Submarine eruptions have also occurred at Kick ‘em Jenny (Devine and Sigurdsson, 1995; Allen et al., 2018); Table 1; Fig 5).

Most volcanoes in the LAA are stratovolcanoes which have often experienced lava domes eruptions, particularly in the Northern and Central arcs (including St Lucia). The Northern and Central volcanoes can produce large explosive eruptions generating pyroclastic flows as well as lava domes and associated block and ash-flows (e.g., Macdonald et al., 2000; Lindsay et al., 2013; Smith et al., 2013; Bezard et al., 2014; Howe et al., 2015). Volcanoes of the Central Arc (Soufrière of Guadeloupe and Montagne Pelée) also have produced numerous explosive Plinian, VEI 4 eruptions (e.g., the P3 79 CE Plinian eruption of Montagne Pelée (Boudon and Balcone, 2021; Carazzo et al., 2020; Michaud-Dubuy et al., 2021), the 1010 CE GDS 3 eruption of La Soufrière (Legendre, 2012; Komorowski et al., 2012; Metcalfe et al., 2021a; 201b) as well as Subplinian VE3 eruptions such as the 1530 CE eruption of Soufrière of Guadeloupe (Boudon et al., 2008; Komorowski et al. 2008; Esposti Ongaro et al., 2020). Larger magnitude VEI 5 Plinian eruptions have also been recorded, for example in Guadeloupe (e.g., the Pintade eruption ca. 42 ka from the Grande Découverte-La Soufrière volcanic complex, Komorowski et al., 2005), Dominica (e.g., the Roseau Tuff; Carey and Sigurdsson, 1980; Solaro et al., 2020) and St Lucia (e.g., the Belfond unit; Lindsay et al., 2013). However, such activity occurred 20,000 to 45,000 years ago, and Plinian eruptions of this scale are not recorded in recent history. As well as stratovolcanoes, the Southern Arc activity has produced a diversity of eruptive phenomena including small explosion craters and lava flows, whereas large dome-forming eruptions are rare due to the magma chemistry (Lindsay and Robertson, 2018).

Some of the most recent sub-aerial, magmatic activity has occurred on the islands of Martinique (e.g., 1902 – 1905; 1929 - 1932), St Vincent (e.g., 1902; 1971-1972; 1979 and 2020 - 2021) and Montserrat

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(e.g., 1995 – present). These eruptions exemplify the multi-phase, transitional activity style that occurs at volcanoes in the Lesser Antilles (Table 1; Fig 5). Here, we discuss some of the most recent and important eruptions in the Lesser Antilles. Therefore, we demonstrate the hazards the volcanoes pose, and the potential impact for the people living on the islands. We also discuss recent unrest at these systems and the implications for future eruptions.

4.1 Soufrière Hills Volcano, Montserrat

Soufrière Hills volcano in Montserrat has experienced a long period of eruptive activity, beginning in 1995 and continuing to the present day. This eruption has provided insights into a range of phenomena regarding andesite dome building eruptions including: [1] cyclicity of magma transport linked with ground deformation, [2] shallow seismicity and eruptive activity, [3] growth and collapse of andesite lava domes, [4] sector collapse and formation of debris avalanches, [5] dynamics of lateral blasts and vulcanian explosions, [6] eruption transitions from effusive to explosive activity triggered by dome collapse, and [7] sub-surface processes and structure of the system, amongst many others.

This eruption is split into five phases with intervals of minimal activity which escalated to the next phase (Cole et al., 2014a). Phases 1 - 3 were long-lived in comparison to Phases 4 - 5 which were shorter but more vigorous (Wadge et al., 2014a). The onset of the eruption was preceded by three years of elevated local seismicity (Robertson et al., 2000; Young et al., 2002; Shepherd et al., 2002; Odbert et al., 2014b) which was followed by an explosive phreatic phase that began on 18 July 1995 and lasted until 15 November 1995. At this point the onset of dome growth was observed marking the transition into the magmatic phases of the eruption.

Phase 1 (15 November 1995 – 10 March 1998) began with a cycle of dome extrusion and Vulcanian explosions leading to dome collapse and pyroclastic density current surges (Voight et al., 1998; Sparks et al., 1998; Cole et al., 1998, 2002; Robertson et al., 1998; Calder et al., 2002; Kokelaar, 2002; Wadge et al., 2014). The explosion on 17 September 1996 produced a column that reached up to 15 km and may have been of Sub-Plinian magnitude and intensity (Robertson et al., 1998). A noticeable pumice deposit was produced with ashfall in the Caribbean region, which was particularly heavy in Guadeloupe. In 1997 activity changed and two series of Vulcanian explosions occurred with a sub-daily cyclic pattern that produced a total of 88 explosions (Druitt et al., 2002b). On 26 December 1997, this phase of the eruption ended with partial edifice collapse of the old pre-eruption crater wall and part of the growing dome. This led to the emplacement of a debris avalanche deposit that reached the South-West coast and triggered a small tsunami with a maximum wave run-up height of 1 m on the western shores of Montserrat (Sparks et al., 2002). The partial edifice collapse also resulted in a sudden decompression at the base of the growing dome, triggering a violent laterally-directed

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explosion. This generated a high-energy, high-velocity, turbulent, dilute pyroclastic density current, and resulted in the most widespread destruction recorded during the eruption (Young et al., 2002; Voight et al., 2002). This first paroxysmal event in Montserrat was followed by an 18 month pause in lava extrusion (Kokelaar, 2002).

Phase 2 (27 November 1999 – 28 July 2003) was preceded by seismic swarms, ash venting and small explosions (Norton et al., 2002). This culminated in pumice-bearing explosions and later initiated the longest period of continuous lava extrusion (Wadge et al., 2014). This phase was characterised by major dome collapse events, with cyclic processes of magma flux and gas flux observed. Dome collapse events in this phase resulted in ash columns up to 9 km high (Carn et al., 2004), which also caused repeated minor ashfalls in the Caribbean region, with much of the collapse material deposited by block-and-ash flows into the sea (e.g., Le Friant et al., 2010). Phase 2 ended with a series of Vulcanian explosions and a total dome collapse of ca. 2×10^6 m³ of lava (Edmonds et al., 2006; Voight et al., 2010b).

Prior to Phase 3 (1 August 2005 – 20 April 2007) a 3.5 month-long series of phreatic episodes was observed, these eruptions resulted in additional fracturing of the shallow regions of the volcanic edifice which facilitated magma transport (Wadge et al., 2014). Dome growth began in August 2005, with the growth rate increasing four-fold by December 2005, allowing rapid growth (Ryan et al., 2010; Wadge et al., 2014). Through the beginning of 2006 dome growth rate and direction varied and by May 2006 a lava spine with an approximate diameter of 50 m had grown. Complete dome collapse, which included the remnants of the Phase 2 dome, occurred in May 2006, producing PDCs. Following this further dome and growth collapses were observed, with magma extrusion varying, some ash venting was observed at the end of 2006 on the rapidly expanding dome. This phase was different to the previous two phases with little explosive activity observed in comparison to Phase 1 - 2 (Wadge et al., 2014).

Phase 4 is split into two periods of activity, 4A (Aug 8th to Oct 8th) and 4B (Dec 2nd to Jan 3rd), with some possible explosive activity preceding this in March 2008. Phase 4A began with escalating seismicity over 48 hours, 12 hours after this seismicity ended, a large explosion occurred producing a 12 km high eruption column and dome collapses (Chardot et al., 2010; Komorowski et al., 2010). Additional Vulcanian explosions that produced pumice fallout deposits and small pumice-rich pyroclastic density currents occurred on 3 December 2008, and 3 January 2009 (Komorowski et al., 2010). Phase 4 was short but was characterised by a high degree of explosivity leading to pyroclastic density currents. Lava extrusion rates were also high and there was a different cyclicity than observed in the previous phases (Wadge et al., 2014).

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Phase 5 (9th October 2009 – 11th February 2010) began with ash venting followed by lava extrusion and pyroclastic density currents. This was followed by large Vulcanian explosions and the termination of phase 5 with a major dome collapse on 11 February 2010 (Stinton et al., 2014b). This generated the second laterally directed explosion of the eruption, which directed high-energy, high-velocity, turbulent, dilute pyroclastic density currents northward. This resulted in significant damage on areas previously affected by pyroclastic density currents in 1997. This paroxysmal event, the last of the ongoing eruption so far, also generated widespread ashfalls in the Caribbean region from Guadeloupe to Saint-Lucia (Komorowski et al., 2011; Cadet et al., 2013). Since 11th February 2010, a pause in the dome growth began, which has continued up to the time of writing, making it the longest pause so far in the eruption.

Overall, the eruption at Soufrière Hills has been characterised by periods of dome-growth, dome collapse and a diversity of Vulcanian to Sub-Plinian explosions, with the eruption becoming increasingly vigorous (Druitt and Kokelaar, 2002; Fig 5). In particular, the cyclicities of this eruption have provided important observations on the nature of eruptions and have allowed a better understanding of the drivers and behaviour of this volcano (Odbert et al., 2014). The cyclicity in eruptive styles and the unique longevity of the eruption, highlight the importance of the continuous monitoring of this system, which remains active and hazardous even during quiescence periods.

Following the end of lava extrusion and the last collapse event which occurred on 11th February 2010, the Hazard Level was lowered from Level 4 to Level 3 and a pause in the eruption began (MVO Scientific Updates, 2010-21). This followed the previous pattern in the eruption where activity was minimal between eruption phases. However, it is unclear how long this pause in activity will last. Through 2010 seismicity and SO₂ emissions continued, as well as minor incandescence on the dome related to hot gasses from a deep source passing through fractures. The observed deformation pattern showed slow inflation of the volcano, with ash venting and degradation of the lava dome observed along with occasional, small pyroclastic flows. This low level of activity continued into 2011, with the Hazard Level reduced again to Level 2 in November 2011 (MVO Scientific Updates, 2010-21).

Activity remained at a low level in 2012 and was characterized by spots of incandescence on the lava dome and continued fumarolic activity and seismic swarms. Peaks in activity were often associated with ash venting, with significant ash venting observed in March 2012, resulting in the formation of a new crater. Large VT events associated with the venting are interpreted to be related to increasing pressure below the volcano from uprising magma, while the ash venting was interpreted to be phreatic in origin (MVO Scientific Updates, 2010-21).

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This low level of activity continued through 2013 and 2014 with periods of elevated seismicity and gas emissions. However, no changes to fumarolic activity or the dome were observed and in August 2014 the Hazard Level was lowered again to Level 1 (MVO Scientific Updates, 2010-21). Since 2017, activity has remained low and consistent with short-lived seismic swarms and SO₂ output generally at 200 - 400 tons per day, though some short-lived peaks in seismic activity and SO₂ output do occur (MVO Scientific Updates, 2010-21). In recent years, lahars, pyroclastic flows and rockfalls pose the most significant threat, particularly following heavy rainfall. Some larger earthquakes continue to be felt by the population, however, most felt earthquakes are now unrelated to Soufriere Hills Volcano (MVO Scientific Updates, 2010-21). The monitoring data is consistent with the previous pauses in activity observed during this eruption, with escalation in activity possible as the eruption has not been declared to be officially over. For the eruption to be considered over, the seismic activity, gas emissions and deformation must all return to substantially lower levels than have so far been observed.

4.2 La Soufrière de Guadeloupe

Though La Soufrière de Guadeloupe has not experienced a magmatic eruption since 1657 CE, the 1976 – 1977 phreatic eruption demonstrates the impact a limited VEI 2, non-magmatic, explosive eruption can have (Komorowski et al., 2005; Legendre, 2012; Komorowski et al., 2012; Rosas-Carbajal et al., 2016; Metcalfe et al., 2021a, b; Komorowski et al., 2022). This eruption also highlights the importance of clear, effective and efficient crisis response management, and risk reduction plans. The eruption posed a significant risk to the population of Guadeloupe and resulted in severe socio-economic difficulties associated with the evacuation of >70,000 people from Basse-Terre, and had a lasting influence on the population of Guadeloupe (Le Guern et al., 1980; Feuillard et al., 1983; Komorowski et al., 2005; Hincks et al., 2014; Chenet et al., 2014; Komorowski et al., 2015). One year prior to the 1976 - 77 event, volcanic seismicity began to increase with three distinct earthquake swarms (Feuillard et al., 1983; Komorowski et al., 2005; Hincks et al., 2014). Despite this increase in seismicity, no large changes in fumarolic activity were observed (Dorel and Feuillard, 1980; Feuillard et al., 1983). One month prior to the eruption a minor landslide occurred and new fractures formed (Feuillard, 2011). Between July 1976 and March 1977, 26 phreatic explosions occurred, which changed the mechanical state of the dome through the opening of new major fractures and widening and deepening of existing fractures and craters.

Some controversy has surrounded the scientific management of this eruption and the resulting crisis response decisions made by authorities (Feuillard, 2011; Komorowski et al., 2005; Beauducel, 2006; Hincks et al., 2014; Chenet et al., 2014; Komorowski et al., 2015; Komorowski et al., 2018). Particularly concerning the report of fresh, unaltered, glassy andesite which was taken as evidence for the

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presence of new magma close to the surface and that the eruption was escalating rapidly (Marinelli, 1976; Brousse et al., 1977; Heiken et al., 1980). This observation had a major influence on the scientific management of the crisis, though ultimately the material was identified as non-juvenile clasts removed from a pumice fallout deposited during the 1530 CE eruption (Boudon et al., 2008; Komorowski et al., 2008).

There is also some debate on the cause of this eruption with two models proposed. The more widely accepted model relates the phreatic eruption to a ‘still-born’ magmatic event (Feuillard et al., 1983). Evidence for this model is provided by the depth of seismicity at ca. 6 km, consistent with the depth of the magma reservoir and SO₂ and Cl signatures which were magmatic in origin. The migration of earthquake hypocentres from depth towards the shallower system over a few hours was also observed and considered to support this model (Feuillard et al., 1983; Villemant et al., 2005; Boichu et al., 2011; Li et al., 2016). Ruzié et al. (2012) suggest new magma is being continuously injected into a shallow magma reservoir at ca. 6 km based on modelling of the flux of noble gas isotopes (⁴He/³He; R/Ra) in hydrothermal fluids and on petrological analysis of the 1530 CE eruption deposits (Semet et al., 1981; Pichavant et al., 2018). Magmatic gases are buffered by the hydrothermal system but an increase in heat flux and migration of chemically reactive fluids in the hydrothermal system are responsible for the sealing of the host-rock and resulting pressurization of the system leading to the phreatic activity observed. The injection of magma into the reservoir is interpreted as the triggering source of several unrest episodes in La Soufrière de Guadeloupe’s history including new pulses in 1992, 1997-1998 (Villemant et al., 2014) and 2018 (Moretti et al., 2020).

However, Zlotnicki et al. (1992) proposed that new magma arrival is not required for a phreatic eruption to occur. Therefore, hydrothermal activity alone, with deposition of impermeable clay minerals that seal of fractures combined with high fluid circulation can result in pressurization of aquifers and phreatic eruption. Moune et al. (2022) also discuss this model for generating phreatic eruptions in the context of the more recent unrest. Independent of the exact mechanism required to generate the phreatic event of 1976-77, all gas data and the hydrothermal fluid C, He and Cl isotopic ratios indicate a persistent magma reservoir degassing and supplying fluids and heat to the shallow hydrothermal system (Brombach et al., 2000; Ruzié et al., 2012; 2013; Allard et al., 2014; Jean-Baptiste et al., 2014; Villemant et al., 2014; Tamburello et al., 2019; Inostroza et al., 2022a, b; Moretti et al., 2020a, b; Moune et al., 2022).

Following the 1976-77 eruption, monitoring of this volcano was increased to include continuous seismic and geodetic monitoring (using global navigation satellite system; GNSS), gas plume monitoring, meteorological networks, periodic extensometric surveys of lava dome fractures, fumarole gas and thermal spring sampling and thermal monitoring (OVSG-IPGP, 1999–2023;

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Komorowski et al., 2005; Jacob et al., 2005; Beauducel, 2006; Villemant et al., 2005; 2014; Moretti et al., 2020a,b; Jessop et al., 2021; Moune et al., 2022; Komorowski et al., 2022). This allowed decreasing activity to be observed from 1984, with minimal activity observed through the 1980s until May 1992 when seismic swarms and a new degassing phase began with summit fumarole reactivation (Zlotnicki et al., 1994; Komorowski et al., 2001; Komorowski et al., 2005; OVSG-IPGP, 1999-2023). Over the next five years fumarolic activity extended at the summit, and then more recently along its eastern flanks, and to the top of the lava dome with the reactivation of three fumaroles (OVSG-IPGP, 1999-2021; Komorowski et al., 2005; Tamburello et al., 2019; Jessop et al., 2021). Late 1997 and early 1998 marked the onset of highly acidic chlorine-rich fluids degassing, which has continued till present day (Komorowski et al., 2005; OVSG-IPGP, 1999-2023; Villemant et al., 2014; Moretti et al., 2020a,b).

Seismic activity is characterised by numerous, low magnitude ($M < 1$), high-frequency volcano-tectonic earthquakes which have been observed to increase in frequency and rate with twenty felt volcanic earthquakes since 1992, including ten since 2013 (Komorowski et al., 2005; Moretti et al., 2020a). Between April 2017-December 2018 unrest reached the highest level since 1976-77 (Moretti et al., 2020a), with a peak in seismicity observed on 27 April 2018 when a M 4.1 earthquake was felt by the Basse-Terre population, the largest earthquake in 42 years (OVSG-IPGP, 1999-2023; Moretti et al., 2020a).

Changes in gas chemistry were also detected before and after the April 2018 earthquake. Low $\text{CO}_2/\text{H}_2\text{S}$ and $\text{SO}_2/\text{H}_2\text{S}$ ratios were observed until November 2017 when a strong chemical perturbation occurred. Perturbations at the time of the earthquake resulted in high $\text{SO}_2/\text{H}_2\text{S}$, CO_2/CH_4 , $\text{H}_2/\text{H}_2\text{O}$ and CO/CO_2 and He/CH_4 recorded by MultiGas and Giggenbach bottles (Moretti et al., 2020a,b; Moune et al., 2022). From June to December 2018 a decrease in $\text{SO}_2/\text{H}_2\text{S}$ and $\text{CO}_2/\text{H}_2\text{S}$ ratios was then observed. These ratios indicate a pulse of fluid with deep hydrothermal/magmatic origin heated and pressurised the hydrothermal system (Moretti et al., 2020a).

This episode period of accelerating unrest is interpreted as a failed phreatic eruption triggered by the ascent of heat and fluids from the deep hydrothermal and magmatic system to shallow depths (Moretti et al. 2020). Two hypotheses are formulated by Moretti et al. (2020a): [1] The deep magmatic fluids, stored at mid- to low-crustal depths in the mush system, generated an upstreaming of magmatic gases through the mush system and the crystallizing magma reservoir. This process leads to cyclic rejuvenation of the magma and exsolution of dissolved volatiles that are released into the hydrothermal system and cause heating and pressurisation. [2] Episodic fresh injections of basaltic magma from mid-crustal depths into the shallow magma reservoir of La Soufrière de Guadeloupe, located between a depth of 5.1 and 8 km triggers the pulsatory ascent of high-temperature and high-

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pressure magmatic fluids into the hydrothermal system. This causes heating and pressurisation that may exceed the tensile strength of the rocks and lead to a phreatic eruption.

However, for both hypotheses, Moretti et al. (2020a) show that the multiparameter unrest data recorded prior to and after the M 4.1 earthquake, indicate that the high degree of fracturing and permeability of the dome allowed fluid pressure to be released laterally, on pre-existing deep tectonic discontinuities of the upper regions of the Grande-Découverte-La Soufrière volcanic complex (Moretti et al., 2020a). Moreover, Moretti et al. (2020a) concluded, regardless of which of these hypotheses is more realistic, the reoccurrence of unrest sequences could escalate to a magmatic eruption. This scenario would result from the presence of either; a rejuvenated magma at shallow depth (ca. 3 km below the summit) resulting from stalled small intrusions (Villemant et al., 2014), or injections of fresh undegassed magma from depth resupplying the shallow magma storage zone of the upper mush system.

Measurements of temperature and mass flux at fumaroles and thermal springs and ground temperature measurements shows that between 2010 and 2020 widespread ground heating of the summit occurred with an increase of the heat output from 0.2 ± 0.1 MW to 5.7 ± 0.9 MW (Jessop et al., 2021). The total heat flux density has also increased over the 2010 – 2020 period from 326 ± 69 W/m² - 403 ± 24 W/m², indicating an increase in thermal intensity at the summit. These changes are related to the spread of fumarolic sites over the dome and also the propagation of thermal anomalies on the summit. The observed increases in temperature, mass and heat flux correlate well with the observed accelerating seismic and degassing unrest of La Soufrière de Guadeloupe over the last decade. On the basis of this measured heat loss per unit (heat flux density, W/m²), La Soufrière de Guadeloupe is one of the most intense heat emitters globally among volcanic domes with an active hydrothermal system (Jessop et al., 2021).

Looking more closely at the long term unrest at La Soufrière de Guadeloupe, particularly at the rate of fracture opening and evolution of the permeability of the system through degassing and gas ratios, could be interpreted as cycles of sealing and reopening of the hydrothermal system (Moune et al., 2022) For example, in March 2017 during a phase of increased permeability, liberation of native S and sulphate and resulting increase of H₂S degassing and a decrease in the CO₂/H₂S ratio could indicate the release of the seal which prevented escalation to a phreatic eruption (Zlotnicki et al., 1992; Moune et al., 2022). Despite this, the cycle of sealing and reopening of fractures and transfer of heat from a lower aquifer, to an upper one (with high fluid circulation) raises concerns for a future phreatic eruption (Zlotnicki et al., 1992; Moune et al., 2022). As the shallow hydrothermal system is hosted in the hydrothermally altered core of the dome (Heap et al., 2021a; 2021b; 2022a; 2022b), any pressurisation may reduce friction and increase pore pressure along mechanically weak pre-existing

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structural discontinuities. This may lead to a partial edifice collapse with potentially devastating consequences (e.g., Peruzzetto et al., 2019; Heap et al., 2021b). Such edifice collapses could occur without the involvement of magma, as observed in the La Soufrière de Guadeloupe stratigraphic record (Komorowski et al., 2005; Rosas-Carbajal et al., 2016). These collapses are associated with moderate laterally directed explosions (up to VEI2) that can generate turbulent pyroclastic density currents of limited extension (< 5 km) as described for the 1797-1798, 1836-1837, and 1976-1977 phreatic eruptions (Komorowski et al., 2005; Rosas-Carbajal et al., 2016).

4.3 Dominica

Dominica has nine active Holocene volcanic centres spread across the island which makes it one of the highest concentrations of active volcanoes in the world (Lindsay et al. 2005; Mayer et al., 2017). These volcanoes have produced the highest erupted magma volume in the LAA (40 km³; Wadge 1984; 1986; Macdonald et al., 2000; Di Napoli et al. 2014; Mayer et al., 2017). Only one recent eruption is recorded on Dominica, but unrest in the form of seismicity and hydrothermal activity has been recorded for the nine volcanic centres. The Valley of Desolation-Boiling Lake area is one of the more active centres and experiences vigorous hydrothermal activity (Sapper 1903; Robson and Wilmore 1955; Lindsay et al., 2005; Mayer et al., 2017) and is monitored regularly by UWI-SRC.

The two most recent eruptions on Dominica have occurred from the Valley of Desolation-Boiling Lake area in 1880 and 1997, with neither eruption witnessed (Lindsay et al., 2005; Fig 1). The 1880 eruption was a large phreatic eruption with ash covering 55 km², resulting in a slight ash fall 19 km out to sea and the generation of lahars (Nicholls et al., 1880b). In comparison, the phreatic eruption occurring on 8th – 9th July 1997 was very small magnitude, however, as this eruption was unobserved little is known about the dynamics. Subsequent investigation reported non-altered lava, hydrothermally altered igneous minerals and hydrothermal minerals present (Komorowski and Hammouya, 1998). This eruption may have been triggered by the occurrence of several landslides. Fumarolic vents were observed to be covered by landslide deposits, which has been hypothesised to have resulted in pressurization of the hydrothermal system and produced a phreatic eruption (James, 1997; Komorowski and Hammouya, 1998; Lindsay et al. 2005; Mayer 2017). The Boiling Lake itself experiences considerable water level and temperature variations and experienced a small explosion of CO₂ and other gases in 1901, killing two people (Ober 1880; James 1988).

Activity at the other complexes varies greatly, with the majority of seismic and hydrothermal activity occurring on the south of the island (Lindsay et al., 2005; Fig 1). The Plat Pays volcanic complex, which is mainly comprised of lava domes, experiences a high number of seismic swarms and fumarolic activity, this complex was active as recently as 450 years ago. Also, in the south of the island are the

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Morne Anglais and Morne Watt complexes which have a history of explosive eruptions and experience regular seismic swarms (Lindsay et al., 2005; Fig 1). One of the potentially most dangerous complexes is the Wotten Waven complex, which has a history of large explosive eruptions and produced the Roseau Tuff. This complex experiences hydrothermal and fumarolic activity and a future explosive eruption has the potential to affect most of Dominica (Lindsay et al., 2005).

In the north of the island the Morne Diablotins and Morne Aux Diabes complexes also experience considerable periods of unrest (Fig 1). Though no historical eruptions are recorded from these complexes, and the majority of seismicity occurs in the south of the island, several periods of increased seismic activity have been recorded in 1841, 1893, 2000 and 2003 (Lindsay et al., 2005).

Finally, the Trois Pitons and Grand Soufriere Hills complexes are the least active complexes and have no evidence for recent eruptions (Lindsay et al., 2005; Fig 1). However, they are poorly studied and do still experience periods of increased seismic activity. The eruptive record is poorly understood, particularly at Grand Soufriere Hills, and until new data is available on the eruptive history, the complex is considered active (Lindsay et al., 2005).

All of Dominica's volcanoes are at the green volcanic alert level and continue to be monitored (UWI SRC Scientific Updates, 2001-22). A background level of seismicity is recorded with occasional swarms of elevated seismicity which, though significant, are considered part of the normal pattern of activity (UWI SRC Scientific Updates, 2001-22). Volcanic unrest has in the past been observed in northern (Morne aux Diabes) and southern Dominica (Plat Pays Volcanic Centre). Swarms occurring in the southern complexes, particularly Morne Plat Pays and Morne Anglais, which are the youngest volcanoes on the island, are particularly significant as this is considered the most likely site of the next eruption (UWI SRC Scientific Updates, 2001-22).

4.4 Montagne Pelée, Martinique

The eruptive record of Montagne Pelée shows a history of sub-Plinian to Plinian and dome-forming eruptions (Westercamp and Traineau, 1983; Carazzo et al., 2012; Michaud-Dubuy, 2019; Michaud-Dubuy et al., 2019; Boudon et al., 2015; Boudon and Balcone-Boissard, 2021). Many of the 8 Plinian eruptions of the last 9000 years at Montagne Pelée (e.g., P1 eruption at ca. 1350 CE; P2 eruption at ca. 330 CE, P3 eruption at ca. 79 CE) are thought to be associated with 20 – 30 km high columns which underwent collapse and produced pyroclastic density currents (Carazzo et al., 2012, 2019, 2020; Michaud-Dubuy et al., 2019). At least 5 of the numerous phases of eruptions in the last 2000 years are associated with lava dome growth and collapse (Carazzo et al., 2019, 2020; Michaud-Dubuy, 2019; Boudon and Balcone-Boissard, 2021) along with violent laterally directed explosions at the base of the dome. These explosions produced devastating, highly energetic and mobile pyroclastic density

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currents with a laterally spreading, dilute flow, as well as valley-channelled denser flows (Lacroix, 1904; Perret, 1935; Boudon et al., 2005; Boudon and Balcone-Boissard, 2021) as exemplified by one of the world's most devastating eruptions that occurred on 8th May 1902.

Prior to this eruption, a repose period of three centuries was observed. This was then interrupted by two short-lived phreatic eruptions in 1792 and 1851-52 (Tanguy, 1994; Boudon et al., 2005). Following these minor eruptions activity, a new period of unrest began in 1889 in the form of renewed fumarolic activity reaching a peak in 1902 (Chrétien, 1983; Chrétien and Brousse, 1988; Boudon et al., 2005). Seismic activity also increased, inferred from the breaking of submarine cables due to submarine landslides and mudflows and the increasing number of earthquakes felt by the population of St Pierre. The onset of eruptive activity began with phreatic eruptions on 23rd – 25th April 1902 which increased in intensity and transitioned to phreatomagmatic activity, perhaps as early as the 2nd or 3rd May. Activity transitioned to magmatic activity on the night of 5th May, and a large glowing dome was seen on the night of 6th – 7th May 1902, and continued to grow rapidly (Lacroix, 1904; Tanguy, 1994; Tanguy, 2004; Boudon et al., 2005; Boudon and Balcone-Boissard, 2021).

The rapid rise of pressurized hot hydrothermal fluids and groundwater led to the sudden drainage of the summit crater lake, causing devastating lahars that killed 23 people on 5th May 1902 in St Pierre. Continuing explosive phreatomagmatic activity triggered new lahars during the night of 7 – 8th May that destroyed the town of Le Prêcheur killing another 400 people (Tanguy, 1994). Sudden decompression of the over-pressurised and rapidly growing lava dome resulted in a small vertical column but a large laterally-directed explosion on 8th May 1902 (Lacroix, 1904; Tanguy, 1994; Boudon and Lajoie, 1989; Boudon et al., 1990; Boudon et al., 2005; Boudon and Balcone-Boissard, 2021). This produced a very mobile pyroclastic density current which expanded rapidly and moved across topographic ridges as a high-energy, low-density pyroclastic surge (Boudon and Lajoie, 1989; Boudon et al., 1990; Bourdier et al., 1989). More concentrated block-and-ash pyroclastic flows rich in solid material from the dome were channelled in numerous valleys down to the coast. These high-velocity and highly turbulent surges caused the complete destruction of St Pierre and the death of 28,000 people on 8th May 1902 (Lacroix, 1904; Tanguy, 1994; Tanguy, 2004; Bourdier et al., 1989; Boudon and Lajoie, 1989; Boudon et al., 1990; Boudon et al., 2005; Brown et al., 2017; Gueugneau et al., 2020; Boudon and Balcone-Boissard, 2021).

Following this deadly phase, activity remained high but intermittent with strong vertical eruptions and further concentrated, high-energy, turbulent pyroclastic density currents which again affected St Pierre on 20th and 26th May, 6th June, 9th July and 30th August 1902. The 30th August flow destroyed the towns of Morne Rouge and Ajoupa Bouillon which caused an additional 1500 - 2000 deaths (Brown et al., 2017). During this explosive phase, a vertical Vulcanian explosion on 9th July occurred on the

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lava dome, producing a small volume pumiceous fallout and an eruption column that collapsed to generate a concentrated, channelised pyroclastic density current (Boudon and Lajoie, 1989; Bourdier et al., 1989; Boudon et al., 1990; 2005; Boudon and Balcone-Boissard, 2021). Following this initial explosive phase, the eruption proceeded until 1905 with effusive dome and lava spine growth (Tanguy, 1994; Boudon et al., 2005; Boudon and Balcone-Boissard, 2021). This continued to produce numerous block-and-ash pyroclastic density currents from gravitational dome collapse. The currents were channelled in valleys to the south and south-west of the volcano without overflowing.

This 1902-1905 eruption highlights the long periods of unrest which can be observed preceding an eruption, in this case unrest began in 1889 and continued to build to a magmatic eruption in 1902, 13 years after the onset of phreatic unrest. In comparison, the less destructive eruption of 1929 had only a couple of months of felt unrest, yet followed a similar sequence of events as the 1902 eruption. The 1929 eruption began with phreatic eruptions followed by the main explosive magmatic phase, which was followed by dome building activity and then a decrease in activity until the eruption ended in 1932 (Bourdier et al., 1989). This eruption was less powerful compared to the 1902 eruption, and as a result the impact on Martinique was much less severe (Bourdier et al., 1989).

As with the current Soufrière Hills Volcano, Montserrat eruption (1995 - ongoing), the 20th century eruptions of Montagne Pelée eruption also provides an account of transition in styles with phreatic and explosive and effusive magmatic activity observed during one eruption. Finally, the 1902 eruption in particular, reminds us of the continuing hazards posed by active LAA volcanoes.

Montagne Pelée is currently experiencing a new phase of volcanic unrest which began in late December 2018. This was observed by the network which recorded a swarm of 17 deep, distal, volcano-tectonic (VT) earthquakes (the hypocentral distance is ≤ 25 km, depth between 10-20 km below sea level) as well as a swarm of shallow-depth (down to 4 km depth) volcano-tectonic earthquakes located on the volcano (Fontaine et al., 2021; 2022). At the time the distal VT seismicity had not been entirely recognized as being of volcanic origin. Therefore the shallow-depth VT seismicity was interpreted as evidence for a reactivation of the hydrothermal systems, which is not necessarily associated with the arrival at depth of magmatic fluids.

In November 2020, the seismic unrest accelerated with respect to the rate and energy of shallow-depth, proximal VT seismicity but also, for the first time, of low frequency, volcanic seismicity. Together this indicated the onset of an increased circulation of pressurized hydrothermal fluids. This unprecedented volcanic unrest prompted the Observatoire Volcanologique et Sismologique de Martinique (OVSM-IPGP) to recommend an increase to the yellow volcanic alert level to the authorities in charge of civil protection. This was implemented by the authorities on 4th December

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7 2020 (OVSM-IPGP, 2020-2022). In addition, 41 hybrid and 25 long-period seismic signals have been
8 recorded up to the time of writing on 1st September 2022 (OVSM, 2020-2022; Fontaine et al., 2021;
9 2022).

10 Signs of the current unrest have been detailed in OVSM, 2020-2022 and Fontaine et al., (2021; 2022),
11 these signals are as follows: [1] an ongoing perturbation of the hydrothermal system, perhaps related
12 to the circulation of hot and pressurized fluids of hydrothermal and/or magmatic origin, observed via
13 the seismicity with high and low frequencies content. No felt earthquakes have been reported since
14 the onset of unrest, though the number of VTs is above the base level recorded between 1st January
15 1999 and 1st January 2019; [2] strongly degraded, browned and dead vegetation zones observed in
16 2021 on the southwest flank of the volcano which show a pulsatory increase and decrease in size; [3]
17 a minor deformation from June 2021 to June 2022 is revealed by continuous GNSS data, best modelled
18 as a source of inflation located about 1 km below and slightly SW of the summit, above the
19 hydrothermal system; [4] the detection of an area of underwater gas emission at shallow depths since
20 June 2021.

21 The ongoing unrest features most likely reflect the ascent of a limited volume of deep magmatic fluids
22 which reinvigorated the shallow hydrothermal circulation. Though no surface fumarolic activity is
23 observed, extensive monitoring of this system is still required given the fact that there is no pre-
24 eruptive unrest monitored by a multiparameter network. Although an extended period of 13 years
25 preceded the major eruption of Montagne Pelée in 1902, we cannot expect that this is a characteristic
26 of magmatic eruptions at Montagne Pelée after extensive times of quiescence.

27 **4.5 Soufrière St Vincent**

28 Soufrière St Vincent is the most active subaerial volcano in the Lesser Antilles due to numerous recent
29 eruptions (Table 1). The major historic eruptions of 1440, 1580, 1718 and 1812, and recent eruptions
30 of 1902-03, 1971-72, 1979 and 2020 - 21 are well documented, however, the prehistoric eruptive
31 history is less well known. The recent eruptive history reveals a range of activity from major explosive
32 phases to periods of completely effusive dome growth (e.g., Anderson and Flett, 1903; Aspinall et al.,
33 1973; Shepard et al., 1979; Robertson, 1992; Robertson, 1995; Cole et al, 2019; Fedele et al., 2021;
34 Joseph et al., 2022). Two additional prehistoric eruptions (1580 and 1440 CE) suggest a cyclicity with
35 repose period ranging between 77 and 140 years that appears to be systematically decreasing (Cole
36 et al 2019).

37 The 1902-3 eruption was the deadliest eruption recorded for St Vincent with 1500 deaths (Pyle et al.,
38 2018; Table 1). As with the 1902 Montagne Pelée eruption, unrest prior to the 1902 Soufrière St
39 Vincent began with an increase in seismic activity. However, unlike Montagne Pelée's extended period
40 of unrest, the 1902-3 eruption was followed by a period of quiescence.

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of unrest, increases in seismic activity are only reported from 1901, one year prior to the eruption (though a longer period of unrest, with earthquakes going unfelt by the population, cannot be discounted (Pyle et al., 2018)). Seismic activity culminated in phreatic eruptions which transitioned into a violent magmatic eruption producing pyroclastic density currents and lahars (Anderson and Flett in 1903; Roobol and Smith, 1975; Pyle et al., 2018). The eruption continued with intermittent vulcanian explosions until March 1903 (Roobol and Smith, 1975; Pyle et al., 2018).

In comparison, a completely effusive eruption was observed in 1971-72. This eruption saw the aseismic emplacement of a basaltic-andesite dome, a direct contrast to the major explosive eruption observed in 1902 (Aspinall et al., 1973; Shepard et al., 1979; Robertson, 1992). Explosive activity began 13th April 1979 following elevated seismicity and lasted for 13 days until 26 April. The explosive phase of the eruption consisted of 11 vulcanian explosions that produced eruptions columns up to 18 km. Pyroclastic flows affected several valleys surrounding the volcano, but only reached the sea at Larikai valley. The explosive phase of the eruption was followed by six months of effusive dome growth (Shepard et al., 1979; Robertson, 1995).

The most recent eruption of Soufrière St Vincent began in December 2020, making this the most recent period of volcanic activity in the Lesser Antilles (Global Volcanism Program, 2021; Joseph et al., 2022). This eruption began with an effusive phase on 27th December 2020 with dome growth in the south-west sector of the crater floor, adjacent to the 1979 dome. Increasing seismicity culminated in April 2021 with a transition from effusive to explosive behaviour, following three months of dome growth. On recommendation from the scientists from the Seismic Research Center (UWI Trinidad) the volcanic alert level was raised by the authorities from orange to red on the afternoon of 8th April 2021. This is the highest level on a four-level scale (UWI-SRC Scientific Updates, 2020-21). The increase in alert level was followed that day by the onset of rapid evacuations from the red zone (Joseph et al., 2022). The first explosion occurred on the morning of 9th April and was vulcanian in nature. This was then followed by a pause of roughly 5 hours after which sub-Plinian activity began, lasting until the 11th of April, when the first pyroclastic density currents appeared. This was followed by a period of pulsing, explosive activity that continued for the next 20 hours. Between 9th April and 22nd April 2021, 32 identifiable explosions (using seismicity, satellite imagery and visual observations) occurred with eruption columns reaching up to 16 km. After the initial intense phase of activity, discrete vulcanian explosions continued with decreasing frequency and energy (UWI SRC scientific updates 2020-21; Global Volcanism Program, 2021; Joseph et al 2022).

The explosions were associated with ashfall and pyroclastic density currents with lahars occurring later after heavy rainfall. The explosions produced significant damage to the northern parts of the island with tephra accumulation leading to collapse of buildings and destruction of crops. In May 2021, the

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7 59 alert level was reduced to orange, though the volcano continued in a state of unrest with escalation
8 60 in activity possible with little warning (UWI SRC scientific updates 2020-21). No casualties occurred
9 61 from the eruption but with >16,000 people evacuated from the North of the island, this eruption will
10 62 no doubt have socio-economic impacts on the population of St Vincent.

11 63 Through June and July 2021 activity continued to decline, thermal anomalies continued to be detected
12 64 on the crater floor and were interpreted to be from a small body of magma remaining close to the
13 65 floor of the Summit Crater (UWI SRC scientific updates 2020-21). Along with thermal anomalies, daily
14 66 seismic activity and substantial degassing continued to occur with persistent steam emissions as the
15 67 dominant observable feature, indicating the system was still in unrest. During August 2021 heavy
16 68 rainfall resulted in lahars in most of the valleys on the volcano, showing the volcano still posed
17 69 significant hazards. In September 2021, the alert level was reduced to yellow given that visual
18 70 observations confirmed that no major changes had occurred in the crater since the end of explosive
19 71 activity in April. At the end of 2021 and beginning of 2022 seismic activity continued at a low level
20 72 with no significant deformation recorded since May 2021. In February 2022 plume chemistry showed
21 73 decreasing SO₂ and CO₂ concentrations and by March the levels had become lower than the detection
22 74 threshold, as observed in 2020 pre-eruption.

23 75 The latest update on the volcano was in March 2022 when seismic activity was reported to have
24 76 reached a minimum, no significant change in deformation signal had been observed and no change in
25 77 gas composition has been observed in the plume. This activity is considered to be at levels similar to
26 78 background activity recorded prior to the 2020 – 2021 eruption and the alert level was lowered to
27 79 green.

30 80 **4.6 Kick ‘em Jenny**

31 81 The submarine volcano Kick ‘em Jenny has erupted 14 times in the 20th and 21st centuries, making this
32 82 the most active volcano in the LAA (Figure 5). The most recent eruption occurred in 2017, following
33 83 this eruption the alert level was reduced to yellow (UWI SRC scientific updates 2020). This alert level
34 84 has remained at yellow due to periods of seismic unrest being recorded (UWI SRC scientific updates
35 85 2020).

36 86 Prior to the 2001 eruption, Kick ‘em Jenny eruptions were not associated with precursory seismicity,
37 87 as far as is known (UWI SRC scientific updates 2020). The 2001 eruption marked a change in behaviour,
38 88 with this and the following 2015 and 2017 eruptions associated with precursory seismicity. Following
39 89 the 2017 eruption several periods of high level seismic unrest have been recorded. The most recent
40 90 period of unrest occurred from April to June 2020, which included >1000 volcanic earthquakes with a
41 91 maximum magnitude 3.1 earthquake recorded (UWI SRC scientific updates 2020).

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This period of unrest is interpreted to relate to magma intrusion or pressure changes (due to crystallisation or degassing) (UWI SRC scientific updates 2020). It is also hypothesised the system may have experienced sufficient magma recharge to generate eruptions without further high level seismicity, as observed for the pre-2001 eruptions. However, further periods of seismic unrest without eruption are also considered likely (UWI SRC scientific updates 2020).

5 Discussion

5.1 How Reliable is the Melt Inclusion Data? 5.1 How Reliable is the Melt Inclusion Data?

MIs can undergo significant post entrapment modifications (PEM) which can result in changes to the composition of the MI (e.g., Steele-Macinnis et al., 2011; Wallace et al., 2015; Cannatelli et al., 2016). In order to ensure the MIs are representative of the original melt care must be taken when choosing the MIs and also when processing the data.

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The majority of studies have avoided or discarded MIs with obvious signs of post entrapment crystallization (PEC), cracks or bubbles (e.g., Mann et al., 2013; Cooper et al., 2020; Metcalfe et al., 2022; Supplementary datasheet 2). This ensures the MIs have experienced minimal PEMs and the composition has not been altered. For studies which report PEC, MIs with >10% PEC have been discarded (e.g., Camejo-Harry et al., 2019; d'Augustin et al., 2020). This leaves some MIs in the dataset presented here which have experienced <10% PEC, however, PEC has been shown to have negligible effects on both major and volatile elements for up to 11% PEC (Moretti et al., 2018). (Supplementary datasheet 2). Bubbles can also have a major impact on the volatile content, particularly CO₂ with up to 90% CO₂ sequestered in th bubble (e.g., Moore et al., 2015; Venugopal et al., 2020). In most studies, mis with bubbles were not analysed (e.g., Balcone-Boissard et al., 2018; d'Augustin et al., 2020). In cases where they were analysed, care has been taken to ensure the bubble size and volatile concentration did not correlate and the volatile contents were comparable to MIs without bubbles present (Supplementary datasheet 2).

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Many studies have also included additional checks on the MIs to ensure the MI has experienced minimal PEM. For example, studies which report bubbles in the MIs ensure lower H₂O and CO₂ values are not associated with larger bubbles (e.g., Melekhova et al., 2017). Balcone-Boissard et al. (2018) ensure their MIs were volatile saturated by comparing CO₂ content of melt inclusions with increasing crystallization, shown through increasing contents of incompatible trace elements (Wallace et al., 1999; Blundy and Cashman, 2008). d'Augustin et al. (2020) also show minimum H⁺ diffusion by comparing H₂O content with Li content. The effect of H⁺ diffusion has also been ensured to be minimal

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through analysis of deuterium (e.g., Metcalfe et al., 2022; Bouvier et al., 2010b), with MIs outside the accepted range of δD values discarded. Finally, ensuring the MIs fall in the same liquid line of descent can provide assurance that there has been minimal PEM affecting the inclusions.

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Corrections and checks to the MIs have been applied by: Toothill et al., 2007, Bouvier et al., 2010 a,b, Melekhova et al., 2017, Balcone-Boissard et al., 2018, Camejo-Harry et al., 2019, Cooper et al., 2019 d'Augustin et al., 2020, and Metcalfe et al., 2022 (Supplementary dataset 2). However, for the Martel et al., 1998, Mann et al., 2013 and Cooper et al., 2020 may have a higher error on pressure estimates (Supplementary dataset 2). Based on the checks performed by the studies included here, and the fact that MIs have minimal PEC and are not associated with cracks or Fe-Ti oxides we can be confident that the MI dataset presented here have experienced minimal PEM. This indicates the MIs provide good preservation of the original major and volatile composition.

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5.22 Controls on Along Arc Variations

Slab structure controls are also hypothesised to influence the central portion of the LAA. Slab age does not vary greatly along strike (Cooper et al., 2020; Braszus et al., 2021) and instead: (1) sediment input (Plank and Langmuir, 1998; Carpentier et al., 2008), (2) subduction of the Tiburon and Baracuda bathymetric ridges (Bouysse and Westercamp, 1990) and, (3) subduction of strongly hydrated fracture zones (Schlaphorst et al., 2019; Cooper et al., 2020; Bie et al., 2022) have been hypothesised to control along arc variations (e.g., Allen et al, 2019). The increased presence of water in the subduction zone lowers the melting point of the rocks promoting melt genesis. Fracture zones have recently been proposed to have a primary control on along arc variations in various subduction zones (e.g., Venugopal et al. 2020), but also in the LAA, in particular the Marathon and Mercurius fracture zones which underlay the Northern and Central LAA (Müller et al., 2019; Cooper et al., 2020; Bie et al., 2020; Fig. 6).

Cooper et al. (2020) demonstrated this through $\delta^{11}B$ studies of melt inclusions across the arc with higher $\delta^{11}B$ values observed in the Central Arc (Fig 6) related to fluids from serpentine dehydration. In comparison, $\delta^{11}B$ values in the Northern and Southern arc are related to fluids from altered oceanic crust and sediment dehydration. The higher $\delta^{11}B$ values in the Central Arc coincide with the location of the Marathon and Mercurius fracture zones subducted beneath the Central Arc. These fracture zones most likely host the serpentine and supply substantial volumes of fluid to the Central Arc (Fig 6). This suggests that the along arc compositional variation and magmatic productivity is controlled by the distribution of subducted serpentine (Cooper et al. 2020). In turn, these variations, will result in

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variations in magma storage due to changes in density contrasts and influence the ascent of the magma, volatiles exsolving will reduce the density further until the neutral buoyancy point is reached. This is particularly important when considering the Central Arc which has the highest magma productivity, shallowest storage depths and the largest number of volcanoes in a state of unrest (Fig 6).

The distribution of serpentine in the slab is hypothesised to control magma productivity and crustal thickness (Fig 6). This could also relate to the variation in composition observed along the arc with the Northern and Central (including St Lucia) arcs erupting generally more evolved compositions and the Southern Arc producing less evolved compositions and dominated by basalts and basaltic andesites (Fig 6). The correlation between the eruption of evolved magmas and the thicker crust is hypothesised to occur due to basaltic magmas becoming more likely to pond as the crust thickens, allowing further fractionation to more evolved compositions (Baker, 1982; Devine, 1995; Macdonald et al., 2000; Fig 6). This suggests that the homogenous composition produced from a narrow range of shallow storage depths in the central arc, is due to the thicker crust which is allowing for the ponding and fractionation of basaltic magmas (Fig 6).

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Subsequently, magma composition, which is at least partly controlled by density contrasts, crustal thickness, magma productivity and serpentine distribution, is a first order control on eruption style. This is observed in the Central arc which produce evolved compositions and large volume, explosive, caldera eruptions, dome building eruptions and associated voluminous pyroclastic density current deposits from stratovolcanoes (Macdonald et al., 2000; Lindsay et al., 2013; Smith et al., 2013; Bezar et al., 2014; Howe et al., 2015a; Lindsay and Robertson, 2018; Fig 6).

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In contrast, the thinner crust in the Southern Arc which allows the eruption of more basaltic magmas erupting in explosion craters and mafic lava flows, with large dome-forming eruptions being less common (e.g., Arculus, 1976; Brown et al., 1977; Thirlwall and Graham, 1984; Robertson, 2003; Lindsay and Robertson, 2018; Fedele et al., 2021).

5.3.2 Magma Composition, Storage and Controls on Eruption Style

5.3.2.1 Magma Chemistry

The Grenada magmas are the most primitive and are particularly silica-poor and could represent a potential parental magma composition which would fractionate to produce the suites of magma observed across the Arc. The whole rock compositions in equilibrium with a primitive magma are shown on Figure 3, and plot below the FeO/MgO ratio corresponding to olivine Fo_{90} , using $K_d^{ol-liq} = 0.3$ (e.g., Melekhova et al., 2019). Using the compositions of the primitive Grenada whole rock we

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provide two models of fractional crystallisation starting at pressures of 300 MPa (White et al., 2017) using RhyoliteMELTS v1.2.0 (Fig 3; Ghiorso and Sack, 1995; Gualda et al., 2012).

In the first model (A), the starting composition is the average of the most primitive whole rock compositions (FeO/MgO <0.7; Fig 3). In the second model (B), we used the average of the most primitive whole rock compositions with the highest CaO/Al₂O₃ ratio. The same input parameters were used for both starting compositions which included a starting pressure of 300 MPa, a starting temperature of 1200°C (liquidus calculated by MELTS) and an oxygen fugacity buffered at NNO (stamper et al., 2014). The melt was then cooled and decompressed in three iterations due to failed convergence. The iterations were completed at increasing dP/dT (details in supplementary datasheet 32), determined by the decompression and cooling intervals. This simulated fractional crystallisation involving phases including olivine, plagioclase, orthopyroxene, clinopyroxene, amphibole, hornblende and spinel (supplementary datasheet 32). This modelled assemblage is coherent with the observed assemblage in whole rock samples across the LAA.

The fractional crystallisation models predict non-linear liquid lines of descent which both reproduce the general trend of the LAA whole rock data. This indicates the LAA magmas are related to one another by fractional crystallisation, and the differentiation trend observed can be explained by fractionation of a primitive melt with a composition similar to the most primitive Grenada eruption products (Fig 3). However, there is a large spread of data, indicating magma mixing and crustal contamination are also occurring to produce the trends within the individual systems. For example, minor assimilation of crust or sediments is observed in St Lucia and Martinique magmas (e.g., Bezar et al., 2015) and mixing textures observed in Guadeloupe (e.g., Boudon et al., 2008).

The magmas from individual volcanoes are often compositionally homogenous across long lasting eruptions (e.g., Montserrat) and across several eruptions (e.g., Guadeloupe). For the volcanoes which have been most recently active (i.e., those in the central arc, Montserrat and St Vincent) the eruption products are described as having restricted bulk compositions and few systematic compositional changes. Together this indicates the ‘active region’ of the magma mush system being tapped and supplying eruptions has remained consistent in composition and also stable. Indeed, in their study of CO₂-SO₂-H₂S gas emissions from Soufrière Hills in Montserrat, Edmonds et al. (2010) demonstrated that released gases must result from the mixing of shallow volatile-depleted rhyolitic gases with deep volatiles supplied by unerupted underplating mafic magma (basaltic andesite).

Water content across the LAA from melt inclusions varies greatly, with large ranges observed at individual systems (Fig 6; Cooper et al., 2020). Lower water contents are generally observed in melt inclusions from the central arc, with higher water contents observed in the northern and southern

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arcs (Fig 6). However, the fact that melt inclusion water contents can often be affected by processes including: magma differentiation, degassing and post-entrapment crystallization or diffusion, suggests that the observed trend may be the result of additional processes following magma differentiation. Therefore, the highest water content recorded at a given volcano is considered the best estimate of the water content (Rasmussen et al., 2022).

5.3.2 Magma Storage

At many systems along the arc the mush model is now the accepted storage model, with a transcrustal mush system often described. Looking closer at the individual systems, we can hypothesise that across the arc there are generally three levels of the mush storage system which varies in depth across the northern, central and southern arcs. Storage depths are often considered to be the depths at which the magma is at neutral buoyancy (Sparks et al., 2019), and due to the rheological properties of the magma (e.g., composition, crystal content, volatile abundance, gas/liquid ratios and temperature) which drive magma emplacement at different depths (e.g., Burov et al., 2003). However, more recently magma H₂O concentration is has been hypothesised as a major control on the storage depth (Rasmussen et al., 2022). Storage depths have in some cases been shown to be consistent with the predicted depths of water degassing and related processes (Rasmussen et al., 2022). The variations in storage depths in the LAA are ~~also~~ not reflected in magma storage temperature variations (Fig 6), with temperature remaining consistent across the arc. This may be due to the different methods and techniques used across the arc to calculate temperature. Bias towards measurement of one phase (e.g., cpx+liquid) and crystal rims + matrix glass, may exacerbate this and contribute to the homogenous temperature profile (Higgins et al., 2022).

Melekhova et al. (2019) use an inversion approach which combines xenolith petrology and seismic receiver functions to present crustal structure along the LAA. This model shows four crustal layers which vary laterally (Figure 2). The magma system depths presented in the literature generally fit within layers 2 (6 - 15 km) and 3 (12 – 32 km), except for St Lucia where the lower crust system fits into layer 4 (26 – 38 km).

5.3.2.a Lower Crust (ca. >18 km, ca. >400 MPa)

Melekhova et al. (2017) infer a depth of 25 km from primitive cumulate composition of ol+cpx+sp from St Kitts which are in equilibrium with the most primitive magmas. This is in agreement with the estimates for cumulates from other islands (>15 - >30 km; e.g., Bissante, 1999; Parkinson and Arculus, 1999; Stamper et al., 2014; Fig 6). Fractionation occurs to produce magmas which ascend into the middle crustal storage zone. However, as cumulates from this lower storage level are rarely observed

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it is unclear whether this is applicable across the arc. Despite this, this does provide a good estimation for the deepest part of the LAA system.

5.32.2.b Mid Crust (ca. 10 – 18 km, ca. 200 - 400 MPa)

Above the lower crustal zone is a zone of mafic-intermediate magma which, depending on the system, is generally described as basalt to basaltic-andesite in composition and associated cumulates (Tollan et al., 2012; Stamper et al., 2014; Melekhova et al., 2019). The magmas stored here are often described as the parental magmas for the more evolved magmas stored in the shallow system. Again, this mid crustal part of the system is less well studied than shallower parts but is commonly reported between 10 – 18 km (Pichavant et al., 2002; Edmonds et al., 2016; Boudon and Balcone-Boissard 2021; e.g., Melekhova et al., 2017; Balcone-Boissard et al., 2018; Fedele et al., 2021; Fig 6). Within the transcrustal mush system, the parental magmas can migrate into the shallow part of the system and undergo processes including fractional crystallisation and magma mixing. For example, during the most recent eruption of Soufriere St Vincent seismicity indicates the initial activity during the explosive phase (23 - 24 March 2021) involved the shallow system (~6 km). This was then followed by deeper seismicity (~10 km depth; 5-6 April 2021) interpreted as fresh magma migrating from the mid system up into the shallow system (Joseph et al., 2022).

5.32.2.c Upper Crust (ca. <10 km, ca. <200 MPa)

The most well-defined parts of the magma systems across the arc are the shallow zones. The shallow area of the system contains evolved magmas which are often described as homogenous (e.g., Dominica, St Vincent, Guadeloupe; e.g., Fedele et al., 2021; Metcalfe et al., 2022). Magmas from this system can display homogeneity from one eruption to the next, even with considerable repose periods (e.g., Martinique, Guadeloupe). Evolved magmas are reported to extend to the mid-crust in the Northern Arc (down to 15 km), but is consistent in the Central Arc, between 6 – 9 km (e.g., Pichavant et al., 2002; d’Augustin et al., 2020; Fedele et al., 2021; Metcalfe et al., 2022; Fig 6). Less data exists for the Southern Arc, but in St Vincent there is some evidence that the shallow system may extend up to <3.5 km (Fedele et al., 2021). However, we do acknowledge that more data is required in order to conclusively observe this trend.

The upper crust system (<10 km) presented here fits into the layer 2 (6 – 15 km) proposed by Melekhova et al. (2019). This is particularly well defined in Guadeloupe, Martinique and St Vincent (Fig 2). Layer 3 (12 – 32 km) proposed by Melekhova et al. (2019) spans a wide range of depths and encompasses both the mid-crust (10 – 18 km) and lower crust (>18 km) systems presented here. This means, for example, in Guadeloupe the lower crust system fits into the Melekhova et al. (2019) layer

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3 (fig 2). This highlights that the magma storage structure which is defined by magma chemistry and petrology provides a snapshot of the crustal structure along the arc.

5.3.3 Processes Occurring in the Storage Zone

The shallow systems are hypothesised to contain an ‘active region’ with eruptible magma which supplies recent eruptions (e.g., Montserrat; Christopher et al., 2015; Edmonds et al., 2016). The active region of mush within the shallow system is generally hypothesised to be remobilised by recharge from the middle basalt to basaltic andesite region (12 – 18 km). Evidence for this is observed in the textures exhibited in lavas and pumices (e.g., banded pumices from Guadeloupe and enclaves from Montserrat; e.g., Boudon et al., 2008; Wadge et al., 2014), which are present in recent eruption products showing this is an active process in triggering eruptions. However, it is important to note mixing textures are generally not consistent through a system’s eruptive history (e.g., Guadeloupe; Metcalfe et al., 2022). This suggests that though recharge by the parental magma from the middle region may occur it does not always result in the eruption of mixed magma compositions. Instead, this may result in the remobilisation of the shallow system and eruption of homogenous magmas and dominantly more evolved liquids in the form of melt inclusions. Also suggested as a hypothesis for remobilisation of the shallow system, is mixing of magmas with a similar composition (Solaro et al., 2019). Or alternatively, through volatiles and heat released from the parental magma migrating up through the system and resulting in remobilisation in the shallower system, with no need for direct mixing (Christopher et al., 2010; Metcalfe et al., 2022).

However, we cannot discount the effect of the well-developed hydrothermal systems present. The hydrothermal systems present at the active volcanoes, including Montserrat, Guadeloupe, Dominica, Martinique and St Vincent. This often complicates the interpretation of monitoring signals due to the scrubbing of magmatic gases, in particular SO₂, which often buffers and suppresses the magmatic signal (Moune et al., 2022). Recent research highlights the potential importance of the hydrothermal systems in generating eruptions (Zlotnicki et al., 1992; Mayer et al., 2017; Moune et al., 2022).

Phreatic eruptions in particular, which can be generated without any change in the magma recharge zone at depth, can be generated due to the sealing of fractures in the hydrothermal system which increases the pressure in deep aquifers to potentially generate an eruption (e.g., Moune et al., 2022). This has been discussed as a potential cause of the 1976-77 La Soufrière phreatic eruption (Zlotnicki et al., 1992) and given the build-up of hydrothermal pressure observed in 2018, it may have led to a phreatic eruption if the fractured state, hence the high permeability of the dome had not allowed for release of the pressure (Moretti et al., 2020a; Moune et al., 2022).

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The effect of the upper system has also been discussed in the context of magmatic eruptions from La Soufriere Guadeloupe, where the eruption style of magma ascending from the active region of the storage zone is affected by the state of the upper system. Though ultimately the main driver of an explosive eruption is the supersaturation of volatiles in the magma and brittle failure of a magma. Metcalfe et al. (2022) hypothesise that eruption style at La Soufriere de Guadeloupe is controlled by magma ascent rate, with faster magma ascents resulting in larger, more explosive eruptions. Along with magma composition and ascent rate, the state of the hydrothermal system can also influence eruption style and could be responsible for the range of eruption styles produced by each individual system. A slow ascending magma experiencing feedbacks between degassing and crystallization will be expected to erupt with low explosivity or effusively, however, sealing of the hydrothermal system and a reduction in permeability may favour the accumulation of gas, leading to over pressurisation and an explosive eruption (e.g., Cassidy et al., 2018; Metcalfe et al., 2022).

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In order to provide an appraisal of the prevalent degassing process (decompression degassing vs CO₂-infiltration), integrated into a fast and comprehensive visualization of the volcano plumbing systems from their source to the surface, we provide pressure vs CO₂ (gas) diagrams (Fig 7). This has been used in literature to: [1] identify degassing processes such as decompression degassing and CO₂-fluxing (Mangiacapra et al., 2008; Barsanti et al., 2009; Arienzo et al., 2010; Moretti et al., 2013, 2019; Caricchi et al., 2018; Papale et al., 2022), and [2] provide a straightforward petrologic/geochemical image of volcano plumbing systems which integrates geophysical information and highlights depth intervals of magma storage.

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The accuracy of this representation is highly dependent on the performance of the adopted model for volatile melt-saturation, in particular, the model's sensitivity to compositional variations from deep to shallow regions (Barsanti et al., 2009; Papale et al., 2022). Among compositional variations, the redox-induced partitioning of iron between FeO and Fe₂O₃ plays a major role in shifting the equilibrium pressure of H₂O-CO₂ saturation by a factor comparable, or even superior to, that of analytical errors on dissolved H₂O and especially CO₂ contents measured in melt inclusions (Moretti, 2022).

For given fO_2 values we calculated the iron oxidation state of each MI by means of the Ottonello et al. (2001) model extended to $P > 1$ bar and hydrous conditions, as in Moretti (2005). We first computed $Fe^{3+}/Fe_{(tot)}$ of MIs, and therefore FeO and Fe₂O₃ contents, by using guess pressure values. Saturation pressures ($P_{(fluid)}$) from the Papale et al. (2006) model were then used to update $Fe^{3+}/Fe_{(tot)}$ values via the redox model of Moretti (2005) and the new FeO and Fe₂O₃ values were entered into the Papale et al. (2006) computation for pressure. Two to three iterations were sufficient to obtain consistent sets of fO_2 - $Fe^{3+}/Fe_{(tot)}$ - $P_{(fluid)}$ values for the two models combined and solve the mutual effects played by FeO-Fe₂O₃ partitioning on equilibrium fluid pressures at H₂O-CO₂ saturation and the role of dissolved

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water on iron oxidation state (Moretti and Papale, 2004; Papale et al., 2006; Moretti, 2022, Moretti and Ottonello, 2022; Papale et al., 2022 for detailed discussions and examples on different melt compositions including basalt and andesite.

When we begin to consider data from across the arc, the P vs CO₂ data (Fig 7) shows the processes of decompression and fluxing are occurring above the Moho, and between ca. 200 and 100 MPa (ca. 4 - 8 km). The data is spread over the whole H₂O-CO₂ range, giving us an estimate of the main levels of emplacement in the crust. Further spread in the gas composition may occur at this crustal level (ca. 200 -100 MPa, ca. 4 - 8 km) due to magma differentiation and magma mixing. There are two main scenarios: (1) the first scenario is a 'static' scenario, where the data is spread over a large XCO₂ (gas) domain for a pressure interval. This is indicative of magma ponding at a given depth interval where H₂O and CO₂ fractionate due to fractional crystallization or CO₂-driven dehydration processes. (2) the second scenario is a 'dynamic' scenario where we identify paths of magma ascent with either XCO₂ increase (CO₂-fluxing) or XCO₂ decrease (decompression degassing).

The P vs CO₂ data indicates in the Northern Arc, St Kitts and Montserrat MI CO₂ decreases with decompression of the magma across pressures of ca. 240 – 100 MPa (Fig 7A). However, MIs from St Kitts are experiencing CO₂ fluxing, shown by the CO₂ increasing with decompression across the same pressure range of ca. 240 – 100 MPa (Fig 7A). There are also a small cluster of St Kitts MIs at a higher pressure (>300 MPa) which are affected by CO₂ fluxing (Fig 7A).

In the Central arc decreasing with decompression is affecting the CO₂ content in the majority of the MIs, this generally occurs across pressures of ca. <200 MPa (Fig 7B). Martinique and St Lucia MIs experience decompression over a narrower range (ca. <130 MPa) (Fig 7B). The Dominica MIs form a bimodal distribution, the data from Balcone-Boissard et al. (2018) (Fig 7B, highlighted by the red circle), has a very hydrous source and experiences CO₂ fluxing over pressures from 375 – 150 MPa. In comparison, the data from d'Augustin et al. (2020) is comparatively CO₂-rich, with CO₂ decreasing with decompression. The majority of these MIs from d'Augustin et al. (2020) experience decompression at <200 MPa, however, there are some MIs which are deeper at >500 MPa.

In the Southern Arc, decompression and CO₂ fluxing are affecting MIs, generally these processes occur at < 200 MPa, however, Grenada has a deeper source down to ca. 400 MPa (Fig 7C). Though there are only a few MIs data points for St Vincent, MI CO₂ is decreasing with decompression across a narrow pressure range (150 – 180 MPa; Fig 7C), however, with more data this range may be extended. The Kick 'em Jenny MIs show a wide spread of data, which also shows CO₂ decreasing with decompression. There are also two high pressure (ca.600 MPa), CO₂-rich data points which are outliers from the

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distribution (Fig 7C). Grenada has a very hydrous source and experiences CO₂ fluxing over a wide range of pressures (60 – 400 MPa; Fig 7C).

5.4.3 Perspectives and Limitations

Despite the wealth of knowledge and understanding of the LAA there is still much to be learned about each magmatic system and about the LAA as a whole. In particular, we need to improve our understanding of the geometry of the deeper parts of the LAA magma storage systems to provide a more complete picture of this system. Understanding the variations in pressure, temperature and depth of the deep magma storage system may help provide links with the variations in mineralogy and chemical composition observed through the system. This could be done, for example, through seismic imaging across larger areas and over longer seismic lines which would extend the data collected by the VOILA project (refs). In addition, deep penetration multichannel seismic (e.g., Laigle et al., 2013), gravity surveys and magnetotellurics imaging at sea using ocean bottom stations (e.g., Darnet et al., 2020) may help to resolve and provide additional information on the structure of the magmatic reservoir. This should include understanding the lateral extent of the storage systems and ‘active regions’ of mush, with particular focus on Dominica which appears to have a much more laterally extensive system in comparison to the other islands (Balcone-Boissard et al., 2018; d’Augustin et al., 2020).

To understand the role of the parental magmas in the different systems and how this varies along the arc could include refining our understanding of how the parental magma remobilises the shallow system. Though remobilisation by a parental magma has been proven in many cases, there are many eruption products, which show no evidence for mixing a less evolved magma demonstrating the parental magma is not always erupted. This creates a bias in the geochemistry dataset as the evolved magmas from the shallow system are more often erupted, with fewer examples of the parental and primitive magmas. This along with the homogenous compositions observed in magmas erupted from the shallow system requires detailed investigation of eruption deposits to better understand if fine-scale mixing textures are present. This will allow us to better understand how the parental magma interacts with the shallow system and why it is not consistently erupted.

Understanding the volatile processes occurring, and in particular how degassing from various depths across the vertically extended magma system contributes to the gas ratios measured at the surface (e.g., Metcalfe et al., [submitted 2023, Frontiers in Earth Science](#); Moretti et al., [submitted, EGU 2023](#)).

This requires an integration of melt inclusion volatile data and gas monitoring data which is available for many systems which have long-term monitoring. In addition, understanding the presence and accumulation of exsolved fluids in the system as the abundance of exsolved volatiles can affect the

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evolution and growth of the magma reservoir (e.g., Degruyter et al., 2017; Rasmussen et al., 2022). Understanding the relationship between the presence of exsolved volatiles to the depth and growth of the magmatic systems in the LAA, may be key to understanding along arc variations in the magmatic systems.

The majority of eruptive activity in the 20th century has occurred in the Central and Southern arcs, with the only eruption in the northern arc being in Montserrat. The majority of eruptions in this period have been vulcanian eruptions, dome eruptions and phreatic eruptions (Fig 5) and it would be reasonable to assume any future activity will also take this form, with less frequent sub-Plinian to Plinian size eruptions interspersed. However, it is also important to consider the large pumiceous pyroclastic flow deposits of Guadeloupe, Dominica and St Lucia. The Roseau and Layou tuffs in Dominica (Howe et al., 2014), the Belfond Unit on St Lucia (Lindsay et al., 2013) and large Plinian eruptions from Grande-Découverte and Carmichaël volcanoes on Guadeloupe (Komorowski et al., 2005; Legendre, 2012) and though much less frequent in the eruptive record, these large Plinian eruptions are important to consider, as such eruptions would impact the entire region (Lindsay and Robertson, 2018)

Volcano monitoring has shown major advancements over the last fifty years, however, the period of monitorable unrest can vary greatly even within the same system. In the case of St Vincent, the two most recent explosive eruptions have had very short periods of monitorable unrest (on the order of days for the 1979 eruption and 2 months for the 2020 eruption (onset in November 2020; Joseph et al., 2022; UWI SRC scientific updates 2020-21). In the case of the St Vincent 2020 eruption, the 2 months of unrest was followed by an effusive phase of eruption with evacuation not taking place until the explosive phase of eruption had begun (Joseph et al., 2022). Other systems such as La Soufrière de Guadeloupe could also potentially have short periods of monitorable unrest prior to eruption, as shown by diffusion timescales (e.g., Metcalfe et al., 2021 a,b). Diffusion timescales allow the calculation of the time between the last magmatic process that destabilizes the magma in the storage zone and eruption, for La Soufrière de Guadeloupe, this is calculated to be on the order of weeks for even large explosive eruptions at this system (Metcalfe et al., 2021a,b). The short time frames of escalation from monitorable unrest to eruption and then an explosive phase requires efficient and effective monitoring plans, to allow for timely and successful crisis management and evacuation, as was demonstrated by the remarkable crisis response to the Soufrière of St Vincent eruption in 2020 (Joseph et al., 2022).

Much can be learnt from the handling of the 2020 Soufrière St Vincent eruption and the support that has been provided for the evacuated people. However, though people begun to return to their homes a few months after the eruption, some people remain displaced and support had to be continued over

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years after the eruption in order to facilitate social and economic recovery (Joseph et al., 2022). This is highlighted by the 1995 Soufrière Hills eruption on Montserrat, where following the evacuation of >6,000 people and a population reduction from ca. 11,000 pre-eruption to ca. 4,000 post-eruption, only 1000 people have returned (Kokelaar, 2002; Aspinall et al., 2009; Barclay et al., 2019).

As previously discussed, a wide variation in eruption styles is observed across the arc as exemplified by the eruptions in the 20th century (Fig 5). The hazards posed by eruptions need to be clearly conveyed to the populations with volcanoes in a state of unrest, particularly on islands where volcanic risk is not necessarily considered high by the population (e.g., Guadeloupe; Chenet et al., 2014). Though the eruptions occurring in the early 20th century (e.g., Montagne Pelée 1902 and Soufrière St Vincent 1902) resulted in a large number of casualties (>30,000 deaths), advancements in science and technology have allowed a better understanding and monitoring of the volcanoes of the LAA. This has undoubtedly prevented large numbers of casualties in eruptions in the later 20th century, however, in most cases, evacuations of people remain unavoidable (Table 1). In some cases, evacuation becomes long term displacement and results in major socio-economic impacts and in the case of the La Soufrière de Guadeloupe 1976 eruption, a loss of trust in scientists and authorities (Hincks et al., 2014; Chenet et al., 2014; Komorowski et al., 2018). This highlights the need for future crises to be managed with care and transparency to limit the socio-economic impacts and maintain the relationship between scientists, authorities and the population.

6 Conclusions

In this review of the LAA we have focused on the magma storage system, geochemistry and recent eruptive history. We have highlighted investigating trends across the arc can be useful to provide more context for each system.

The distribution of subducted serpentine has a first order control on magmatic productivity (Fig 6), compositional variation and variation in magma storage depth as serpentine can supply substantial volumes of fluid to the arc (Cooper et al., 2020). This is particularly important to consider in the Central Arc where evolved magmas are stored at shallow depths and are likely to erupt explosively in the near future, given the unrest observed at many systems in LAA.

Through this review we show that there appears to be three different levels of magma storage within a vertical and laterally extensive, transcrustal magma mush. The deepest storage area is the least well defined but is most likely at >18 km depth. Above this, at ca. 10 – 18 km are the parental magmas which undergo fractional crystallization to produce more evolved magmas which are stored in the shallow system at ca. <10 km. Within the shallow mush system there is an ‘active region’ which has supplied the most recent eruptions, and is most likely remobilised by magma, heat and volatiles from

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the parental magmas at 10 – 18 km. Eruption products from recent eruptions are often described as having restricted bulk compositions with few systematic compositional changes. This indicates the ‘active region’ of the magma systems supplying eruptions can remain homogenous in composition and also stable over long periods of time, perhaps by buffering recharge from depth.

Relating the sub-surface processes occurring leading up to an eruption, and the monitorable surface observables remains a challenge. Adapting current monitoring strategies to detect, in a timely manner, the entire eruption timeline from the point at which sub-surface processes effectively generate recordable signals on the surface would greatly enhance monitoring and crisis management strategies. However, many uncertainties must be addressed within this including understanding the surface observables in terms of the sub-surface processes, timescales, and dynamics. In addition, understanding the modulation, abortion and escalation of sub-surface processes and the probability it can lead to an eruption are also important and complicated factors to consider for unrest thresholds, alert levels and early-warning systems.

Each island in the LAA has an active volcano with eruptions occurring in the Holocene, and 24 eruptions since 1900. Recent eruptions demonstrate the volcanoes of the LAA can result in thousands of deaths, large scale and long-term displacements and severe economic impacts. Though advancements in science, technology and our understanding of LAA volcanoes has undoubtedly prevented large numbers of casualties in the late 20th and 21st centuries, large evacuations of people remain unavoidable. Long term displacements of people require continued support to facilitate social and economic recovery.

The majority of recent eruptive activity has occurred in the Central and Southern arcs, and though the northern arc should not be neglected, research focus should be based on where the next eruption is most likely to occur. This should include understanding the lateral extent of the storage systems and ‘active regions’ of mush, and as mentioned previously, with particular focus on Dominica which appears to have a much more laterally extensive system in comparison to the other islands. This also provides an area where monitoring efforts should be concentrated, in particular renewed unrest in Martinique and ongoing unrest in Guadeloupe should continue to be monitored carefully. This should include further studies of past eruptions to provide further constraints on the magma storage and behaviour and also careful monitoring of the hydrothermal systems.

Overall, further research on the LAA must continue to be underpinned by the hazardous nature of the system and the tens of thousands of people that can be affected by even a small eruption. Continuing our advancement in understanding of the LAA is essential to continue to improve monitoring techniques, reduce the impact of eruption and provide clear crisis management.

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Data Availability

Our compiled whole rock and melt inclusion data set are available (supplementary material dataset 1), along with the MELTS models (supplementary dataset 32).

Author Contribution Statement

AM collated the data and drafted the manuscript and figures. J-CK and SM supervised AM during the PhD project and fieldwork this paper stemmed from. RM conceptualized the Pressure-CO₂ gas diagram, performed calculations and with AM discussed and interpreted the results. All authors discussed the data, wrote and revised the manuscript and the figures.

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Figure Captions

Figure 1 Schematic map of the Lesser Antilles (adapted from Wadge 1986 and Leone et al., 2019). The Lesser Antilles arc bifurcates at Martinique with the outer Eo-Oligocene arc making up the Limestone

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Caraĩbes and the inner Pliocene and Pleistocene making up the current active arc. The major volcanic centres which are <2 Ma, are split into three areas of the arc: Northern, Central and Southern (Wadge and Shepherd, 1984) including the submarine volcano Kick ‘em Jenny. Also shown are the active volcanoes which have had eruptions since 1270 CE (red triangles) and eruptions in the Holocene (orange triangles). These twenty-one Holocene active volcanoes have the potential to impact tens of thousands of people living in the Lesser Antilles. The photographs show the variation in morphology: St Kitts – the northern flank of Mt Liamuiga; Montserrat – Soufriere Hills Volcano morphology following eruption, with collapse features and lahar deposits; Guadeloupe – the La Soufriere de Guadeloupe lava dome emplaced in 1530 CE (Boudon et al., 2008) and surrounding scoria cones (right of the image); Dominica – view of the lava domes within the Plat Pays Complex; Martinique – the edifice of Mont Pelee; St Lucia – the Pitons which make up the Soufriere Volcanic Centre; St Vincent Soufriere St Vincent prior to the recent eruption, visible is the crater structure in which the 2020 dome grew. Images from: Guadeloupe and Dominica (Metcalf, 2020); St Kitts, St Lucia and St Vincent (UWI-SRC, 2015 – 2022); Martinique (Michaud-Dubuy, 2017), Montserrat (Fournier, 2005; Hickson et al. 2013).

Figure 2 Schematic summary diagram of magma storage across the arc adapted from Melekhova et al. (2019). Storage models from along the arc are superimposed over the crustal structure (see key). Laterally and vertically extensive mush systems are now proposed for many systems across the arc, which can be divided broadly into three levels: (1) >18 km; (2) 10 – 18 km; (3) <10 km. The location of the mush systems, generally fit with the location of crustal layers (e.g., Guadeloupe, Martinique, St Vincent). The majority of research has focused on the shallow and middle systems. Many systems are hypothesised to have an active region of mush which has supplied recent eruptions in the shallow area of the system (<10 km). The shallow system is remobilised by less evolved magmas from >10 km. This highlights the need to refine storage models, particularly at depths greater than 10 km. Magmas may be produced from parental magmas fractionating at >18 km, as hypothesised for St Kitts, Guadeloupe and St Lucia. (Bissante, 1999 ; Melekhova et al., 2015 ; Melekhova et al., 2017 ; Paulatto et al., 2012 ; Christopher et al., 2015 ; Edmonds et al., 2016 ; Pichavant et al., 2018 ; Metcalfe et al., 2022 ; Balcone-Boissard et al., 2018 ; d’Augustin et al., 2020 ; Martel et al., 1998 ; Fedele et al., 2021 ; Camejo-Harry et al., 2019 ; Boudon and Balcone-Boissard, 2021).

Figure 3 Plot of $\text{CaO}/\text{Al}_2\text{O}_3$ vs $\text{FeO}_{(\text{total})}/\text{MgO}$ using whole rock data compilation from across the arc. As well as data from the northern (circle symbols), central (square symbols) and southern (diamond symbol) arcs, we present two MELTS models of fractional crystallisation (see discussion). This shows that the whole rock data is related through fractional crystallisation, with the variability observed related to mixing. (Saba (Defant et al., 2001; Roobol and Smith, 2004); Statia (Roobol and Smith, 2004);

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St Kitts (Baker, 1968, 1980, 1984; Toothill et al., 2007; Higgins et al., 2021); Montserrat (Murphy et al., 2000; Zellmer et al., 2003; Christopher et al., 2015; Plail et al., 2018); Guadeloupe (including La Soufrière and Mont Caraïbes, Toubol et al., 2007; Boudon et al., 2008; Samper et al., 2009; Ricci et al., 2017; Pichavant et al., 2018; Metcalfe et al., 2021, 2022); Dominica (Lindsay et al., 2005; Howe et al., 2015; Boudon et al., 2017; Balcone-Boissard et al., 2018; d’Augustin et al., 2020); Martinique (Gourgand et al., 1989; Pichavant et al., 2002); St Lucia (Wright et al., 1984; Vidal et al., 1991; Lindsay et al., 2013; Bezaud et al., 2015); St Vincent (Graham and Thirwall, 1981; Heath et al., 1988; Fedele et al., 2021; Cole et al., 2019; Joseph et al., 2022); Kick ‘em Jenny (Devine and Sigurdsson, 1996; Camejo-Harry et al., 2019); Grenada (Arculus et al., 1976; Arculus, 1978; Cawthorn et al., 1973; Thirwall and Graham, 1984; Thirwall et al., 1994, 1996; Woodland et al., 2002; Rojas-Agramonte et al., 2017).

Figure 4 Melt inclusion data from across the Lesser Antilles (A) SiO₂; (B) Na₂O + K₂O; (C) Cl; (D) H₂O; (E) CO₂ vs volcano latitude. The most SiO₂ rich magmas are observed in the Central Arc, while the least evolved magmas are found in the Southern Arc. Though the most evolved magmas are in the Central Arc, the most alkali rich magmas are found on Montserrat in the Northern Arc. Volatile content varies greatly across the arc, particularly H₂O, however, magmas from the Central Arc notably more CO₂ rich than the Northern and Southern Arcs. For full melt inclusion dataset see supplementary dataset 1. (Saba (Roobol and Smith, 2004); Stacia (Cooper et al., 2020); St Kitts (Toothill et al., 2007; Melekhova et al., 2017; Cooper et al., 2020); Redonda (Cooper et al., 2020); Montserrat (Devine et al., 1998; Humphreys et al., 2010; Mann et al., 2013; Cooper et al., 2020); Guadeloupe (Poussineau, 2005; Boudon et al., 2008; Pichavant et al., 2018; Cooper et al., 2020; Metcalfe et al., 2022; Metcalfe, 2022); Dominica (Balcone-Boissard et al., 2018; Cooper et al., 2020; d’Augustin et al., 2020; Ostorero et al., 2021); Martinique (Martel et al., 1998; Cooper et al., 2020); St Lucia (Cooper et al., 2020); St Vincent (Devine and Sigurdsson, 1983; Heath et al., 1998; Bouvier et al., 2010b; Cooper et al., 2020); Petite Mustique (Cooper et al., 2020); Kick ‘em Jenny (Camejo-Harry et al., 2019); Grenada (Bouvier et al., 2010a; Cooper et al., 2020).

Figure 5 Timeline showing the distribution of LAA 20th and 21st Century eruption styles. Grey bars indicate years an eruption was occurring in the Lesser Antilles. For each eruption we show the main eruptive styles which occurred and indicate the maximum VEI. Where the data is available, we show the average bulk rock SiO₂. Smaller and less explosive eruptions are more common (e.g., phreatic, vulcanian and dome) than larger explosive eruptions. Submarine eruptions are also common due to the high level of activity observed at Kick ‘em Jenny.

Figure 6 Panels showing: The Lesser Antilles and the location of the Marathon and Mercurius fault zones, 20th century eruptions, crustal thickness, magma storage depth, magma temperature, MI SiO₂ content, MI H₂O content and MI δ¹³B content variation. Crustal thickness is highest in the central arc,

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which coincides with the shallow storage magma zones, the most evolved magma compositions, the highest MI $\delta^{11}\text{B}$ content and the location of the Marathon and Mercurius fault zones.

Figure 7 XCO₂ in the gas phase v pressure for **(A)** The Northern Arc; **(B)** The Central Arc; **(C)** The Southern Arc. From this we can see the process of decompression and CO₂ fluxing occurring across the arc, the direction of these processes are shown by the arrows. We highlight that in Dominica we see both process of CO₂ fluxing (increasing CO₂ with decreasing pressure) and decompression (decreasing CO₂ with decreasing CO₂). For full melt inclusion dataset see supplementary dataset 1. (Saba (Roobol and Smith, 2004); Statia (Cooper et al., 2020); St Kitts (Toothill et al., 2007; Melekhova et al., 2017; Cooper et al., 2020); Redonda (Cooper et al., 2020); Montserrat (Devine et al., 1998; Humphreys et al., 2010; Mann et al., 2013; Cooper et al., 2020); Guadeloupe (Poussineau, 2005; Boudon et al., 2008; Pichavant et al., 2018; Cooper et al., 2020; Metcalfe et al., 2022; Metcalfe, 2022); Dominica (Baalcone-Boissard et al., 2018; Cooper et al., 2020; d’Augustin et al., 2020; Ostorero et al., 2021); Martinique (Martel et al., 1998; Cooper et al., 2020); St Lucia (Cooper et al., 2020); St Vincent (Devine and Sigurdsson, 1983; Heath et al., 1998; Bouvier et al., 2010b; Cooper et al., 2020); Petite Mustique (Cooper et al., 2020); Kick ‘em Jenny (Camejo-Harry et al., 2019); Grenada (Bouvier et al., 2010a; Cooper et al., 2020).

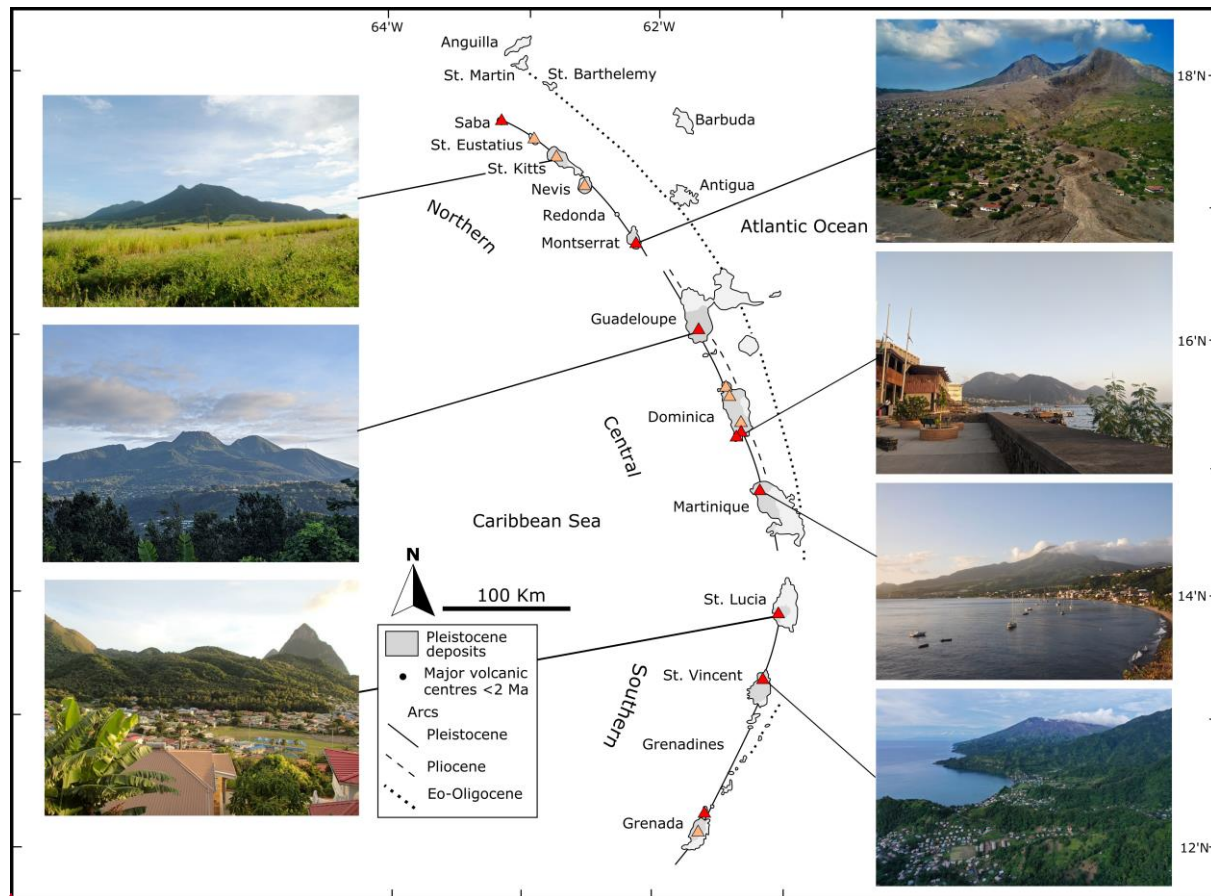
Table 1 Sub-aerial eruptions in the LAA occurring in the 20th and 21st centuries (Lacroix, 1904; Jolivet, 1958; Aspinall et al., 1973; Shepard et al., 1979; Feuillard et al., 1983; Fournier et al., 2009; Wadge et al., 2014; Komorowski et al., 2005, 2015, 2022; Pyle et al., 2018; Barclay et al., 2019; Joseph et al., 2022; Global Volcanism Program, 2023). 14 submarine eruptions have also occurred from Kick ‘em Jenny (Lindsay et al., 2005a). *Current day summit temperatures Lindsay et al., 2005; Moretti et al., 2020a,b; Basan et al., 2022). **UWI-SRC 2020-22; Global Volcanism Program, 2020. ***some ‘self-evacuation’ occurred but no numbers are available (Pyle et al., 2018).

Figures and Tables

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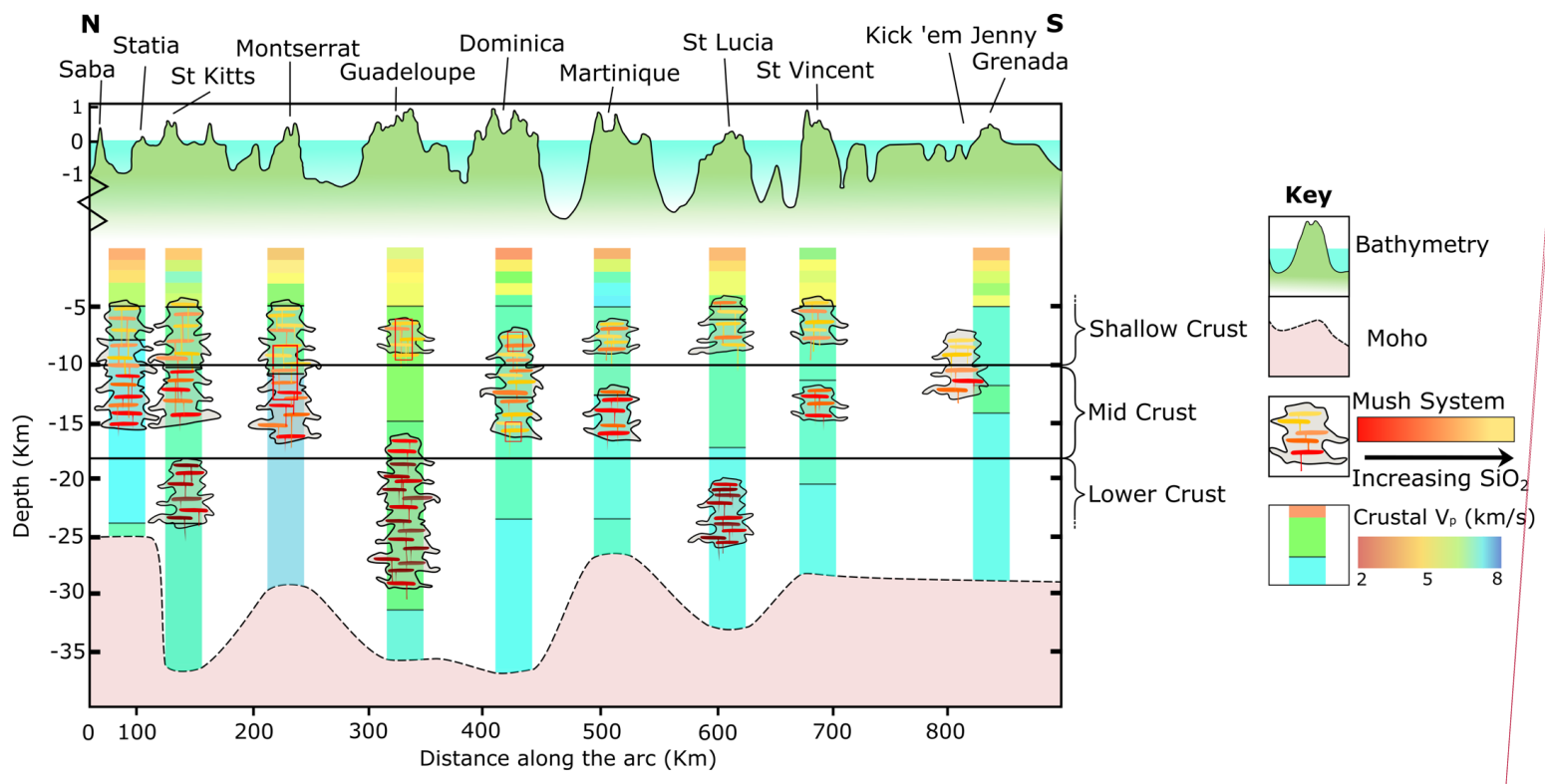


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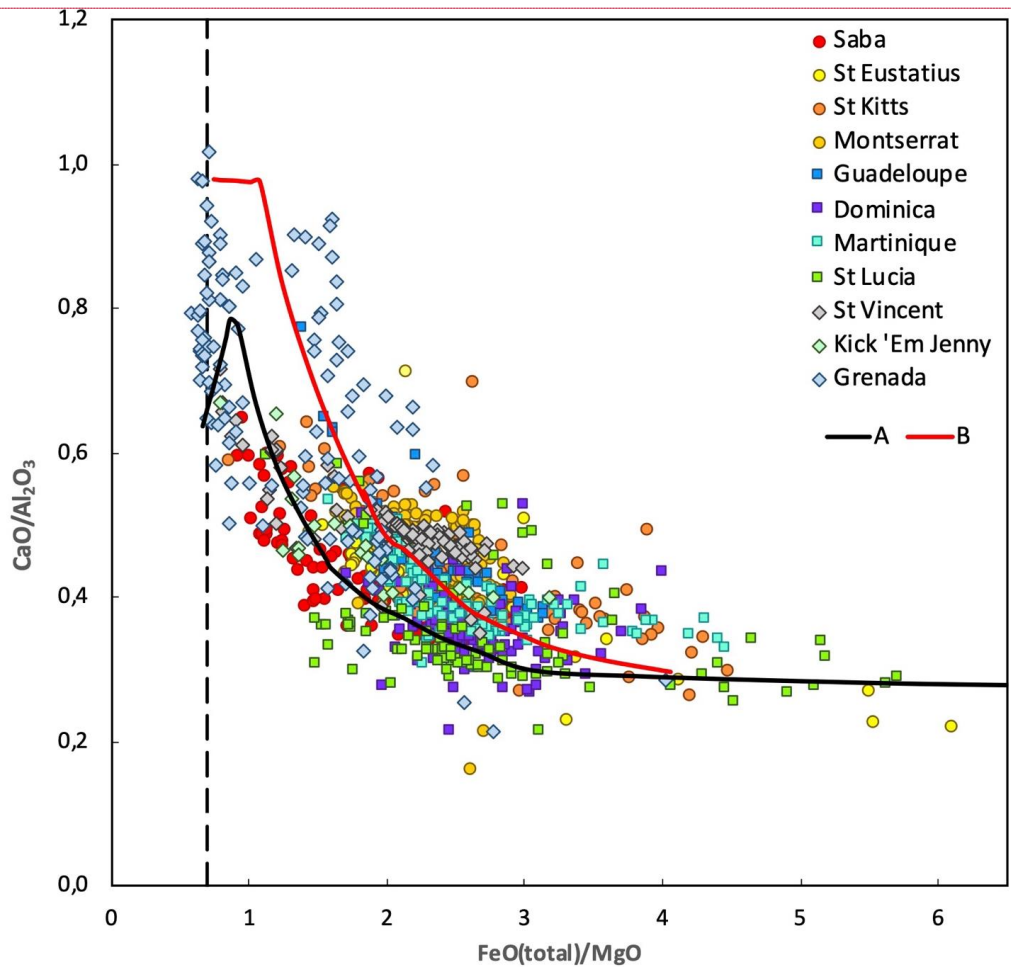


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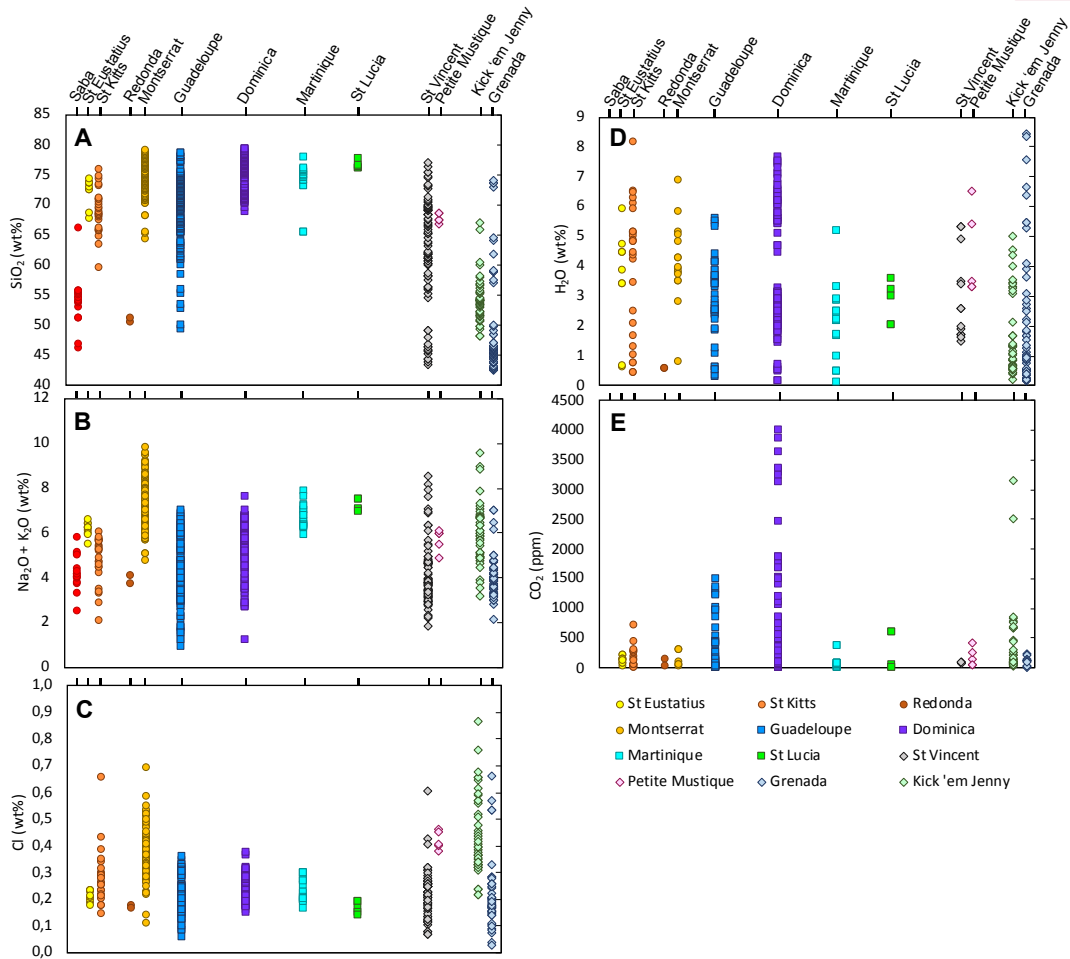
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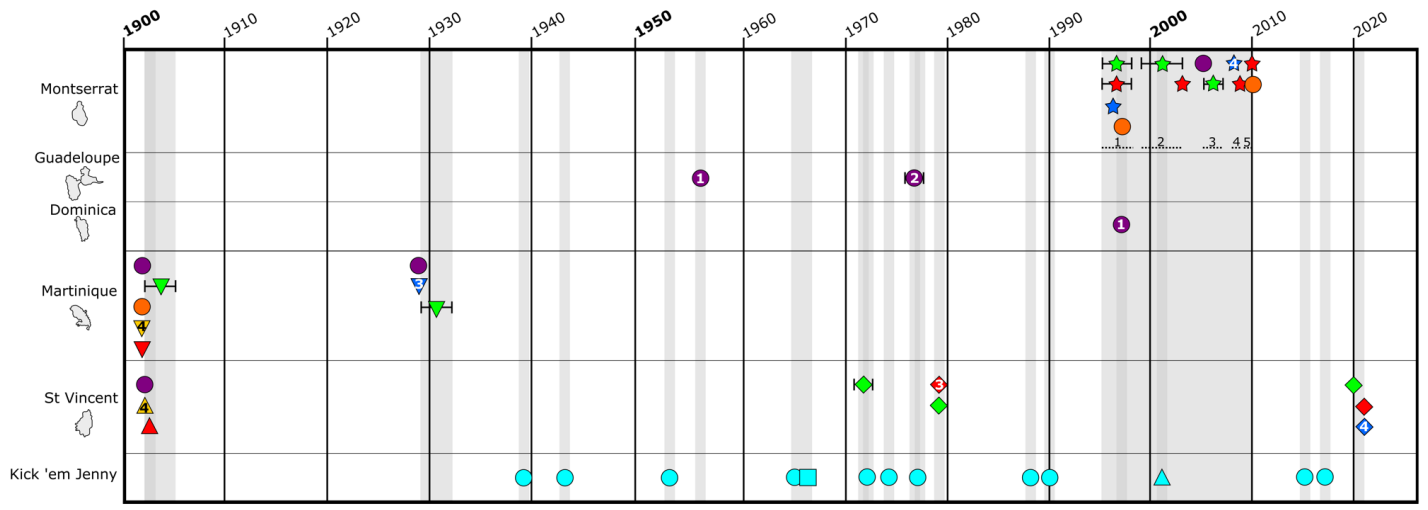


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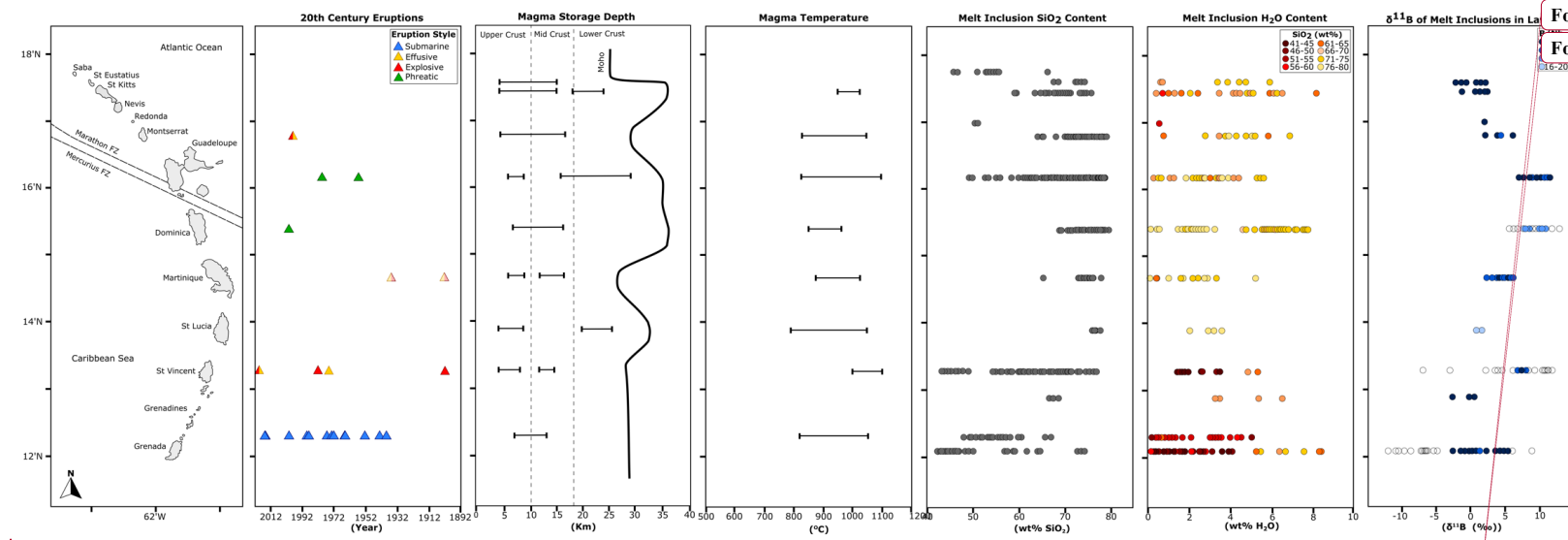


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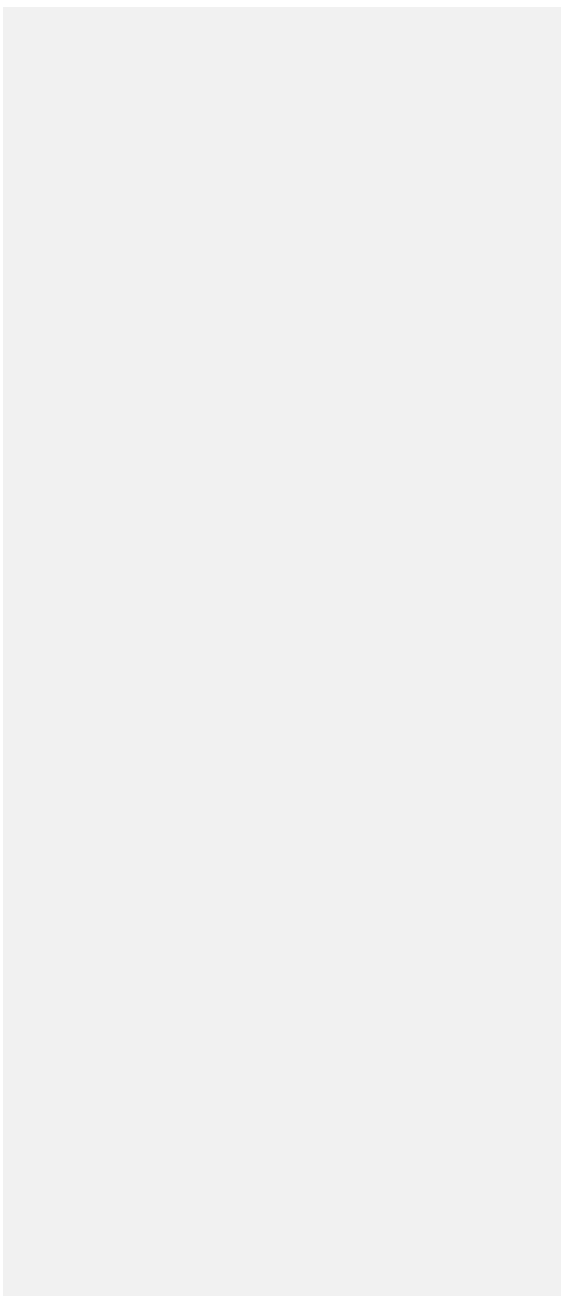


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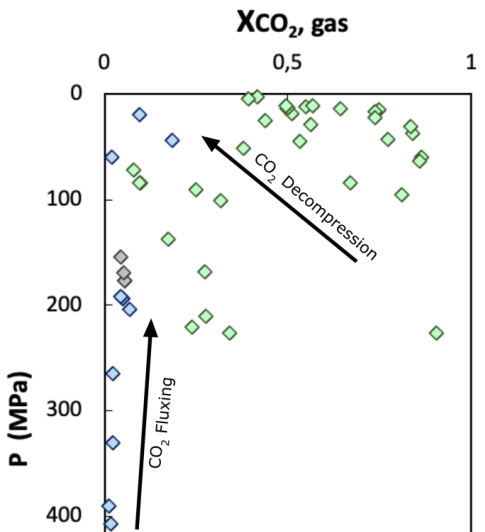
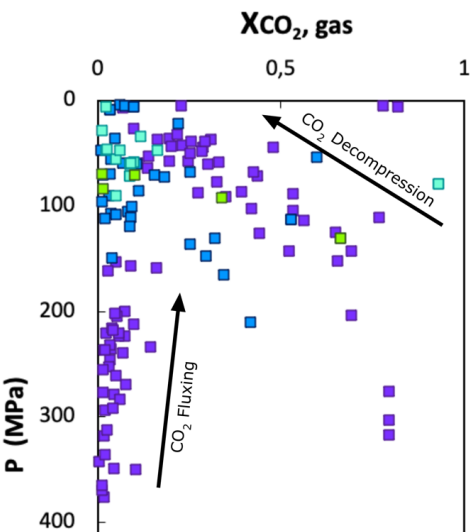
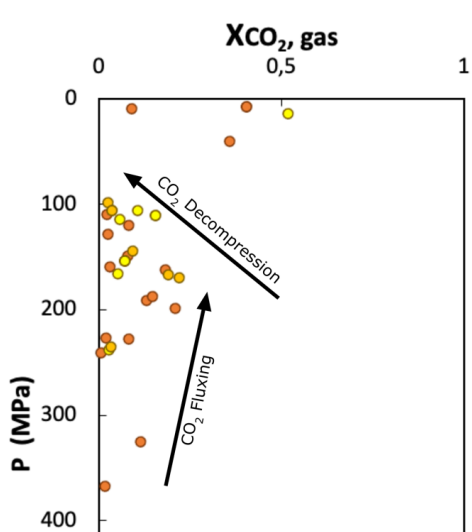
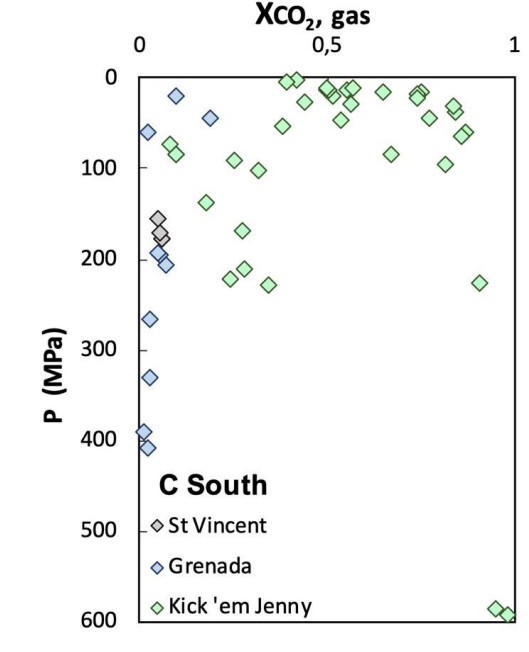
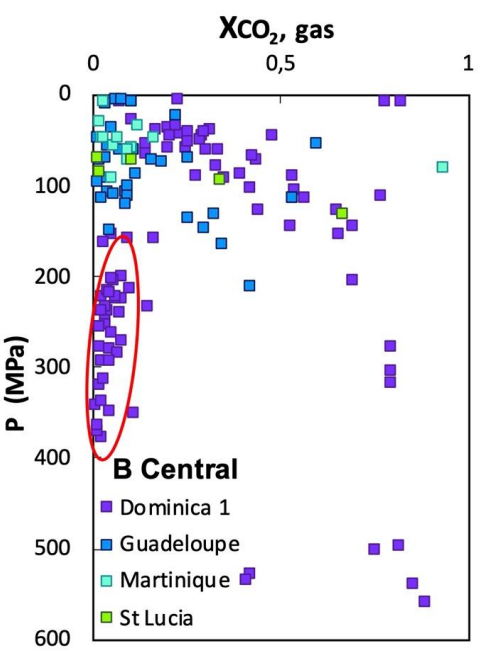
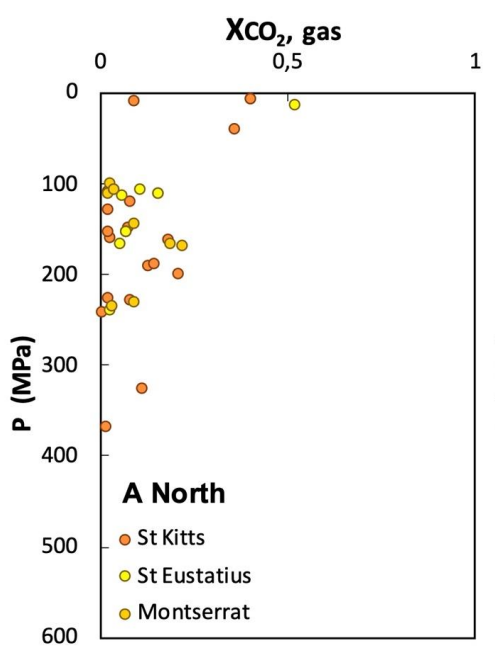
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Table 1										
Island	Volcano	Summit Temperature (°C)*	Eruption start	Eruption end	Main Eruptive Style	Maximum VEI	Eruptive Phases	Unrest Period Reported Prior to Eruption	Number of Deaths	People Evacuated
St Vincent	Soufrière St Vincent	>95	2020	2021	Explosive	4	- Dome growth - Cyclic vulcanian explosions - Sub-Plinian activity	~2 months**	0	>16,000
Dominica	Morne Watt	91 - 99	1997	1997	Phreatic	1	- Unobserved phreatic explosion	N/A	0	0
Montserrat	Soufrière Hills	98 - 99	1995	Ongoing	Explosive	3 - 4	- Cyclic dome growth and collapse, - Cyclic vulcanian explosions - Sub-Plinian activity - Lateral blasts	3 years	19	>6,000
St Vincent	Soufrière St Vincent	-	1979	1979	Explosive	3	- Vulcanian explosions - Dome growth	~days	2	20,000
Guadeloupe	La Soufrière de Guadeloupe	~100	1976	1977	Phreatic	2	- 26 phreatic explosions	13 months	0	70,000

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St Vincent	Soufrière St Vincent	80-100	1971	1972	Effusive	0	- Aseismic dome growth	N/A	0	0
Guadeloupe	La Soufrière de Guadeloupe	~100	1956	1956	Phreatic	1	- Phreatic explosion	~months	0	0
Martinique	Montagne Pelée	-	1929	1932	Explosive	3	-Phreatic eruptions - Pelean activity - Dome growth	~months	N/A	N/A
Martinique	Montagne Pelée	-	1902	1907	Explosive	4	- Vulcanian explosions - Sub-Plinian and Pelean activity - Dome growth and collapse	~13 years	~30,000	N/A
St Vincent	Soufrière St Vincent	-	1902	1903	Explosive	4	- Vulcanian explosions - Sub-Plinian activity	1 year	~1500	N/A***

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