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1 Alteration-induced volcano instability at La Soufrière de Guadeloupe
2 (Eastern Caribbean)

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22

23 **Abstract**

24 Volcanoes are unstable structures that deform laterally and frequently experience mass
25 wasting events. Hydrothermal alteration is often invoked as a mechanism that contributes

26 significantly to volcano instability. We present a study that combines laboratory experiments,
27 geophysical data, and large-scale numerical modeling to better understand the influence of
28 alteration on volcano stability, using La Soufrière de Guadeloupe (Eastern Caribbean) as a case
29 study. Laboratory experiments on variably-altered (advanced argillic alteration) blocks show
30 that uniaxial compressive strength, Young's modulus, and cohesion decrease as a function of
31 increasing alteration, but that the internal friction angle does not change systematically.
32 Simplified volcano cross sections were prepared (a homogenous volcano, a volcano containing
33 the alteration zone identified by a recent electrical survey, and a volcano with an artificially
34 enlarged area of alteration) and mechanical properties were assigned to zones corresponding to
35 unaltered and altered rock. Numerical modeling performed on these cross sections, using a
36 hydro-thermo-mechanical modeling code, show (1) the importance of using upscaled values in
37 large-scale models and (2) that alteration significantly increases volcano deformation and
38 collapse volume. Finally, we combined published muon tomography data with our laboratory
39 data to create a 3D strength map, exposing a low-strength zone beneath the southern flank of
40 the volcano coincident with the hydrothermal system. We conclude that hydrothermal alteration
41 decreases volcano stability and thus expedites volcano spreading and increases the likelihood
42 of mass wasting events and associated volcanic hazards. Hydrothermal alteration, and its
43 evolution, should therefore be monitored at active volcanoes worldwide.

44

45 **Key points**

46 • Laboratory experiments show that hydrothermal alteration reduces the strength of
47 volcanic rock from La Soufrière.

48

49 • Numerical modeling shows that hydrothermal alteration significantly increases volcano
50 deformation and collapse volume.

51

- 52 • We provide a 3D strength map of La Soufrière that exposes a low-strength zone
53 coincident with the hydrothermal system.

54

55 **Plain language summary**

56 The rocks forming a volcanic edifice can be altered by circulating hydrothermal fluids.
57 This alteration can influence the physical and mechanical properties of these rocks, which could
58 jeopardize volcano stability. The stability of a volcanic edifice is an important consideration in
59 volcanic hazards and risk assessments due to the potentially dire consequences of partial
60 volcanic flank collapse. Using a combination of experimental data, geophysical data, and
61 modeling, and La Soufrière de Guadeloupe (Eastern Caribbean, France) as a case study, we
62 find that hydrothermal alteration decreases volcano stability and thus promotes volcano
63 instability and associated volcanic hazards. As a result, we conclude that hydrothermal
64 alteration, and its evolution, should be monitored at active volcanoes worldwide.

65

66 **Keywords:** hydrothermal alteration; uniaxial compressive strength; Young's modulus;
67 upscaling; flank collapse

68

69 **1 Introduction**

70 Volcanoes are inherently unstable structures that are built haphazardly, in both space
71 and time, from the products of successive effusive and explosive eruptions and endogenous
72 growth. These materials have highly variable physical and mechanical properties (Heap and
73 Violay, 2021) and often form oversteepened and unstable slopes. As a result, volcano
74 deformation (such as volcano spreading; Borgia et al., 2000) and mass wasting events (such as
75 debris avalanches resulting from partial flank collapse, lahars, and rockfalls; Roverato et al.,
76 2021) are commonplace at many volcanoes worldwide. Catastrophic collapse resulting from
77 volcano spreading (van Wyk de Vries and Francis, 1997) and mass wasting events present a
78 significant volcanic hazard, and can also trigger hazardous laterally-directed explosions and
79 devastating—both economically and in terms of loss of life—pyroclastic density currents (Cole
80 et al., 2015; Glicken, 1996; Sparks et al., 2020; Voight et al., 1981). Indeed, partial flank
81 collapses at about 200 volcanoes have resulted in at least 20,000 fatalities in the last 10,000
82 years (Siebert et al., 2010). Although about 52% of these flank collapses are associated with
83 magmatic eruptions, about 22% are associated with non-magmatic phreatic and/or
84 hydrothermal eruptions (Siebert, 1984; Siebert et al., 1987, 2010). As a result, stability
85 assessments at volcanoes are an essential component of volcano monitoring and volcanic
86 hazard mitigation.

87 Hydrothermal alteration, common to many volcanoes worldwide, is often invoked as a
88 mechanism that contributes significantly to volcano instability. Indeed, studies using geological
89 evidence, geochemical or geophysical data, numerical modeling, laboratory experiments, or a
90 combination of these approaches have highlighted that hydrothermal alteration has or can
91 weaken a volcanic slope sufficiently to promote collapse (Ball et al., 2013, 2015, 2018; Cecchi
92 et al., 2004; del Potro and Hürlimann, 2009; Finn et al., 2018; John et al., 2008; López and
93 Williams, 1993; Opfergelt et al., 2006; Reid et al., 2001; Rosas-Carbajal et al., 2016; Salaün et

94 al., 2011; van Wyk de Vries et al., 2000; Voight et al., 2002; Watters et al., 2000). Fluid and
95 pore fluid pressure re-distributions caused by alteration are also thought to promote instability
96 (Ball et al., 2018; Day, 1996; Heap et al., 2021a; Reid, 2004) and erratic explosive behavior (de
97 Moor et al., 2019; Heap et al., 2019; Mick et al., 2021).

98 Geological evidence is provided by the abundance of hydrothermally altered materials
99 typically found in debris avalanche deposits resulting from partial flank collapse. For example,
100 in their analysis of the 26 December 1997 partial edifice collapse at Soufrière Hills volcano
101 (Montserrat, Eastern Caribbean), Voight et al. (2002) found varicolored, hydrothermally-
102 altered materials within the avalanche deposits, highlighting that alteration could have
103 contributed to the collapse of the southern sector of the volcano. Avalanche deposits and
104 material ejected during phreatic explosions at La Soufrière de Guadeloupe (Eastern Caribbean;
105 France) were also found to contain various parts of the active and ancient hydrothermal systems
106 of the volcano (Salaün et al., 2011). Debris avalanche deposits at Tutupaca volcano (Peru) also
107 contain hydrothermally altered materials (Samaniego et al., 2015). Extensive fumarole activity
108 and hydrothermal alteration on the unstable flanks of Casita volcano (Nicaragua) was
109 documented by van Wyk de Vries et al. (2000), a volcano that experienced a partial flank
110 collapse in 1998 with associated debris avalanches that killed about 2,500 people (Kerle and
111 van Wyk de Vries, 2001; van Wyk de Vries et al., 2000). Opfergelt et al. (2006) further
112 suggested that the presence of abundant smectite (up to 50 wt.%) in the hydrothermally-altered
113 core of Casita volcano contributed to slope instability by acting as a barrier to water infiltration
114 thereby promoting fluid circulation along marked discontinuities, gradually decreasing the
115 shear strength of the host-rock. Finally, more than 55 debris flows have originated from Mt
116 Rainer (USA) in the Holocene, and hydrothermally derived clays have been found in some of
117 the most widespread lahar deposits (Reid et al., 2001).

118 Geophysical and geochemical evidence has been provided by a wide range of different
119 techniques. For example, electrical tomography data exposed the extent of the hydrothermally
120 altered zone, the location and geometry of low-strength detachment planes, and potential
121 collapse volumes at La Soufrière de Guadeloupe (Rosas-Carbajal et al., 2016). Hyperspectral
122 remote sensing (Kereszturi et al., 2021) and helicopter electromagnetic and magnetic
123 measurements (Finn et al., 2018) have been used to map hydrothermal alteration zones to
124 inform slope stability modeling at Mt Ruapehu (New Zealand) and Mount Baker (USA),
125 respectively, highlighting the most likely location of future collapses. Finally, geochemical data
126 and modeling provided by López and Williams (1993) suggested that hydrothermal alteration
127 along faults can create weak sliding planes that can facilitate slope failure.

128 Numerical and analog modeling has also provided key insights into how hydrothermal
129 alteration can influence volcano stability. For example, using analogue experiments in which
130 sand-plaster mixtures and silicone were used to represent fresh and altered rock, respectively,
131 van Wyk de Vries et al. (2000) showed that a weak and ductile (hydrothermally altered) core
132 can reproduce the deformation structures observed at Casita volcano, suggesting that
133 hydrothermal alteration destabilized the volcano. Motivated by intensive alteration on the
134 flanks and summit of Mt Rainer, Reid et al. (2001) examined the influence of alteration on
135 volcano stability using a geotechnical method of 3D column limit-equilibrium slope stability
136 analysis. Using a best-estimate distribution of the alteration, and typical values of cohesion,
137 internal friction angle, and rock unit weight for fresh, lightly altered, and highly altered rocks,
138 Reid et al. (2001) concluded that steep slopes and large volumes of altered and weak rock can
139 promote potentially large gravitational failures.

140 Although these aforementioned studies, and others, have improved our understanding
141 of alteration-induced volcano instability, the shape and size of the alteration zones and/or the
142 geomechanical properties used in large-scale stability modeling are often poorly constrained or

143 arbitrarily chosen. Indeed, there is a paucity of mechanical data for variably altered volcanic
144 rocks for use in large-scale numerical models designed to better understand flank stability. As
145 noted by Ball et al. (2018), the scarcity of rock physical and mechanical properties for altered
146 volcanic rocks limits the investigation of volcano stability using large-scale models. We present
147 here a multidisciplinary study in which we first provide physical and mechanical properties for
148 variably altered andesites from La Soufrière de Guadeloupe. We then assign laboratory-scale
149 and upscaled (i.e. modified to account for large-scale discontinuities that are not captured in
150 laboratory-scale samples) physical and mechanical properties to zones identified by a recent
151 electrical survey of the dome of La Soufrière de Guadeloupe (Rosas-Carbajal et al., 2016) and
152 perform large-scale numerical modeling to showcase the importance of hydrothermal alteration
153 on volcano stability. Finally, we combine recent muon tomography data (Rosas-Carbajal et al.,
154 2017) with our laboratory data to create a 3D compressive strength map of the volcano.
155 Although we use La Soufrière de Guadeloupe as a case study, we consider our salient
156 conclusions relevant for andesitic stratovolcanoes worldwide.

157

158 **2 La Soufrière de Guadeloupe (Eastern Caribbean, France)**

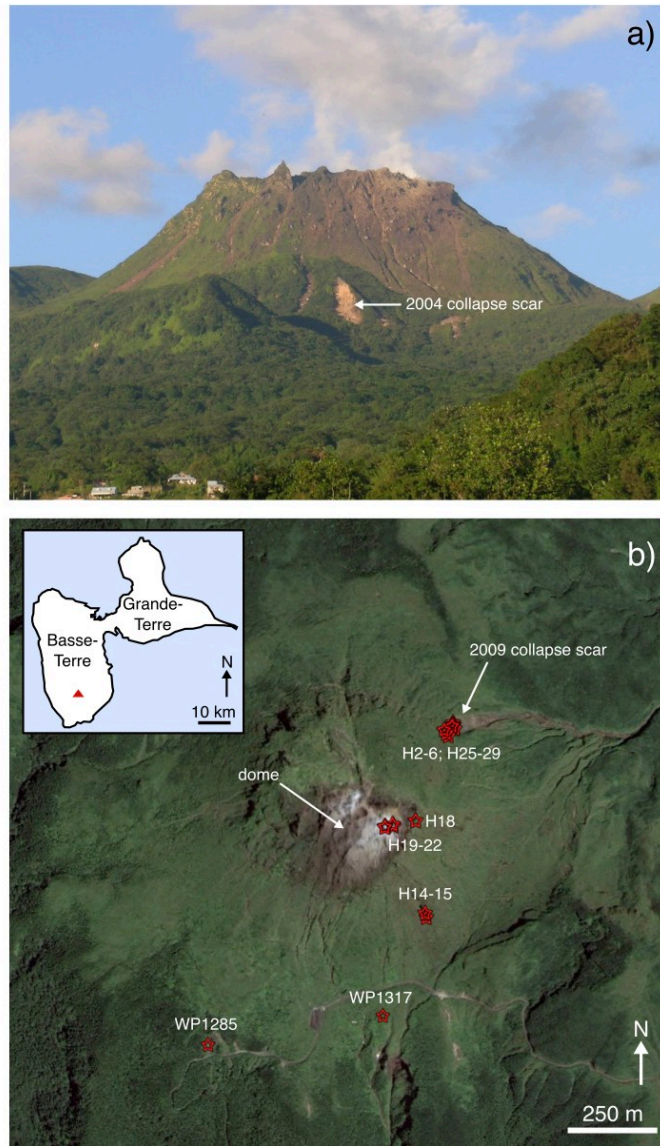
159 La Soufrière de Guadeloupe (Figure 1a), henceforth called La Soufrière, is located on
160 the island of Guadeloupe in the Eastern Caribbean (Figure 1b) and is an archetype of the
161 hazardous andesitic volcanoes that occur in many subduction zones. The current lava dome
162 formed in 1530 CE and, since 1635 CE, there have been six non-magmatic phreatic or
163 hydrothermal eruptions (Komorowski et al., 2005). The largest and most-recent eruption
164 occurred in 1976–1977. After months of seismic unrest, the volcano exploded suddenly and
165 without precursory signs in 1976–1977, resulting in a volcanic crisis that led to the evacuation
166 of about 70,000 people and serious socioeconomic consequences for the island (Feuillard et al.,
167 1983; Hincks et al., 2014; Komorowski et al., 2005, 2015).

168 Volcanic unrest has been increasing since the volcano reawakened in 1992, marked by
169 the restart of fumarolic and seismic activity (Komorowski et al., 2005) and, in April 2018, the
170 largest felt volcano-tectonic earthquake since the 1976–1977 crisis (M_L 4.1 or M_W 3.7; Moretti
171 et al., 2020). The 2018 episode of accelerated unrest was interpreted as a failed phreatic eruption
172 and raised serious concerns for the stability of the southwest flank of the dome, which shows
173 the largest displacements (up to 9 mm/year towards the southwest) over the last 20 years
174 (Moretti et al., 2020). The main features of the ongoing unrest include the continuous expansion
175 of the outgassing area on top of the lava dome and the appearance of new steam-dominated
176 fumaroles in July 2014 and February 2016 that are characterized by a deep magmatic gas
177 component (Brombach et al., 2000; Moretti et al., 2020; Villemant et al., 2014). Such expansion
178 was accompanied by an increase in the heat output from the entire dome, which evolved from
179 ~ 1.2 MW in 2010 to ~ 7.6 MW in 2020 (Jessop et al., 2021). This increase in heat output is of
180 great concern for La Soufrière as it has been recently demonstrated that volcanoes with a high
181 heat flux are more likely to experience magmatic or phreatic eruptions (Girona et al., 2021).
182 High-flow rate thermal acid sulfate-chloride springs, associated with degassing of the
183 underlying magmatic body, are observed on the slopes and at the base of the dome and supply
184 hot acid fluids and CO_2 , but also SO_2 , H_2S , HCl , and HF , to the above hydrothermal system
185 (Villemant et al., 2005; 2014). These acidic fluids are (1) transferred through the dome and feed
186 the steam-rich summit fumaroles and (2) circulate through the dome structure along listric,
187 concave-upward structural discontinuities that drain to the southwest (Brombach et al., 2000;
188 Rosas-Carbajal et al., 2016; Salaün et al., 2011; Villemant et al., 2014). Evidence for the
189 preferential circulation of acidic thermal fluids along these structures is provided by the 3D
190 geometry of thermal springs (Moretti et al., 2020; Villemant et al., 2005) and geophysical
191 imaging (Bouligand et al., 2016; Brothelande et al., 2014; Nicollin et al., 2006; Rosas-Carbajal
192 et al., 2016). The shallow hydrothermal system is considered responsible for the recent

193 observed shallow deformation and seismic activity (at depths at or above sea level, with
194 magnitudes less than one) (Moretti et al., 2020). Indeed, the monthly bulletins from the
195 observatory (Observatoire Volcanologique et Sismologique de Guadeloupe, Institut de
196 Physique du Globe de Paris (OVSG-IPGP), 1999-2021; available at
197 <http://www.ipgp.fr/fr/ovsg/bulletins-mensuels-de-lovsg>) document hundreds of shallow, low-
198 magnitude ($M < 1$) earthquakes each month. Deeper and higher magnitude off-axis seismic
199 activity are thought to be related to increases in pore pressure in response to the pulsatory arrival
200 of hot magmatic fluids along the larger faults that cut the dome (Moretti et al., 2020).

201 The recent unrest, suggesting that La Soufrière is undergoing a renewed period of
202 structural and chemical weakening, raises concern for the stability of the volcano. Indeed, the
203 most-recent unrest and instability at La Soufrière was associated with hydrothermal activity.
204 For example, the materials ejected during the explosive hydrothermal eruption of 1976–1977
205 were largely hydrothermally altered (Feuillard et al., 1983) and geological studies have also
206 shown that La Soufrière has an exceptional record of partial edifice collapse (at least eight
207 collapses in the past 9,150 years) that have produced extensive debris avalanche deposits
208 dominantly composed of hydrothermally altered materials (Boudon et al., 1987; Komorowski
209 et al., 2005; Le Friant et al., 2006; Peruzzetto et al., 2019; Rosas-Carbajal et al., 2016; Salaün
210 et al., 2011). The simulations of Le Friant et al. (2006) and Peruzzetto et al. (2019) suggest that
211 the northern and eastern parts of the town of Saint Claude could be impacted by small-volume
212 partial dome collapses and that, in the worst-case scenario of a major dome collapse, a large
213 part of Basse-Terre could be affected. As a result, not only is La Soufrière an ideal natural
214 laboratory to study the influence of hydrothermal alteration on volcano stability, but such a
215 study is also timely because of the ongoing, and increasing, unrest at the volcano.

216



217

218 **Figure 1.** (a) Photograph of La Soufrière de Guadeloupe (Eastern Caribbean, France). This

219 photograph also shows the collapse scar (indicated by the white arrow) triggered by the 21

220 November 2004 Les Saintes magnitude Mw 6.3 regional earthquake (Feuillet et al., 2011). (b)

221 Map of La Soufrière de Guadeloupe (taken from Google Maps®) showing the sampling

222 locations for the 17 blocks acquired for this study. This image also shows the location of the

223 dome and the 2009 collapse scar (indicated by white arrows). Inset shows a map of

224 Guadeloupe in which the location of La Soufrière de Guadeloupe is indicated by a red

225 triangle.

226

227 **3 Materials and Methods**

228 A total of 17 blocks were collected from La Soufrière during a field campaign in 2019
229 (sampling locations shown as red stars in Figure 1b). Our aim was to sample a range of materials
230 from different locations on the volcano that best represent the variability in porosity and
231 hydrothermal alteration of the rocks forming the edifice. A large proportion of the blocks (eight
232 out of 17) were collected from the collapse scar of the 19 November 2009 landslide, triggered
233 by extreme rainfall (Moretti et al., 2020; Peruzzetto et al., 2019), which was seen as an
234 opportunity to sample blocks that are more representative of the interior of the volcano.

235 Eight blocks were taken from the collapse scar of the 2009 landslide (H2A, H2B, H3,
236 H4A, H5A, H6, H25, and H29), one block was taken from the collapse scar of the landslide
237 triggered by the 21 November 2004 *Les Saintes* magnitude Mw 6.3 regional earthquake
238 (Feuillet et al., 2011; Figure 1b) (WP1285), two blocks were taken from the West wall of the
239 fault “*Faïlle 30 août*” (H14 and H15), and one block, a volcanic bomb from the 1976–1977
240 eruption, was taken from the roof of a small disused thermal bathhouse to the South of the dome
241 (WP1317). The remaining samples were taken from the dome: one block was taken from the
242 “*Lacroix Supérieur*” outgassing fracture (H18), and four blocks were taken from the lava spines
243 of the 1530 CE dome: two blocks from “*Cratère Sud Central*” (H19 and H20) and two blocks
244 from an adjacent site (H21 and H22). Site descriptions and Global Positioning System (GPS)
245 coordinates are available for each block in the Supplementary Information.

246 Cylindrical samples were prepared from each of the blocks to a diameter of 20 mm and
247 then cut and precision-ground to a nominal length of 40 mm. The samples were washed and
248 then dried in a vacuum-oven at 40 °C for at least 48 h. The connected porosity of each sample
249 was calculated using the bulk sample volume and the skeletal (solid) sample volume measured
250 by a helium pycnometer. Dry and wet bulk sample densities were calculated using the bulk
251 sample volume and the dry and wet mass, respectively. Polished thin sections were prepared

252 from offcuts of the samples for microstructural analysis, performed using a Tescan Vega 2
253 XMU scanning electron microscope (SEM). The mineral content of the 17 blocks was analyzed
254 and quantified using X-ray powder diffraction (XRPD) on powdered offcuts of the samples,
255 assisted by Raman spectroscopy on the prepared thin sections (see Supplementary Information
256 for more details).

257 Dry uniaxial compressive strength was measured on oven-dry samples from all 17
258 blocks in a uniaxial load frame. Samples were deformed under ambient laboratory pressure and
259 temperature at a constant axial strain rate of 10^{-5} s^{-1} until macroscopic failure. The static
260 Young's modulus was determined from the elastic portion of the uniaxial stress-strain curves
261 (Heap et al., 2020a). Triaxial compression experiments were performed on blocks selected to
262 best represent the variability in observed alteration (samples H2B, H3, and H18). These samples
263 were vacuum-saturated in deionized water, inserted into a rubber jacket, placed inside a
264 pressure vessel, and then taken to the target confining (from 10.5 to 25 MPa) and pore fluid
265 pressures (10 MPa for all experiments) using servo-controlled pumps. Samples were deformed
266 under ambient laboratory temperature at a constant axial strain rate of 10^{-5} s^{-1} until either
267 macroscopic failure (for brittle experiments) or until 3% axial strain (for ductile experiments).
268 The triaxial apparatus was also used to measure the water-saturated uniaxial compressive
269 strength of samples H2B, H3, and H18. For both uniaxial and triaxial experiments, axial
270 displacement and axial load were measured using a linear variable differential transducer and a
271 load cell, respectively. Axial displacement (minus the displacement accumulated within the
272 load chain) and axial load were converted to axial strain and axial stress using the sample
273 dimensions. Here, we define the effective pressure, P_{eff} , as the confining pressure, P_c , minus
274 the pore fluid pressure, P_p . We adopt the convention that compressive stresses and strains are
275 positive.

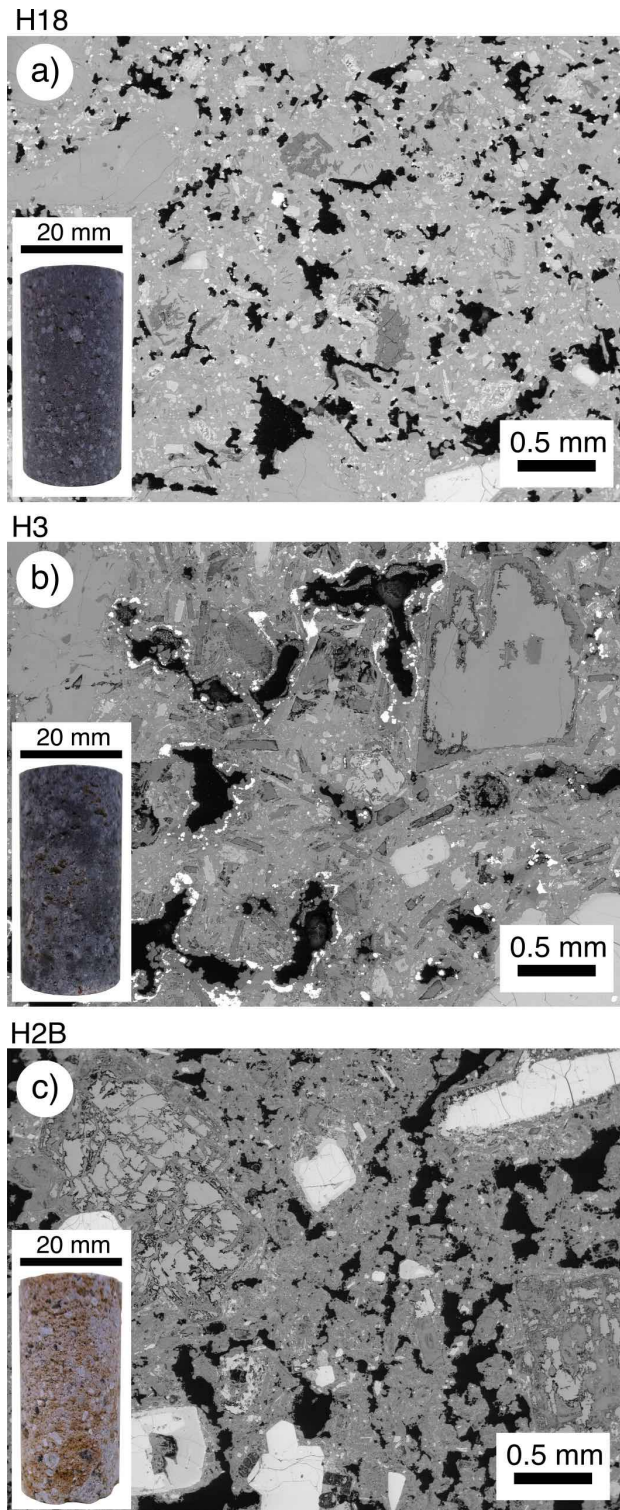
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277 **4 Results**

278 4.1 Microstructure, mineralogy, and alteration

279 The mineral contents for each of the 17 blocks are given in Table 1, and Figure 2 shows
280 backscattered SEM images of three blocks selected to represent the range of alteration intensity
281 (defined by their percentage of secondary minerals) observed in the rocks collected for this
282 study (relatively unaltered – H18; moderately altered – H3; and highly altered – H2B). SEM
283 images of all the blocks are available in the Supplementary Information. All of the andesite
284 blocks are characterized by a porphyritic texture comprising phenocrysts of dominantly
285 plagioclase and pyroxene (and high-density oxides) within a crystallized groundmass (Figure
286 2). All of the samples also contain secondary minerals, including silica polymorphs (quartz,
287 cristobalite, tridymite, and opal-A), hematite, pyrite, alunite or natroalunite, gypsum, kaolinite,
288 and talc (Table 1). Plagioclase phenocrysts are often pervasively altered and partially replaced
289 by kaolinite (e.g., Figure 2c) or opal-A. Secondary minerals (natroalunite, alunite, cristobalite,
290 tridymite, pyrite, and kaolinite) are also found precipitated within pores and microcracks (see
291 Supplementary Information for additional SEM images).

292



293

294

295

296

Figure 2. Backscattered scanning electron microscope images of three andesites from La Soufrière de Guadeloupe (Eastern Caribbean, France) selected to represent the range of observed alteration: (a) relatively unaltered – H18, (b) moderately altered – H3, and (c) highly

297 altered – H2B. Black – porosity; grayscale – rock groundmass. Insets show photographs of
 298 the 20 mm-diameter cylindrical samples.

299

Mineral	H2A	H2B	H3	H4A	H5A	H6	H14	H15	H18	H19	H20	H21	H22	H25	H29	WP1285	WP1317
Plagioclase	56.7	12.3	46.6	23.3	41.3	30.0	60.7	22.5	61.2	22.0	28.7	24.2	59.5	38.7	62.4	64.7	61.6
Clinopyroxene	8.7	3.4	5.6	4.9	5.2	6.4	6.3	7.3	8.4	5.0	8.9	12.4	8.9	5.3	7.8	5.2	5.9
Orthopyroxene	10.8	9.5	11.8	11.8	11.1	10.8	8.6	9.2	12.2	10.2	15.0	19.3	13.6	10.2	11.2	13.2	15.6
(Ti-) Magnetite	0.7	-	0.8	-	-	-	0.8	-	2.9	-	2.4	3.1	0.8	-	2.7	3.5	0.7
Quartz*	1.0	0.5	0.6	0.6	0.5	0.5	1.7	0.7	0.7	1.7	0.3	0.2	0.6	0.3	0.4	0.2	0.7
Cristobalite*	11.3	12.8	10.6	11.8	13.0	11.1	13.5	10.2	11.7	9.5	11.4	11.7	10.6	9.8	12.4	-	-
Tridymite*	-	-	-	-	-	-	-	0.7	-	-	-	-	-	-	-	13.2	13.2
Hematite*	-	-	-	-	-	-	3.4	-	2.8	2.4	-	-	-	-	3.1	-	-
Pyrite*	3.5	-	3.8	2.3	-	-	-	-	-	-	-	0.4	3.1	0.6	-	-	-
Alunite*	-	-	-	-	-	-	-	-	-	-	-	-	-	-	-	-	2.4
Na-Alunite*	1.4	1.6	2.8	1.3	5.4	5.1	5.1	15.0	-	14.2	0.5	0.5	-	9.8	-	-	-
Gypsum*	-	-	-	0.7	-	-	-	-	-	-	0.8	1.2	-	-	-	-	-
Kaolinite*	6	59.7	17.4	43.3	23.5	36.0	<1	34.3	-	2.0	2.0	2.0	<1	25.3	-	-	-
Talc*	-	-	-	-	-	-	-	-	-	-	-	-	2.9	-	-	-	-
Opal-A*	-	-	-	-	-	-	-	-	-	33.0	30.0	25.0	-	-	10.0	-	-

300

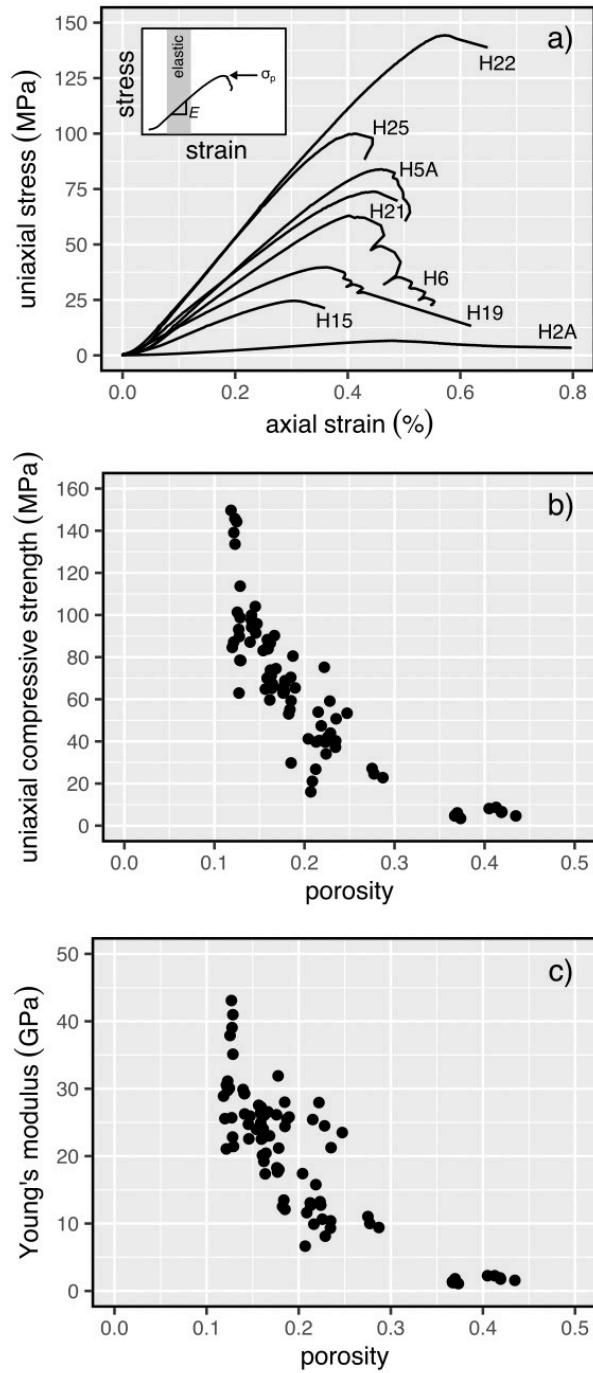
301 **Table 1.** Mineral contents, measured by X-ray powder diffraction and refined using Raman
 302 spectroscopy and optical microscopy, of the 17 blocks collected for this study (sampling
 303 locations shown in Figure 1b). Values in wt.%. Asterisk denotes a secondary/alteration
 304 mineral.

305

306 4.2 Mechanical data

307 Representative uniaxial stress-strain curves for selected samples are shown in Figure
 308 3a. Stress is first a non-linearly increasing function of strain, often attributed to the closure of
 309 pre-existing microcracks. This stage is followed by a quasi-linear elastic stage. In the next stage,
 310 stress is a non-linearly decreasing function of strain, a result of the initiation, propagation, and
 311 coalescence of microcracks. Finally, a stress drop signals the formation of a macroscopic shear
 312 fracture. Uniaxial compressive strength and Young's modulus are plotted as a function of
 313 connected porosity in Figures 3b and 3c, respectively (data available in Table 2). We first note

314 that the connected porosity of our samples varies from 0.12 to 0.43. The uniaxial compressive
315 strength and Young's modulus of our sample suite varies from ~3.5 to ~150 MPa and from ~1
316 to ~43 GPa, respectively (Figure 3). Our data show that uniaxial compressive strength and
317 Young's modulus both decrease as a function of increasing porosity (Figure 3).
318



319

320 **Figure 3.** (a) Representative stress-strain curves for selected andesite samples from La
321 Soufrière de Guadeloupe (Eastern Caribbean, France). Inset shows an isolated curve
322 indicating from where the Young's modulus (the slope of the stress-strain curve), E , was
323 determined (the quasi-linear elastic stage is indicated by the gray zone). Also labeled is the
324 uniaxial compressive strength, σ_p , which is the maximum axial stress the sample attained
325 prior to macroscopic failure (signaled by the stress drop). Uniaxial compressive strength (b)
326 and Young's modulus (c) as a function of connected porosity (data available in Table 2).
327

Sample	Connected porosity	Uniaxial compressive strength (MPa)	Young's modulus (GPa)	Percentage of secondary minerals
H2A_2	0.18	63.7	31.9	23.2
H2A_5	0.19	80.5	25.5	23.2
H2A_6	0.19	65.4	25.8	23.2
H2A_11	0.19	59.3	24.4	23.2
H2A_12	0.18	70.4	28.0	23.2
H2B_3	0.42	6.6	1.8	74.6
H2B_10	0.42	6.4	1.9	74.6
H2B_11	0.41	8.7	2.3	74.6
H2B_12	0.41	8.1	2.3	74.6
H2B_15	0.43	4.6	1.6	74.6
H3_3	0.16	70.3	27.3	35.2
H3_7	0.16	69.9	24.9	35.2
H3_8	0.16	64.8	27.5	35.2
H3_11	0.16	59.7	24.0	35.2
H3_13	0.16	71.2	26.2	35.2
H4A_2	0.23	40.3	10.4	60.0
H4A_4	0.23	42.0	10.6	60.0
H4A_6	0.23	43.9	8.1	60.0
H4A_8	0.23	37.2	9.3	60.0
H4A_9	0.22	40.4	9.9	60.0
H5A_2	0.16	88.3	26.4	42.4
H5A_3	0.16	86.6	22.8	42.4
H5A_5	0.16	83.9	22.5	42.4
H5A_8	0.18	68.8	21.2	42.4
H5A_10	0.17	74.5	23.0	42.4
H6_6	0.18	55.1	13.5	52.7
H6_9	0.18	68.0	18.0	52.7
H6_9	0.18	53.0	12.5	52.7
H6_12	0.18	64.2	17.6	52.7
H6_13	0.18	63.0	18.3	52.7
H14_2	0.18	65.4	26.1	23.7

H14_3	0.21	21.1	11.6	23.7
H14_5	0.21	26.8	13.1	23.7
H14_6	0.19	29.8	12.1	23.7
H14_10	0.21	16.1	6.6	23.7
H15_3	0.28	24.6	10.0	60.9
H15_4	0.28	27.1	11.0	60.9
H15_5	0.29	22.8	9.4	60.9
H18_3	0.13	101.3	37.9	15.2
H18_4	0.13	89.8	39.1	15.2
H18_5	0.13	98.8	41.0	15.2
H18_6	0.13	93.2	43.1	15.2
H18_7	0.13	113.7	35.1	15.2
H19_3	0.21	39.7	12.7	62.8
H19_8	0.22	47.4	15.8	62.8
H19_9	0.22	34.1	12.7	62.8
H19_12	0.20	41.2	17.4	62.8
H19_15	0.22	39.6	13.2	62.8
H20_2	0.37	4.8	1.2	45.0
H20_3	0.37	4.4	1.3	45.0
H20_7	0.37	4.6	1.4	45.0
H20_8	0.37	6.1	1.8	45.0
H20_10	0.37	3.4	1.1	45.0
H21_3	0.17	90.2	26.6	41.0
H21_8	0.16	65.2	17.4	41.0
H21_11	0.16	86.9	25.8	41.0
H21_12	0.16	73.8	19.2	41.0
H21_13	0.16	87.2	20.1	41.0
H22_2	0.12	145.7	29.7	17.2
H22_3	0.12	144.3	30.1	17.2
H22_4	0.12	139.1	30.5	17.2
H22_5	0.12	149.7	28.9	17.2
H22_6	0.12	133.6	31.1	17.2
H25_2	0.14	99.9	29.3	45.8
H25_4	0.14	97.7	29.3	45.8
H25_5	0.14	94.2	26.2	45.8
H25_9	0.14	87.1	29.9	45.8
H25_12	0.15	83.1	24.0	45.8
H29_2	0.22	53.9	25.4	25.9
H29_8	0.23	59.1	24.5	25.9
H29_9	0.22	75.2	27.9	25.9
H29_12	0.25	53.4	23.5	25.9
H29_16	0.24	50.7	21.3	25.9
WP1285_2	0.13	78.5	22.8	13.4
WP1285_8	0.13	63.0	25.7	13.4
WP1285_10	0.13	78.3	21.4	13.4
WP1285_11	0.12	87.2	21.1	13.4
WP1285_15	0.12	84.6	25.6	13.4
WP1317_2	0.16	67.3	20.4	16.3
WP1317_6	0.16	84.5	24.7	16.3
WP1317_7	0.15	91.5	22.6	16.3

WP1317_8	0.15	95.8	25.9	16.3
WP1317_12	0.15	104.0	24.7	16.3

328

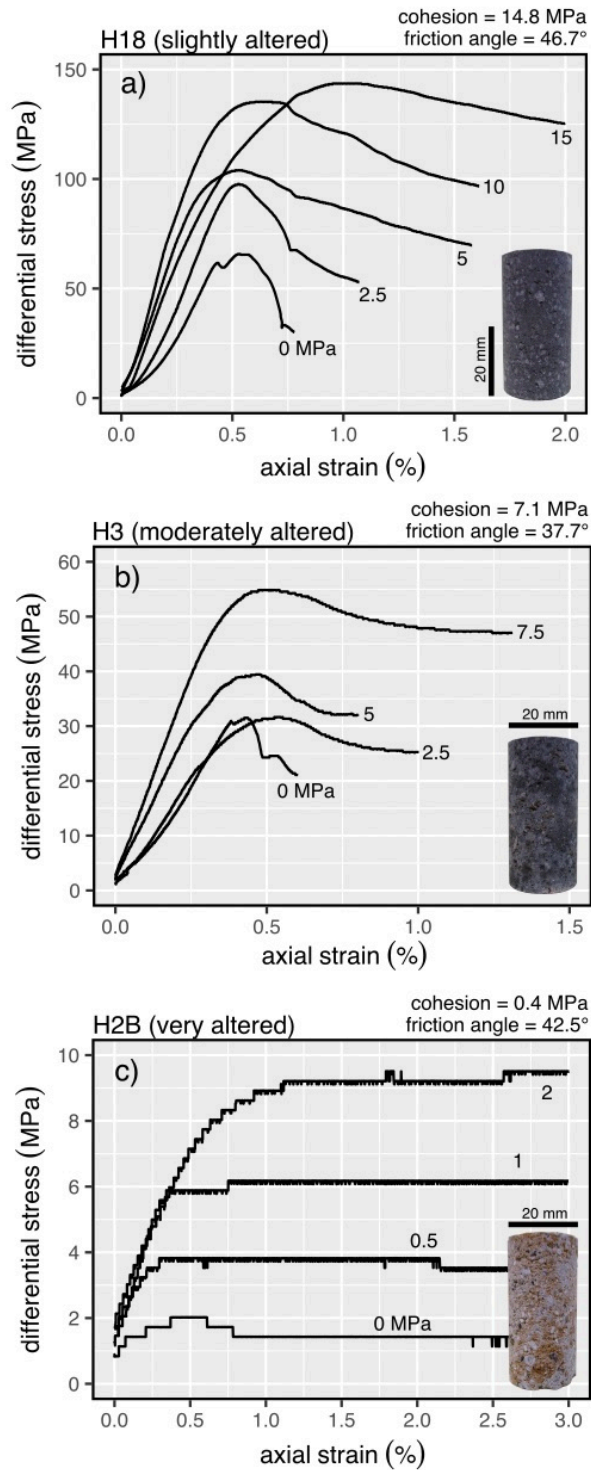
329 **Table 2.** Connected porosity, uniaxial compressive strength, Young's modulus, and the
330 percentage of secondary minerals (i.e. alteration) for the samples prepared for this study.

331 Sample locations are provided in Figure 1b (site descriptions and GPS coordinates are
332 available in the Supplementary Information).

333

334 Samples from three blocks were selected for triaxial compression experiments: a
335 relatively unaltered block (H18), a moderately altered block (H3), and a very altered block
336 (H2B). The microstructure of these samples is provided in Figure 2 and their mineral
337 componentry in Table 1. The percentages of secondary minerals (determined using the XRPD
338 data; Table 1) for these rocks are 15.2, 35.2, and 74.6%, respectively. Triaxial experiments were
339 performed on H18 samples at effective pressures of 0, 2.5, 5, 10, and 15 MPa (Figure 4a), on
340 H3 samples at effective pressures of 0, 2.5, 5, and 7.5 MPa (Figure 4b), and on H2B samples at
341 effective pressures of 0, 0.5, 1, and 2 MPa (Figure 4c) (data available in Table 3). Different
342 effective pressures were chosen for the different samples in an attempt to ensure the brittle
343 behavior required to calculate the cohesion and angle of internal friction. Blocks H18 (Figure
344 4a) and H3 (Figure 4b) were brittle over the chosen range of effective pressures. The
345 compressive strength of blocks H18 and H3 increased as a function of increasing effective
346 pressure: H18 increased from 65.4 to 143.8 MPa and H3 increased from 31.5 to 54.8 MPa as
347 effective pressure was increased from 0 to 15 and 7.5 MPa, respectively (Figure 4). The stress-
348 strain curves for block H2B are very different to those for blocks H18 and H3 (Figure 4). Subtle
349 strain softening associated with brittle behavior was observed for the experiments performed at
350 effective pressures of 0 and 0.5 MPa, but no strain softening was seen at 1 and 2 MPa,
351 suggestive of ductile behavior (Figure 4c).

352 The cohesion and angle of internal friction calculated using these triaxial data
353 (excluding the two ductile H2B experiments), necessary for the large-scale numerical modeling,
354 were 14.8 MPa and 46.7° for H18, 7.1 MPa and 37.7° for H3, and 0.4 MPa and 42.5° for H2B,
355 respectively (Figure 4). The method for determining these values is explained in the
356 Supplementary Information.
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Figure 4. Stress-strain curves for andesites from La Soufrière de Guadeloupe (Eastern Caribbean, France) deformed under different effective pressures (given by the number adjacent to each curve). (a) Relatively unaltered sample H18. (b) Moderately altered sample H3. (c) Very altered sample H2B. Differential stress is the axial stress minus the confining

363 pressure. Insets show photographs of the 20 mm-diameter cylindrical samples. The values of
 364 cohesion and internal friction angle, calculated using these data, are also provided for each of
 365 the three rocks.

366

Sample	Connected porosity	Percentage of secondary minerals	Pore fluid pressure (MPa)	Confining pressure (MPa)	Effective pressure (MPa)	Peak differential stress (MPa)
H18_11	0.13	15.2	0 (wet)	0	0	65.7
H18_10	0.14	15.2	10	12.5	2.5	97.7
H18_2	0.12	15.2	10	15	5	104.0
H18_8	0.12	15.2	10	20	10	135.2
H3_14	0.16	35.2	0 (wet)	0	0	31.5
H3_12	0.15	35.2	10	12.5	2.5	31.7
H3_4	0.15	35.2	10	15	5	39.4
H3_5	0.16	35.2	10	17.5	7.5	54.9
H2B_7	0.44	74.6	0 (wet)	0	0	2.0
H2B_8	0.44	74.6	10	10.5	0.5	3.8
H2B_6	0.39	74.6	10	11	1	-
H2B_5	0.38	74.6	10	12	2	-

367

368 **Table 3.** Experimental summary for the triaxial deformation experiments (and wet uniaxial
 369 experiments) performed on relatively unaltered (H18), moderately altered (H3), and very
 370 altered (H2B) andesites from La Soufrière de Guadeloupe (Eastern Caribbean, France).

371 Sample locations are provided in Figure 1b (site descriptions and GPS coordinates are
 372 available in the Supplementary Information) and microstructural images are provided in

373 Figure 2.

374

375 **5 Discussion**

376 5.1 Hydrothermal alteration and mineral precipitation

377 Volcanic andesitic systems discharging acid-chloride-sulfate hydrothermal fluids
 378 (Browne, 1978; Ellis and Mahon, 1977; Taran and Kalacheva, 2020) are typically characterized
 379 by either propylitic alteration, resulting from the rapid ascension of acidic fluids and the lack

380 of time available for neutralization (Taran and Kalacheva, 2020), or intermediate to advanced
381 argillic alteration in zones where fluids can circulate. The samples measured for this study were
382 sourced from (1) zones proximal to summit fumarolic vents that tap the central zone
383 characterized by the rapid ascent of steam separated from a boiling hydrothermal aquifer
384 (Brombach et al., 2000; Moretti et al., 2020) (e.g., H18–22; Table 1) and (2) from zones far
385 from currently active fumarolic zones, where the shallow groundwaters are heated via
386 conduction and mixed with Na-Cl liquids from the hydrothermal aquifer (Brombach et al.,
387 2000; Villemant et al., 2014) (e.g., those sampled from the collapse scar of the 2009 landslide:
388 H2A, H2B, H3, H4A, H5A, H6, H25, and H29; Table 1) (Figure 1b).

389 The secondary mineral assemblage in our samples (Table 1) is essentially the same as
390 that found in the debris avalanche deposits to the southwest of La Soufrière (Salaün et al., 2011).
391 However, the debris-avalanche deposits investigated by Salaün et al. (2011) contain abundant
392 smectite, whereas the samples collected for this study contain kaolinite and no smectite (Table
393 1). The presence of kaolinite, a key mineral in the argillic alteration facies, typically indicates
394 fluid-rock interaction at relatively low fluid temperatures (< 150–200 °C) and a pH of ~4.5–6
395 (Fulignati, 2020; Inoue, 1995), but can also occur in the advanced argillic alteration facies (pH
396 < 3; temperatures up to about 300 °C). The altered samples analyzed in this study therefore
397 represent a more advanced stage of hydrolysis reactions than the smectite-bearing products
398 from the Holocene debris-avalanche deposits described in Salaün et al. (2011), indicating that
399 the rocks analyzed here were altered by hydrothermal fluids with a higher H⁺ activity that led
400 to complete alkali leaching.

401 The most abundant secondary minerals are cristobalite, amorphous silica, natroalunite,
402 and kaolinite (Table 1). Cristobalite is a silica polymorph stable at low-pressure and high-
403 temperature in its cubic β-form (Schipper et al., 2020). Because it is kinetically favored,
404 metastable cristobalite is commonly found in preference to quartz in lava domes, where it

405 precipitates from a vapor-phase enriched in magmatic components or results from groundmass
406 devitrification (Horwell et al., 2013; Martel et al., 2021; Schipper et al., 2020). Our samples
407 contain tetragonal α -cristobalite, suggesting a final temperature lower than 240 °C, the nominal
408 temperature for the transition from β - to α -cristobalite (Damby et al., 2014). Schipper et al.
409 (2020) identified a lower and broader range for the β - α transition in volcanic cristobalites
410 (~150–200 °C), which overlaps with the temperature interval inferred by the presence of
411 kaolinite. The cristobalite in the studied samples precipitated onto pore walls (see Figure 2 and
412 the Supplementary Information) and so we consider that it predominately formed by vapor
413 deposition. Amorphous silica (opal-A) partially to completely replaced the plagioclase
414 phenocrysts (see Figure 2 and the Supplementary Information), as often observed in acid-sulfate
415 alteration zones (McCollom et al., 2013).

416 Alunite is a characteristic family of sulfate minerals produced by hypogene magmatic-
417 hydrothermal and steam-heated alteration or supergene processes (Rye et al., 1992). As such it
418 is commonly found in the alteration products of rhyolitic to andesitic rock compositions,
419 including surficial solfataric or acid hot spring environments. Alunite and natroalunite are the
420 two most common minerals of the alunite group (Stoffregen et al., 2000). Only the volcanic
421 bomb from the 1976–1977 eruption (sample WP1317) contains alunite, whereas natroalunite is
422 commonly observed (Table 1). Natroalunite is the high temperature end-member of the alunite
423 group and typically forms at 200–350 °C, but is also stable at 380–450 °C (Henley and Berger,
424 2011; Meyer and Hemley 1967). Natroalunite precipitates from an acidic (pH < 4) fluid with a
425 high sulfate content and high [Na]/[K] ratio (Hemley et al., 1969; Stoffregen and Cygan, 1990).
426 Natroalunite precipitation requires fluids with [Na]/[K] about 100 (at pH = 3; Deyell and
427 Dipple, 2005), much higher than measured in circulating fluids and in whole rock and
428 groundmass compositions at La Soufrière (Boudon et al., 2008; Metcalfe et al., 2021; Villemant
429 et al., 2005), suggesting that equilibrium was not obtained or that the fluids responsible are

430 different from those discharged at the surface (Deyell and Dipple, 2005). If the fluids are
431 different, acidic Na-Cl-SO₄ fluids would be required, which form H₂S and H₂SO₄ following the
432 interaction between Na-Cl waters and elemental sulfur at depth (Taran and Kalacheva, 2020).
433 This hypothesis is compatible with the observation that highly acidic chlorine-sulfate rich fluids
434 (pH < 1) have been rising and circulating within the dome as of early 1998, forming acid ponds
435 within summit craters (Komorowski et al., 2005, Rosas-Carbajal et al., 2016; Villemant et al.,
436 2014). Alternative hypotheses include the interaction of Na-Cl neutral water (water/rock
437 volume ratio >> 1) with rocks already showing advanced argillic alteration, and that the rocks
438 collected represent old and buried solfataric fields (Marini et al., 2003) resulting from previous
439 volcanic cycles.

440 We conclude that the observed secondary mineral assemblages (Table 1) are the result
441 of intense fluid-rock interactions. Advanced argillic alteration of the shallow portion of the
442 edifice was promoted by the efficient circulation of cooled (below 350 °C, and down to 150°C–
443 200 °C), acidic (pH < 4) hydrothermal fluids, possibly mixed with meteoric waters. The
444 alteration assemblages observed at La Soufrière are similar to those commonly identified at
445 domes and craters of active volcanoes worldwide (Heap et al., 2019; Zimbelman et al., 2005;
446 Yilmaz et al., 2021), although the rocks from Mt Unzen volcano (Japan) were overprinted by
447 propylitic alteration (Yilmaz et al., 2021), not observed here. Multiple reaction paths and
448 variable alteration intensity reflect a variable degree of reaction progress. For a given alteration
449 rate, reaction progress will increase as exposure to altering fluids and pathway length increase,
450 or as water flow and water/rock volume ratio decrease. Variations in these environmental
451 parameters simply reflect the heterogeneity and structural complexity of the dome and thus the
452 complex architecture of the hydrothermal system, as revealed by geological and geophysical
453 data (Brothelande et al., 2014; Jessop et al., 2021; Moretti et al., 2020; Nicollin et al., 2006;
454 Rosas-Carbajal et al., 2016; Tamburello et al., 2019). In particular, the complex 3D geometry

455 of large faults and fractures in the dome allow for the efficient drainage of fluids and the
456 compartmentalization of supergene and hypogene alteration. In such an environment: (1)
457 intermediate to advanced argillic alteration will develop preferentially along discontinuities
458 favoring fluid circulation, leading to the formation of low-strength layers dominated by clay
459 and sulfate minerals; (2) partial edifice collapse or erosion favored in these weakened zones
460 will expose pyrite-rich zones of the hydrothermal system below the water table and promote
461 the deepening of supergene oxidation and the alteration aureole (as seen in the rain-triggered
462 landslide in 2009; Rosas-Carbajal et al., 2016); and (3) H₂S will be easily released by deeper,
463 boiling fluids or by the disproportionation of magmatic SO₂ to H₂S and aqueous sulfate during
464 the condensation of magmatic vapor plume at intermediate depths.

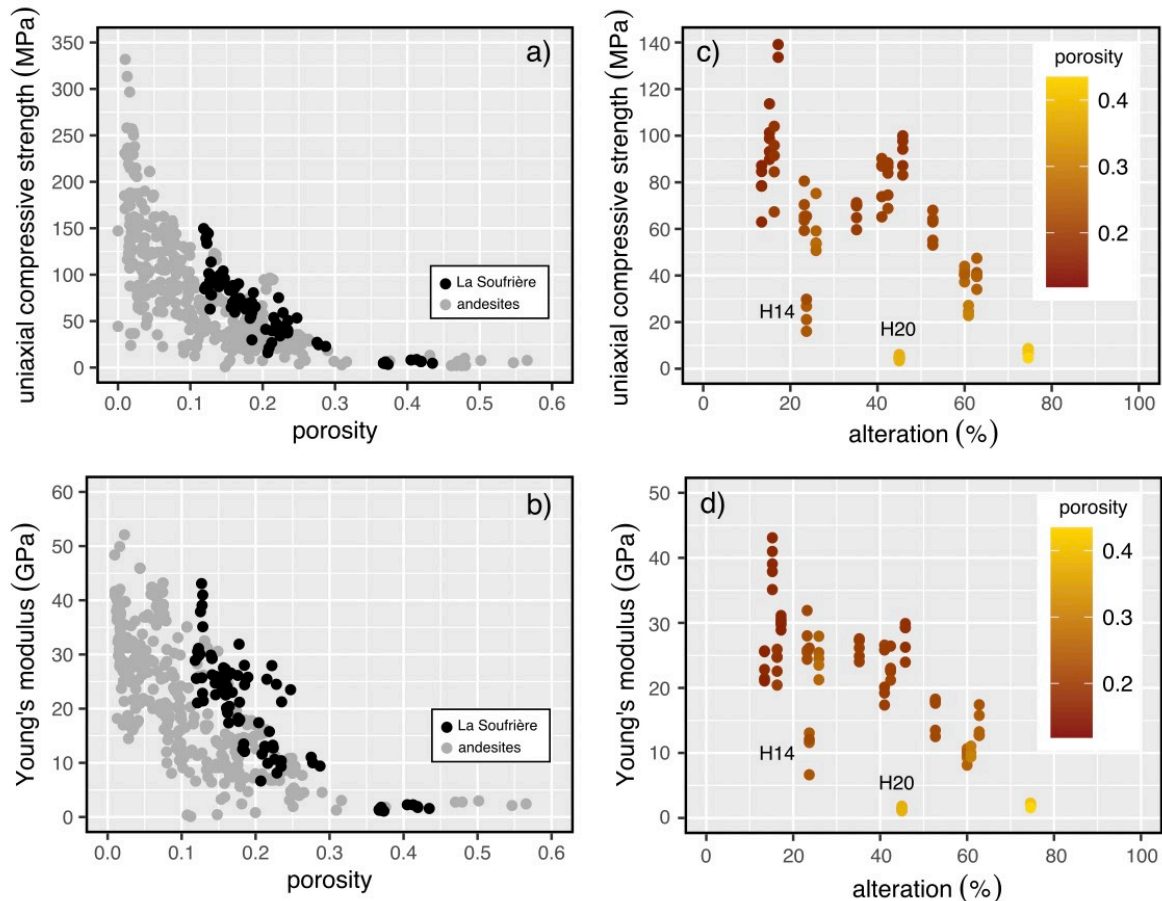
465

466 5.2 The influence of porosity and alteration on uniaxial compressive strength

467 New mechanical data show that the uniaxial compressive strength and Young's
468 modulus of variably-altered andesites from La Soufrière both decrease as a function of
469 increasing porosity (Figure 3), in accordance with previous studies on volcanic rocks (Heap et
470 al., 2014a, 2020a; Heap and Violay, 2021; Mordensky et al., 2018; Schaefer et al., 2015; Zhu
471 et al., 2016). We compare our new data with those previously published for andesites in Figure
472 5a and 5b, which show that the data for the rocks from La Soufrière are in broad agreement
473 with those previously published. This comparison shows that the andesites forming La
474 Soufrière are generally more porous, and therefore weaker (Figure 5a) and softer (Figure 5b),
475 than andesites from other stratovolcanoes (the block with the lowest porosity, H22, has a
476 porosity of 0.12). Indeed, muon tomography data have shown that the material forming the
477 dome at La Soufrière is typically low-density (typically $\leq 1500 \text{ kg.m}^{-3}$; Lesparre et al., 2012;
478 Rosas-Carbajal et al., 2017) and, therefore, likely comprises high-porosity rocks. However, we
479 note that the relative abundance of high-porosity rocks with a low strength and a low Young's

480 modulus from La Soufrière may be a consequence of sampling only 17 blocks. In other words,
481 low-porosity rocks may exist at La Soufrière and, if many more blocks were sampled, maybe
482 the data would encompass a similar range to the compiled dataset for other volcanoes.

483



484

485 **Figure 5.** (a) Uniaxial compressive strength and (b) Young's modulus as a function of
486 porosity for andesites from La Soufrière de Guadeloupe (Eastern Caribbean, France) and from
487 volcanoes worldwide (data from: Karaman and Kesimal, 2015; Heap et al., 2014b; Siratovich
488 et al., 2014; Wyring et al., 2014; Heap et al., 2015a; Mordensky et al., 2018; Harnett et al.,
489 2019; Kennedy et al., 2020; Heap and Violay, 2021). Uniaxial compressive strength (c) and
490 Young's modulus (d) as a function of alteration (the percentage of secondary minerals).
491 Symbol color indicates the connected porosity (red colors indicate low porosity and yellow

492 colors indicate high porosity). Outliers—blocks H14 and H20—are labelled on panels (c) and
493 (d).

494

495 Of interest to the goal of this study is whether hydrothermal alteration has influenced
496 the compressive strength and Young's modulus of the materials collected. Our triaxial data
497 highlight that alteration (percent of secondary minerals) reduces rock cohesion, but does not
498 systematically influence the angle of internal friction (Figure 4). Experimental studies have
499 shown that alteration can influence the strength of volcanic rock (del Potro and Hürlimann,
500 2009; Farquharson et al., 2019; Frolova et al., 2014; Heap et al., 2020b, 2021b, Mordensky et
501 al., 2018, 2019; Wyering et al., 2014). The type of alteration, porosity increasing dissolution
502 and alteration to clay (del Potro and Hürlimann, 2009; Farquharson et al., 2019; Opfergelt et
503 al., 2006; Watters and Delahaut, 1995) or porosity decreasing mineral precipitation (Heap et
504 al., 2020b, 2021), dictates whether the alteration decreases or increases the strength and
505 Young's modulus of the rock. Figure 5c and 5d show plots of uniaxial compressive strength
506 and Young's modulus, respectively, as a function of the percentage of secondary minerals for
507 the rocks from La Soufrière. The symbols in Figure 5c and 5d are color-coded to show their
508 porosity: red colors indicate low porosity and yellow colors indicate high porosity. Although
509 there are two notable outliers, Figure 5c and 5d, respectively, show that uniaxial compressive
510 strength and Young's modulus decrease as a function of increasing hydrothermal alteration.
511 Uniaxial compressive strength and Young's modulus are reduced from ~60–140 to < 10 MPa
512 and from 20–45 to < 5 GPa as alteration increases from ~16–18 to ~75% (Figure 5c and 5d).
513 The low uniaxial compressive strengths and Young's moduli of the two relatively unaltered
514 blocks that form the most obvious outliers (H14 and H20) can be explained by meso-scale
515 fractures found within sample H14 (sample photographs are provided in the Supplementary
516 Information) and the high porosity of sample H20, respectively. Although the data in Figure 5c

517 and 5d suggest that an increase in alteration results in a decrease in strength and Young's
518 modulus, it is difficult to untangle the influence of porosity. In other words, are the weak
519 samples weak because of alteration or are the more porous samples simply more altered because
520 of their higher surface area available for rock-fluid interactions? Indeed, our data do show that,
521 in general, alteration increases as a function of increasing porosity. Although laboratory-
522 controlled hydrothermal alteration experiments (Farquharson et al., 2019) are required to
523 definitively answer this question, we contend here that the observed alteration served to reduce
524 the strength and Young's modulus of these andesites from La Soufrière, regardless of their
525 porosity, due to the presence of abundant clay minerals, phases considered to reduce the
526 strength of volcanic rocks (del Potro and Hürlimann, 2009; Nicolas et al., 2020; Opfergelt et
527 al., 2006; Watters and Delahaut, 1995).

528

529 5.3 Influence of alteration on the stability of La Soufrière de Guadeloupe: large-scale numerical
530 modeling

531 We used the open-source hydro-thermo-mechanical modeling code LaMEM
532 (Lithosphere and Mantle Evolution Model; <https://bitbucket.org/bkaus/lamem/src/master/>;
533 Kaus et al., 2016), which models the non-linear, visco-elastoplastic deformation of rocks, to
534 investigate the influence of upscaling and alteration on the stability of La Soufrière. This model
535 has recently been used to model large-scale volcano deformation (Heap et al., 2021a).
536 Mathematically, the model solves a coupled system of conservation equations, the conservation
537 equations of momentum and mass. The various deformation mechanisms are connected in an
538 additive constitutive relationship, where the total strain rate is the sum of the individual strain
539 rates of elastic, viscous, and plastic deformation. LaMEM uses the Drucker-Prager failure
540 criterion to determine the magnitude of the plastic strain rate. Numerically, LaMEM uses a
541 staggered-grid finite difference method for discretization and a marker-and-cell method to

542 assign and track rock properties within a Eulerian advection framework (Harlow and Welch,
543 1965). This approach enables the model to attain the large deformations required in, for
544 example, salt tectonics, continental collision, and magmatic systems (Baumann et al., 2014;
545 Pusok and Kaus, 2015; Reuber et al., 2018). The ability to model large deformations is seen
546 here as an advantage of LaMEM over commonly-used commercial packages such as FLAC (by
547 Itasca Consulting Group), RS2 (by Rocscience), and PLAXIS 2D (by Virtuosity). We refer the
548 reader to Kaus et al. (2016) and the Supplementary Information for a more detailed description
549 of LaMEM.

550 Models were run on simplified 2D North-South cross-sections of the volcano (643 km
551 Easting), built using the geomIO tool (Bauville and Baumann, 2019). The model domain was
552 first separated into deformable and non-deformable domains, guided by geological
553 observations and the recent electrical survey of Rosas-Carbajal et al. (2016). The models have
554 a dimension of 2000×800 m, a model resolution of 4×3 m, free-slip boundary conditions at
555 the sides, and a free surface at the top. All our models assumed an entirely liquid saturated
556 volcano. The average topographic height was used as the reference surface for determining the
557 hydrostatic pressure, p_h , using $p_f = p_h + \lambda_i (p_{lith} - p_h)$, where p_f and p_{lith} are the pore
558 fluid and lithostatic pressure, respectively, and λ_i is the pore fluid pressure ratio. Following
559 Heap et al. (2021a), we assumed a value of 0.2 for λ_i (Table 4). To account for the uncertainty
560 in λ_i , additional models using $\lambda_i = 0.4$ are provided in the Supplementary Information. We
561 highlight that, although the assumption of complete volcano saturation is an oversimplification
562 (Ball et al., 2015, 2018; Hurwitz et al., 2003), the main goal of our modeling was to understand,
563 all else being equal, whether hydrothermal alteration is a contributing factor to the deformation
564 observed at La Soufrière. Further, the distribution of liquid-saturated and dry zones within the
565 volcano is currently poorly constrained. However, using the average topographic height as the

566 reference surface resulted in a “dry” volcano summit with a pore fluid pressure of zero (see
 567 Supplementary Information for further details).

568 By modifying the segmentation of the deformable domain, we prepared three cross-
 569 sections: (1) a homogeneous volcano, (2) a volcano containing alteration zones identified by
 570 the electrical conductivity data of Rosas-Carbajal et al. (2016), and (3) a volcano in which we
 571 artificially enlarged the highly altered zone identified by Rosas-Carbajal et al. (2016). Zones
 572 within the volcano cross-sections were designated as either slightly altered (V3), moderately
 573 altered (V2), or highly altered (V1) (Figures 6a-d). Zones V3, V2, and V1 were assigned the
 574 mechanical properties of samples H18, H6, and H2B, respectively, the three blocks chosen to
 575 represent relatively unaltered, moderately altered, and highly altered rock (Figure 2; Table 1).
 576 The laboratory data used in the modeling were: water-saturated bulk density, Young’ modulus,
 577 cohesion, and the angle of internal friction. Poisson’s ratio, not measured here, was assumed to
 578 be 0.2 (as recommended by Heap et al., 2020a). The model parameters for V3, V2, and V1 are
 579 provided in Table 4. As the models are dynamic, we must decide when the models reach a
 580 quasi-steady state and can be compared (see Baumann and Kaus, 2015). Here, we stop the
 581 models after an initial deformation phase that is dominated by an elastic response. For each of
 582 the three modeled scenarios, we ran one model in which we used the laboratory data and one
 583 model in which we used upscaled values of the laboratory data (Table 4).

584

	V1 (highly altered)	V2 (moderately altered)	V3 (relatively unaltered)
Laboratory-scale input parameters			
Bulk density(kg/m ³)	1900	2400	2500
Young’s modulus (GPa)	0.4	9.6	20.3
Poisson’s ratio	0.2	0.2	0.2
Cohesion (MPa)	0.4	7.1	14.8
Internal friction angle (°)	42.5	37.7	46.7
Pore fluid pressure ratio	0.2	0.2	0.2
Upscaled input parameters			
Bulk density(kg/m ³)	1900	2400	2500
Young’s modulus (GPa)	0.2	3.9	8.3
Poisson’s ratio	0.3	0.3	0.3

Cohesion (MPa)	0.67	1.09	2.02
Internal friction angle (°)	30.1	33.5	46.4
Pore fluid pressure ratio	0.2	0.2	0.2

585

586

Table 4. Laboratory-scale and upscaled input parameters for the numerical modeling

587

performed using hydro-thermo-mechanical modeling code LaMEM (Lithosphere and Mantle

588

Evolution Model).

589

590

The laboratory-scale Young's modulus and Poisson's ratio were upscaled using

591

empirical relationships for Young's modulus (Hoek and Diederichs, 2006) and Poisson's ratio

592

(Vásárhelyi, 2009) that account for rock-mass structure using the Geological Strength Index

593

(GSI), as described in Heap et al. (2020a). Following Heap et al. (2020a), GSI was assumed to

594

be 55 and the damage parameter in the Hoek-Diederichs equation was considered zero. The

595

cohesion and angle of internal friction were upscaled by transforming the generalized form of

596

the Hoek-Brown failure criterion to the Mohr-Coulomb failure criterion (as in Hoek et al.,

597

2002). We used the mean uniaxial compressive strength of samples H18, H6, and H2B for

598

layers V3, V2, and V1, respectively. The empirical fitting parameter m_i required for the

599

generalized Hoek-Brown failure criterion was determined from the triaxial data (Figure 4) using

600

RocData (Rocscience; <https://www.rocscience.com>). To do so, we again assumed a GSI of 55

601

and a damage parameter of zero. Since a depth is required to upscale the cohesion and angle of

602

internal friction, we selected a depth half of that between the summit of the volcano and the top

603

of the undeformable zone in the model (275 m). The upscaled model parameters for V3, V2,

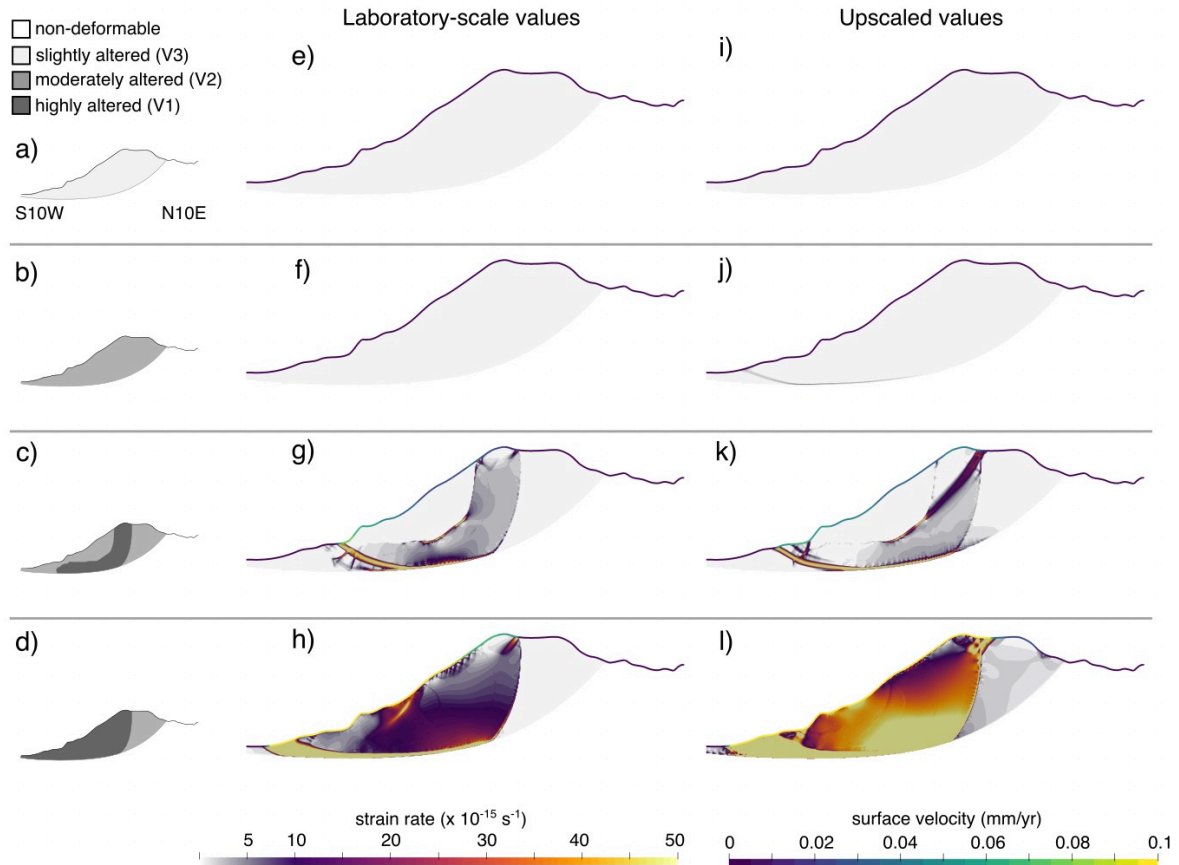
604

and V1 are provided in Table 4. A more detailed description of the upscaling methods can be

605

found in the Supplementary Information.

606



607

608 **Figure 6.** Results of the large-scale 2D numerical modeling designed to better understand the

609 influence of alteration on the stability of La Soufrière de Guadeloupe Eastern (Caribbean,

610 France). Panels (a-d) show the model setup, in which white, light grey, grey, and dark grey

611 indicate non-deformable, slightly altered (V3), moderately altered (V2), and highly altered

612 (V1) zones. Panels (e-h) and (i-l) show the model results for the models run using laboratory-

613 scale and upscaled mechanical properties, respectively. The color of the line representing the

614 surface of the volcano indicates the surface velocity (mm/yr) and the color inside the volcano

615 indicates the strain rate (s^{-1}).

616

617 The model results, the velocity magnitude at the surface and the 2nd invariant of the

618 deviatoric strain rate tensor, which we refer to as the strain rate, within the deformable domain,

619 are presented in Figure 6. The strain rate allows us to understand whether the deformation is

620 widespread or localized, and the velocity at the surface can be compared to geophysical data.
621 Figures 6e-h and 6i-l show model runs using the laboratory-scale and upscaled values,
622 respectively. We first highlight that our results emphasize the importance of using upscaled
623 values for large-scale modeling. Not only are the strain rates and velocities much lower when
624 using the laboratory-scale values, but the deformation also manifests slightly differently (Figure
625 6).

626 To discuss the influence of alteration, we will restrict our discussion to the upscaled
627 models (Figures 6i-l). We first note that strain rates and velocities are low in the homogeneously
628 altered volcanoes, that deformation is largely restricted to the surface and the base of southern
629 slope, and that there is little difference in strain rates and velocities when the alteration is
630 increased from slightly to moderately altered (Figures 6i and 6j). When the highly-altered zone
631 guided by the recent electrical survey of Rosas-Carbajal et al. (2016) is included, the strain rates
632 and velocities increase significantly (Figure 6k). Surface velocities increase from < 0.01 to
633 ~ 0.05 mm/yr and a large listric, concave-upward sliding surface emerges, which stretches from
634 the top of the dome to near the base of the southern flank (Figure 6k). A localized slip surface
635 at the base of the deformable domain and two smaller slip surfaces also emerge near the base
636 of the southern flank (Figure 6k). Therefore, the inclusion of the highly altered zone, the
637 consequence of pervasive alteration resulting from the circulation of acid hydrothermal fluids
638 (Rosas-Carbajal et al., 2016), has significantly increased volcano deformation rates and
639 therefore decreased the stability of the volcano. As outlined above, recent manifestations of
640 increased hydrothermal activity, such as fumaroles and hot springs (Moretti et al., 2020; Jessop
641 et al., 2021), also provide evidence that the measured ground deformation is, in part, due to the
642 alteration of the rock forming the dome. Continuous and campaign Global Navigation Satellite
643 System (GNSS) measurements over ~ 20 years show that the dome is sliding at a rate of 0.3–
644 0.7 cm/yr (Moretti et al., 2020). Therefore, although our modeling captures the dynamics of the

645 deformation of the southern flank, our simplified 2D model setup (Figure 6) underestimates the
646 deformation measured at the volcano. This underestimation is likely the result of the presence
647 of large-scale faults, discontinuities, extremely weak clay-rich layers, and zones of high pore
648 pressure, not included in our modeling. Indeed, the strain rate in the southern flank increases
649 when we increase the pore fluid pressure ratio, λ_i , from 0.2 to 0.4 (see additional models
650 presented in the Supplementary Information). Future models will aim to incorporate these
651 features.

652 In a final step, and to further explore how alteration can influence volcano stability, we
653 modeled a scenario in which the highly altered zone identified by Rosas-Carbajal et al. (2016)
654 was artificially enlarged to incorporate the entire southern flank (Figure 6d). Increasing the size
655 of the alteration zone increases the surface velocities from ~ 0.05 to ~ 0.1 mm/yr and
656 significantly increases the strain rate within the flank (Figure 6l). The slip surface that at the
657 base of the deformable domain is now larger and has a higher strain rate, and the entire southern
658 flank is now characterized by higher strain rates (Figure 6l). In addition, we also observe that
659 expanding the alteration zone likely increases the potential collapse volume (Figure 6l). Based
660 on these models, increasing the size of the highly altered zone (i.e. to a size larger than the
661 present-day altered zone) will significantly increase the risk posed by the volcano. Deformation
662 snapshots from additional models, presented in the Supplementary Information, show the
663 progression of deformation for the present-day and increased-alteration scenarios from 5 to 100
664 kyrs (using upscaled physical and mechanical properties). These additional models also show
665 that increasing the extent of the alteration increases the rate of volcano spreading.

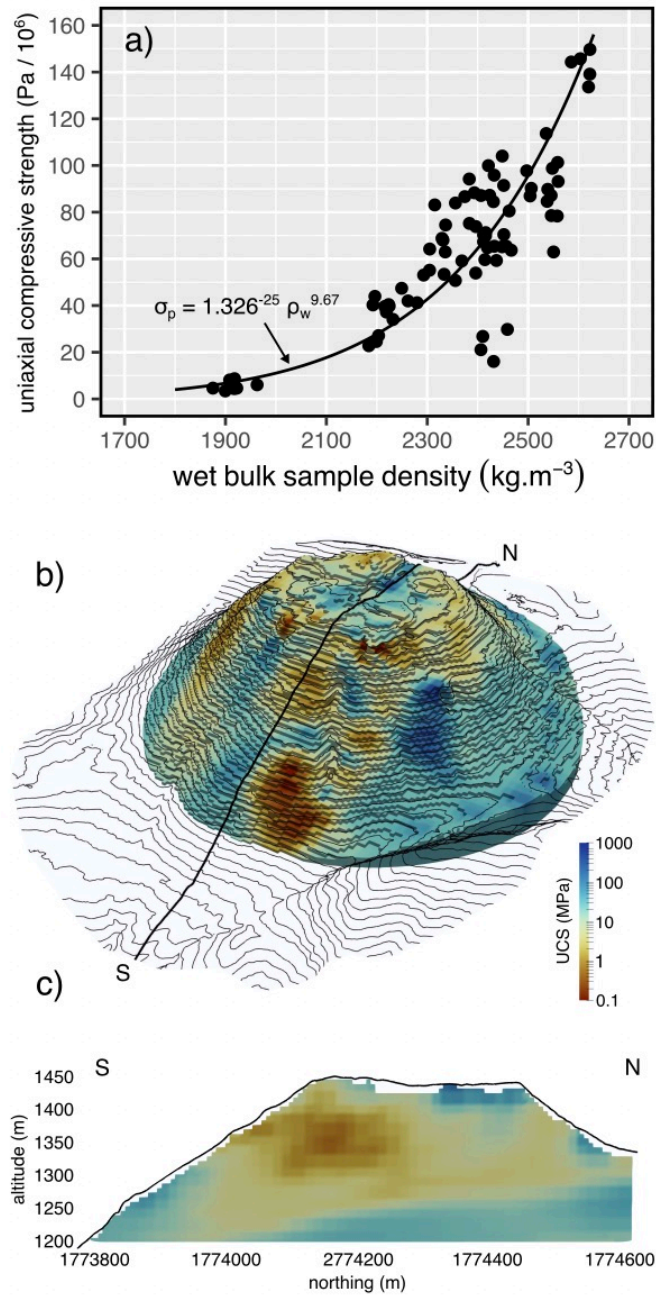
666

667 5.4 Three-dimensional strength model for La Soufrière de Guadeloupe

668 The results of the previous section highlight the importance of combining geophysical
669 surveys, laboratory data, and large-scale modeling to evaluate the influence of alteration on

670 volcanic flank instability. Our 2D model geometry is based on qualitative interpretations from
671 an electrical conductivity model to delimit different regions based on their degree of alteration.
672 A quantitative approach would require a petrophysical relation to convert the conductivity
673 values to rock strength and cohesion, but such relations are highly non-unique as electrical
674 conductivity depends on many factors including clay content, fluid temperature, and
675 composition. Instead, rock strength is controlled to a first order by porosity, as shown in Figure
676 3b, and thus to the bulk rock density.

677 Here, we combine a 3D model of the bulk density of the present-day lava dome (Rosas-
678 Carbajal et al., 2017) with an empirical power-law function relating wet bulk density and
679 uniaxial compressive strength, σ_p , where $\sigma_p = 1.326^{-25} \rho_w^{9.67}$ (σ_p in Pa and ρ_w in kg/m^3) to
680 create a 3D strength map of the volcano (Figure 7). The empirical function was specifically
681 fitted to our new data for La Soufrière (Table 2; Figure 7a). Because the distribution of liquid-
682 saturated and dry zones within the volcano is currently poorly constrained, we chose to use the
683 bulk density of our samples saturated with liquid water, a scenario we consider better represents
684 the natural state of the volcano than a completely dry volcano. The 3D bulk density model of
685 Rosas-Carbajal et al. (2017) was obtained by jointly inverting gravity data and muon data
686 acquired using three simultaneous muon detectors that scanned the lava dome from different
687 points of view. Muon tomography is a novel technique, which uses the attenuation experienced
688 by cosmic muons as they pass through the volcano to estimate its density (Tanaka et al., 2010).
689 Whereas a single detector can provide 2D profiles of the average density of the volcano in the
690 different directions sampled, multiple simultaneous detectors from different points of view can
691 be used to create a 3D model of the volcano (Bonechi et al., 2020).



692

693 **Figure 7.** (a) Uniaxial compressive strength as a function of wet bulk sample density for
 694 andesites from La Soufrière de Guadeloupe (Eastern Caribbean, France). Curve is the best-fit
 695 empirical power-law function. (b) 3D uniaxial compressive strength map of La Soufrière de
 696 Guadeloupe (Eastern Caribbean, France). The black line, which runs North-South, indicates
 697 the position of the cross-section in panel (c). (c) North-South cross-section through La
 698 Soufrière showing the distribution of uniaxial compressive strength. Color scale is the Roma

699 (seismic tomography) color scale of Fabio Cramerì (Cramerì et al., 2020). UCS – uniaxial
700 compressive strength.

701

702 Figure 7 highlights that large portions of the dome are characterized by very low values
703 of strength (< 5 MPa). In particular, the southern flank of the volcano hosts a significant weak
704 zone (Figures 7b and 7c). This low-strength zone is coincident with the location of the main
705 hydrothermal reservoir at La Soufrière, as exposed by electrical conductivity measurements
706 (Lesparre et al., 2012; Rosas-Carbajal et al., 2016). Based on (1) the influence of
707 hydrothermally altered rock on edifice stability and spreading rates at La Soufrière revealed by
708 the numerical modeling in the previous section, and (2) the large volume of low-strength rock
709 in the La Soufrière dome highlighted by our 3D strength model, we conclude that the stability
710 of La Soufrière is compromised by its high degree of rock alteration, and that the rate of volcano
711 spreading and the probability of an eventual partial edifice collapse may increase solely due to
712 the ongoing alteration caused by circulating hydrothermal fluids, and may not necessarily
713 require intensified hydrothermal and/or magmatic activity.

714 Based on our 3D strength model, we can provide estimates for the volumes of rock
715 within the lava dome below a certain strength threshold. For example, we estimate that rock
716 with a strength lower than 1 and 10 MPa comprises minimum dome volumes of 2.7×10^6 and
717 $36 \times 10^6 \text{ m}^3$, respectively, corresponding to at least 4% and 55% of the dome above 1200 m. In
718 other words, more than half of the La Soufrière dome could be considered to have a low
719 strength. We highlight that these volume and percentage estimates should be considered
720 conservative due to the positioning of the three telescopes at La Soufrière, which provides better
721 coverage of the southern flank of the volcano (Rosas-Carbajal et al., 2017). It is likely,
722 therefore, that the volumes and percentages of weak rock within the dome at La Soufrière are
723 higher than estimated here.

724 Although the strength map shown in Figure 7 is currently simplistic (for example, it is
725 well known that strength increases with depth in the brittle field; Heap et al., 2015b; Figure 4),
726 this 3D model permits the quick identification of zones within the volcano that could promote
727 instability and more experimental data, when available, will allow us to account for depth and
728 other factors that can influence rock mechanical behavior, such as temperature and saturation
729 state (Heap and Violay, 2021).

730

731 **6 Conclusions**

732 Hydrothermal alteration is often invoked as a mechanism that greatly contributes to
733 volcano instability (Reid et al., 2001; van Wyk de Vries et al., 2000). However, owing to the
734 paucity of experimental data and the difficulty in accurately imaging subsurface alteration
735 zones, the rock properties and the shape and size of the alteration zones required for large-scale
736 modeling are often poorly constrained or even assumed. Here we use a combination of
737 laboratory and geophysical data to numerically model the influence of hydrothermal alteration
738 on volcano stability, using La Soufrière de Guadeloupe as a case study. Our numerical modeling
739 suggests that hydrothermal alteration can significantly increase the deformation rate within the
740 flanks of the volcano, and also increase the potential collapse volume.

741 The volcano stability modeling and 3D strength map presented herein, firsts for La
742 Soufriere, help to validate the hypothesis forwarded by the numerous geophysical studies
743 performed at the volcano: that the identified hydrothermal reservoir characterized by altered
744 materials promotes volcano instability that could result in collapse (Bouligand et al., 2016;
745 Brothelande et al., 2014; Lesparre et al., 2012; Nicollin et al., 2006; Rosas-Carbajal et al., 2016,
746 2017). As discussed above, eventual partial edifice collapse at La Soufrière may not necessarily
747 require intensified hydrothermal and/or magmatic activity. Not only would the debris avalanche
748 resulting from partial or major dome collapse threaten the inhabited areas surrounding the

749 volcano (Le Friant, 2006; Peruzzetto et al., 2019), but collapse could also depressurize the
750 hydrothermal system leading to steam-driven eruptions, dangerous laterally-directed
751 explosions, and associated high energy pyroclastic density currents, similar to the activity
752 during the 1976–1977 volcanic crisis (Feuillard et al., 1983; Komorowski et al., 2005). We also
753 highlight that the fragmentation threshold (the overpressure required for complete
754 fragmentation upon depressurization), the proportion of fines produced as a result of
755 fragmentation, and the ejection speed are also influenced by hydrothermal alteration (Mayer et
756 al., 2015, 2016; 2017; Montanaro et al., 2016), reinforcing the need for a multidisciplinary
757 approach to unravel how hydrothermal alteration affects volcanic hazards (see also de Moor et
758 al., 2019; Mick et al., 2021).

759 Our modeling and 3D strength map therefore highlight the importance of monitoring
760 the extent and evolution of hydrothermal alteration, and the formation of low-strength altered
761 layers, at La Soufrière and at other active volcanoes worldwide. Such monitoring can be
762 achieved, for example, using geophysical methods such as electrical tomography (Ahmed et
763 al., 2018; Byrdina et al., 2017; Ghorbani et al., 2018; Rosas-Carbajal et al., 2016) and muon
764 tomography (Lesparre et al., 2012; Rosas-Carbajal et al., 2017), near-surface seismic imaging
765 (Amoroso et al., 2018), thermal and gas monitoring (de Moor et al., 2019; Edmonds et al., 2003;
766 Jessop et al., 2021; Moretti et al., 2020; Tamburello et al., 2019), geological mapping (van Wyk
767 de Vries et al., 2000), deformation monitoring (Moretti et al., 2020), magnetic methods (Finn
768 et al., 2007), and remote and/or ground-based optical and spectroscopic methods (Crowley and
769 Zimbelman, 1997; Darmawan et al., 2018; John et al., 2008; Kereszturi et al., 2020; Mueller et
770 al., 2021). Further, although we document a reduction in strength as a function of alteration
771 (Figure 5c), alteration can also increase strength and potentially promote volcanic instability by
772 creating zones of high pore fluid pressure (Heap et al., 2021a; Reid, 2004). As noted in Heap
773 et al. (2021a), it is important not only to map the extent and evolution of hydrothermal

774 alteration, but also the locations in the volcano prone to porosity increasing and porosity
775 decreasing alteration.

776 We have also shown that microtextural and mineralogical analysis of host-rock material
777 erupted during phreatic explosive eruptions, as well as material from landslides and debris-
778 avalanche deposits, can provide highly valuable insights into the nature, extent, and dynamics
779 of alteration processes and fluid circulation in the pre-collapse edifice (see also Boudon et al.,
780 1998; Komorowski et al., 2010; Salaün et al., 2011). Hence, we highlight that it is also important
781 to quantify the mineral content of new material exposed by landslides and collapses, as well as
782 their deposits, and to continuously monitor fluid compositions and water table fluctuations
783 (Hurwitz et al., 2003) with respect to the geometry of weakened low-strength structures that are
784 characteristic of the internal framework volcanic edifices hosting hydrothermal systems. A
785 better understanding of the spatio-temporal evolution of hydrothermal alteration, and how it
786 can influence volcano stability using large-scale modeling, will help improve the mitigation of
787 volcanic hazards at active volcanoes worldwide.

788

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804

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