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# Consequences of volcano sector collapse on magmatic storage zones: insights from numerical modeling

V. Pinel<sup>a</sup>, F. Albino<sup>b</sup>

<sup>a</sup>*ISTerre, Université de Savoie, IRD, CNRS, F73376 Le Bourget du Lac, France*

<sup>b</sup>*Nordic Volcanological Center, Institute of Earth Sciences, University of Iceland, 101 Reykjavik, Iceland*

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## Abstract

Major volcano flank collapses strongly affect the underlying magmatic plumbing system. Here, we consider the magma storage zone as a liquid pocket embedded in an elastic medium, and we perform numerical simulations in two-dimensional axisymmetric geometry as well as in three dimensions in order to evaluate the consequences of a major collapse event. We quantify the pressure decrease induced within and around a magma reservoir by a volcano flank collapse. This pressure reduction is expected to favour replenishment with less evolved magma from deeper sources. We also estimate the impact of the magma pressure decrease, together with the stress field variations around the reservoir, on the eruptive event associated with the edifice failure. We show that, for a given magma reservoir geometry, the collapse of a large strato-volcano tends to reduce the volume of the simultaneous eruption; destabilization of large edifices may even suppress magma emission, resulting in phreatic eruptions instead. This effect is greater for shallow reservoirs, and is more pronounced for spherical reservoirs than for

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*Email address: [Virginie.Pinel@univ-savoie.fr](mailto:Virginie.Pinel@univ-savoie.fr) (V. Pinel)*

vertically-elongated ones. It is reduced for compressible magmas containing a large amount of volatiles. Over a longer time scale, the modification of pressure conditions for dyke initiation at the chamber wall may also explain an increase in eruption rate as well as an apparent change of magma storage location.

*Keywords:* Edifice flank collapse, numerical modeling

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## 1. Introduction

Large flank collapses have been recognized as common phenomena in the long-lived evolution of volcanic edifices. A large number of studies focus on the causes of, and/or triggers for, these destabilization events. They show that the origin of the destabilization can be related to exogenous processes such as weathering, but in most cases volcanic activity itself is involved (Mc Guire, 1996). In particular, the ability of magmatic intrusions to favour large flank collapse either during vertical dyke emplacement or during sill formation has been observed in the field (Famin and Michon, 2010), and investigated through modeling (Paul et al., 1987; Iverson, 1995). Siebert (1992) emphasized the potential hazards represented by sector collapses, which fully justify studies investigating volcano stability (Voight and Elsworth, 1997; Borselli et al., 2011). From a risk assessment perspective, the direct impact of a sudden and drastic sector collapse is also investigated through studies or modelling related to the volume and extension of the associated debris avalanche deposits (Borselli et al., 2011).

Another field of study encompasses describing and quantifying the consequences of such an event on the magmatic plumbing system evolution. The

19 long-term history of volcanic edifices reveals that partial destruction of an  
20 edifice is usually followed by a change in eruption rate and/or magma com-  
21 position (Presley et al., 1997; Hildenbrand et al., 2004; Hora et al., 2007;  
22 Longpré et al., 2010; Boulesteix et al., 2012). For oceanic volcanoes, this  
23 observation has usually been related to an increase in decompression melting  
24 subsequent to collapse (Presley et al., 1997; Hildenbrand et al., 2004), al-  
25 though Manconi et al. (2009) also evoked the depressurisation of a magmatic  
26 storage zone. For continental volcanoes, Pinel and Jaupart (2005), using an  
27 analytical elastic model for the two-dimensional plane strain approximation,  
28 quantified the pressure decrease induced within a magmatic reservoir by the  
29 partial destruction of an edifice. They also detailed the influence of such an  
30 event on the volume of magma erupted during the failure event.

31 Meanwhile, other surface load variations, occurring over a larger time  
32 scale, have been proven to have a significant impact on eruptive behaviour. In  
33 particular, a temporal correlation is observed between ice retreat induced by  
34 climate warming and volume of magma erupted, with an increase of eruption  
35 rates during postglacial periods (Jellinek et al., 2004; Sinton et al., 2005). The  
36 effect of ice retreat on both magma melting and storage has been investigated  
37 (see Sigmundsson et al. (2010) for a review). More recently, new modeling  
38 has shown that magma propagation within the upper crust is also affected  
39 by ice unloading, with an increased likelihood of magma storage within the  
40 crust during transport towards the surface. This is in good agreement with  
41 some geodetic observations performed around Vatnajökull ice cap in Iceland  
42 (Hooper et al., 2011).

43 In this study, we calculate the pressure reduction induced by a sudden

44 flank collapse event within and around a magma storage zone located beneath  
45 a volcanic edifice. We only consider one-shot catastrophic flank collapses,  
46 rather than the effects of large, progressive landslides. We then quantify  
47 how the flank collapse affects the volume of erupted magma during the as-  
48 sociated eruption, resulting from the storage zone withdrawal. Results are  
49 derived from numerical simulations incorporating the equation of elasticity,  
50 performed with the commercial software COMSOL both in axisymmetric ge-  
51 ometry and in three dimensions. We also discuss the potential impact of  
52 large flank collapses on the long-term eruptive history, based on petrological  
53 observations.

## 54 **2. Pressure decrease induced by a volcano flank collapse**

55 Broadly speaking, a major sector collapse is equivalent to a surface un-  
56 loading event. In reality, the edifice portion which fails is not removed from  
57 the Earth's surface, but is redistributed over a larger area. As previously  
58 shown by Pinel and Jaupart (2005), using analytical solutions, and by Al-  
59 bino et al. (2010), using numerical models in axisymmetric geometry, an  
60 unloading event always induces a pressure decrease within the underlying  
61 crust. This pressure reduction is of the same order of magnitude as the load  
62 removed from the surface.

### 63 *2.1. A conical load removed over an elastic half-space*

64 If we consider the crust to be an elastic, homogeneous medium charac-  
65 terized by its Young's modulus,  $E$ , and Poisson's ratio,  $\nu$ , the stress changes  
66 induced at depth by a conical load can be derived by integration of the point  
67 load solution. At the axis, the vertical stress due to a cone of radius,  $R_e$ , and

68 maximum height,  $H_e$ , as a function of depth below the surface,  $z$ , is given  
 69 by:

$$\sigma_{zz} = P_e \left[ 1 - \frac{z}{\sqrt{R_e^2 + z^2}} \right], \quad (1)$$

70 with  $P_e = \rho_c g H_e$ , where  $\rho_c$  is the load density. The horizontal components  
 71 are equal and given by Pinel and Jaupart (2000):

$$\begin{aligned} \sigma_{rr} = \sigma_{\theta\theta} &= \frac{P_e}{2} \left[ (1 + 2\nu) - 2(1 + \nu) \frac{z}{R_e} \ln \left( \frac{R_e + \sqrt{R_e^2 + z^2}}{z} \right) + \frac{z}{\sqrt{R_e^2 + z^2}} \right] & \text{for } z > 0 \\ \sigma_{rr} = \sigma_{\theta\theta} &= \frac{P_e}{2} (1 + 2\nu) & \text{for } z = 0, \end{aligned} \quad (2)$$

72 It follows that the pressure,  $P$ , defined as one third on the stress tensor  
 73 trace, induced by a conical load, is equal to:

$$\begin{aligned} P &= \frac{2P_e}{3} (1 + \nu) \left[ 1 - \frac{z}{R_e} \ln \left( \frac{R_e + \sqrt{R_e^2 + z^2}}{z} \right) \right] & \text{for } z > 0 \\ P &= \frac{2P_e}{3} (1 + \nu) & \text{for } z = 0 \end{aligned} \quad (3)$$

74 Stress and pressure reduction induced by the removal of conical load are  
 75 shown in figure 1. The stress component most affected by the load is, as  
 76 expected, the vertical one  $\sigma_{zz}$ . The amplitude of the perturbation is greatest  
 77 at the surface, and is directly related to the height of the load removed. The  
 78 pressure reduction decreases with depth and becomes negligible at depths  
 79 greater than three times the radius of the load.

## 80 2.2. A conical load removed from above a magmatic reservoir

81 Most tomographic studies performed on volcanoes (Monteiller et al., 2005;  
 82 Prôno et al., 2009) reveal that the crust is far from being homogeneous around  
 83 a magmatic system. In particular, shallow magma storage zones have been  
 84 detected in many locations by either petrologic, seismic or geodetic studies  
 85 (Gardner et al., 1995; Sturkell et al., 2006; Peltier et al., 2008). The pressure

86 reduction induced by an unloading event within these magma pockets will  
87 depend both on the crustal deformation and on the equation of state of  
88 the melt embedded in the crust. Here we consider an ellipsoidal magmatic  
89 reservoir filled with fluid, embedded in an homogeneous elastic crust. Initially  
90 the liquid has the same density as the surrounding crust and is characterized  
91 by its bulk modulus,  $K$ . We only deal with the perturbation induced by  
92 a conical load removed from the Earth's surface, on which the initial stress  
93 field has no influence.

94 Within the magma reservoir the pressure change,  $\Delta P$ , is related to the  
95 reservoir volume change,  $\Delta V$ , through the bulk modulus definition:

$$\Delta P = -K \frac{\Delta V}{V}, \quad (4)$$

96 with  $V$  being the initial volume of the reservoir.

97 The change in reservoir volume is also a function of the chamber wall dis-  
98 placement, which depends on both the conical load and the magma pressure  
99 change. This volume change is calculated numerically, using the equations  
100 of elasticity with the commercial software COMSOL. The domain of calcu-  
101 lation is a 100\*100 km square box with a mesh of about 100 000 triangular  
102 units that is refined around the volcanic edifice and magma reservoir. No  
103 displacement perpendicular to the boundary is allowed at the basal and lat-  
104 eral boundaries; the upper boundary is considered as being a free surface.  
105 The load is modeled with a normal stress applied at the upper surface, and  
106 a normal stress equal to the magma overpressure is applied at the reservoir  
107 walls. Numerical solutions have been validated using well-known analytical  
108 solutions as detailed in Albino et al. (2010). Pressure reduction within and  
109 around the magma reservoir induced by a conical load of 2 km radius, 1 km

110 height, and density  $2800 \text{ kg/m}^3$ , is shown in figure 2 for two different chamber  
111 geometries: a spherical chamber, and a vertically-elongated chamber (pro-  
112 late shape). Calculations are performed for a chamber top at 1 km depth  
113 and with a maximum chamber radius of 1 km. The pressure variation within  
114 the crust departs strongly from the homogeneous case (figure 1) in the vicin-  
115 ity of the reservoir, this difference being more pronounced for the spherical  
116 reservoir than for the prolate one. An increase in pressure is even observed  
117 (negative values of the pressure reduction) at the chamber margins, the effect  
118 being most extreme at the chamber top. This is due to the deformation of  
119 the reservoir wall induced by the unloading event and partially counterbal-  
120 anced by pressure variations within the magma reservoir. Within the magma  
121 reservoir, pressure always decreases as a consequence of the unloading event,  
122 the effect being, once again, larger in the case of the spherical reservoir than  
123 in the case of the prolate one. The amplitude of the magma pressure reduc-  
124 tion increases with the value of the bulk modulus. This can be explained by  
125 the fact that for incompressible magmas (larger value of  $K$ ) no reservoir vol-  
126 ume change occurs, thus all the volume reduction induced by the unloading  
127 event has to be compensated by a pressure reduction within the chamber.  
128 The effect of compressibility is shown in figure 3. In a compressible magma,  
129 buoyancy forces appear due to magma density variation; however since the  
130 model used here is valid in most of natural cases, as discussed by Pinel and  
131 Jaupart (2005), these forces have not been included here.

132 Figure 5 shows the pressure reduction within a spherical reservoir with  
133 a top at 1 km depth, induced by the removal of the upper 20 % of the  
134 edifice, which corresponds to a mean value based on field observations (figure



135 4a). This pressure reduction is more marked when considering larger edifices  
136 and smaller magma reservoirs. Calculations performed also show that the  
137 amplitude of the pressure reduction decreases for deeper magma reservoirs  
138 and is less marked in the case of a prolate reservoir than in a spherical one.

### 139 **3. Effect of volcano flank collapse on an ongoing eruption**

140 As described above, partial destruction of an edifice always induces a  
141 pressure decrease in the underlying storage zone. Simultaneously, this drastic  
142 change in surface load strongly modifies the stress field at the reservoir walls,  
143 and thus presumably dyke initiation and closure in this zone. While long-  
144 term conduit systems develop at the surface of silicic strato-volcanoes, this  
145 shallow level conduits are connected to deeper reservoirs via dykes as shown  
146 by deformation data (Mattioli et al., 1998), seismicity (Roman et al., 2006),  
147 and magma flow studies (Costa et al., 2007). Thus it is important to consider  
148 dyke opening and closure at the reservoir walls when dealing with magma  
149 transport to the surface.

#### 150 *3.1. Principle*

151 For any given state of stress, one may define a threshold pressure,  $P_r$ ,  
152 required for dyke initiation at the chamber wall. Here, we consider that  
153 once this pressure threshold is reached within the magma reservoir, a dyke  
154 initiates and magma leaves the storage zone to reach the surface and feed  
155 an eruption, such that pressure within the magma chamber cannot exceed  
156 this threshold value. Following Albino et al. (2010), we consider that tensile  
157 rupture of the reservoir wall occurs when the deviatoric part of the hoop stress  
158 exceeds the tensile strength,  $T_s$ , of host rocks. This criterion is consistent

159 with the rupture criterion used in dyke propagation studies (Lister, 1990).  
160 Once a dyke has initiated, magma pressure within the reservoir decreases  
161 until it reaches a second threshold value,  $P_{cl}$ , at which point the dyke closes  
162 (Pinel et al., 2010). It follows that the volume of erupted magma is directly  
163 proportional to the pressure difference:  $\Delta P_e = P_r - P_{cl}$ .

164 We consider an initial state, corresponding to the situation just before the  
165 edifice partial destruction, with a pressurized storage zone. For this initial  
166 state, we can define the two threshold values,  $P_r(i)$  and  $P_{cl}(i)$  (see figure  
167 6). As magma intrusions often acts as the triggers for edifice destabilization  
168 (Mc Guire, 1996), we make the assumption that the system is about to  
169 erupt, such that the magma pressure,  $P(i)$ , is equal to the threshold value  
170 required for dyke initiation,  $P_r(i)$ . A major collapse of the edifice then occurs,  
171 and our final state is a truncated edifice. The unloading event results in a  
172 pressure reduction ( $\Delta P$ ) within the magma reservoir, such that the final  
173 magma pressure within the reservoir is  $P(f) < P(i)$ . In the final state, the  
174 two threshold pressures, here denoted  $P_r(f)$  and  $P_{cl}(f)$  are different from the  
175 ones in the initial state (see figure 6). Based on the evolution of the threshold  
176 pressures, which can either increase or decrease, three different scenarios can  
177 result. Where the pressure difference,  $P - P_{cl}$ , increases with the edifice  
178 partial destruction (case 1 of figure 6, where  $P(f) - P_{cl}(f) > P(i) - P_{cl}(i)$ ), the  
179 edifice collapse is followed by an eruption with a volume of erupted magma  
180 larger that it would have been in the absence of edifice destruction. Where  
181 the pressure difference,  $P - P_{cl}$ , decreases with the edifice partial destruction  
182 (case 2 of figure 6, where  $P(f) - P_{cl}(f) < P(i) - P_{cl}(i)$ ), the erupted volume  
183 associated with the edifice collapse is smaller that it would have been in

184 absence of edifice destruction. In the event of the magma pressure within  
185 the reservoir dropping below the threshold pressure for dyke closure (case 3  
186 of figure 6, where  $P(f) < P_{cl}(f)$ ), the incipient eruption is aborted and no  
187 magma is erupted at the surface as a consequence of the edifice collapse.

188 Factors determining which scenario occurs are the initial edifice size and  
189 geometry, the amount of edifice destruction, the size, shape and depth of the  
190 magma reservoir, and magma gas content.

### 191 *3.2. Numerical results for axisymmetric models*

192 We performed numerical simulations for edifices of different initial sizes,  
193 with various reservoir depths and sizes, and two different reservoir shapes:  
194 a spherical one and a vertically-elongated one (prolate shape). We consid-  
195 ered the effect of partial destruction, corresponding to the removal of the  
196 upper 20 % of a strato-volcano with a conical shape and a slope of 30 de-  
197 grees (figure 9 B). A slope of 30 degrees is an upper limit for the upper part  
198 of strato-volcanoes based on a compilation of Digital Elevation Models, and  
199 corresponds to the maximum frequency of major slope failure events on Qua-  
200 ternary volcanoes (Voight and Elsworth, 1997); collapse of 20 % of the initial  
201 volcano can be considered as a mean value based on a compilation of field  
202 observations (figure 4a). In all simulations we consider an incompressible  
203 magma; effects of compressibility and gas content are discussed at the end  
204 of this section, and in section 5.

205 For a spherical reservoir located at 1 km depth beneath the volcanic ed-  
206 ifice, each of the three scenarios previously described (figure 6) can occur  
207 (see figure 7 b). When the collapse affects small edifices, the erupted vol-  
208 ume is larger than, but still close to, that expected in the absence of edifice

209 destruction. As the edifice size increases, so the volume of erupted magma  
210 after the edifice collapse tends to decrease. This volume reduces to zero when  
211 large strato-volcanoes are partially destroyed by flank collapse, resulting in  
212 the abortion of any incipient eruption. For a shallower magma reservoir, a  
213 smaller edifice size is required to reach the point of aborted eruption, (see  
214 figure 7 a), whereas all effects of edifice collapse on erupted magma volume  
215 are reduced with a deeper chamber (see figure 7 c).

216 Effects of an edifice collapse event on the subsequent eruption also depend  
217 on the magma reservoir shape. Figure 8 shows that, for a prolate reservoir,  
218 the influence of the collapse is smaller, and a larger edifice size is required  
219 in order to decrease the amount of erupted magma, than for a spherical  
220 reservoir at the same depth. Above a prolate reservoir with a top at one  
221 kilometer depth, magma eruption is only aborted when the edifice radius is  
222 greater than 6 km.

223 For storage zones located at a few kilometers depth beneath the volcanic  
224 edifice, it is possible to include compressibility effects (see section 5), which  
225 can be important when volatiles are present. The inclusion of compressibility  
226 effects mainly acts to reduce the magma pressure decrease within the storage  
227 zone following the unloading event, as shown in figure 3. It follows that  
228 the volume of erupted magma will thus be larger than in the case of an  
229 incompressible magma. Larger edifice size is required to counteract this, and  
230 reduce the amount of erupted magma.

### 231 *3.3. 3D effects: Influence of the shape of the load removed*

232 To be more realistic, we also carry out 3D models in order to simulate  
233 asymmetric flank collapses and mimic the resulting horseshoe- shaped craters

234 of the final edifice geometry that are observed at many strato-volcanoes, such  
235 as Mount St Helens (Cascades, USA), Bezymianni and Shiveluch (Kamtchatka,  
236 Russia) or Galunggung (Indonesia). The volume of the landslide remains  
237 fixed at 20 % of the initial volcanic edifice, but the collapse now occurs only  
238 at one side of the volcano (figure 9 C). The unloading associated with the ed-  
239 ifice collapse is thus asymmetric, with potential consequences for failure con-  
240 ditions at the reservoir wall. The removed part of the edifice is re-distributed  
241 as a thin deposit layer around the volcano. This layer has a constant thick-  
242 ness from the base of the edifice to a distance of ten times the edifice radius,  
243 and is only emplaced in a sector of  $\pm 30^\circ$  from the collapse flank. The runout  
244 and sector angle are based on field observations, as shown in figure 4b.

245 For comparison, elastic parameters are the same as in the previous ax-  
246 isymmetric models. We run models for the same reservoir geometry: a sphere  
247 with a 3 km radius, with a top situated at two different depths (1 and 3 km).  
248 Three different values for the initial edifice radius (1, 3 and 6 km) are con-  
249 sidered. Table 1 gives the comparison between symmetric and asymmetric  
250 collapses. Here the symmetric case is recalculated so that the removed part  
251 of the edifice is re-distributed as a deposit layer of constant thickness from  
252 the edifice to a distance equal to 5 times the edifice radius. This distance  
253 is chosen in order to obtain the same order of magnitude for the deposit  
254 thickness as for the asymmetric case.

255 Values of the ratio of erupted volume after collapse to erupted volume  
256 without collapse are almost the same for both collapse geometries. The  
257 erupted volume after collapse only remains the same as that without any  
258 edifice destruction for small edifices (case 1, for an edifice of 1 km radius).

259 However, this erupted volume decreases when the initial edifice size increases  
260 (case 2, for an edifice of 3 km radius and a chamber top at 1 km depth, or  
261 for an edifice of 6 km radius and a chamber top at 3 km depth). For large  
262 edifices and shallow storage zones, no magma is erupted, since eruption is  
263 aborted by the collapse event (case 3 for an edifice of 6 km radius and a  
264 chamber depth of 1 km).

265 The closure pressure tends to decrease for asymmetric failure compared  
266 to symmetric failure. This effect produces a small increase in erupted vol-  
267 ume compared to the symmetrical model for case 2 (Table 1). Calculated  
268 differences are minimal between the two failure geometries, with a difference  
269 in total erupted volume of only a few percent. In the same way, our results  
270 for erupted volume do not change significantly when taking into account  
271 the mass load redistribution due to runout deposits at the periphery of the  
272 volcano.

273 From these calculations, it appears that the results are not significantly  
274 affected by the collapse geometry; thus the less time-consuming axisymmetric  
275 models, can be used to perform parametric studies.

#### 276 *3.4. Phreatic eruptions*

277 Several large volcanic failure events have not resulted in emission of juve-  
278 nile material, but instead lead to a phreatic eruption. Such phreatic events  
279 were defined by Siebert et al. (1987) as Bandai-type eruptions after the  
280 name of the Bandai-san volcano in northeast Japan, which produced large  
281 phreatic explosions associated with a major debris avalanche in 1888 (Ya-  
282 mamoto et al., 1999). At Bandai-san, the volcanic failure was triggered by  
283 an earthquake and there is no direct evidence of magma involvement in this

284 catastrophic event. As no juvenile products were erupted, the straightfor-  
285 ward conclusion is that no magma was present at shallow level. However,  
286 pressurized fluids had been stored at a shallow depth, and a heat source is  
287 necessary to explain the vigorous hydrothermal system. Our results provide  
288 an alternative framework to interpret such phreatic events, in which magma  
289 might be trapped in reservoirs beneath the edifice. We show that for shallow  
290 reservoirs, even if magma was present, a large edifice collapse could poten-  
291 tially abort the incipient eruption.

#### 292 **4. Effect of edifice destabilization on the long-term eruptive history**

293 The three main modifications documented with regard to the erupted  
294 magma after a major flank collapse are: an increase in eruption rate (Beget  
295 and Kienle, 1992; Siebert et al., 1995; Boulesteix et al., 2012), a change in  
296 erupted magma composition towards less evolved and denser magmas (Man-  
297 conic et al., 2009; Longpré et al., 2010; Boulesteix et al., 2012), and a change  
298 in magma storage pressure (Rutherford and Devine, 2008).

299 As shown in section 2, an edifice flank collapse always causes a pressure  
300 decrease within underlying storage zones. Where the shallow reservoir is  
301 still connected to a deeper source of magma, this reservoir depressurisation  
302 should induce a rapid replenishment as observed after a reservoir withdrawal  
303 due to an eruptive event (Lu et al., 2010). The deeper source is expected to  
304 contain more primitive magmas, and the replenishment should thus increase  
305 the amount of less differentiated magmas within the shallow storage zone. A  
306 reduced edifice size also allows the eruption of denser products which would  
307 otherwise have stalled at shallow depth (Pinel and Jaupart, 2000).

308 The evolution of the threshold pressure required for dyke initiation, which  
309 corresponds to the maximum pressure within the reservoir, should have an  
310 impact over a longer time scale. At Mount St Helens, the pressure evolution  
311 of the storage zone can be seen through petrological studies of rocks span-  
312 ning the last thousand years (Gardner et al., 1995; Rutherford and Devine,  
313 2008). The observed pressure range is between 130 MPa and 300 MPa, with  
314 several episodes of storage pressure reduction occurring over periods of a few  
315 years. The most recent episode of storage pressure decrease corresponds to  
316 the renewed activity in 2004-2006, with a pressure decrease in the order of  
317 50 MPa compared to the magma erupted in 1980. Such episodes of pressure  
318 decrease have been interpreted as being due to a rise of the storage zone  
319 (Gardner et al., 1995). However, it has been shown that each episode of  
320 pressure decrease follows directly on from a large edifice destabilization, such  
321 as the one which occurred in 1980 (Hopson and Melson, 1980; Hausback and  
322 Swanson, 1990). Pinel and Jaupart (2003) and Pinel et al. (2010) propose  
323 that this pressure variation could be explained by a magma pressure decrease  
324 within a fixed storage zone induced by the edifice partial destruction rather  
325 than by upward migration of the storage location. At Mount St Helens, the  
326 petrological data are thus consistent with a decrease of the threshold pres-  
327 sure required for dyke initiation following flank failure. Where the shallow  
328 reservoir is fed by a source of constant pressure at depth, such a decrease  
329 should result in an increase of the eruption rate (Pinel et al., 2010), in good  
330 agreement with the observations.



## 331 5. Discussion

332 We investigate the influence of the reservoir shape by considering only  
333 vertically-elongated ellipsoids. Calculations with horizontally-elongated reser-  
334 voirs (oblate shape) could have been performed, but the edifice collapse  
335 impact is only significant for shallow reservoirs, and it has been shown  
336 previously that shallow, oblate reservoirs strongly favour caldera formation  
337 (Roche, 2000; Geyer et al., 2006). Caldera collapse formation is a complex  
338 and specific phenomenon, already studied elsewhere (Pinel, 2011), and be-  
339 yond the scope of this paper in relation to major flank collapse. Thus we  
340 choose to ignore oblate-shaped chambers here.

341 In this study we assume that rocks encasing the reservoir behave elasti-  
342 cally. Volcano flank collapses are often sudden events. For instance, detach-  
343 ment of the northern flank of Mount St Helens, USA, in May 1980, occurred  
344 in a few seconds, as testified by eye-witnesses. Besides, we show that an un-  
345 loading event significantly affects the stress field only at depths of less than  
346 three times the radius of the removed load (figure 1). At the time scale of  
347 an eruptive cycle, geodetic measurements recorded during replenishment or  
348 eruptive events (Sturkell et al., 2006; Lu et al., 2010; Lu and Dzurisin, 2010)  
349 prove that the elastic assumption is valid, at least for shallow reservoirs.  
350 This geophysical observation justifies our elastic assumption when looking  
351 at the impact of a major flank collapse on the subsequent eruptive event for  
352 continental volcanoes, for which the lateral extension of the edifice remains  
353 small compared to the elastic crustal thickness. Estimations of the Maxwell  
354 relaxation time for upper crustal rocks in volcanic areas are around 30-80  
355 ky (Jellinek et al., 2004), and the usual duration for cone-building episodes

356 is less than 100 ky for continental volcanoes (Davidson and DeSilva, 2000).  
357 Based on this consideration, our elastic model can still be used to discuss  
358 the influence of a major flank collapse on the long-term eruptive history of a  
359 given continental volcanic system. However, the elastic assumption becomes  
360 less valid when looking at the effect of destabilization on large oceanic volca-  
361 noes. The largest debris avalanche deposits (reaching 5000 km<sup>3</sup>) are observed  
362 around large ocean-island volcanoes (Mc Guire, 1996). The lateral extension  
363 of the removed load is then close to, or even larger than, the elastic crustal  
364 thickness. For instance, the El Golfo landslide, which affected El Hierro is-  
365 land in the Canary Islands, had an inferred lateral extension of more than  
366 10 km, for an elastic crustal thickness in this area of around 20 km, based on  
367 seismic and gravity data (Watts et al., 1997). To investigate the impact of  
368 these large-scale flank collapses on the magmatic plumbing system of oceanic  
369 volcanoes, it would be necessary to take the viscous response of the mantle  
370 into consideration, too, as proposed by Sigmundsson et al. (2012) in their  
371 study of the influence of long-term ice retreat on magma storage zones.

372 The model developed in this paper deals with a magma storage zone be-  
373 neath the volcanic edifice such that it cannot account for decompression of  
374 the magma emplaced at shallow depth within the volcanic edifice. This effect  
375 should be taken into account to fully describe what occurred during the May  
376 1980 eruption of Mount St Helens, where the volcano collapse was followed  
377 by a 1 km<sup>3</sup> Plinian ash eruption (Bradley and Myers, 2000). However the ef-  
378 fect of magma compressibility within the chamber can also be included when  
379 buoyancy forces induced by magma density variations remain small relative  
380 to magma pressure changes (see Pinel and Jaupart (2005) for a complete dis-

381 cussion). This condition is verified for magma with a few percent of volatiles  
382 stored at depths greater than a few kilometers.

## 383 **6. Conclusion**

384 Numerical simulations using the elasticity equations help to constrain the  
385 potential impact of a major volcano flank collapse on the ongoing eruption,  
386 as well as the longer term eruptive history, of continental volcanoes. Develop-  
387 ment of models taking into account the viscous response of the mantle would  
388 be necessary in order to model more precisely the potential consequences of  
389 the larger flank collapses affecting oceanic volcanoes.

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Table 1: Comparison between results obtained with a truncated edifice (2D axisymmetric calculation, figure 9 B) and with a sector collapse (3D calculation, figure 9 C). Mass is redistributed at the volcano base: all around the volcano for the axisymmetric case, and in a wedge-shaped sector for the asymmetric flank collapse. Calculations are performed for a spherical reservoir with a 3 km radius, and a top at 1 km or 3 km depth, for three different initial edifice radii: 1, 3 and 6 km. The same experiments are represented by crosses in figure 7.

Figure 1: Stress reduction under the centre of a conical load (2 km radius, 1 km height, density of 2800 kg/m<sup>3</sup>) which is removed from the surface. Calculation is for an elastic half-space with Poisson’s ratio equal to 0.25. The dotted line is for the vertical component of the stress tensor,  $\sigma_{zz}$ , the dashed line is for the horizontal components,  $\sigma_{rr} = \sigma_{\theta\theta}$ , and the solid line is for the pressure reduction,  $P = (1/3)(\sigma_{rr} + \sigma_{\theta\theta} + \sigma_{zz})$ .

Figure 2: Pressure decrease within and around a magmatic reservoir, under the centre of a conical load (2 km radius, 1 km height, density of 2800 kg/m<sup>3</sup>) which is removed from the surface. The chamber resides in an otherwise elastic homogeneous half-space with Young’s modulus  $E = 30$  GPa and Poisson’s ratio  $\nu = 0.25$ . In the absence of a magma chamber (black dashed curve), the pressure profile obtained is equivalent to the one given by equation 3 (solid line in figure 1). Other lines are for pressure profiles where there is a magma reservoir, with varying values of the magma bulk modulus (blue line, incompressible magma; purple line,  $K = 20$  GPa; red line,  $K = 10$  GPa; orange line,  $K = 1$  GPa; yellow line,  $K=0$  GPa). a) The chamber is a sphere (radius 1 km, depth to chamber top 1 km). b) The chamber is a prolate ellipsoid (half-height 1 km, half-width 0.25 km, depth to chamber top 1 km).

Figure 3: Pressure reduction within the magma chamber ( $\Delta P(K)$ ) induced by the removal of a surface conical load (radius 2km, height 1km, density 2800 kg/m<sup>3</sup>) as a function of the bulk modulus, K, of the magma. Crustal Young's modulus and Poisson's ratio are equal to 30 GPa and 0.25, respectively. The pressure change is normalized by the pressure change in incompressible magma ( $\Delta P_\infty$ ). The shaded area shows the range of values characteristic of dry magmas (K between 1 and 20 GPa; Tait et al. (1989)). The solid curve is for the spherical reservoir and the dashed curve for the prolate one.

Figure 4: a) Volume of major collapse versus edifice volume. Failure volumes are taken from Table 1 in Mc Guire (1996), except for Parinacota volcano (Hora et al., 2007). Edifice volumes before collapse are estimated using the present topography of the volcano. When no collapse scar is visible in the topography, corresponding to old events, the volume before collapse is taken to be equal to the present volume; when a large scar is visible, the volume before collapse is obtained by adding the present volume to the failure volume. The total current volume of each volcano, at present time, has been calculated by the numerical integration of SRTM (3" arc) elevation. For volume calculation, we assume that edifice extension stops when the slope becomes small ( $< 10^\circ$ ), and we subtract the mean elevation of the basement to deduce volcano height. The ratio between the volcano collapse volume and the total edifice volume is between 10 % and 30 % (dashed lines). For all collapse models, we use the mean value of 20 % (solid line) for this ratio. b) Runout distances of debris avalanches versus volcano radii. Runout distances are taken from Table 1 in Mc Guire (1996). Most volcanoes produce avalanches which travel to distances of around 6 times the radius of the edifice (lower dashed line). However, in certain conditions, such as Colima volcano, the deposits can travel up to 12 times the radius of the edifice (upper dashed line). A value of 10 is used in the 3D asymmetric modelling (solid line).

Figure 5: Pressure reduction within the magma reservoir induced by the removal of the upper 20% of a conical edifice with a slope of 30 degrees. Results are presented as a function of the reservoir and edifice radius. Calculations are performed for a spherical reservoir with a top at 1km depth, and filled with incompressible magma. Crustal Poisson's ratio is equal to 0.25.

Figure 6: a) Evolution of magma pressure,  $P$ , and threshold pressure for failure,  $P_r$ , induced by volcanic collapse. (b) Evolution of the erupted volume of magma following edifice destabilization. For a given reservoir geometry, the erupted volume is proportional to the pressure difference,  $\Delta P_e = P - P_{cl}$ , where  $P$  is the magma pressure when the eruption starts and  $P_{cl}$  is the threshold pressure value under which dykes close at the chamber wall, ending the eruptive event. Before edifice destabilization, the system is about to erupt, such that this difference is given by  $\Delta P_e(i) = P_r(i) - P_{cl}(i)$ , with  $P_r$  the threshold pressure for dyke initiation at the chamber wall. After the major flank collapse, magma pressure within the chamber decreases by an amount  $\Delta P$  to the value  $P(f)$ , and the pressure difference becomes  $\Delta P_e(f) = P(f) - P_{cl}(f)$ . Three different cases can occur: case 1, when  $\Delta P_e(f)$  is greater than  $\Delta P_e(i)$ , where the volume of magma erupted is larger than it would have been with no edifice collapse; case 2, when  $\Delta P_e(f)$  is less than  $\Delta P_e(i)$ , such that the volume of magma erupted is smaller than it would have been with no edifice collapse and, case 3, when  $\Delta P_e(f)$  becomes negative, which means that the incipient eruption is aborted and there is no magma erupted.

Figure 7: Evolution of the erupted volume of magma following the removal of the upper 20 % of a conical edifice with a slope of 30 degrees. Results are presented as a function of the reservoir and edifice radius. Calculations are performed for a spherical reservoir filled with incompressible magma. Crustal Poisson's ratio is equal to 0.25. Three different values for depth to the top of the magma reservoir are considered: a) 0.5 km depth, b) 1 km depth, c) 3 km. The dashed lines approximately define the limits between the various cases defined in figure 6. Crosses are for calculations also performed in 3D with an asymmetric flank collapse.

Figure 8: Evolution of the erupted volume of magma following the removal of the upper 20 % of a conical edifice with a slope of 30 degrees. Results are presented as a function of the reservoir vertical semi-axis and edifice radius. Calculations are performed for a prolate reservoir with a top at 1 km depth, filled with incompressible magma. Crustal Poisson's ratio is equal to 0.25. After the edifice partial collapse, the volume of magma erupted is always smaller than it would have been in absence of edifice collapse (case 2 of figure 6).

Figure 9: Edifice geometries considered. A) The initial edifice shape before the sector collapse. B) The truncated final edifice shape when the upper 20 % of the original edifice has been removed. C) The horseshoe-shaped crater of the final edifice when the 20 % of the original edifice has been removed by a sector collapse. The edifice represented has a radius of 1 km and a slope of 30 degrees. The scale bar shows edifice height in meters.

Magma chamber depth (km)	Edifice radius (km)	Erupted Vol/ Erupted Vol without collapse		Case
		<i>Symmetric collapse</i>	<i>Asymmetric collapse</i>	
1	1	1	1	1
1	3	0.59	0.61	2
1	6	0	0	3
3	1	1	1	1
3	3	0.92	0.92	2
3	6	0.58	0.63	2

**Table 1**  
[Click here to download Table1 Table.pdf](#)



Figure1  
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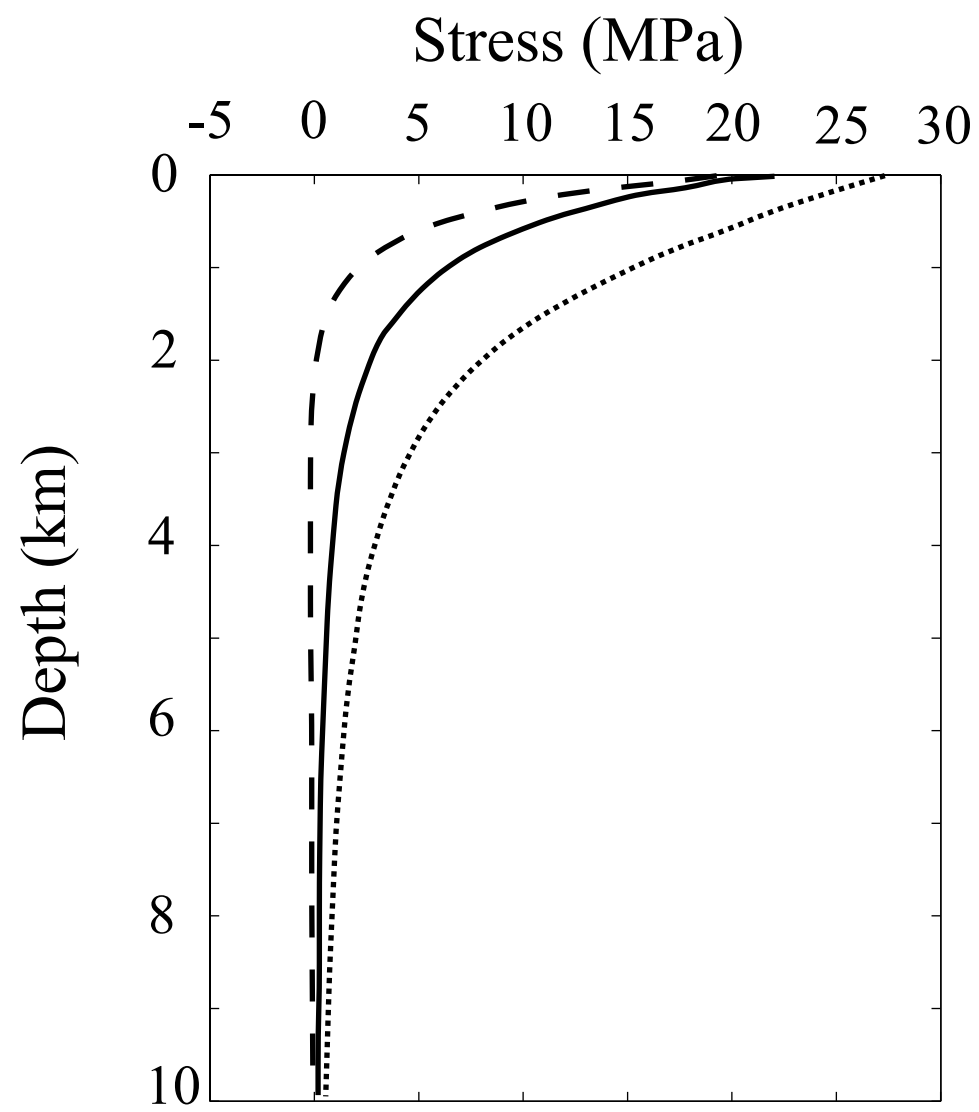


Figure2

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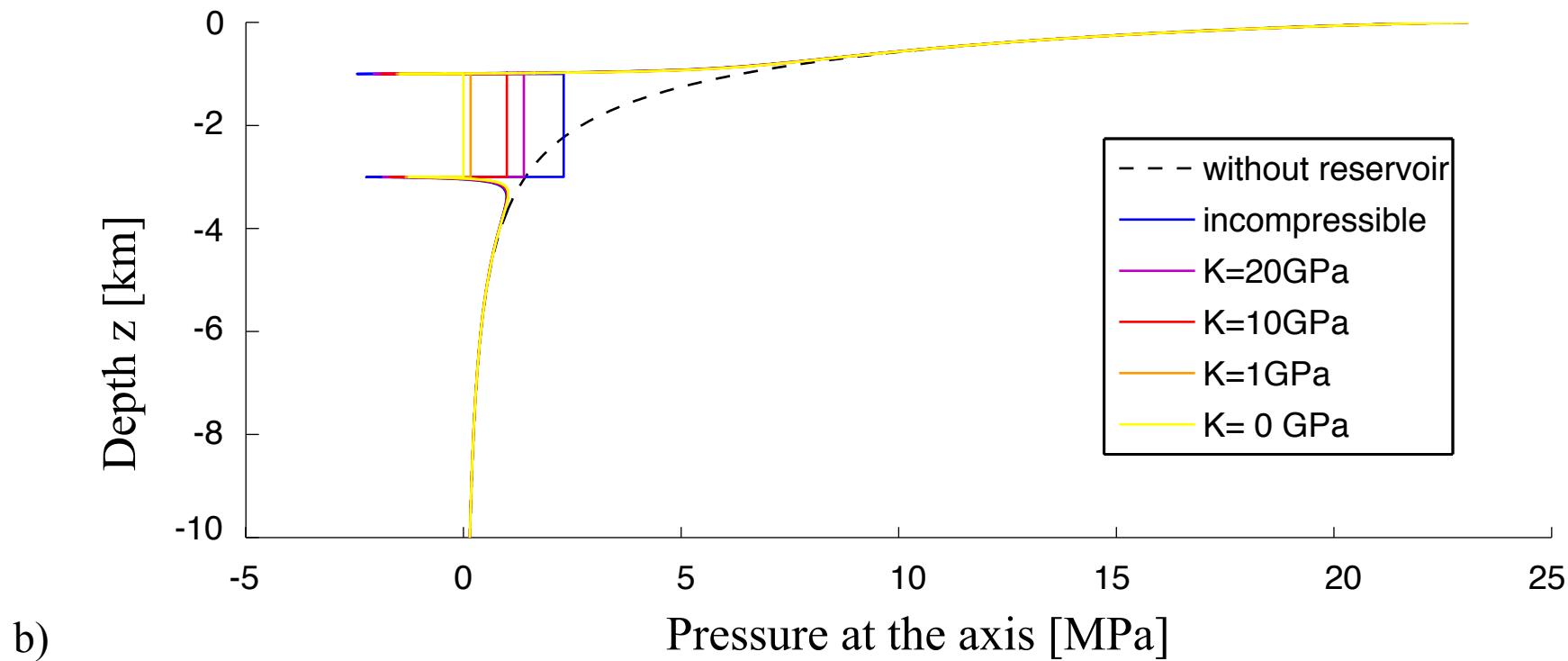
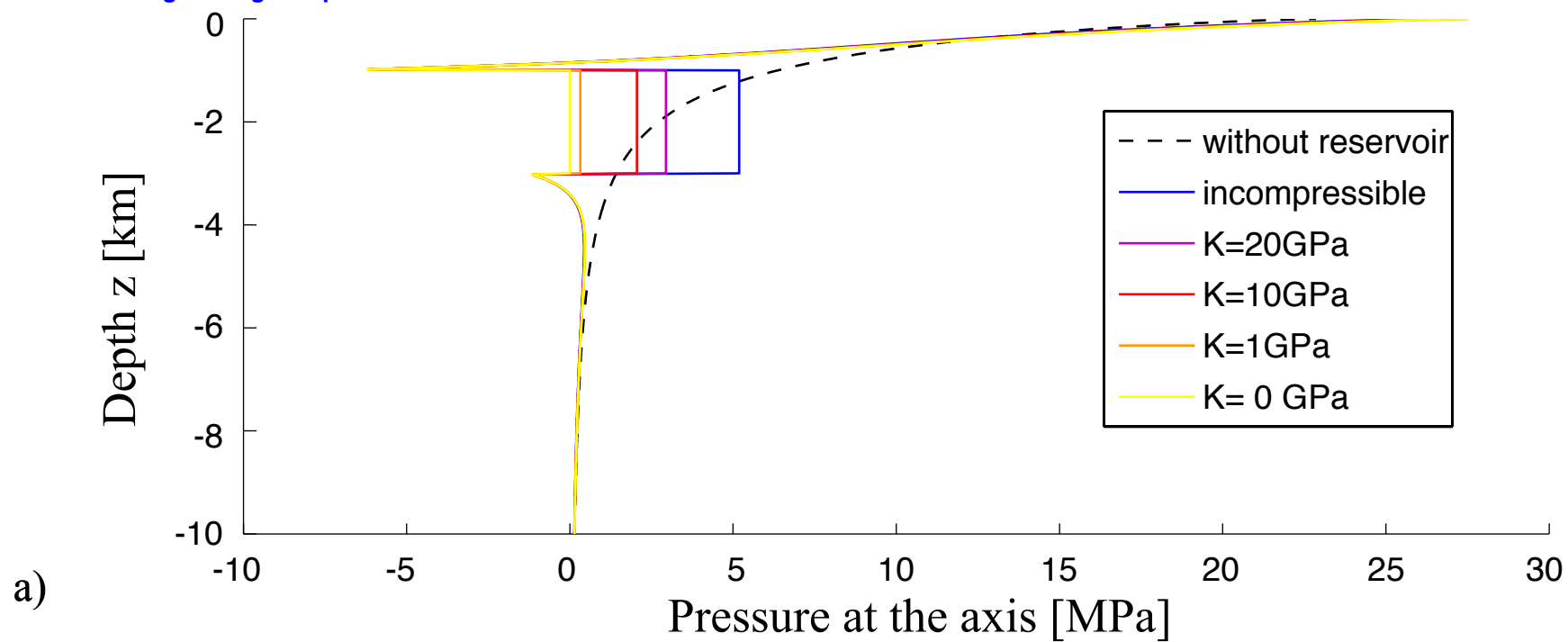
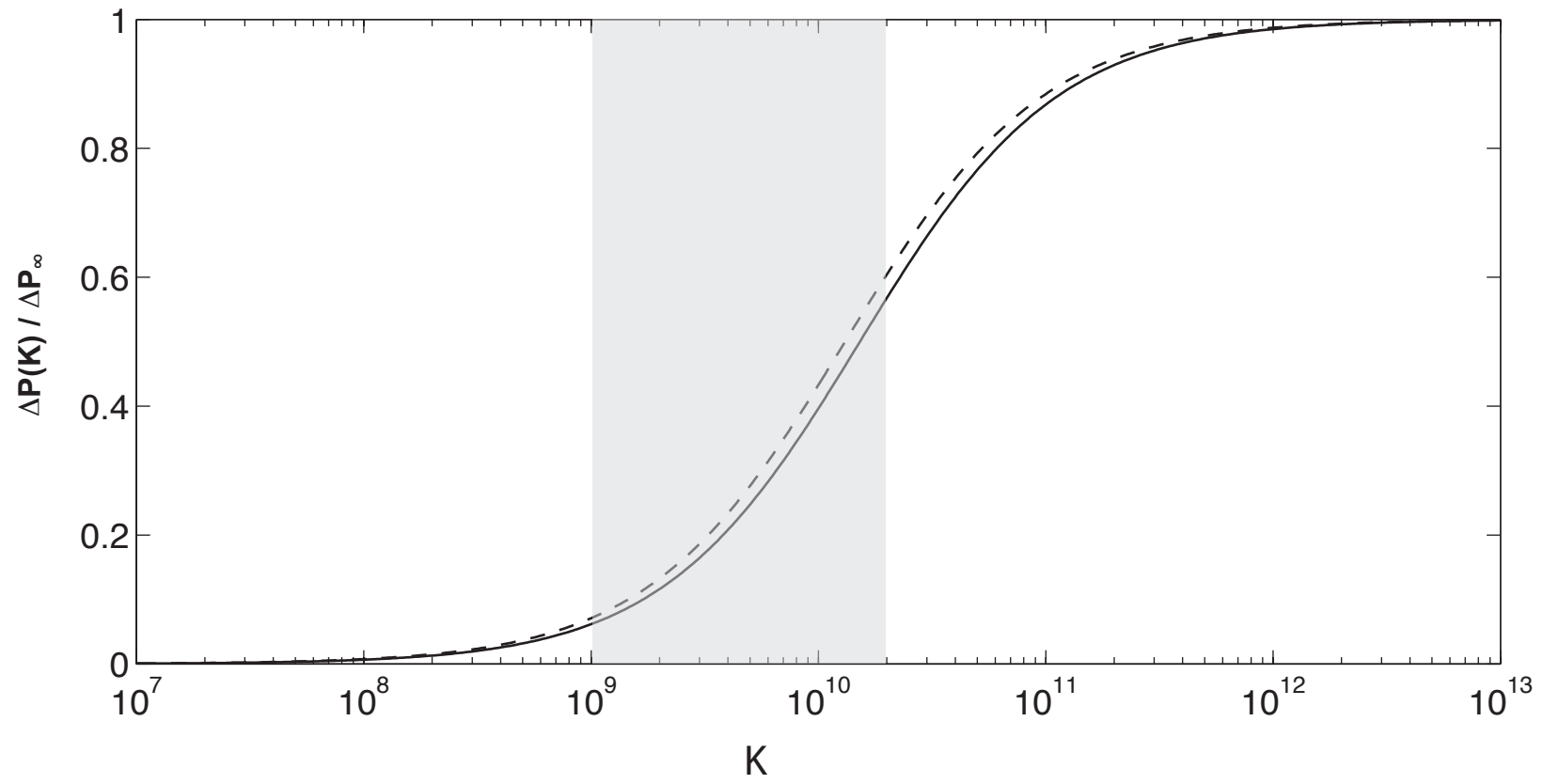


Figure3  
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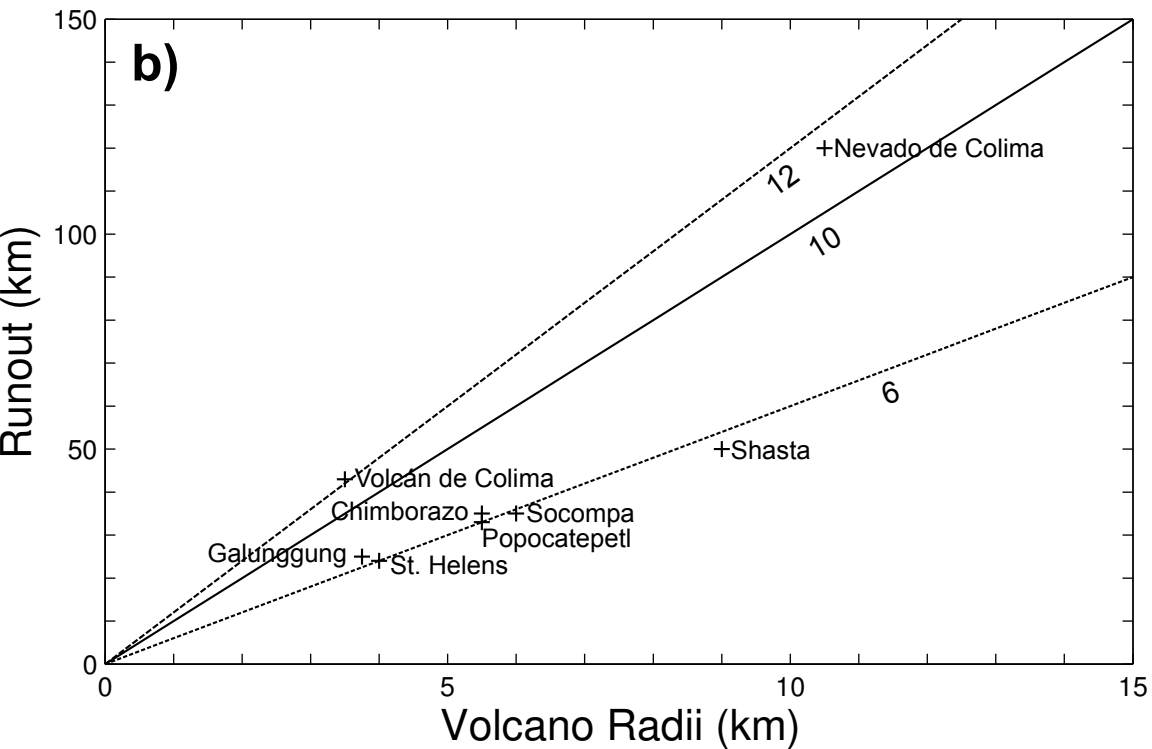
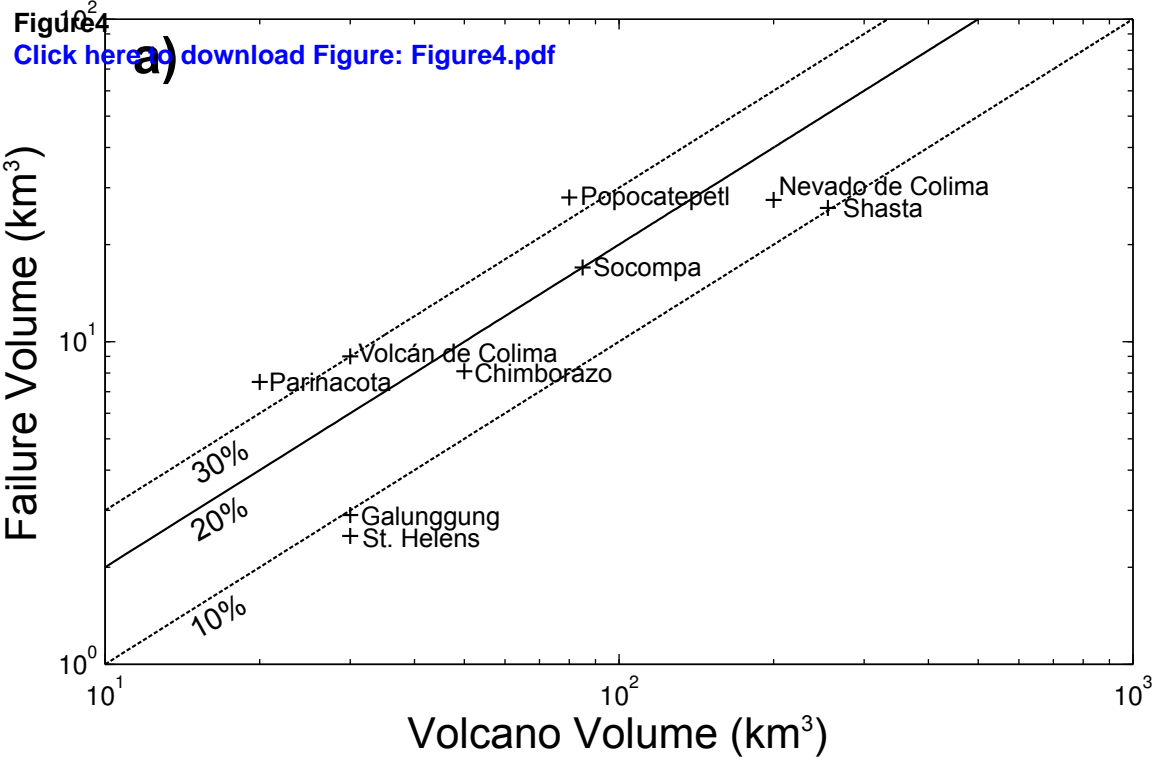
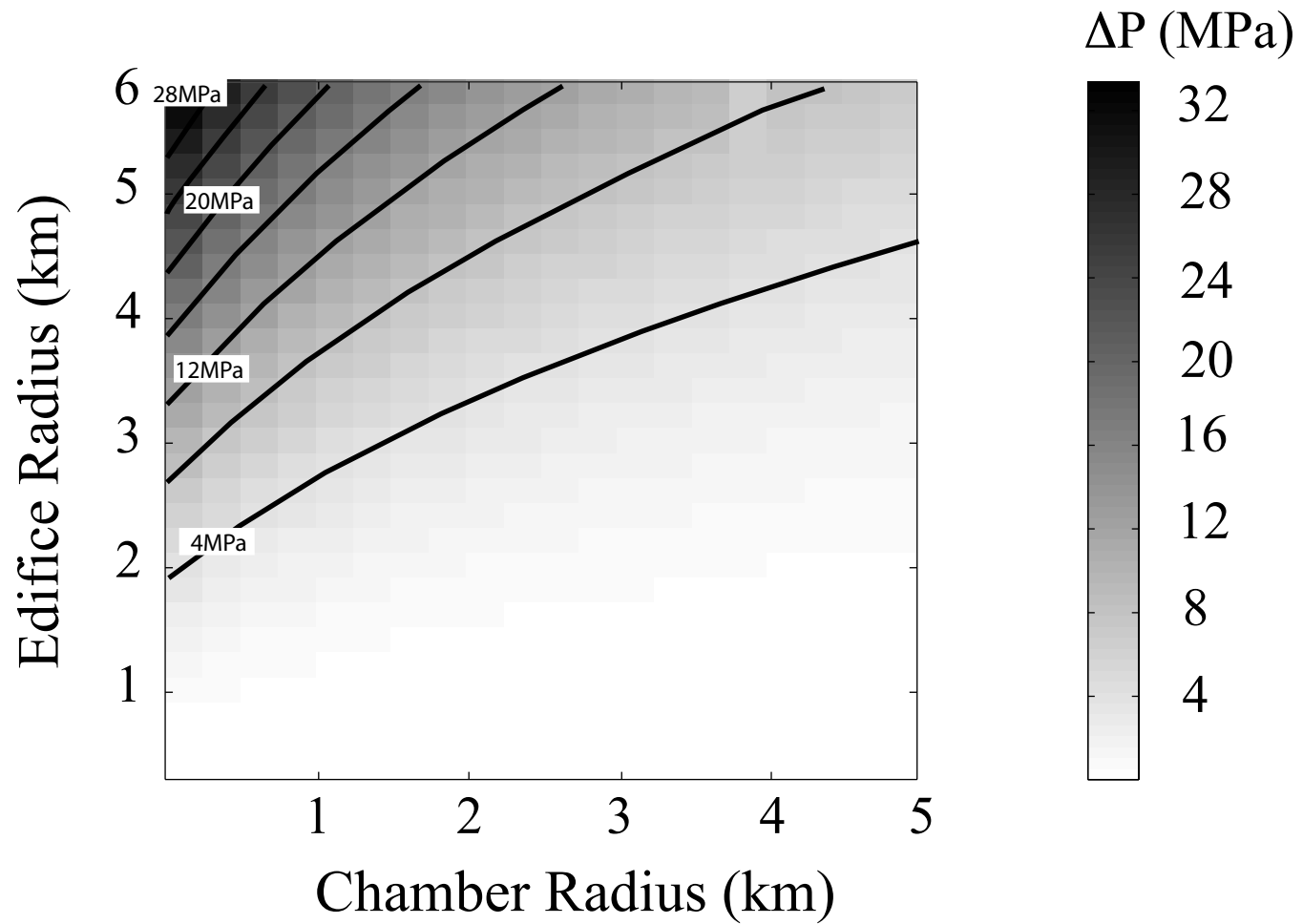


Figure5

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**Figure6**

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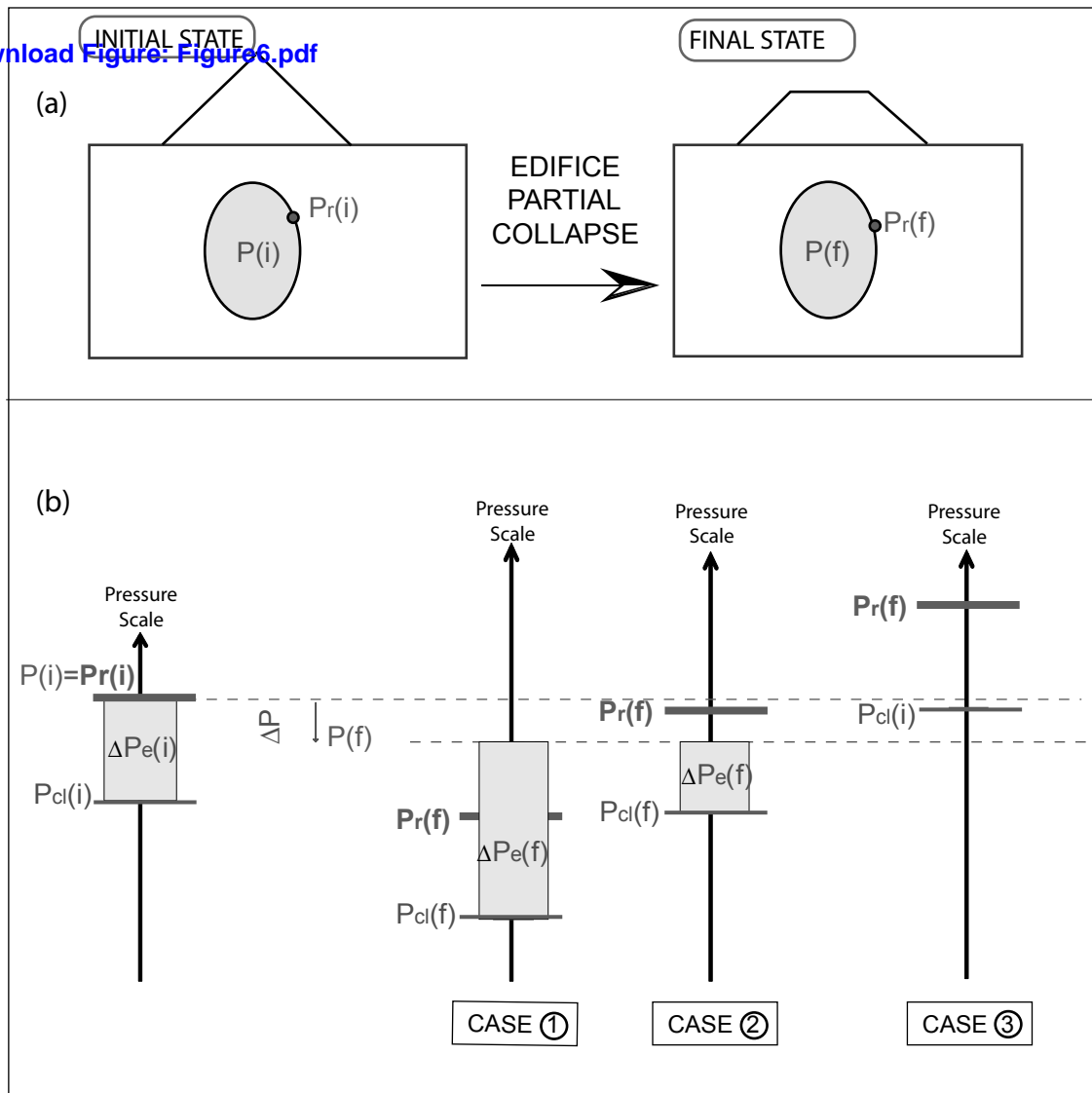
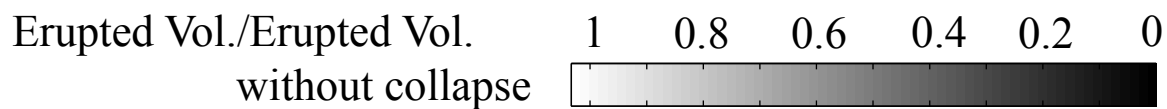


Figure7

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Chamber top at 0.5 km depth

Chamber top at 1 km depth

Chamber top at 3 km depth

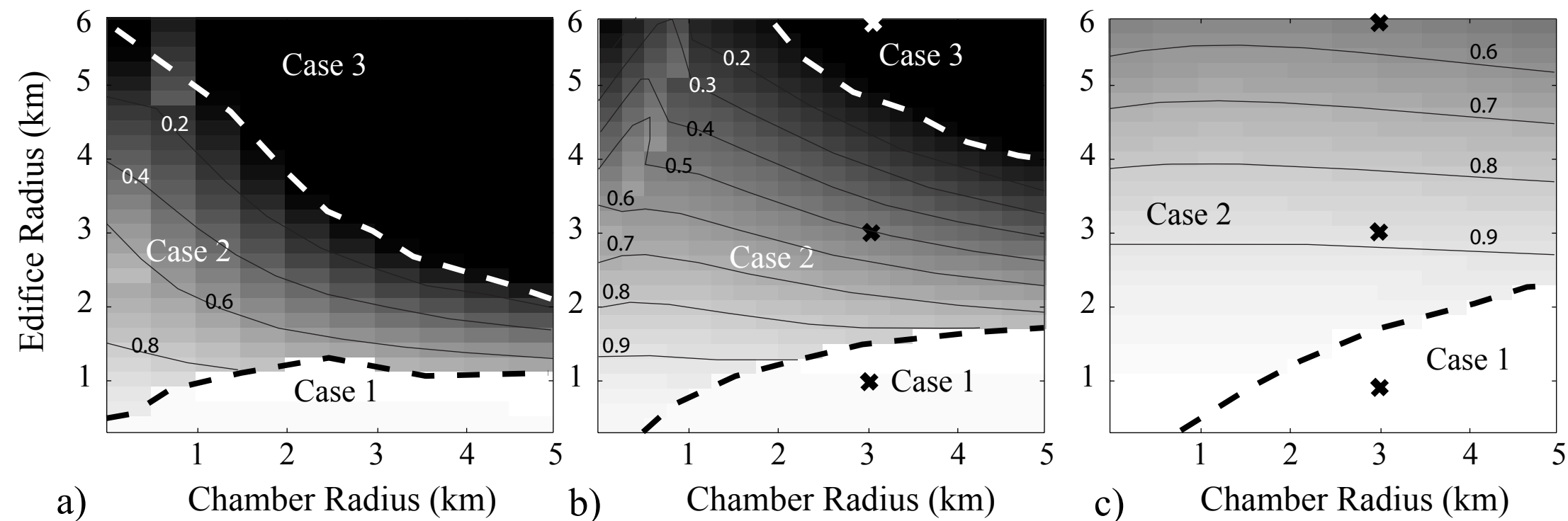
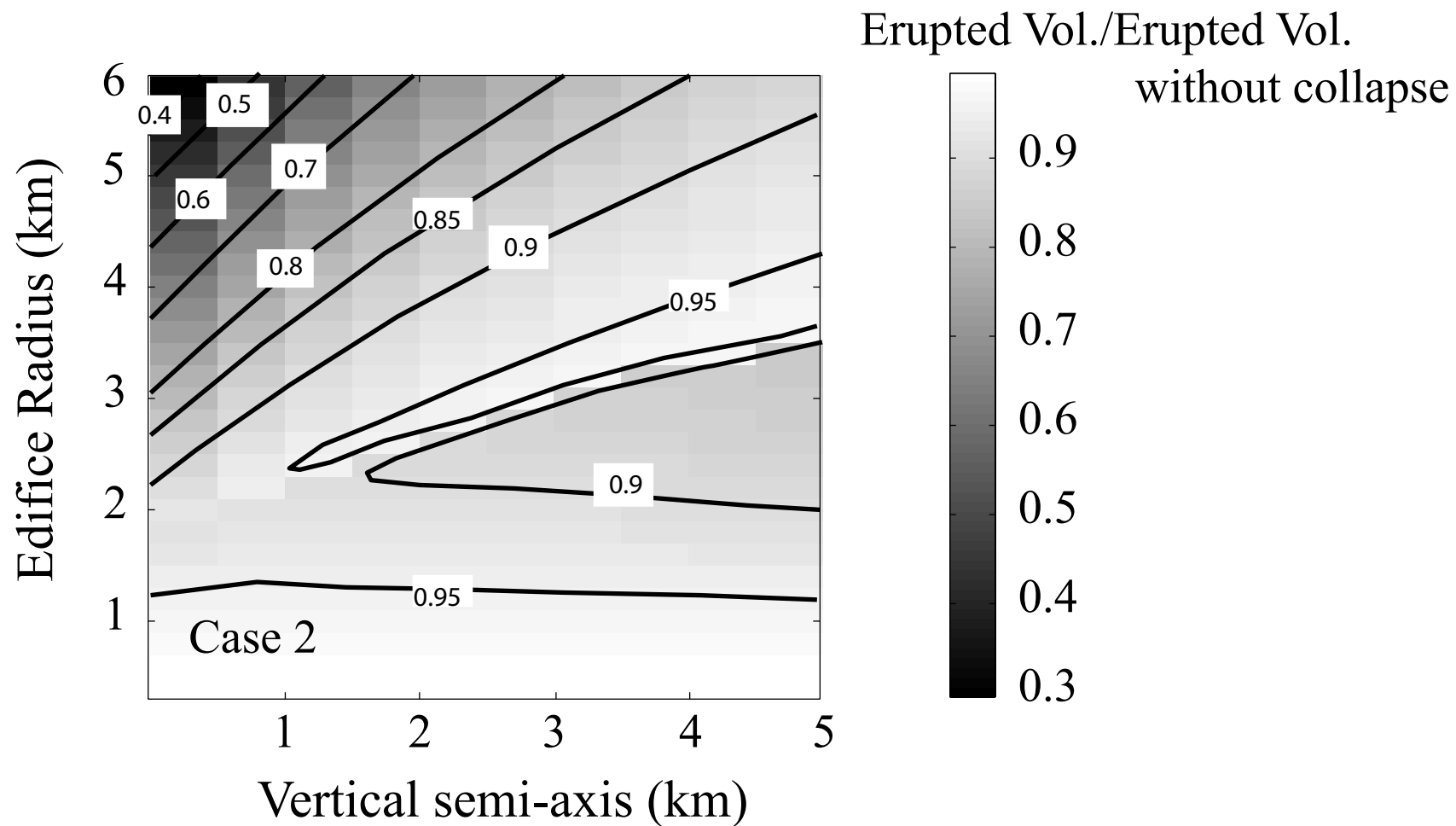


Figure8

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**Figure9**

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