



HAL
open science

Transition from stable column to partial collapse during the 79 cal CE P3 Plinian eruption of Mt. Pelée volcano (Lesser Antilles)

Guillaume Carazzo, Stephen Tait, Audrey Michaud-Dubuy, Allan Fries, Edouard Kaminski

► To cite this version:

Guillaume Carazzo, Stephen Tait, Audrey Michaud-Dubuy, Allan Fries, Edouard Kaminski. Transition from stable column to partial collapse during the 79 cal CE P3 Plinian eruption of Mt. Pelée volcano (Lesser Antilles). *Journal of Volcanology and Geothermal Research*, 2020, 392, pp.106764. 10.1016/j.jvolgeores.2019.106764 . insu-02920453

HAL Id: insu-02920453

<https://insu.hal.science/insu-02920453v1>

Submitted on 21 Jul 2022

HAL is a multi-disciplinary open access archive for the deposit and dissemination of scientific research documents, whether they are published or not. The documents may come from teaching and research institutions in France or abroad, or from public or private research centers.

L'archive ouverte pluridisciplinaire **HAL**, est destinée au dépôt et à la diffusion de documents scientifiques de niveau recherche, publiés ou non, émanant des établissements d'enseignement et de recherche français ou étrangers, des laboratoires publics ou privés.



Distributed under a Creative Commons Attribution - NonCommercial 4.0 International License

1 **Transition from stable column to partial collapse during the 79 cal CE P3**

2 **Plinian eruption of Mt Pelée volcano (Lesser Antilles)**

3

4 Guillaume Carazzo^{1,*}, Stephen Tait¹, Audrey Michaud-Dubuy¹, Allan Fries^{2,3}, Edouard
5 Kaminski¹

6

7 ¹Université de Paris, Institut de physique du globe de Paris, CNRS, F-75005 Paris, France

8 ²Laboratoire Magmas et Volcans, Université Clermont Auvergne-CNRS-IRD, OPGC,
9 Clermont-Ferrand, France

10 ³Now at: Université de Genève, Section of Earth and Environmental Sciences, Genève,
11 Switzerland

12 *Corresponding author: Guillaume Carazzo, carazzo@ipgp.fr

13 ORCID Number: orcid.org/0000-0003-0395-2595

14 **Highlights**

- 15 - We determine the eruptive sequence of the P3 Plinian eruption of Mt Pelée volcano
- 16 - New radiocarbon dating measurements provide a refined age of 79 ± 21 cal CE
- 17 - We estimate key eruptive parameters of this VEI 5 eruption in the transitional regime
- 18 - Increasing eruption rate and decreasing gas content led to partial column collapse
- 19 - The transition occurred at conditions well predicted by our 1D theoretical model

20

21 **Keywords**

22 Mt Pelée volcano, Plinian eruption, tephra fallout, partial collapse, tephrostratigraphy,
23 radiocarbon dating

24

25 **Abstract**

26 Explosive volcanic eruptions commonly form sustained Plinian columns that collapse at some
27 stage producing dangerous pyroclastic density currents (PDC) on the ground. Numerical and
28 laboratory models of volcanic plumes show that the conditions leading to total column
29 collapse are strongly controlled by the amount of exsolved gas at the source and the mass
30 eruption rate. However, column collapse is rarely total and the volcanic jet often separates in
31 a dense collapsing part feeding PDC and a buoyant rising plume spreading volcanic gases and
32 pyroclasts in the atmosphere. This transitional regime has been directly observed and/or
33 inferred from the structure of the deposits for several past eruptions, but the number of cases
34 for which the partial collapse regime is described in detail, including the 79 CE Vesuvius, and
35 the 186 CE Taupo eruptions, remains too small to fully constrain physical models. Here, we
36 present a detailed reconstruction of the time evolution of the P3 eruption at Mt Pelée volcano
37 (Martinique, Lesser Antilles) that underwent partial column collapse in order to discuss the
38 mechanisms controlling the eruption dynamics and improve the volcanological database on

39 transitional eruptions. The P3 eruptive succession consists of seven major phases that
40 produced a total of 1 km^3 dense rock equivalent (DRE) of deposits (i.e., VEI 5 event), starting
41 with a thick pumice fall deposit (0.1 km^3 DRE) overlain by alternating pyroclastic density
42 current (PDC) (0.7 km^3 DRE) and pumice fall deposits (0.2 km^3 DRE). We use physical
43 models together with field data on deposit dispersal, thickness, and grain-size distribution to
44 reconstruct the dynamical evolution of the volcanic column. Our results show that the mass
45 eruption rate (MER) increased from 1.2×10^8 to $1.7 \times 10^8 \text{ kg s}^{-1}$ during the initial phase
46 producing a 28 to 30 km-high Plinian plume. The MER later reached up to $2.5 \times 10^8 \text{ kg s}^{-1}$
47 and the column entered the partial collapse regime characterized by the formation of a small
48 (i.e., 12 to 17 km-high) ash plume and contemporaneous PDCs mainly channelized in three
49 paleo-valleys. These estimates are used together with published data on magmatic water
50 contents in glass inclusions to decipher the mechanisms leading to partial collapse. The P3
51 eruption column collapsed due to an increase in mass eruption rate and a decrease in gas
52 content. A similar evolution was also inferred for the 1300 CE P1 and 280 CE P2 eruption
53 deposits, revealing a systematic behavior in the recent Plinian eruptions of Mt Pelée volcano.
54 The comparison of model predictions of column collapse and field data reveals a good
55 agreement for the P1, P2, P3 and Taupo eruptions, but not for the 79 CE Vesuvius eruption
56 where thermal disequilibrium between gas and pyroclasts most likely strongly affected the
57 column dynamics.

58

59 **1. Introduction**

60 Plinian eruptions are amongst the most powerful and destructive volcanic events on Earth.
61 These explosive eruptions generally form a sustained stable column that can rise up to 40 km
62 in the atmosphere (e.g., the 1991 Pinatubo eruption) and spread out laterally as an umbrella
63 cloud carrying pyroclasts at large distances from the volcano ([Koyaguchi and Tokuno, 1993](#)).

64 Abrupt or progressive changes in eruptive conditions at the base of a Plinian column can lead
65 to a radically different behavior in which the column collapses as a turbulent fountain
66 producing hazardous pyroclastic density currents (PDC) that may reach and destroy populated
67 areas ([Wilson et al., 2014](#)). One important goal in physical volcanology is to identify the key
68 processes affecting the column stability in order to improve hazard assessment.

69 Recent real-time observations of volcanic plumes and field-based studies on fallout
70 tephra from past eruptions show that explosive eruptions commonly have a complex behavior
71 where both a stable Plinian column and a collapsing fountain feeding PDC are
72 contemporaneous ([Carazzo et al., 2015 and references therein](#)). This transitional regime of
73 partial column collapse occurs for Plinian eruptions ranging in size from small to large (e.g.,
74 the 79 CE Vesuvius ([Shea et al., 2011](#)), 186 CE Taupo ([Wilson and Walker, 1985](#)), 1150
75 Quilotoa ([Di Muro et al., 2008](#)), 1912 Novarupta-Katmai ([Fierstein and Hildreth, 1992](#)), and
76 1963 Agung ([Self and Rampino, 2012](#)) eruptions). Detailed mapping and analysis of eruptive
77 products emplaced during the transitional regime provide valuable data to better understand
78 the mechanisms controlling the partitioning of a Plinian column, and can be used to robustly
79 test numerical and analog models of explosive volcanic eruptions ([Di Muro et al., 2004](#);
80 [Carazzo et al., 2015](#); [Suzuki et al., 2016](#)). However, the number of natural cases for which the
81 partial collapse of an initially stable column is described in detail (i.e., in terms of time
82 variation in mass eruption rate, gas content, grain-size distribution and exit velocity) remains
83 too small to fully constrain theoretical models of volcanic plumes ([Michaud-Dubuy et al.,](#)
84 [2018](#)). With an aim of increasing the inventory of well-constrained past eruptions, we have
85 undertaken a long-term study of the Plinian eruptions of Mt Pelée volcano (Lesser Antilles) as
86 they systematically produced sustained columns that ultimately collapsed ([Traineau et al.,](#)
87 [1989](#)). Because the Mt Pelée eruptions share many characteristics including similar pre-
88 eruptive magma storage conditions ([Martel et al., 1998](#)), they can be used to decipher the

89 mechanisms controlling the behavior of a Plinian plume. In previous studies, we reconstructed
90 the dynamics of the 1300 CE P1 (Carazzo et al., 2012), and 280 CE P2 Plinian eruptions
91 (Carazzo et al., 2019) focusing on the causes for column collapse. We now extend this work
92 to study the P3 eruption deposits whose structure suggests the entrance into the partial
93 collapse regime at some stage of the eruption (Traineau et al., 1989; Wright et al., 2016),
94 making it an excellent candidate to improve the volcanological database on transitional
95 eruptions.

96 The P3 eruption is described in the literature (Roobol and Smith, 1976; Westercamp
97 and Traineau, 1983; Traineau et al., 1989; Wright et al., 2016) but the exact sequence of
98 events and the eruptive parameters are currently poorly known. Here, we use data from a new
99 comprehensive field study together with physical models to reconstruct the detailed time
100 evolution of this eruption and to discuss the mechanisms responsible for the transition from
101 stable column to partial collapse. We identify the eruptive conditions leading to this
102 transitional regime and compare the results to previous work on P3, and on younger Plinian
103 eruptions in Martinique (P1 and P2). We then compare our estimated eruptive parameters for
104 the column at collapse with theoretical predictions from a model of volcanic plumes.

105

106 **2. Overview of the P3 eruption**

107 The P3 deposits were first identified by Roobol and Smith (1976) as a thick pumice fallout
108 unit on the western flank of Mt Pelée volcano standing above the 2406 and 2447 yr BP
109 horizons of their stratigraphic sections. Westercamp and Traineau (1983) provided the first
110 age, synthetic stratigraphic section and distribution map of the P3 deposits (Fig. 1a). The P3
111 Plinian eruption is currently dated at 2010 ± 140 yr BP based on the average of five ^{14}C ages
112 (Westercamp and Traineau, 1983; Traineau et al., 1989). According to these studies, the
113 eruption began with the formation of a sustained Plinian column that covered the western

114 flank of the volcano with a meter of coarse lapilli, lithic-rich, pumice fallout deposit (phase
115 P3₁). This initial stage was immediately followed by the production of high-concentration
116 PDC in the Grande Savane, Habitation Depaz and Falaise River paleo-valleys (Fig. 1a). The
117 second eruptive phase identified by Westercamp and Traineau (1983) consisted of a sustained
118 Plinian column that spread fine lapilli pumice to the south with a striking elongated axis of
119 dispersion (phase P3₂). The P3 eruption ended up with the formation of a pumice ash-rich bed
120 (phase P3₃) interbedded with high-concentration PDC deposits (Fig. 1a). Later, Traineau et al.
121 (1989) excluded the second phase (P3₂) whose distal deposits were found below tephra from
122 an eruption older than P3₁. Recent work by Michaud-Dubuy et al. (2019) confirmed that the
123 deposits named P3₂ by Westercamp and Traineau (1983) are not interbedded between P3₁ and
124 P3₃ but belong to a much older event, the 13.5 ka cal BP Bellefontaine Plinian eruption.

125 Wright et al. (2016) studied the distal products of a Plinian eruption dated between
126 1800 and 2200 yr BP (their subunit d), which corresponds to the P3 eruption identified by
127 Westercamp and Traineau (1983). Their synthetic stratigraphic section includes a basal
128 pumice-rich low-concentration PDC (named 'surge') deposit overlain by a pumice fallout
129 deposit interbedded with high-concentration PDC (named 'flow') and low-concentration PDC
130 (named 'ash hurricane') deposits. According to their distribution map of the deposits (Fig. 1b),
131 the high-concentration PDC flowed in the Fond Canonville valley, Grande River, Bijou River,
132 and Falaise River, whereas the low-concentration PDC that elutriated from the top of the
133 current spread on hills at distances up to 20 km from the vent. Isopachs of the pumice fallout
134 deposit exhibit a SSW elongation (Fig. 1b), in good agreement with the one inferred by
135 Westercamp and Traineau (1983) (Fig. 1a). Wright et al. (2016) interpreted their complete
136 stratigraphic succession as resulting from the partial and final collapse of an initially stable
137 Plinian column.

138 The eruptive parameters of the P3 event are currently poorly constrained. Volume
139 estimates for the fallout tephra of the P3 eruption yield a relatively small value of 0.19 km³
140 dense rock equivalent (DRE) (Traineau et al., 1989). The direction of tephra dispersal during
141 the initial phase (P3₁) suggests that the column was affected by low altitude tropospheric
142 winds, and thus was certainly more than 6 km (Traineau et al., 1989), but no more than 12 km
143 in height (Wright et al., 2016). Such a large uncertainty on the maximum column height
144 makes it difficult to estimate its mass eruption rate. Overall, these sparse and loosely
145 constrained data cannot be used to understand the conditions leading to column collapse.

146

147 **3. Methodology**

148 3.1. Fieldwork

149 We identified the P3 eruption deposits at 102 locations in Martinique (Fig. 2). In most cases,
150 these deposits are easily recognizable by their strong internal layering, the abundance of
151 juvenile and accidental lithic fragments in the pumice fallout units, and the presence of the P1
152 and/or P2 deposits above them (Carazzo et al., 2012; 2019). The P3 outcrops are distributed
153 all around the volcano except to the northwest where exposure is very limited due to dense
154 tropical forest and difficult conditions of access. The thickness of each layer of P3 was
155 measured at every site in order to construct isopach maps. We also excavated a standard 25 x
156 20 cm area of each fall layer and measured the major axes of the five largest lithic fragments
157 found in order to produce isopleth maps.

158

159 3.2. Radiocarbon dating

160 We dated four charcoals sampled within the P3 deposits at sites 6 (unit A-G), 97 (unit A), 178
161 (unit A), and 180 (unit A-G), and a paleosol sampled beneath unit A at site 145 (see locations
162 in Fig. 2 and unit descriptions in section 4.1) in order to refine the age of the P3 eruption.

163 Ages were determined using accelerator mass spectrometry at the LMC14 (Artemis,
164 Laboratoire de Mesure du Carbone 14, CEA Saclay, France), and calibrated using the free
165 software OxCal 4.3 (Bronk Ramsey, 2009) with the atmospheric IntCal13 calibration curve
166 commonly used for the Northern hemisphere (Reimer, 2013). The radiocarbon ages obtained
167 for our stratigraphically-constrained samples were combined with those of Westercamp and
168 Traineau (1983) and validated using the R_combine function of Oxcal and χ^2 test prior to
169 calibration (Ward and Wilson, 1978).

170

171 3.3. Grain-size analysis

172 We carried out grain-size analyses on twenty samples from ten locations representative of the
173 seven different units (A to G) (see Table 1). The samples were dried for 24 h in an oven and
174 sieved by hand down to 6ϕ . We separated the lithic fragments from the pumices by hand in
175 the size range -6ϕ to -4ϕ , and used a binocular microscope to discard crystals and lithic
176 fragments in the size range -4ϕ to -2ϕ . Volume calculations for isomass maps for each ϕ
177 interval were then used to determine the grain-size distributions of single sub-layers. The total
178 grain-size distribution was estimated from the total mass in sieve class ϕ , M_ϕ , by the volume
179 integral (Kaminski and Jaupart, 1998)

$$180 \quad M_\phi = \int_0^L h(l) C_\phi(l) A(l) dl, \quad (1)$$

181 where $h(l)$ is the deposit thickness, $C_\phi(l)$ is the concentration of class ϕ at distance l from the
182 vent, $A(l)dl$ is the area bounded by isopachs at distances l and $l+dl$, and L is the distance
183 where h or C_ϕ drop to zero. We used linear interpolations for h and C_ϕ between the localities.

184 The cumulative frequency curves were determined assuming a power-law size
185 distribution of the rock fragments (Turcotte, 1986; Kaminski and Jaupart, 1998), where the
186 number of particles with a radius larger than r is given by:

187 $N(R_p \geq r_p) = \lambda r^{-D},$ (2)

188 with D the power-law exponent that commonly varies between 2.9 and 3.9 for Plinian
189 eruptions (Girault et al., 2014). Because fragmentation is a size-invariant process, the
190 exponent D can be estimated accurately using any sufficiently large range of sizes. Thus, the
191 lack of fine particles lost out at sea does not affect our estimations of D .

192

193 3.4. Eruptive parameters

194 The volume of tephra fallout produced during an explosive eruption can be inferred using
195 several methods based on the thinning trend of the deposit with distance from the source.
196 Here, we used the exponential, power-law and Weibull fits (Pyle, 1989; Bonadonna and
197 Houghton, 2005; Bonadonna and Costa, 2012) computed using the AshCalc software (Daggit
198 et al., 2014). Bearing in mind that only proximal and furthermore incompletely preserved
199 deposits are available in Martinique, the calculated volumes of fall deposits should be taken
200 as minimum estimates. The volume of dilute PDC deposits was calculated using the observed
201 relationship between the volume and area covered by dilute PDC (Dade and Huppert, 1998;
202 Calder et al., 1999), as in Carazzo et al. (2012, 2019) for the P1 and P2 eruptions. The volume
203 of dense PDC was calculated using the product of the average thickness and area covered by
204 the dense PDC. This method is bound to provide minimum estimates since only proximal and
205 furthermore incompletely preserved deposits are available in Martinique.

206 The maximum column heights associated with the fall deposits were estimated from
207 the distribution of lithic fragments on our isopleth maps using the method of Carey and
208 Sigurdsson (1986). Error bars were calculated using the three values of maximum height
209 inferred from the 8, 16, and 32-mm isopleths (see Fig. 17 in Carey and Sigurdsson, 1986).
210 The alternative method of Bonadonna and Costa (2013) based on variations of grain size with
211 the distance from the source was also used to confirm the robustness of the estimates.

212 The mass discharge rate feeding the eruption during phases of stable column was
213 inferred from a given column height by using the empirical relationship from [Mastin et al.](#)
214 [\(2009\)](#) and the predictions of the model of [Girault et al. \(2016\)](#), which explicitly includes the
215 effect of total grain-size distribution on the plume dynamics. Calculations were made for
216 tropical atmospheric conditions and for a crystal-bearing rhyolitic magma (andesitic bulk
217 composition) with an initial temperature of 1,150 K ([Martel and Poussineau, 2007](#)) and a
218 dense rock density of 2,400 kg m⁻³ ([Traineau et al., 1989](#)). The mass discharge rates for the
219 collapsing phases were calculated using the model of [Bursik and Woods \(1996\)](#), which treats
220 pyroclastic density currents as supercritical dilute suspensions entraining air as they
221 propagate. [Doyle et al. \(2010\)](#) showed that this assumption is valid for tall, fine-grained
222 column collapses, for which the flow slowly transfers its mass to the dense basal flow. The
223 model of [Bursik and Woods \(1996\)](#) is thus relevant for collapsing phases characterized by the
224 formation of dilute PDC deposits, and was used here.

225

226 **4. The P3 eruptive succession**

227 4.1. Stratigraphy

228 [Fig. 3a](#) shows a composite stratigraphic section of the P3 deposits, divided into seven major
229 phases based on diagnostic stratigraphic and lithofacies associations, which we describe in
230 more detail below.

231

232 4.1.1. Unit A

233 The stratigraphic succession begins with a layer of clast-supported, coarse white pumice
234 lapilli ([Fig. 4a-d](#)). Unit A contains juvenile and accidental lithic fragments in a total amount
235 that typically increases from ≈ 20 wt.% at base to ≈ 25 wt.% at top ([Fig. 3c](#)). At most sites,

236 median pumice size remains relatively constant, whereas both the maximum lithic size and
237 the sorting increase from base to top (Fig. 3b-c).

238 In the downwind direction (to the southwest), unit A has a maximum thickness of 180
239 cm at 4.5 km from the crater (Fig. 4c). In the crosswind direction, unit A is 130 cm thick at
240 4.6 km, and steadily thins to 50 cm within 6.1 km of the crater (Fig. 4d). The relatively
241 widespread nature of this deposit, its uniformly decreasing thickness with distance from the
242 source, its clast-supported framework, and its pumice and juvenile lithic fragments
243 characteristics identify unit A as a fall deposit.

244

245 4.1.2. Unit B

246 Unit B consists of light brown fine ash material in which pumice, lithic fragments and crystals
247 are dispersed in a matrix of dense, angular, glass fragments (Fig. 4b, c, e). Most clasts range
248 in size from fine to coarse ash but a few large pumices are present (Fig. 3b). The total amount
249 of lithic fragments in unit B commonly reaches ≈ 28 wt.%. Where unit B is present, the
250 contact with the underlying layer is sharp, which suggests that there was no time break
251 between deposition of the two units.

252 Unit B varies irregularly in thickness with distance from the source, and ranges from
253 18 to 35 cm at 3.8 km from the eruptive center (Fig. 4), but is absent beyond 7 km from the
254 vent. Where thick, unit B contains a few lenses of coarse lapilli, angular to sub-rounded
255 pumices. At some sites in valleys, pumice lapilli and blocks are rounded and account for a
256 large proportion in the matrix of white fine ash. The relatively limited dispersal of this unit, its
257 irregular thickness, matrix-supported framework and grain types (juvenile and accidental
258 lithic fragments), as well as the nature of the lower contact, indicate that unit B is a low-
259 concentration pyroclastic density current deposit (Branney and Kokelaar, 2002).

260

261 4.1.3. Unit C

262 Unit C is a relatively thin blanket of clast-supported, white pumice lapilli (Fig. 4b, c, e)
263 containing juvenile and accidental lithic fragments in a total amount that reaches ≈ 25 wt.% in
264 most studied outcrops (Fig. 3c). At most field sites, unit C is easily distinguishable from unit
265 A, the former being distinctly richer in lithic fragments and coarse lapilli pumice. The
266 maximum lithic size in unit C is, however, always lower than the maximum lithic size at the
267 top of unit A (Fig. 3c).

268 Unit C was dispersed in approximately the same direction as unit A, and is the thickest
269 ≈ 3.7 km southwest of the volcano in the downwind direction, where it reaches 22 cm. This
270 layer steadily thins with distance from the vent, and is still 10 cm thick at 6.2 km in the
271 crosswind direction. At most sites, the upper part of unit C was eroded by the pyroclastic
272 density current that overrode it and deposited unit D (see below). The relatively widespread
273 nature of unit C, its uniformly decreasing thickness with distance from the source, its clast-
274 supported framework, and its pumice and juvenile lithic fragments characteristics identify it
275 as a fall deposit.

276

277 4.1.4. Unit D

278 Unit D is a relatively thin layer of fine ash and small rounded pumice lapilli dispersed in a
279 matrix of dense, angular, glass fragments (Fig. 4a, c, f). At distal sites, this layer is thin,
280 stratified and unconsolidated. The total amount of lithic fragments in unit D commonly
281 reaches ≈ 20 wt.% (Fig. 3c). Where unit D is present, the contact with the underlying layer is
282 sharp and exhibits some erosion of unit C.

283 Unit D varies irregularly in thickness with distance from the vent, and ranges from 5 to
284 17 cm at 3.8 km from the volcano (Fig. 4). This layer is, however, absent beyond 7 km from
285 the vent. These irregular thicknesses, the relatively limited dispersal of this deposit, the clast

286 and matrix types, and the nature of the lower contact indicate that unit D is a low-
287 concentration pyroclastic density current deposit.

288

289 4.1.5. Unit E

290 Unit E is a layer of clast-supported, white pumice lapilli (Fig. 4b, c, e) containing juvenile and
291 accidental lithic fragments in a total amount that typically reaches ≈ 18 wt.% (Fig. 3c). At
292 most sites, both the median grain-size and the maximum lithic size are lower than those
293 measured in unit C (Fig. 3c).

294 Material was dispersed in approximately the same direction as units A and C to the
295 southwest of the volcano. The maximum thickness is 13 cm at 3.7 km, and steadily thins to 9
296 cm within 4.8 km from the vent in the downwind direction. The full thickness of unit E is
297 however poorly preserved due to significant erosion by the overlying pyroclastic density
298 current that overrode it and deposited unit F (see below). The relatively widespread nature of
299 this deposit, its uniformly decreasing thickness with distance from the source, its clast-
300 supported framework, and its pumice and juvenile lithic fragments characteristics identify unit
301 E as a fall deposit.

302

303 4.1.6. Unit F

304 Unit F is a matrix-supported layer of fine to coarse ash containing rounded pumice lapilli and
305 lithic fragments (Fig. 4b, f), the latter of which typically reaches ≈ 15 wt.% (Fig. 3c). Where
306 unit F is present, the contact with the underlying layer is sharp and exhibits some erosion of
307 unit E.

308 Unit F varies irregularly in thickness with distance from the vent depending on the
309 topography, ranges from 6 to 75 cm at 3.8 km from the volcano, and is absent 7 km from the
310 vent (Fig. 4). The relatively limited dispersal of this unit, its irregular thickness, matrix-

311 supported framework and grain types (juvenile and accidental lithic fragments), as well as the
312 nature of the lower contact, indicate that unit F is a low-concentration pyroclastic density
313 current deposit.

314

315 4.1.7. Unit G

316 Unit G is the uppermost layer in the P3 sequence and consists of a blanket of fine to coarse
317 white pumice lapilli (Fig. 4a, b, f). This unit contains juvenile and accidental lithic fragments
318 in a total amount that reaches ≈ 20 wt.% at most studied outcrops (Fig. 3c) with a maximum
319 lithic size similar to that of unit C (Fig. 3c).

320 In the downwind direction (to the southwest), unit G has a maximum thickness of 80
321 cm at 1.4 km from the crater, and steadily thins to 20 cm within 4.8 km of the vent. In the
322 crosswind direction, unit G is 30 cm thick at 5 km, and 15 cm at 6.4 km from the vent. The
323 relatively widespread nature of this deposit, its uniformly decreasing thickness with distance
324 from the source, its clast-supported framework, and its pumice and juvenile lithic fragment
325 characteristics identify unit G as a fall deposit.

326

327 4.2. Age of the P3 eruption

328 Radiocarbon dates and calibrated ages are reported in [Supplementary material Table S1](#). Two
329 samples provide the exact same radiocarbon age of $1,870 \pm 30$ yr BP (sites 145 and 178 on
330 the western flank). A third sample taken near location 178 gives a radiocarbon age of $1,915 \pm$
331 30 yr BP (site 97). The oldest age is found near the town of Le Carbet where the charcoal
332 sampled within the pumice fallout deposit provides an age of $2,030 \pm 30$ yr BP. On the
333 eastern flank of the volcano, the charcoal sampled at location 6 has an age of $1,795 \pm 30$ yr
334 BP. These five new radiocarbon ages are in good agreement with a compilation of twelve ^{14}C
335 ages determined in previous studies ([Walker, 1973](#); [Roobol and Smith, 1976](#); [Traineau, 1982](#);

336 [Westercamp and Traineau, 1983](#)) ranging from 1,800 to 2,150 yr BP ([Supplementary material](#)
337 [Table S1](#)). Combining these seventeen values using the R_combine function of OxCal ([Bronk](#)
338 [Ramsey, 2009](#)) provides an age of $1,926 \pm 11$ yr BP for the P3 eruption. We note that the χ^2
339 test fails here most likely due to the large number of individual measurements. However, a
340 reasonable correction on the error bar of each dating by a factor of 2 is enough to successfully
341 pass the χ^2 test, reinforcing the confidence that all these otherwise stratigraphically-
342 constrained samples belong to the same eruption. Our new refined radiocarbon age for the P3
343 eruption stands in the lower error bars of the $2,010 \pm 140$ yr BP proposed by [Traineau et al.](#)
344 [\(1989\)](#). Correcting our new estimate with the IntCal13 calibration curve ([Reimer, 2013](#))
345 available in OxCal gives a mean calendar age of 79 ± 21 cal CE (i.e., 31-37 cal CE at 1.7%
346 probability and 51-125 cal CE at 93.7% probability).

347

348 4.3. Grain-size distribution of selected samples

349 Twenty samples from the P3 units were analyzed ([Table 1](#)) in order to discuss the evolution of
350 the grain-size distribution during the entire P3 eruption at a single outcrop ([Fig. 3](#)), and to
351 determine the total grain-size distribution for the lapilli fraction of unit A deposits (see section
352 4.6). The objective is twofold: to confirm the fall or flow nature of the P3 units, and to
353 estimate the amount of gas trapped in pumice fragments at the beginning of the eruption (see
354 section 6.3). A detailed analysis of the pumice textures is beyond the scope of this paper and
355 can be found in [Martel and Poussineau \(2007\)](#).

356 The fourteen samples from unit A display the typical fallout characteristics with
357 median diameter ranging from 0.7ϕ to -3.3ϕ , and sorting ranging from 1.4 to 2.7. These
358 values are relatively close to those obtained by [Bardintzeff et al. \(1989\)](#) who measured
359 median diameter and sorting ranging from -0.3ϕ to -3ϕ , and 1.7 to 2.8, respectively. Most of
360 the samples collected on land belong to proximal deposits since the medial and distal deposits

361 were lost at sea. The grain-size distribution of individual samples is systematically bimodal
362 (Fig. 3b) and the amount of ash particles (< 2 mm) always increases from base (≈ 9 wt.%) to
363 top (≈ 36 wt.%). The top of unit A is thus always more poorly sorted than the base, but the
364 median grain-size remains approximately constant throughout the unit (Fig. 3c).

365 The six samples from units B to G collected at outcrop 163 show that units C, E, and
366 G have typical fallout characteristics similar to unit A. The median diameter for these units
367 ranges between -0.1ϕ and -3ϕ , and they have moderate sorting values from 2.5 to 2.7 (Fig.
368 3c). The amount of ash-sized particles is similar to or larger than in unit A, ranging from ≈ 28
369 wt.% to ≈ 55 wt.%. Units B, D, and F have flow features with median diameter ranging from
370 1.7ϕ and 0.4ϕ , and moderate sorting ranging from 2.2 to 2.6 (Fig. 3c). The amount of ash-
371 sized particles is larger than in the fall deposits, ranging from ≈ 62 wt.% to ≈ 85 wt.%.

372

373 4.4. Isopach maps

374 Fig. 5 shows a stratigraphic correlation of P3 outcrops along three different dispersal axes
375 (i.e., north to south, northwest to southeast, and northeast to southwest, see Fig. 2 for
376 localization). The complete sequence can be found up to 5 km from the vent (sites 13, 163,
377 101, 78 in Fig. 5). Thickness measurements at each location are reported on isopach maps for
378 the cumulative pumice fallout phases A+C+E+G (Fig. 6a), the pumice fallout phase A (Fig.
379 6b), the cumulative pyroclastic density current phases B+D+F (Fig. 6c), and the cumulative
380 pumice fallout phases C+E+G (Fig. 6d). These maps are later used to calculate the total grain-
381 size distribution of unit A (see section 4.6), and the volume of deposits (see section 5.1).

382 The cumulative isopach map of the pumice fallout phases A+C+E+G shows
383 ellipsoidal contour patterns indicating fallout dispersion towards the southwest (Fig. 6a). We
384 find a similar direction of dispersion for the pumice fallout phase A, and the cumulative
385 pumice fallout phases C+E+G (Fig. 6b, d). This direction of dispersion is in relatively good

386 agreement with the results of [Traineau et al. \(1989\)](#) and [Wright et al. \(2016\)](#) ([Fig. 1](#)), and is
387 characteristic of the wind profiles in the Lesser Antilles during the wet season ([Michaud-
388 Dubuy et al., 2019](#)). In the upwind direction to the northeast and in the crosswind direction to
389 the southeast, the entire P3 succession is reduced to a single layer of clast-supported, white,
390 angular, fine pumice lapilli containing abundant juvenile and accidental lithic fragments,
391 making it difficult to distinguish the pumice fallout units A, C, E, and G. At locations near the
392 vent, unit A is however clearly visible under the cumulative pumice fallout of units C+E+G
393 ([Fig. 4d](#)), which allows to build an individual isopach map for the pumice fallout phase A
394 ([Fig. 6b](#)). We note that the northwestern arms of our isopach maps are not well constrained
395 due to a lack of outcrops (in our study as in previous ones) in a region extremely difficult to
396 access due to the presence of dense tropical forest on steep hills.

397 Units B, D, and F are widespread on the southwest flank of Mt Pelée volcano and the
398 low-concentration PDC deposits almost reached the sea there ([Fig. 6c](#)). The high-
399 concentration PDC deposits are channelized in paleo-valleys and now form large plateaux in
400 the Grande Savane, Habitation Depaz and Ajoupa Bouillon areas (see locations in [Fig. 1a](#)), an
401 observation consistent with the results of [Westercamp and Traineau \(1983\)](#) and [Traineau et al.
402 \(1989\)](#). However, contrary to [Wright et al. \(2016\)](#), we found no evidence of high-
403 concentration PDC deposits in the Bijou River, Grande River, and Fond Canonville areas (see
404 locations in [Fig. 1b](#)).

405

406 4.5. Isopleth maps of air fall deposits

407 [Fig. 7](#) shows the isopleth maps built from the measurements of the major axes of the five
408 largest fragments found at the base and top of unit A, and at the base of unit C. The isopleth
409 map of the base of unit A is well constrained thanks to the good preservation of the deposit
410 ([Fig. 7a](#)). However, isopleth contours for the top of unit A and the base of unit C are

411 constrained by only a few points from localities where the two units are separated by the
412 presence of unit B and/or by clear variation in grain-size characteristics (Fig. 7b, c). Isopleth
413 maps for units E and G could not be constructed due to the poor preservation of these
414 deposits. However, in both cases, we were able to draw a single isopleth curve (i.e., the 15
415 mm-isopleth for unit E and 25 mm-isopleth for unit G) that is not reported in Fig. 7. The main
416 direction of dispersion to the southwest (Fig. 7) is consistent with the one inferred from our
417 isopach maps (Fig. 6). Isopleth contours of the top of unit A are clearly more extended in the
418 crosswind direction than those of the base of units A and C.

419

420 4.6. Total grain-size distribution (*Unit A*)

421 We calculated the total grain-size distribution (TGSD) of unit A using the method of
422 Kaminski and Jaupart (1998) (see section 3.3 for calculation details). The interbedded nature
423 and important erosion of the subsequent units made the reconstruction of the TGSD for units
424 B to G impossible. The TGSD of unit A is bimodal with a primary fine mode at 2ϕ and a
425 secondary coarse mode at -2ϕ (Supplementary material Figure S1). The median diameter and
426 sorting for the TGSD of unit A are -0.2ϕ and 2.3, respectively. The amounts of ash ($< 2 \mu\text{m}$)
427 and fine ash particles ($< 63 \mu\text{m}$) reach 59 wt.% and less than 1 wt.%, respectively. The latter
428 value should be taken with caution because most of the fine ash particles settled at sea and
429 cannot be taken into account in our grain-size analyses. From these results, we infer that the
430 power law coefficient D that fully characterizes the TGSD at the beginning of the P3 eruption
431 (i.e., unit A) is $D = 3.3 \pm 0.1$ (Supplementary material Figure S1). The time evolution of the
432 value of D after the phase A remains unknown, but this lack of information will not affect our
433 calculations of the stable plume / collapsing fountain transition that occurred at the end of
434 phase A (see section 6.3).

435

436 5. Eruptive dynamics

437 5.1. Erupted volumes

438 The estimates of volume of tephra fall from phase A using the three integration
439 techniques introduced in section 3 give 0.282 km^3 with the exponential method, 0.316 km^3
440 with the power law, and 0.306 km^3 with the Weibull function (Fig. 8a). Integration of the
441 exponential, power-law, and Weibull fits for the cumulative pumice fallout C+E+G yields
442 volumes of 0.334 , 0.317 , and 0.355 km^3 , respectively (Fig. 8b). Because the latter estimates
443 are based on three proximal isopach contours only they remain relatively poorly constrained
444 (Fig. 6d). We thus calculated the volume of the cumulative pumice fallout A+C+E+G in order
445 to achieve a better estimate of the total volume of tephra fall. Integrating the exponential,
446 power-law, and Weibull fits yields 0.735 , 0.692 , and 0.712 km^3 , respectively (Fig. 8c). We
447 thus retain a minimum volume of $0.30 \pm 0.01 \text{ km}^3$ for unit A, and $0.41 \pm 0.02 \text{ km}^3$ for the
448 cumulative pumice fall units C+E+G, and $0.71 \pm 0.02 \text{ km}^3$ for the total volume of tephra fall.
449 Based on fallout deposit and magma densities of $1,000$ and $2,400 \text{ kg m}^{-3}$, respectively
450 (Traineau et al., 1989), we infer the cumulative DRE volume of fall units to be 0.3 km^3 .

451 The volume of PDC deposits is calculated using the area covered by the dense and
452 dilute PDC deposits. The deposits corresponding to the dilute PDC of phases B, D, and F
453 cover 42 , 20 , and 36 km^2 , respectively (Fig. 6c), which yield bulk volumes of 0.025 km^3 for
454 unit B, 0.010 km^3 for unit D, and 0.016 km^3 for unit G (Calder et al., 1999). The cumulative
455 dense PDC deposits of the phases B, D, and F cover 7 km^2 in Grande Savane, 9.5 km^2 in
456 Habitation Depaz, and 18 km^2 in Ajoupa Bouillon (see locations in Fig. 1a), with an average
457 thickness of 15 , 30 , and 2 m , respectively (Traineau, 1982). This yields a minimum volume of
458 $0.67 \pm 0.02 \text{ km}^3$ DRE for the dense PDC deposits, and gives a total volume of (dense + dilute)
459 PDC deposits of $0.72 \pm 0.02 \text{ km}^3$ DRE.

460 The final estimate of the total volume of the P3 eruption (units A+B+C+D+E+F+G) is
461 thus $1.02 \pm 0.06 \text{ km}^3$ DRE, and the total mass of tephra emitted is estimated to be $2.4 \pm 0.1 \times$
462 10^{12} kg. The P3 eruption thus stands as a VEI 5 event (Newhall and Self, 1984) with a
463 magnitude of 5.4 (Pyle, 2000).

464

465 5.2. Column heights and exit velocities (*airfall units*)

466 The estimates of maximum height using our isopleth maps (Fig. 7) with the model of Carey
467 and Sigurdsson (1986) give 28.3 ± 3.3 km for the base of unit A, 29.7 ± 1.9 km for the top of
468 unit A, and 16.9 ± 0.4 km for the base of unit C. Following the approach of Bonadonna and
469 Costa (2013), a Weibull fit gives a maximum height of 28.1 ± 1 km for the base of unit A,
470 30.0 ± 0.5 km for the top of unit A, and 17.2 ± 0.5 km for the base of unit C (Supplementary
471 material Figure S2). The lack of information about the distribution of lithic sizes in unit E and
472 G prevents us from providing a robust estimate for the maximum height. Based on a single
473 isopleth curve in each case, we calculate maximum column heights of 12 ± 4 km for unit E
474 and 13 ± 4 km for unit G using the model of Carey and Sigurdsson (1986).

475 We also use data on the decrease of lithic sizes with distance from the vent in
476 Supplementary material Figure S2 to estimate the minimum exit velocity of the volcanic
477 plume at the vent. Extrapolating the Weibull fits down to $A^{1/2} = 0.01$ km to find the maximum
478 lithic size at the vent, we infer minimum exit velocities of $210 \pm 10 \text{ m s}^{-1}$ for the base of unit
479 A, $220 \pm 10 \text{ m s}^{-1}$ for the top of unit A, and $220 \pm 10 \text{ m s}^{-1}$ for the base of unit C.

480

481 5.3. Mass eruption rates and duration

482 We use our new estimates of maximum column height to calculate the mass eruption rate
483 (MER) feeding the plume during the P3 eruption. The model of Girault et al. (2016) used with
484 their complex wind profile, which is close to the average wind profiles in the Lesser Antilles,

485 yields maximum MERs of 10^8 kg s^{-1} for the beginning of phase A, $1.5 \times 10^8 \text{ kg s}^{-1}$ for the end
486 of phase A, $3 \times 10^7 \text{ kg s}^{-1}$ for phase C, $1.4 \times 10^7 \text{ kg s}^{-1}$ for phase E, and $1.7 \times 10^7 \text{ kg s}^{-1}$ for
487 phase G. The empirical formula of [Mastin et al. \(2009\)](#) built on observations of 28 Plinian
488 eruptions yields maximum MERs of $1.4 \times 10^8 \text{ kg s}^{-1}$ for the beginning of phase A, 1.9×10^8
489 kg s^{-1} for the end of phase A, $1.8 \times 10^7 \text{ kg s}^{-1}$ for phase C, $4.2 \times 10^6 \text{ kg s}^{-1}$ for phase E, and 5.9
490 $\times 10^6 \text{ kg s}^{-1}$ for phase G. Based on these two series of estimates, we retain peak MERs of 1.2
491 $\pm 0.2 \times 10^8 \text{ kg s}^{-1}$ for the beginning of phase A, $1.7 \pm 0.2 \times 10^8 \text{ kg s}^{-1}$ for the end of phase A,
492 $2.4 \pm 0.6 \times 10^7 \text{ kg s}^{-1}$ for phase C, $9.1 \pm 4.9 \times 10^6 \text{ kg s}^{-1}$ for phase E, and $1.2 \pm 0.5 \times 10^7 \text{ kg s}^{-1}$
493 for phase G. Combined with masses of fallout deposits (section 5.1), these MER provide
494 minimum durations of ~ 45 min for phase A, $\sim 1\text{h}45$ for phase C, $\sim 3\text{h}$ for phase E, and $\sim 2\text{h}45$
495 for phase G.

496 The MER during the collapsing phases B, D, and F is more difficult to assess. The
497 model of [Bursik and Woods \(1996\)](#) yields a mass eruption rate of $\sim 2.3 \pm 0.8 \times 10^8 \text{ kg s}^{-1}$ for
498 the three phases B, D, and F. We note that this value does not take into account topographic
499 slopes but [Doyle et al. \(2010\)](#) showed that the runout distance of such a flow mostly depends
500 on the initial collapse height, which is imposed by the MER rather than by ground slopes.
501 Combined with masses of PDC deposits (section 5.1), this MER provides minimum durations
502 of ~ 50 min for phase B, ~ 25 min for phase D, and ~ 50 min for phase F.

503

504 **6. Discussion**

505 Our revisit of the P3 deposits shows that the eruption started with the formation of a stable
506 column that spread pyroclasts to the southwest of the volcano (phase A). This initial phase
507 was immediately followed by a partial collapse of the column producing alternating tephra
508 fallout (phases C, E, G) and pyroclastic density currents (phases B, D, E). We now compare
509 our results to previous work on the P3 eruption, and on the younger P1 and P2 eruptions of

510 Mt Pelée volcano. We then focus our discussion on the causes of partial column collapse
511 during the P3 eruption by using a physical model of volcanic plumes.

512

513 6.1 Comparison with previous studies of P3 eruption

514 Our new stratigraphic data regarding the P3 succession largely differs from those of previous
515 studies (Westercamp and Traineau, 1983; Traineau et al., 1989; Wright et al., 2016). The P3₁
516 phase identified by Westercamp and Traineau (1983) corresponds to our eruptive unit A, and
517 their P3₃ phase most likely corresponds to our entire sequence from unit B to unit G. The
518 distal products identified by Wright et al. (2016) were attributed to the P3 eruption based on a
519 single radiocarbon age determined from a charcoal sampled at the top of their stratigraphic
520 section (see their figure 6). We revisited their outcrop locations 2 and 3 along a main road,
521 and found that their stratigraphic section lies beneath the 13.5 ka cal BP Bellefontaine Plinian
522 eruption (Michaud-Dubuy et al., 2019). To further reinforce our observation, we sampled a
523 paleosol at the base of their stratigraphic section and obtained a radiocarbon age of $17,750 \pm$
524 100 yr BP, which indicates that the distal deposits studied by Wright et al. (2016) are much
525 older than the P3 eruption.

526 The total volume of the P3 eruption (units A+B+C+D+E+F+G) is calculated to be
527 1.02 ± 0.06 km³ DRE, which is one order of magnitude larger than the 0.19 km³ DRE
528 previously estimated (Traineau et al., 1989). We attribute this important discrepancy to the
529 large volume of PDC deposits that were not taken into account in Traineau et al. (1989), and
530 to the improvement of the reconstruction techniques used here to calculate the total volume of
531 a pumice fall deposit (Bonadonna and Costa, 2013). The maximum column height is
532 estimated to be ~ 30 km at the end of phase A, which is much higher than the previous
533 estimates (> 6 km by Traineau et al., 1989 and < 12 km by Wright et al., 2016). Again we
534 attribute this important difference to the better quality of our isopleth maps and to the

535 improvement of the reconstruction techniques used to estimate the maximum height from the
536 distribution of lithic fragments in the field.

537

538 6.2 Comparison with younger Plinian eruptions in Martinique (P1 and P2)

539 The three latest Plinian eruptions at Mt Pelée volcano (P1, P2 and P3) have similar
540 depositional successions that can be interpreted to result from the partial collapse of an
541 initially stable Plinian column (Traineau et al., 1989; Carazzo et al., 2012, 2019; Wright et al.,
542 2016). The total DRE volume estimated for P3 ($\sim 1 \text{ km}^3$) is larger than those of the P1 (~ 0.2
543 km^3) and P2 ($\sim 0.8 \text{ km}^3$) eruptions. The maximum column height during the phase of sustained
544 Plinian column is also larger for P3 (28-30 km) than for the P1 (19-22 km) and P2 (22-26 km)
545 eruptions. Consequently, the mass eruption rate is estimated to be larger for P3 ($1.4 \times 10^8 \text{ kg}$
546 s^{-1}) than for P1 ($3.6 \times 10^7 \text{ kg s}^{-1}$) and P2 ($1.1 \times 10^8 \text{ kg s}^{-1}$). The minimum eruption durations
547 are 11h for P3, 7h for P2, and 5h for P1. The total grain-size distribution is however relatively
548 similar for the three events, with power law exponent D of 3.3, 3.4-3.5, and 3.2-3.3 for P3,
549 P2, and P1, respectively. Our estimates of exsolved gas contents for P3 (2-2.9 wt.% - see
550 section 6.3 for calculations details) are larger than those of P1 (1.6-2.1 wt.%) and P2 (1.7-2.1
551 wt.%). The minimum exit velocities inferred from the distribution of lithic fragments for P3
552 ($210\text{-}220 \text{ m s}^{-1}$) are also larger than those of P1 ($150\text{-}165 \text{ m s}^{-1}$) and P2 ($180\text{-}200 \text{ m s}^{-1}$), a
553 result consistent with our estimates of free gas contents. Table 2 summarizes the eruptive
554 parameters retrieved for the last three Plinian eruptions at Mt Pelée volcano. P3 stands as a
555 large Plinian eruption more powerful and voluminous (VEI 5, M 5.4) than the P1 (VEI 4, M
556 4.6) and P2 (VEI 4, M 5.2) eruptions. However, because these three eruptions have much in
557 common, in particular similar pre-eruptive conditions (Martel et al., 1998; Martel and
558 Poussineau, 2007), they can be used to test the performance of theoretical models in
559 predicting the conditions for column collapse.

560

561 6.3 The stable plume / pyroclastic fountain transition during a sustained Plinian eruption

562 The P3 eruption underwent partial column collapse after an initial phase of stable Plinian
563 plume. [Wilson et al. \(1980\)](#) showed that the mass eruption rate and the amount of free gas in
564 the volcanic mixture at the vent both strongly control the transition between the stable Plinian
565 plume and the collapsing fountain regimes. Our estimates of the MER show a progressive
566 increase over time from $\sim 1.2 \times 10^8 \text{ kg s}^{-1}$ (base of unit A) to $\sim 2.5 \times 10^8 \text{ kg s}^{-1}$ (unit B+C) ([Fig.](#)
567 [9](#)), an evolution consistent with the increase in maximum lithic sizes throughout unit A ([Fig.](#)
568 [3c](#)). Other powerful eruptions underwent a similar evolution leading to column collapse, such
569 as the 79 CE Vesuvius ([Shea et al., 2011](#)), and the 186 CE Taupo ([Walker, 1980](#)) eruptions.

570 The time evolution of total gas content feeding the column at the vent is more difficult
571 to assess. Petro-geochemical measurements provide quantitative estimates of pre-eruptive
572 magma storage conditions, including temperature (875-900°C), pressure ($\sim 200 \pm 50 \text{ MPa}$),
573 and water content ($\sim 5.8 \pm 0.5 \text{ wt.}\%$) ([Martel et al., 1998](#); [Martel and Poussineau, 2007](#)). Here,
574 we use the magmatic water content measured in glass inclusions available in the literature as a
575 total volatile content in the melt (n_0). Correcting this value from the presence of crystals and
576 lithic fragments ([Table 3](#)), we find that the initial gas content (x_0) in the magma decreased
577 from 4.1 to 2.9 wt.% during the eruption ([Table 3](#)). As in [Kaminski and Jaupart \(1998\)](#) and
578 [Carazzo et al. \(2012; 2019\)](#), we calculate the mass fraction of exsolved gas in the mixture at
579 fragmentation level (x_f) for a threshold vesicularity of 70 %, which assumes closed-system
580 conditions consistent with degassing models built on U-series measurements in the P3
581 eruption products ([Villemant et al., 1996](#)). We find that x_f decreases from 2.9 to 2.1 wt.%
582 during the eruption ([Fig. 9](#)). To strengthen these estimates, we calculate the theoretical exit
583 velocities at the vent assuming that the volcanic mixture decompresses freely into the
584 atmosphere ([Woods and Bower, 1995](#)). For this, we infer the effective amount of free gas in

585 the volcanic mixture at the base of the column that is modulated by gas entrapment by
586 pumices as a function of their size distribution (Kaminski and Jaupart, 1998). We use the
587 power law exponent $D = 3.3$ (for unit A) estimated in section 4.6, and a typical open porosity
588 of 60-70 % (Michaud-Dubuy et al., 2018) to calculate an amount of free gas of 2.4-2.7 wt.%
589 for phase A. These estimates provide exit velocities after decompression in the range 215-230
590 m s^{-1} (Table 3) that are consistent with those estimated from the distribution of the lithic
591 fragments collected at the base and top of unit A (see section 5.2). This relatively good
592 agreement at least for phase A reinforces the confidence that our estimates of x_f provide
593 reasonable values of free gas content during the P3 eruption.

594 [Fig. 10](#) compares the estimated values of MER and exsolved gas content in the column
595 at collapse with the theoretical 1D predictions for tropical conditions (Michaud-Dubuy et al.,
596 2018). The conditions at the beginning of the P3 eruption formed a stable plume as predicted
597 by the model. During phase A, the combined effect of a slightly increasing MER and a
598 decreasing gas content led the eruption to conditions close to the plume/fountain transition.
599 The eruption reached a peak in MER during the concomitant phases B and C, which most
600 likely induced the partial column collapse. During the partial column collapse stage of the
601 eruption (units B to G), the MER did not change significantly but the gas content slowly
602 decreased until the end of the eruption. The fairly good agreement between our field data and
603 the theoretical model can be taken as an indirect confirmation of the proposed evolution of the
604 total gas content feeding the column, even if a direct confirmation is not possible.
605 Furthermore, our results suggest that the stable column underwent partial collapse soon after
606 the beginning of the eruption (i.e., within less than a hour), an evolution that was also inferred
607 from the analysis of the P1 and P2 Plinian eruption deposits (Carazzo et al., 2012; 2019). In
608 these three eruptions, the transition occurred at conditions well predicted by our theoretical
609 model of volcanic plumes (Fig. 10).

610 Comparing these results with other historical eruptions requires gathering well-
611 constrained values of exsolved gas contents, mass eruption rates and power-law exponent D
612 of the TGSD (Michaud-Dubuy et al., 2018). There are however only a few examples of such
613 well-documented events, and we choose here to compare our results to the 186 CE Taupo
614 (VEI 7) and 79 CE Vesuvius (VEI 5) eruptions. Fig. 10 shows that the model successfully
615 predicts the transition for the Taupo eruption but fails to explain the transition for the
616 Vesuvius eruption. Furthermore, one can note that the collapsing phase of the 79 CE Vesuvius
617 eruption is characterized by the same amount of gas and MER at the source as the stable
618 phase (end of phase A) of the P3 eruption. The only difference between the two cases is the
619 total grain-size distribution. Indeed, as discussed in Michaud-Dubuy et al. (2018), the 79 CE
620 Vesuvius eruption is characterized by a specific total grain-size distribution ($D = 3.0$) where
621 pyroclasts are in average not large enough to be lost by sedimentation but not small enough to
622 be in thermal equilibrium with the gas. The hypothesis of thermal equilibrium between the
623 gas and pyroclasts that is used in our 1D model of volcanic plumes, valid for the fine grain-
624 size distribution of the beginning of the P3 eruption ($D = 3.3$), may therefore not hold
625 anymore for this eruption and may explain the peculiar collapse conditions.

626

627 **7. Conclusion**

628 We have presented a new comprehensive field study of the P3 eruption, and a detailed
629 reconstruction of the mechanisms controlling the eruptive dynamics. New radiocarbon
630 measurements, averaged with those previously available, provide a refined age of 79 ± 21 cal
631 CE for this event. The eruption started with the formation of a 28 to 30 km-high stable
632 column (phase A), which partially collapsed due to the combined effect of increasing MER
633 and decreasing gas content (phase B). In the partial collapse regime, the eruptive mixture
634 formed a 12 to 17 km-high marginally stable column (phases C, E, and G) punctuated with

635 the production of PDC fed by the partially collapsing column (phases B, D, and F). Our
636 calculations show that P3 was a large Plinian eruption (VEI 5, $M = 5.4$) evolving close to the
637 plume/fountain transition. The total volume of tephra reached 1 km^3 DRE, and the mass
638 eruption rate was of the order of 10^8 kg s^{-1} during a period of at least 11h.

639 The P3 eruption is more powerful than the last two Plinian eruptions in Martinique but
640 the three eruptions underwent similar evolution towards a marginally stable volcanic column.
641 This recurrent phenomenon provides a great opportunity to look carefully at the parameters
642 affecting column stability, to constrain theoretical models of volcanic plumes, and thus
643 improve volcanic disaster forecasting. Petrological and geochemical measurements on the P3
644 eruptive products may help to further investigate the complexity of magma degassing,
645 crystallization and fragmentation in the volcanic conduit ([Gurioli et al., 2005](#); [Martel and](#)
646 [Poussineau, 2007](#); [Shea et al., 2012; 2014](#)), and thus refine the estimated values of exsolved
647 gas content at the vent. The global consistency between our theoretical predictions and field
648 observations of column collapse for a number of Plinian eruptions confirms that current
649 theoretical 1D models of volcanic plumes predict correctly the conditions of collapse when
650 the total grain-size distribution is dominated by fine fragments (i.e., obeys a power law
651 distribution with an exponent strictly larger than 3). Our results can thus be used to assess the
652 dominant hazard during a Plinian scale event, which may pass from pumice fall to PDC as
653 eruptive conditions vary.

654 Mt Pelée volcano is currently quiescent but the reconstruction of its eruptive history
655 over the past 5,000 years shows that a Plinian eruption is a major potential scenario in the
656 future. Such an event with the magnitude of the P3 eruption would disrupt the lives of more
657 than 376,000 people living in Martinique and possibly impact other volcanic islands
658 depending on the wind speed and direction.

659

660 **Acknowledgements**

661 We warmly thank the staff of the Mt Pelée volcanological observatory (OVSM) for field and
662 administrative assistance, J.-P. Dumoulin, L. Beck, E. Delque-Kolic and C. Moreau (LMC14,
663 CNRS UMS2572) who performed the ¹⁴C dating, G. Delaviel-Anger and T. Crouzal for their
664 help in the field, and A. Jegouzo for enthusiastic discussions about the P3 eruption. This work
665 was partially funded by the Institut National des Sciences de l'Univers – Centre National de la
666 Recherche Scientifique programme CT3-ALEA, INSU-CNRS Artemis 2016 and 2017 for ¹⁴C
667 dating, and CASAVA (ANR contract ANR-09-ANR-RISK-002) and RAVEX (ANR contract
668 ANR-16-CE03-0002). This study contributes to the IdEx project "Université de Paris" (ANR-
669 18-IDEX-0001).

670

671 **References**

672 Bardintzeff, J.M., Miskovsky, J.-C., Traineau, H., Westercamp, D., 1989. The recent pumice
673 eruptions of Mt Pelée, Martinique. Part II: Grain-size studies and modelling the last Plinian
674 phase P1. *J. Volcanol. Geotherm. Res.* 38: 35-48.

675

676 Bonadonna, C., Costa, A. 2012. Estimating the volume of tephra deposits: a new simple
677 strategy. *Geology* 40(5): 415-418.

678

679 Bonadonna, C., Costa, A., 2013. Plume height, volume, and classification of explosive
680 volcanic eruptions based on the Weibull function. *Bull. Volcanol.* 75: 742-761.

681 Bonadonna, C., Houghton, B.F., 2005. Total grain-size distribution and volume of tephra fall
682 deposits. *Bull. Volcanol.* 67: 441-456.
683

684 Branney, M.J., Kokelaar, B.P., 2002. Pyroclastic density currents and the sedimentation of
685 ignimbrites. *Geol. Soc. Mem.* 27, 143 p.
686

687 Bronk Ramsey, C., 2009. Bayesian analysis of radiocarbon dates. *Radiocarbon* 51: 337–360.
688

689 Bursik, M.I., Woods, A.W., 1996. The dynamics and thermodynamics of large ash flows.
690 *Bull. Volcanol.* 58: 175– 193.
691

692 Calder, E.S., Cole, P.D., Dade, W.B., Druitt, T.H., Hoblitt, R.P., Huppert, H.E., Ritchie, L.,
693 Sparks, R.S.J., Young, S.R., 1999. Mobility of pyroclastic flows and surges at the Soufriere
694 Hills Volcano, Montserrat. *Geophys. Res. Lett.* 26: 537–540.
695

696 Carazzo, G., Tait, S., Kaminski, E., Gardner, J.E., 2012. The recent Plinian explosive activity
697 of Mt. Pelée volcano (Lesser Antilles): The P1 AD 1300 eruption, *Bull. Volcanol.* 74: 2187-
698 2203.
699

700 Carazzo, G., Kaminski, E., Tait S., 2015. The timing and intensity of column collapse during
701 explosive volcanic eruptions. *Earth Planet. Sci. Lett.* 411: 208-217.
702

703 Carazzo, G., Tait, S., Kaminski, E., 2019. Marginally stable recent plinian eruptions of Mt.
704 Pelée volcano (Lesser Antilles): The P2 AD 280 eruption, *Bull. Volcanol.* 81: 1-17.
705

706 Carey, S., Sigurdsson, H., 1986. The 1982 eruptions of El Chichon volcano, Mexico (2):
707 Observations and numerical modelling of tephra-fall distribution. *Bull. Volcanol.* 48: 127-
708 141.

709

710 Dade, W.B., Huppert, H.E., 1998. Long runout rockfalls. *Geology* 26: 803-806.

711

712 Daggit, M.L., Mather, T.A., Pyle, D.M., Page, S., 2014. AshCalc – a new tool for the
713 comparison of the exponential, power-law and Weibull models of tephra deposition. *J. Appl.*
714 *Volcanol.* 3: 7.

715

716 Di Muro, A., Neri, A., Rosi, M., 2004. Contemporaneous convective and collapsing eruptive
717 dynamics: the transitional regime of explosive eruptions. *Geophys. Res. Lett.* 31: L10607.

718

719 Di Muro, A., Rosi, M., Aguilera, E., Barbieri, R., Massa, G., Mundula, F., Pieri, F., 2008.
720 Transport and sedimentation dynamics of transitional explosive eruption columns: the
721 example of the 800 BP Quilotoa eruption (Ecuador). *J. Volcanol. Geotherm. Res.* 174: 307-
722 324.

723

724 Doyle, E.E., Hogg, A.J., Mader, H.M., Sparks, R.S.J., 2010. A two-layer model for the
725 evolution and propagation of dense and dilute regions of pyroclastic currents. *J. Volcanol.*
726 *Geotherm. Res.* 190: 365–378.

727

728 Fierstein, J., Hildreth, W., 1992. The Plinian eruptions of 1912 at Novarupta, Katmai National
729 Park, Alaska. *Bull. Volcanol.* 54: 646-684.

730

731 Girault, F., Carazzo, G., Tait, S., Ferrucci, F., Kaminski, E., 2014. The effects of total grain-
732 size distribution on the dynamics of turbulent volcanic plumes. *Earth Plan. Sci. Lett.* 394:
733 124-134.
734

735 Girault, F., Carazzo, G., Tait, S., Kaminski, E., 2016. Combined effects of total grain-size
736 distribution and crosswind on the rise of explosive volcanic columns. *J. Volcanol. Geotherm.*
737 *Res.* 326: 103-113.
738

739 Gurioli, L., Houghton, B.F., Chasman, K.V., Cioni, R., 2005. Complex changes in eruption
740 dynamics during the 79 AD eruption of Vesuvius. *Bull. Volcanol.* 67; 144-159.
741

742 Kaminski, E., Jaupart, C., 1998. The size distribution of pyroclasts and the fragmentation
743 sequence in explosive volcanic eruptions. *J. Geophys. Res.* 103: 29,759-29,779.
744

745 Koyaguchi, T., Tokuno, M., 1993. Origin of the giant eruption cloud of Pinatubo, June 15,
746 1991. *J. Volcanol. Geotherm. Res.* 55: 85-96.
747

748 Mastin, L.G., Guffanti, M., Seizvrancx, R., Webley, P., Barsotti, S., dean, K., Durant, A.,
749 Ewert, J.W., Neri, A., Rose, W.I., Schneider, D., Siebert, L., Stunder, B., Swanson, G.,
750 Tupper, A., Volentik, A., Waythomas, C.F., 2009. A multidisciplinary effort to assign
751 realistic source parameters to models of volcanic ash-cloud transport and dispersion during
752 eruptions. *J. Volcanol. Geotherm. Res.* 186: 10-21.
753

754 Martel, C., Pichavant, M., Bourdier, J.-L., Traineau, H., Holtz, F., Scaillet, B., 1998. Magma

755 storage conditions and control of eruption regime in silicic volcanoes: experimental evidence
756 from Mt Pelée. *Earth Planet. Sci. Lett.* 156: 89–99.

757

758 Martel, C., Poussineau, S., 2007. Diversity of eruptive styles inferred from the microlites of
759 Mt. Pelée andesite (Martinique, Lesser Antilles). *J. Volcanol. Geotherm. Res.* 166: 233–254.

760

761 Michaud-Dubuy, A., Carazzo, G., Kaminski, E., Girault, F., 2018. A revisit of the role of gas
762 entrapment on the stability conditions of explosive volcanic columns. *J. Volcanol. Geotherm.*
763 *Res.* 357: 349-361.

764

765 Michaud-Dubuy, A., Carazzo, G., Tait, S., Le Hir, G., Fluteau, F., Kaminski, E., 2019. Impact
766 of wind direction variability on hazard assessment in Martinique (Lesser Antilles): the
767 example of the 13.5 ka cal BP Bellefontaine Plinian eruption of Mount Pelée volcano. *J.*
768 *Volcanol. Geotherm. Res.* 381: 193-208.

769

770 Newhall, C.G., Self, S., 1982. The volcanic explosivity index (VEI): an estimate of explosive
771 magnitude for historical volcanism. *J. Geophys. Res.* 87: 1231-1238.

772

773 Pyle, D.M., 1989. The thickness, volume and grainsize of tephra fall deposits. *Bull. Volcanol.*
774 51:1–15.

775

776 Pyle, D.M., 2000. Sizes of volcanic eruptions. In: Sigurdsson, H., Houghton, B., Reimer, H.,
777 Stix, J., McNutt, S. (eds) *Encyclopedia of volcanoes*. Academic, San Diego, pp 263-269.

778

779 Reimer, P., 2013. Selection and treatment of data for radiocarbon calibration: an update to the
780 international calibration (IntCal) criteria. *Radiocarbon* 55: 1923–1945.
781

782 Roobol M.J., Smith A.L., 1976. Stratigraphic studies of Mount Pelée, Martinique. *Bull.*
783 *B.R.G.M.*, sect. IV, 4: 297–304.
784

785 Self, S., Rampino, M.R., 2012. The 1963-1964 eruption of Agung volcano (Bali, Indonesia).
786 *Bull. Volcanol.* 74: 1521-1536.
787

788 Shea, T., Gurioli, L., Houghton, B.F., Cioni, R., Cashman, K.V., 2011. Column collapse and
789 generation of pyroclastic density currents during the A.D. 79 eruption of Vesuvius: The role
790 of pyroclast density. *Geology* 39: 695-698.
791

792 Shea, T., Gurioli, L., Houghton, B.F., 2012. Transitions between fall phases and pyroclastic
793 density currents during the AD 79 eruption at Vesuvius: Building a transient conduit model
794 from the textural and volatile record. *Bull. Volcanol.* 74: 2363-2381.
795

796 Shea, T., Hellerbrand, E., Gurioli, L., Tuffen, H., 2014. Conduit- to localized-scale degassing
797 during Plinian eruptions: Insights from major element and volatile (Cl and H₂O) analyses
798 within Vesuvius AD 79 pumice. *J. Petrol.* 55(2): 315-344.
799

800 Sigurdsson, H., Carey, S., Fisher, R.V., 1984. The 1982 eruptions of El Chichon volcano,
801 Mexico: stratigraphy of pyroclastic deposits. *J. Volcanol. Geotherm. Res.* 23: 11-37.
802

803 Suzuki, Y.J., Costa, A., Cerminara, M., Esposti Ongaro, T., Herzog, M., Van Eaton, A.R.,
804 Denby, L.C., 2016. Inter-comparison of three dimensional models of volcanic plumes. *J.*
805 *Volcanol. Geotherm. Res.* 326: 26-42.

806

807 Traineau, H., 1982. Contribution à l'étude géologique de la Montagne Pelée, Martinique.
808 Evolution de l'activité éruptive au cours de la période récente. Thèse de 3ième cycle,
809 Université de Paris Sud, Orsay, 209 p.

810

811 Traineau, H., Westercamp, D., Bardintzeff, J.M., Miskovsky, J.C., 1989. The recent pumice
812 eruptions of Mt. Pelée volcano, Martinique. Part I: Depositional sequences, description of
813 pumiceous deposits. *J. Volcanol. Geotherm. Res.* 38: 17–33.

814

815 Turcotte, D.L., 1986. Fractals and fragmentation. *J. Geophys. Res.* 91: 1921-1926.

816

817 Villemant, B., Boudon, G., Komorowski J.-C., 1996. U-series disequilibrium in arc magmas
818 induced by water-magma interaction. *Earth Planet. Sci. Lett.* 140: 259-267.

819

820 Walker, G.P.L., 1980. The Taupo pumice: product of the most powerful known (ultraplinian)
821 eruption? *J. Volcanol. Geotherm. Res.* 8: 69-94.

822

823 Walker, G.P.L., 1983. Explosive volcanic eruptions - a new classification scheme. *Geol.*
824 *Rundsch.* 62: 431-446.

825

826 Ward, G.K., Wilson, S.R., 1978. Procedures for comparing and combining radiocarbon age
827 determinations: a critique. *Archaeometry* 20: 19–31.

828

829 Westercamp D., Traineau H., 1983. The past 5,000 years of volcanic activity at Mt. Pelée
830 Martinique (F.W.I.): implications for assessment of volcanic hazards. *J. Volcanol. Geotherm.*
831 *Res.* 17: 159–185.

832

833 Wilson, L., Sparks, R.S.J., Walker, G.P.L., 1980. Explosive volcanic eruptions-IV. The
834 control of magma properties and conduit geometry on eruption column behavior. *Geophys. J.*
835 *R. Astron. Soc.* 63: 117-148.

836

837 Wilson, L., Walker, G.P.L., 1985. The Taupo eruption, New Zealand. I General aspects.
838 *Philos. Trans. R. Soc. Lond. A* 314: 199-228.

839

840 Wilson, G., Wilson, T.W., Deligne, N.I., Cole, J.W., 2014. Volcanic hazard impacts to critical
841 infrastructure: a review. *J. Volcanol. Geotherm. Res.* 286: 148-182.

842

843 Woods, A.W., Bower, S.M., 1995. The decompression of volcanic jets in a crater during
844 explosive volcanic eruptions. *Earth Planet. Sci. Lett.* 131: 189-205.

845

846 Wright, J.V., Smith A.L., Roobol M.J., Mattioli G.S., Fryxell J.E., 2016. Distal ash hurricane
847 (pyroclastic density current) deposits from a ca. 2000 yr B.P. Plinian-style eruption of Mount
848 Pelée, Martinique: Distribution, grain-size characteristics, and implications for future hazards,
849 *Geol. Soc. Am. Bull.* 128: 777–791.

850

851 **Table captions**

852

Sample	Site	Unit	Subunit	Altitude (m)	Distance from the vent (km)	Thickness (cm)
1	163	A	(0-40 cm)	269	3.7	160
2	163	A	(40-60 cm)	269	3.7	160
3	163	A	(60-100 cm)	269	3.7	160
4	163	A	(100-140 cm)	269	3.7	160
5	163	A	(140-160 cm)	269	3.7	160
6	163	B	Bulk	269	3.7	18
7	163	C	Bulk	269	3.7	20
8	163	D	Bulk	269	3.7	15
9	163	E	Bulk	269	3.7	13
10	163	F	Bulk	269	3.7	6
11	163	G	Bulk	269	3.7	37
12	88	A-G	Bulk	466	4.3	45
13	66b	A-G	Bulk	371	4.9	65
14	142	A-G	Bulk	234	5.8	60
15	194	A-G	Bulk	280	5.8	55
16	114	A-G	Bulk	195	6.3	105
17	137	A-G	Bulk	19	6.5	100
18	191	A-G	Bulk	35	6.5	46
19	144	A-G	Bulk	47	6.6	80
20	145	A-G	Bulk	540	11.9	25

853 **Table 1:** Sampling of the P3 deposits for grain-size analysis.

854

855

856

857

858

859

860

861

862

863

864

Parameters	P3 (79 CE)	P2 (280 CE)	P1 (1300 CE)
Total DRE volume	1.02 km³	0.77 km ³	0.16 km ³
Erupted mass	2.4 x 10¹² kg	1.7 x 10 ¹² kg	4 x 10 ¹¹ kg
VEI	5	4	4
Magnitude	5.4	5.2	4.6
Maximum column height	30 km	26 km	22 km
Maximum PDC runout	7 – 10.3 km	7 – 11.5 km	4.5 – 8 km
Maximum MER (fall)	1.4 x 10⁸ kg s⁻¹	1.1 x 10 ⁸ kg s ⁻¹	3.6 x 10 ⁷ kg s ⁻¹
Maximum MER (PDC)	2.5 x 10⁸ kg s⁻¹	3.1 x 10 ⁸ kg s ⁻¹	9 x 10 ⁷ kg s ⁻¹
Maximum intensity	11.4	11.5	11
Duration	> 11 h	> 7 h	> 5 h
TGSD	D = 3.3	D = 3.4 - 3.5	D = 3.2 – 3.3
Minimum exit velocity	210 - 220 m s⁻¹	180 – 200 m s ⁻¹	150 – 165 m s ⁻¹
Total gas content	2 - 2.9 wt.%	1.8 – 2.2 wt.%	1.6 – 2.1 wt.%

865 **Table 2:** Summary of the estimated eruptive parameters for the P3 eruption, and comparison
866 with P1 and P2 ([Carazzo et al., 2012; 2019](#)).

867

868

869

870

871

872

873

874

Phase	n_0 (wt.%)	α_c	x_0 (wt.%)	x_f (wt.%)	U_{free} (m s ⁻¹)	U_{min} (m s ⁻¹)
Early A	5.8	29	4.1	2.9	230	210
Late A	5.8	38	3.6	2.5	215	220
B-C	5.8	42	3.4	2.3	-	220
D-E	5.8	45	3.2	2.2	-	-
F-G	5.8	49	2.9	2.1	-	-

875 **Table 3:** Gas contents and deduced exit velocities. n_0 is the total volatile content of the melt
876 ([Martel et al., 1998](#)), α_c is the percentage of crystals in the melt plus lithics in the flow, x_0 is
877 the mass fraction of gas in the magma assuming complete degassing, x_f is the mass fraction of
878 exsolved gas in the gas+pyroclasts mixture at fragmentation for a threshold vesicularity of
879 70%, U_{free} is the supersonic velocity after decompression calculated using x_0 , $D = 3.3$ and an
880 open porosity of 60-70% ([Michaud-Dubuy et al., 2018](#)), and U_{min} is the eruptive velocity
881 deduced from the distribution of the lithic fragments (section 5.2).

882

883

884

885

886

887

888

889

890

891

892

893 **Figure captions**

894

895 **Fig. 1:** Current maps for the P3 eruption. **a** Isopach map (in centimeters) from [Westercamp](#)
896 [and Traineau \(1983\)](#). (1) Dashed and (2) dotted lines correspond to isopachs for the pumice
897 fallout deposit P3₁ and P3₂, respectively. (3) The light grey and (4) dark areas give the
898 distribution of the low-concentration PDC (surge), and high-concentration PDC (flow)
899 deposits P3₃, respectively. **b** Isopach map (in centimeters) from [Wright et al. \(2016\)](#). (1)
900 Dashed lines correspond to isopachs for a pumice fallout deposit. (2) The light and (3) dark
901 grey areas give the distribution of a low-concentration PDC (surge) deposit, and a low-
902 concentration PDC (ash hurricane) deposit, respectively. (4) The dark areas correspond to
903 high-concentration PDC (flow) deposits.

904

905 **Fig. 2:** Overview of our field area in Martinique (inset). White circles and numbers refer to
906 localities where P3 deposits are present. Black dots show outcrops where P3 deposits are
907 absent (due to erosion) and/or too deeply buried under recent eruption deposits. The dotted
908 lines give the stratigraphic correlation reported in [Fig. 5](#).

909

910 **Fig. 3:** **a** Composite stratigraphic section showing typical unit thickness and grain-size
911 characteristics of the P3 eruptive sequence. **b** Grain-size distribution selected samples
912 representing the different eruptive units, including five stratigraphic heights in unit A, at
913 location 163. **c** Variations of the median grain-size (blue circles), sorting (green circles),
914 juvenile lithic content (grey squares), accidental lithic content (dark squares), total lithic
915 content (white squares), and lithic size (red squares) along the stratigraphic height at location
916 163.

917

918 **Fig. 4:** Representative photographs of outcrops of the P3 deposits in Martinique at sites **a** 57,
919 **b** 78, **c** 13, **d** 65, and **e**, **f** 163. See [Fig. 2](#) for location and distance from the source. All scale
920 bars are 20 cm long.

921

922 **Fig. 5:** Stratigraphic logs of representative sections of deposits of the P3 eruption. See [Fig. 2](#)
923 for outcrop locations.

924

925 **Fig. 6:** Isopach maps (in centimeters) for **a** the cumulative pumice fallout (units A+C+E+G),
926 **b** the pumice fallout unit A, **c** the cumulative PDC deposits (units B+D+F) and **d** the
927 cumulative pumice fallout (units C+E+G). Open circles indicate measured sample locations.
928 Bold numbers in panel (c) indicate the thicknesses of high-concentration PDC deposits.

929

930 **Fig. 7:** Isopleth maps (in millimeters) for the lithic fragments sampled **a** at the base of unit A,
931 **b** at the top of unit A, and **c** the base of unit C. Open circles indicate measured sample
932 locations. Direction of dispersal axes is consistent with those inferred from the isopach maps.

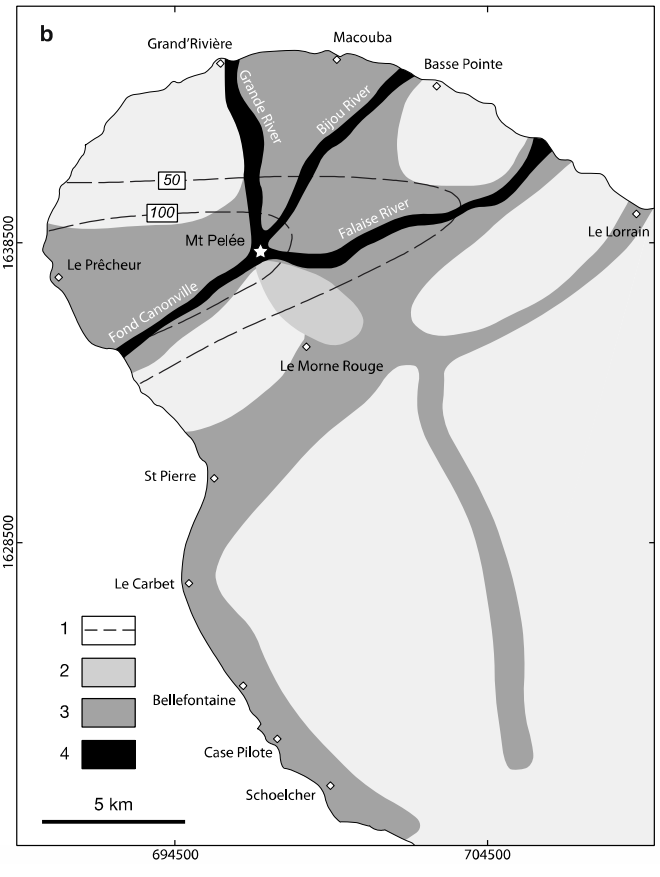
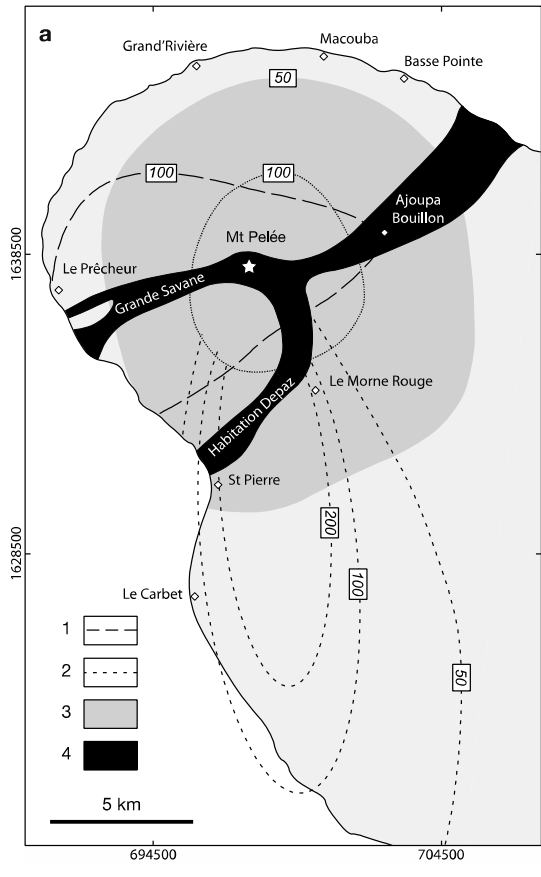
933

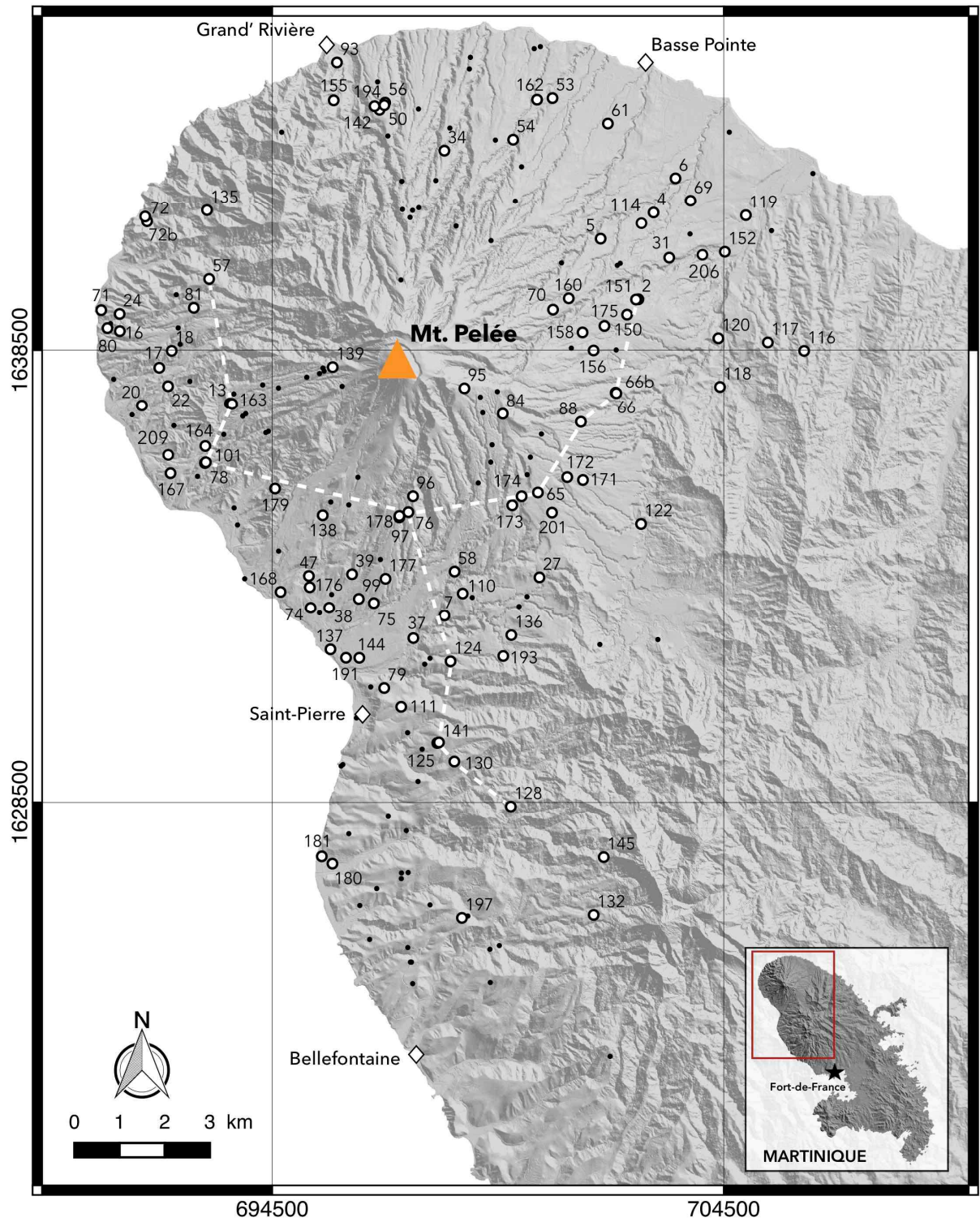
934 **Fig. 8:** Deposit thinning profiles generated from the isopach maps for **a** unit A, **b** the
935 cumulative pumice fallout C+E+G, and **c** the cumulative pumice fallout A+C+E+G
936 represented by semi-log plots of square root of isopach areas (in kilometers) versus thickness
937 (in meters). Thinning trends are approximated by exponential (purple dashed line), power-law
938 (blue dotted line), and Weibull (red solid line) fits.

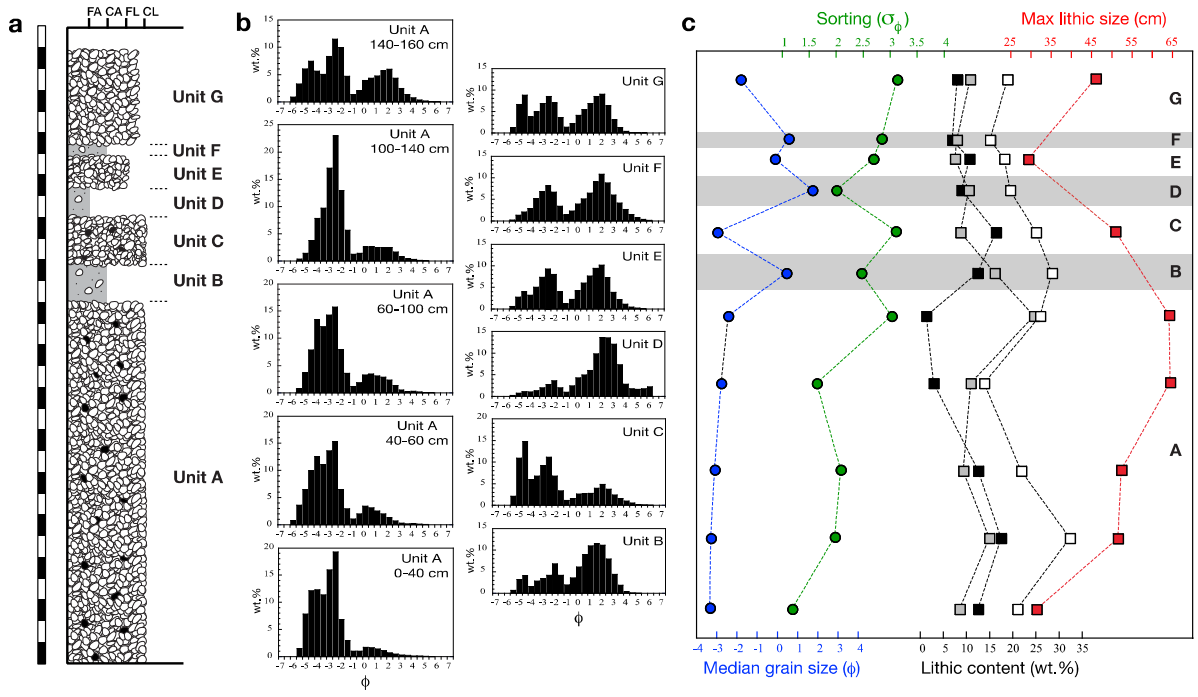
939

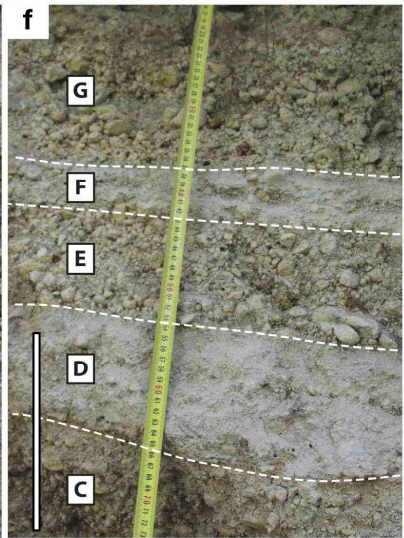
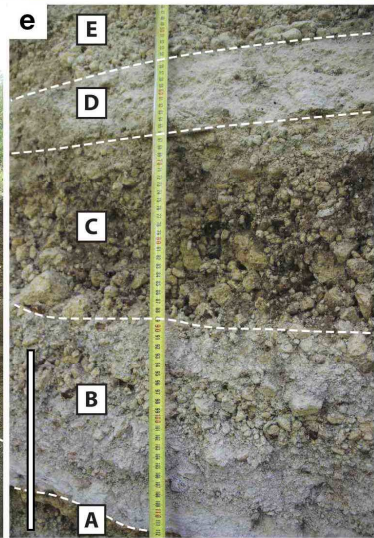
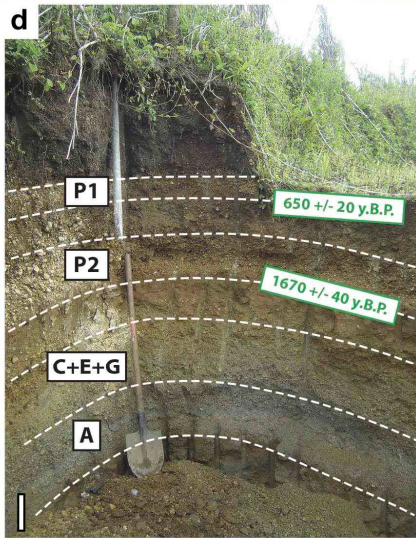
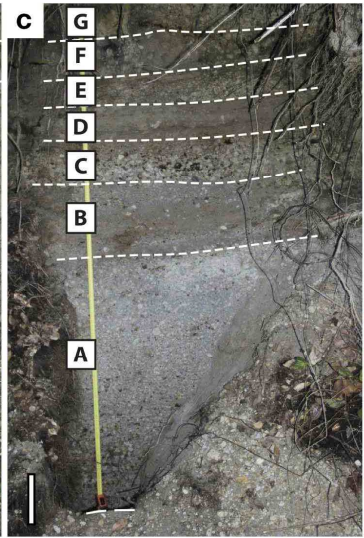
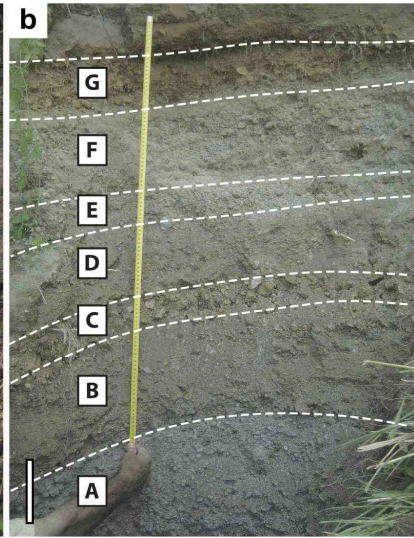
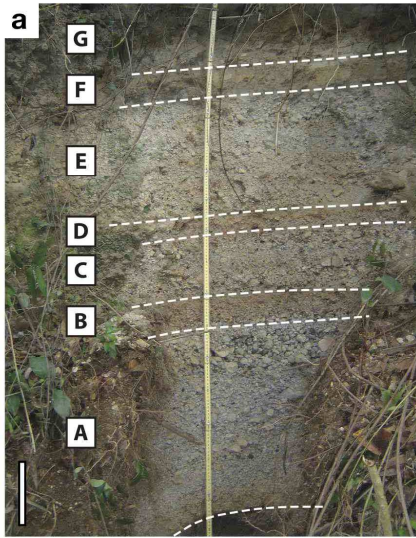
940 **Fig. 9:** Summary of the time evolution of the P3 eruptive parameters. **a** Cumulative erupted
941 mass (in kg). **b** Maximum column height (in km). **c** Mass eruption rate (in kg s^{-1}). **d** Total
942 exsolved gas content, x_f (in wt.%).

943 **Fig. 10:** Transition diagram depicting the P3 eruptive events and depositional units
944 (diamonds), and comparison with the P1 (squares), P2 (triangles), Taupo (circles) and
945 Vesuvius (inverted triangles) eruptions. Numbers associated with diamonds correspond to
946 different phases of the P3 eruption: 1: beginning of phase A, 2: end of phase A, 3:
947 simultaneous phases B and C, 4: simultaneous phases D and E, and 5: simultaneous phases F
948 and G. Dashed line corresponds to the maximum mass eruption rate feeding the column
949 before collapse as a function of the total gas content at the column base (using [Michaud-](#)
950 [Dubuy et al. \(2018\)](#) model for tropical conditions and $D = 3.3$). Open, gray, and dark symbols
951 correspond to geological data inferred for the stable column phase, partial column collapse
952 phase, and total column collapse phase, respectively. Data for the P1, P2, Taupo and Vesuvius
953 eruptions are compiled in [Carazzo et al. \(2012, 2019\)](#) and [Michaud-Dubuy et al. \(2018\)](#),
954 respectively.









North West

South

South

North East

