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# The hydrothermal alteration of cooling lava domes

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23 **ABSTRACT**

24 Hydrothermal alteration is a recognized cause of volcanic instability and edifice collapse,  
25 including that of lava domes or dome complexes. Alteration by percolating fluids transforms  
26 primary minerals in dome lavas to weaker secondary products such as clay minerals; moreover,  
27 secondary mineral precipitation can affect the porosity and permeability of dome lithologies. The  
28 location and intensity of alteration in a dome depend heavily on fluid pathways and availability  
29 in conjunction with heat supply. Here we investigate post-emplacment lava dome weakening by  
30 hydrothermal alteration using a finite element numerical model of water migration in simplified  
31 dome geometries. This is combined with the Rock Alteration Index (RAI) to predict zones of  
32 alteration and secondary mineral precipitation. Our results show that alteration potential is  
33 highest at the interface between the hot core of a lava dome and its clastic talus carapace. The  
34 longest-lived alteration-potential fields occur in domes with persistent heat sources and  
35 permeabilities that allow sufficient infiltration of water for alteration processes, but not so much  
36 that domes cool quickly. This leads us to conclude that alteration-induced collapses are most  
37 likely to be shallow-seated and originate in the talus or talus/core interface in domes which have  
38 a sustained supply of magmatic heat. Mineral precipitation at these zones of permeability  
39 contrast could create barriers to fluid flow, potentially causing gas pressurization which might  
40 promote deeper-seated and larger-volume collapses. This study contributes to our knowledge of  
41 how hydrothermal alteration can affect lava domes and provides constraints on potential sites for  
42 alteration-related collapses, which can be used to target hazard monitoring.

43

## 44 INTRODUCTION

45 Hydrothermal alteration of volcanic edifices promotes weakening and instability,  
46 increases the propensity for collapse and can lead to significant volcanic hazards (Voight et al.  
47 2002; Reid et al. 2002b; McGuire 2003; Carrasco-Núñez et al. 2006; John et al. 2008; del Potro  
48 and Hürlimann 2009). The extent of alteration, and magnitude of any resultant collapses ranges  
49 from those that involve parts of individual lava domes through to major volcanic sector collapses  
50 (Siebert 2002). Here we focus on modeling hydrothermal alteration at the scale of an individual  
51 lava dome. Our motivation is that we know that hydrothermal systems in and around lava domes  
52 are pervasive, hazardous and poorly understood. Lava dome internal structure can also be  
53 simplified for modeling in a reasonably valid way, which is significantly more difficult for  
54 larger, heterogeneous edifices.

55 Direct evidence of important hydrothermal systems associated with lava domes include  
56 (i) *soufrière* systems, which are commonplace around the bases of domes (Boudon et al. 1998;  
57 Walker et al. 2006; Bedrosian et al. 2007; Aizawa et al. 2009) and (ii) mass-flow deposits  
58 (including debris-avalanches and debris-flow deposits) sourced from old domes, or dome  
59 complexes which contain a high proportion of altered material and clay-rich matrix (Opfergelt et  
60 al. 2006; Devoli et al. 2009). In fact much of what we know about alteration in domes comes  
61 from the study of these mass-flow deposits, where the hydrothermally altered components are no  
62 longer *in situ*, and their original position in the edifice can only be inferred indirectly from  
63 stratigraphy and flow reconstructions.. Lava domes sit as variably permeable caps often directly  
64 above the volcanic conduits from which they were extruded. Such conduits may be the source of  
65 thermal and/or volatile fluxes for extended periods of time after an eruption has ceased  
66 (Bedrosian et al. 2007; Salaün et al. 2011). This configuration may render lava domes more

67 susceptible to post-emplacement alteration than other more widely dispersed erupted units.  
68 Indeed, a conduit-capping dome can host a hydrothermal system that would not otherwise exist  
69 had an eruptive phase resulted in a more exposed upper conduit.

70 Lavas within domes can be altered to secondary mineral assemblages (e.g. smectite clays,  
71 kaolinite and alunite), weakening the edifice, reducing slope stability, and ultimately resulting in  
72 slope failure (Boudon et al. 1998; Opfergelt et al. 2006). Clay-rich alteration materials are not  
73 only low-strength but have the potential to absorb and channel groundwater, locally increasing  
74 fluid pore pressure and promoting the expansion and/or formation of low-strength zones and  
75 exacerbating the risk of slope failure (Voight and Elsworth 1997). Secondary mineral formation  
76 (including hydrous silica) may also contribute to internal pressurization of lava domes by  
77 reducing gas permeability, thus provoking collapse through explosive decompression (Voight  
78 and Elsworth 2000). Upon collapse, clay-hosted pore water can lubricate mass-flows, resulting in  
79 the generation of more-mobile and cohesive debris-flows than would be generated by collapse of  
80 dry material (Boudon et al. 1998; Reid et al. 2002a; Opfergelt et al. 2006; John et al. 2008).

81 Two classic examples of edifice collapse involving hydrothermally altered lava  
82 domes/dome complexes are the 1998 event at Casita in Nicaragua (Scott et al. 2005; Opfergelt et  
83 al. 2006; Devoli et al. 2009) and the 1997 debris-avalanche at Soufriere Hills, Montserrat  
84 (Sparks et al. 2002; Voight et al. 2002). In both cases low-strength, low-permeability alteration  
85 products are thought to have hosted water, which reduced the effective stresses in, and shear  
86 strength of, the rocks and which ultimately led to catastrophic destabilization of the edifices. At  
87 Casita, a *ca.* 8 ka dacite lava dome complex, a 1.6 million m<sup>3</sup> collapse on 30 October 1998 was  
88 triggered by intense rainfall associated with Hurricane Mitch. The collapse generated a debris-  
89 flow and lahar that resulted in more than 2500 fatalities (Sheridan et al. 1999; Kerle 2002). The

90 collapsed material was rich in smectite clays formed by intense hydrothermal alteration of the  
91 original dacites (Opfergelt et al. 2006). A recent re-analysis of the collapse source area and  
92 deposits suggests that all failure surfaces formed at or near an interface between units of  
93 overlying volcanic breccia and underlying units of altered, clay-rich pyroclastic deposits and  
94 lavas. Clay contents in the altered units were estimated to be 38-50 wt. % of the whole mass and  
95 more than 90% of the fine fraction, with water contents in the remaining undisturbed clay-rich  
96 material ranging from 56-81% (Devoli et al. 2009). For the 26 December 1997 debris-avalanche  
97 at Soufrière Hills Volcano, Montserrat, hydrothermal alteration of the dome-retaining crater wall,  
98 itself made up of an older dome and pyroclastic deposits, (Galway's Mountain/Soufriere, ~113  
99 ka; Harford et al., 2002) was implicated as a major contributor to the destabilization and  
100 subsequent collapse and depressurization of the active lava dome (Sparks et al. 2002; Voight et  
101 al. 2002). The resultant debris-avalanche deposits contained portions of the new dome but also  
102 between 6-15 wt. % kaolinite and smectite group clays, alteration products typical of unsealed  
103 acid-sulfate hydrothermal systems. Intact avalanche blocks showed repeating layers of these  
104 alteration suites and suggested that the collapse slip surface intersected a layered hydrothermal  
105 system. The authors suggested that collapse mechanisms could have included an increase in  
106 pore-fluid pressure in the older dome materials due to the presence of low-permeability clay  
107 layers (Voight et al. 2002).

108         While collapses from a given, actively extruding (fresh), lava dome are common during  
109 an eruptive phase, those that result from post-emplacement dome weakening by hydrothermal  
110 alteration are relatively infrequent, yet may pose very significant hazards that are harder to  
111 anticipate. The work reported here aims to increase our understanding of the collapse potential of  
112 young but inactive domes. Only sparse information exists on the collapse frequency of inactive,

113 altered lava domes, either young or old. But the increasing number of mapped debris-avalanche  
114 deposits (many of which source from lava dome complexes) and the similar post-eruptive  
115 processes at work in both lava domes and larger volcanic edifices suggests that it is important to  
116 understand alteration in these systems. Salaün et al. (2011) and Friant et al. (2006) have mapped  
117 debris-avalanche deposits and potential source areas on the Grande Découverte–Soufrière  
118 volcano and lava dome in Guadeloupe, which indicate that the recurrence interval of such  
119 collapses may be as high as one per 1000 years over the last 8 ky. These collapses have resulted  
120 in debris-avalanche deposits rich in hydrothermally-altered material and Salaün et al. (2011)  
121 suggest that hydrothermal alteration in the domes and flows that erupted after each collapse was  
122 rapid and widespread..

123         Geophysical and geochemical investigations have been used to characterize the  
124 hydrothermal system of specific lava domes at given points in time (Bedrosian et al. 2007; Finn  
125 et al. 2007; Finn and Deszcz-Pan 2011; Brothelande et al. 2014). Such studies are logistically  
126 challenging and cannot provide information about temporal variations of a system unless they  
127 are repeated. At the Mount St. Helens lava dome (Bedrosian et al. 2007), electrical resistivity  
128 surveys revealed that meteoric water circulated in the young dome due to heat input from a near-  
129 surface magmatic source, but did not capture longer-term changes occurring in the hydrothermal  
130 system as the dome subsequently cooled. Aeromagnetic and electromagnetic surveys of Cascade  
131 volcanoes (Finn et al. 2007; Finn and Deszcz-Pan 2011) and the La Soufriere volcano  
132 (Brothelande et al. 2014) have indicated the presence of water and altered material in specific  
133 locations in the edifices at the times of the surveys, but provide limited information about flow  
134 pathways within the hydrothermal systems and how they might be expected to have evolved or  
135 evolve in the future..

136 In this paper we present a physics-based numerical model of heat and fluid flow in a  
137 generic lava dome combined with assessments of alteration potential, to determine where  
138 alteration is most likely to occur within a lava dome. The dynamics of hydrothermal flow and  
139 alteration processes in domes can be highly complex, and in detail each dome is unique; we do  
140 not address all possible complexities but make a first step in quantitative modeling of major  
141 aspects of the systems. While we do not seek to address the specific type of alteration in this  
142 study, it is possible to distinguish likely regions of alteration based on knowledge of temperature  
143 gradients and fluid flux. We use this understanding to make some inferences about the different  
144 collapse styles and source areas that may occur in different hydrothermal settings. This work lays  
145 the groundwork for future investigations to identify likely alteration minerals, in order to  
146 distinguish whether individual collapses are related to weak alteration minerals (such as clays) or  
147 precipitation that reduce the porosity/permeability of the upper dome (such as silica).

148

#### 149 **FINITE ELEMENT HEAT AND MASS TRANSFER CODE (FEHM)**

150 The effect of meteorically-derived water on the hydrothermal system of cooling lava  
151 domes is modeled using The Finite Element Heat and Mass transfer (FEHM) code. FEHM is an  
152 extensively validated (Zyvoloski et al. 1999; Dash et al. 2003; Dash 2003) porous flow simulator  
153 capable of modeling the flow of heat, water, air, and water vapor in a variably saturated porous  
154 and/or fractured medium at temperatures up to 1500°C and fluid pressures of up to 1000 MPa.  
155 The code employs a Newton-Raphson scheme to iteratively solve discretized conservation  
156 equations for mass, energy and momentum for fluid and vapor on a Voronoi-conforming finite-  
157 volume computational mesh (Miller et al. 2007; Zyvoloski 2007). Basic governing equations of  
158 state for conservation of mass and energy and modified Darcy flux are shown in Figure 1; for a



159 detailed derivation, see the supplemental material. The temperature and pressure ranges of  
160 FEHM are ideal for modeling volcanic systems and although it has previously been used to  
161 model cooling pyroclastic deposits (Keating 2005), volcanic seamount discharge and recharge  
162 (Hutnak et al. 2006), and mineral alteration in hydrothermal fault systems (Chaudhuri et al.  
163 2009), it has never been applied to volcanic edifices or lava dome systems.

164 Active emplacement of a dome, whether endogenous or exogenous, is a principle forcing  
165 mechanism for collapse (Calder et al. 2002; Calder et al. 2005);. However in systems where  
166 emplacement has paused or ceased, other forcing mechanisms, including environmental ones,  
167 come into play (Calder et al. 2005; Barclay et al. 2006). Our focus here is on young but  
168 inactive/cooling lava dome systems. The rationale for this choice includes a number of  
169 considerations. First, only in inactive domes is hydrothermal alteration likely to play a significant  
170 role in collapses, and the relevant alteration minerals are most stable at temperatures of 200°C or  
171 less (Ball et al., 2013; ;Giggenbach 1992). Second, in this initial application of FEHM, we  
172 required a relatively simple system where lava extrusion, and its associated heterogeneous mass  
173 and thermal fluxes can be negated. Furthermore, restricting the simulations to lower temperatures  
174 (< 200°C) allows us to neglect high-temperature heat sources, which cause sudden short-term  
175 phase changes that increase model instability and prevent convergence on a solution for the  
176 governing equations (Ingebritsen et al. 2010).

177

## 178 **Modeling methods**

179 Our simulations are run for 100 years on a simplified lava dome geometry consisting of  
180 five material regions: A narrow conduit, a solid dome core, a clastic dome carapace or talus  
181 (covering and also tapering into a wedge away from the dome), a solid substrate, and colluvium

182 overlying the substrate (Figure 2). The conduit radius was selected based on inferences made for  
183 actual domes as well as rheological models of dome eruptions (Fink and Pollard 1983; Costa et  
184 al. 2007). Previous models of dome growth and evolution (Fink and Griffiths 1998a; Hale 2008;  
185 Hale et al. 2009a; Hale et al. 2009b) and structural relationships of existing lava domes (Wadge  
186 et al. 2009) were used to design the configuration of the material regions used in our models.  
187 Two dome geometries are represented: a “crater-confined” dome similar to the domes at the  
188 Santiaguito lava dome complex in Guatemala (Ball et al. 2013), which provided some of the  
189 original motivation for this study, and a “perched” dome, essentially erupting on a sloping  
190 substrate whose core and talus are unconfined by a crater wall, similar to that of the Unzen or  
191 Merapi lava domes (Smithsonian Institution 1991; Anderson et al. 1999; Nakada et al. 1999;  
192 Walter et al. 2013). These dome geometries were investigated to determine if there was a  
193 difference in fluid migration when the dome was confined by a crater wall vs. unconfined.

194 Voronoi-conforming finite-volume computational meshes of these geometries were  
195 generated with the LaGriT Grid Generation Toolset, which was developed at Los Alamos  
196 National Laboratory (Fields et al. 1996; Miller et al. 2007). LaGriT accepts input files consisting  
197 of coordinate data defining material regions, as well as commands choosing the level of  
198 discretization in those regions (or in other subregions specified by the user). The dome meshes  
199 for this study use a two-dimensional axisymmetric coordinate system; each mesh is a slice from  
200 the center of a dome, including a thick substrate that is extended down to several thousand  
201 meters in order to avoid boundary effects. The crater-confined and perched dome models are  
202 triangulated to (i.e. to have a spatial resolution of) 20 meters within the top 1500 meters of the  
203 dome and substrate; the remaining (lower) 1000 meters of the mesh are triangulated to 40 meters

204 to save computational time and allow FEHM to calculate processes in the dome on a more  
205 detailed scale.

206 Material properties (density, porosity, initial permeability, specific heat, and thermal  
207 conductivity) of each dome region were taken from ranges given in the literature, summarized in  
208 Table 1. All values were chosen from lithologies commonly associated with lava domes (dome  
209 rocks, lava flows, and block-and-ash-flow deposits) of andesitic and dacitic composition. In most  
210 of these studies the values were determined from hand sample and drill core analyses. Where  
211 literature values had a wide range, a restricted range was chosen for modeling based on the most  
212 commonly found values. A complete record of the values from which these ranges were defined  
213 may be found in the supplementary material.

214 Although the boundaries of large lava dome structures such as shear lobes are well-  
215 described (Fink and Griffiths 1998b; Watts et al. 2002), there is a limited amount of structural  
216 information available on the geometry of smaller, denser fracture networks in domes or their  
217 carapaces (i.e., the orientation, depth of penetration, concentration of fractures in a given area,  
218 aperture width of the fractures). Therefore, in this study we have chosen to treat material regions  
219 as continua where the properties of the porous and/or fractured medium are averaged to account  
220 for variations that are not captured in our mesh. The appropriateness of a continuum approach  
221 depends on the ratio of the fracture density scale to the flow region scale. Khaleel (1989)  
222 modeled two-dimensional planar laminar flow through the columnar-fractured Columbia River  
223 Basalt Group lava flows and determined that for interconnected networks of filled/unfilled  
224 fractures of uniform aperture and column diameters of 1 m, continuum models were appropriate  
225 for length scales of at least 6 times the column diameter. That author also indicated that a  
226 continuum approach could be appropriate for other fractured rock masses if the entire rock mass

227 is at least 6 times the smallest spacing of the fractures, and preferably ~20-30 times the spacing  
228 in the case of variably sized and distributed fracture networks. The scale of fracturing and  
229 structural features on a lava dome is much smaller than the scale of dome and immediate  
230 substrate (even shear lobes of 10's to a few 100's m can be considered close to the 1/6th cutoff  
231 point), and so we feel that the continuum approach is reasonable for our simplified domes.

232

### 233 **Precipitation/recharge**

234 Because the actual elevation of water tables within volcanic edifices is not well  
235 constrained (Hurwitz 2003), we initiate model runs for these domes by assigning complete  
236 saturation.. Variable saturation is then allowed to develop as the simulation runs. This  
237 assumption is considered reasonable for volcanic systems/domes located in a tropical region that  
238 receive significant (i.e. > 1000 mm/yr) amounts of precipitation, such as the lava dome  
239 volcanoes detailed in Table 3. Previous models of groundwater in volcanoes using similar  
240 recharge and permeability values (Join et al., 2005; Hurwitz et al., 2003) show that water tables  
241 may rise to high levels within an edifice. As such, it is a reasonable assumption that there may be  
242 conditions under which a lower-temperature (cooling) lava dome could become almost entirely  
243 saturated (for example during an intense precipitation event). Even if there is an unsaturated zone  
244 in the immediate rubbly surface of the dome, the models would be unable to represent it if it was  
245 smaller than the 20 m mesh spacing. Given the paucity of information on water tables associated  
246 with lava domes, water-saturation is taken as a reasonable first-order approach.

247 Ground surface recharge in the models (1300 mm/yr, or  $\sim 4.16 \times 10^{-7} \text{ m}^3/[\text{m}^2 \text{ s}]$ ) is  
248 approximated using yearly precipitation rates in tropical regions. Actual infiltration rates into a  
249 dome would be expected to be less than precipitation due to evaporation, vegetation, localized

250 impermeable surfaces, and runoff, but such variations would be site-specific, localized and  
251 impractical to represent in these simulations. Additionally, Rad et al. (2007) state that infiltration  
252 in volcanic settings with exposed fresh lavas and pyroclastic flows may be as high as 80%. Thus  
253 we consider that using a recharge value at the low end of measured precipitation rates for five  
254 existing domes in tropical to temperate climate regions is a reasonable proxy for infiltration  
255 (Table 3).

256

### 257 **Boundary and initial conditions**

258 For simplicity of the simulations, a number of material properties and boundary  
259 conditions were held constant throughout the simulations (Table 3). These include porosity,  
260 density, thermal conductivity, and specific heat of the material region, and pressure, temperature  
261 and recharge along the atmospheric surface (Zone 6, Figure 2). FEHM automatically treats the  
262 boundaries of the modeling mesh as impermeable unless otherwise specified, so left side (Zone  
263 7, Figure 2) and basal (Zone 8, Figure 2) no-flow boundary conditions are assigned in these  
264 simulations. This reflects a radial dome geometry and a dome-topped volcanic edifice resting on  
265 an impermeable base (equivalent to plutonic or metamorphic rock, which can have  
266 permeabilities as low as  $1 \times 10^{-16}$  to  $1 \times 10^{-18} \text{ m}^2$ ; Brace, 1984).

267 Initial conditions for the material regions in all simulations are detailed in Table 4. Initial  
268 rock permeabilities are divided into ‘low’ permeability systems ( $1.0 \times 10^{-15} \text{ m}^2$  core and  
269 substrate,  $1.0 \times 10^{-13} \text{ m}^2$  talus), ‘intermediate’ permeability systems ( $1.0 \times 10^{-14} \text{ m}^2$  core and  
270 substrate,  $1.0 \times 10^{-12} \text{ m}^2$  talus) and ‘high’ permeability systems ( $1.0 \times 10^{-13} \text{ m}^2$  core and substrate,  
271  $1.0 \times 10^{-11} \text{ m}^2$  talus) values. Two thermal conditions are used for simulations. One condition  
272 allows the dome and conduit to cool over time from their initial temperature conditions

273 (specified in Table 4 for each material region). The other maintains a constant 200°C heat source  
274 in the conduit, simulating a recharging magmatic heat source such as a dike or other intrusion  
275 providing heat to the system, slow solidification with release of latent heat, or heat fluxing from  
276 outgassing. Relative permeabilities vary with time and saturation according to a linear  
277 formulation (Zyvoloski et al. 1999; Zyvoloski 2007) which uses a residual liquid saturation of  
278 0.2 and a residual vapor saturations at 0.1 (see supplementary material for linear function  
279 equations). Again, we focus here on low-temperature domes because adding fluids to the model  
280 under high temperature conditions results in the model attempting to simulate abrupt phase  
281 changes and substantially slowing the modeling process. High temperatures may also result in  
282 extremely low saturations in the modeling domain, which again slow calculations. We also note  
283 that the temperature of formation of many of the minerals of concern with respect to edifice  
284 weakening and instability (particularly smectite clay) fall within the 100-300°C temperature  
285 range, which makes this a logical point in the temperature evolution of a cooling dome to  
286 investigate. At higher temperatures, the alteration mineral suite changes significantly and  
287 becomes dominated by minerals like biotite, actinolite, chlorite, and silica polymorphs (Henley  
288 and Ellis 1983; Reyes 1990), which are not usually implicated in alteration-related edifice  
289 collapse.

290

### 291 **Alteration potential determined with the rock alteration index (RAI)**

292 Volumetric liquid and vapor fluxes are calculated directly within FEHM, while  
293 temperature gradients are post-processed based on the FEHM temperature fields. The  
294 temperature field and liquid volumetric flux vectors are then used to calculate alteration potential  
295 with the Rock Alteration Index (RAI) for all nodes in the models at regular time intervals for

296 each simulation. The RAI was developed to describe the likelihood of alteration processes  
297 occurring based on temperature gradients and patterns of fluid migration in subsurface aqueous  
298 reservoirs. Steep temperature gradients along flow paths promote alteration by inducing  
299 thermodynamic instability in the system and encouraging the formation of secondary minerals in  
300 equilibrium with thermal conditions (Raffensperger and Vlassopoulos 1999; Wetzel et al. 2001).

301 The RAI is calculated from the temperature gradient and the fluid volumetric flux:

$$302 \quad \text{RAI} = \mathbf{q} \cdot \nabla T \quad (1)$$

303 where  $\mathbf{q}$  is the fluid volumetric flux (volume per cross-sectional area per time ( $\text{m}^3/(\text{m}^2 \text{ s})$ ), and  $T$   
304 is the temperature ( $^{\circ}\text{C}$ ). High positive values of the RAI coincide with areas of higher fluid flux  
305 and/or flow paths of rapidly increasing temperature and would promote mineral dissolution.  
306 High negative values are found in areas of higher fluid flux and/or flow paths of rapidly  
307 decreasing temperature and would promote mineral precipitation. An RAI value of zero does not  
308 necessarily reflect zero fluid flow, but can also indicate flow along isotherms. It should be noted  
309 that the RAI as calculated is an indicator of alteration potential only; it can be combined with  
310 geochemical species models to determine the likelihood of specific mineral formation, but for  
311 this initial study we focus on alteration potential only.

312

## 313 **RESULTS**

314 Twelve simulations were undertaken, varying the thermal and permeability conditions for  
315 each of the two dome geometries as described previously (Table 4). Temperature results  
316 highlight the differing evolution of temperature profiles in domes with a heat source and without,  
317 as in dome eruptions where magma either remains in some reservoir beneath the dome or has  
318 withdrawn. Variable permeability conditions were investigated because permeability is a primary

319 control on fluid flux (both direction and intensity); we are also trying to represent differences in  
320 permeability between materials as well as cover the overall range of reasonable permeabilities  
321 based on rock property measurements (see previous section). As temperature distribution and  
322 fluid flux are the dependent variables by which the alteration indices are calculated, knowing  
323 how they vary in space and time allows us to comment on why certain patterns of RAIs develop.

324

### 325 **Temperature gradients and liquid/vapor flux patterns**

326         The highest temperatures in the simulations persist in areas that are initially warm (dome  
327 cores and conduits) and decrease over time; higher-than-background temperatures progressively  
328 migrate downward into the substrate (Figure 3). In both dome geometries, domes with lower  
329 initial permeabilities tend to cool slower and do not achieve background temperatures by the end  
330 of the simulations, while initially higher permeability domes cool more rapidly, reaching  
331 background temperatures by or before the 100-year time limit. Likewise, in domes without a  
332 maintained heat source, both the onset of cooling as well as reaching background temperatures  
333 occurs earlier than for domes with a heat source. A persistent 200°C heat source in the conduit  
334 generally allowed the lower portions of the domes to remain hotter for longer periods of time,  
335 provided permeabilities were low. The inner cores of the domes remain hot longer in simulations  
336 with conduit heat, but there is little effect on outlying areas of the dome and talus. Crater-  
337 confined domes cool more slowly than perched domes, with perched domes cooling completely  
338 by the 50-year mark whereas crater-confined domes still retain zones of elevated temperatures in  
339 their cores (Figures 3e and 3f.) One persistent feature in every simulation is the development of a  
340 zone of warmer temperatures (usually about 40-70°C) at the base of the dome (Figures 3c and d,  
341 initial development), which migrates toward the break in slope created by the intersection of the



342 dome talus and slope talus. These zones develop faster in domes with medium or high  
343 permeabilities, but persist longer in lower permeability domes.

344 Liquid flux patterns are dominated by gravitational flow and show no evidence of  
345 convection (Figure 4). The highest liquid fluxes (up to an order of magnitude higher than the  
346 recharge flux) are found in both the talus of both the dome and slopes throughout the 100-year  
347 simulations; liquid fluxes decrease where liquid saturations are low. In most simulations, this  
348 decrease occurs at a front which migrates from the top of the dome core (Figures 5a and b) to its  
349 center, and around the head of the conduit (Figure 5g). Over time, high fluid fluxes decrease to  
350 background levels. Overall, fluxes vary between much less than the meteoric influx (min.  $\sim 1.0$   
351  $\times 10^{-8}$   $\text{m}^3/\text{m}^2 \text{ s}$ ) to an order of magnitude higher. Maximum fluxes occur in simulations where the  
352 initial permeabilities are high ( $1.0 \times 10^{-12}$   $\text{m}^2$  for talus,  $1.0 \times 10^{-13}$   $\text{m}^2$  for dome core and  
353 substrate). Perched domes contain larger areas of higher fluxes (Figure 4b) than crater-confined  
354 domes (Figure 4a), which coincide with the location of talus layers beneath the domes. These  
355 talus layers divert flow noticeably under the perched domes (Figures 4d-h), while in crater-  
356 confined domes this effect is much less pronounced and not as long-lived (Figure 4e). (Adding a  
357  $200^\circ\text{C}$  heat source to the models does not appear to have a significant effect on fluid fluxes (i.e.  
358 through increasing buoyancy of fluids, etc.), but indirect factors like low saturation zones related  
359 to a heat source (which reduce relative permeability and thus fluid flux), may be masking lesser  
360 effects.

361 Vapor fluxes (Figure 5) appear at the beginning of a simulation and a boiling point front  
362 (Figures 5a and b) migrates from the core/talus interface at the summit of the dome into the core  
363 of the dome as time progresses (Figures 5c - f). This front follows the high temperature gradient  
364 between the hot core of the dome and the recharge-cooled carapace. Once the dome has cooled,

365 vapor fluxes are confined to the area around the conduit head (Figures 5e, f and g), although they  
366 only persist in simulations where the conduit is maintained at high temperature. The addition of a  
367 persistent heat source does not increase the strength of the vapor fluxes but does affect the fluxes  
368 indirectly (by increasing the areas that are not fully saturated and thus increasing the area over  
369 which vapor fluxes are produced). Vapor volume fluxes range from  $1.0 \times 10^{-6}$  to  $9.0 \times 10^{-6} \text{ m}^3 \text{ m}^{-2} \text{ s}^{-1}$   
370 and are higher when overall permeabilities are higher. Vapor fluxes are initially higher in  
371 crater-confined domes than perched domes (Figures 5a and 5b), but more widespread in perched  
372 domes. By the 50-year mark, vapor fluxes in both domes are approximately the same magnitude  
373 (Figures 5e and 5f).

374

### 375 **Resulting alteration potential**

376 Alteration potential (Figure 6) in both domes is controlled primarily by the magnitude  
377 and direction of fluid flux and the temperature gradients of the cooling domes, and is thus  
378 intimately related to permeability contrasts in materials and the availability of heat. Because the  
379 highest liquid fluxes occur in the higher-permeability talus layers (Figures 4a and b), and the  
380 greatest temperature gradients are generated between the dome cores and the cooler substrates  
381 (Figures 3a and b), the highest positive and negative RAI values occur at the interface between  
382 the core and talus of the dome (Figures 6a and b). High positive values at the dome summits  
383 suggest potential for mineral dissolution (alteration). Conversely, high negative values at the  
384 base of the dome and around the slope break in the talus indicate potential for mineral  
385 precipitation. Crater-confined domes show more intense positive and negative RAI values  
386 initially (Figures 6a and 6b), and are the only geometry that still shows non-zero RAI values in  
387 the core of the dome by the 10-year simulation time (Figures 6c and 6d). Neither geometry has  
388 non-zero RAI values in the dome after the 50-year simulation time (Figures 6e – h).

389 Permeability and thermal conditions also control the strength and longevity of alteration  
390 zones. Potential alteration is more intense with low to moderately high permeabilities, but is not  
391 sustained for long unless a conduit heat source is present. Alteration potential decreases  
392 dramatically in these models within even the first 10 years (Figures 6c and d). However, very  
393 high permeabilities preclude strong alteration at the talus/core interface and instead, high RAI  
394 values are only generated around the conduit. Lower permeabilities combined with a heat source  
395 sustain both positive and negative RAI values longer, but at lower intensities than those  
396 developed at higher permeabilities. In both dome geometries, negative RAI values at the base of  
397 the dome tend to persist longer than the positive ones in the upper dome (Figures 6c and d). The  
398 ideal combination for forming sustained, localized areas with high RAI values (positive or  
399 negative) appears to be a maintained conduit heat source combined with low to intermediate  
400 permeabilities ( $1 \times 10^{-14} \text{ m}^2$  for core and substrate and  $1 \times 10^{-13} \text{ m}^2$  for talus). This enables water  
401 flux across strong temperature gradients without allowing the dome to cool too quickly.

402

## 403 **DISCUSSION**

404 Domes cool from 150 to  $\sim 30^\circ\text{C}$  within 100 years in these simulations, suggesting that the  
405 lifetime of a low-temperature hydrothermal system in a tropical lava dome is only years to  
406 decades if the dome lacks a sustained high-temperature magmatic heat source. A sustained  
407  $200^\circ\text{C}$  heat source prolongs cooling, but  $200^\circ\text{C}$  appears to be insufficiently hot to effectively  
408 drive hydrothermal circulation within the domes. Perched dome geometries cool faster than  
409 crater-confined domes, likely due to the presence of high-permeability talus layers beneath  
410 portions of the dome that allow more water transport around the dome core.

411 Overall, higher permeabilities and fluid fluxes promote more-intense early-onset RAIs;  
412 however, these values are not as persistent as those in domes of lower permeabilities. Crater-  
413 confined geometries retain intense RAIs somewhat longer than perched domes because of the  
414 presence of high-permeability talus layers beneath perched domes (which helps cool the domes  
415 faster), but geometry exerts a much weaker control than permeability. Therefore early-onset  
416 RAIs might be expected in pervasively fractured domes, but longer-lasting RAIs would be found  
417 in more-coherent domes. For both modeled dome geometries, high positive RAIs are formed  
418 near the summit while negative RAIs occur at the base of the domes. This indicates that  
419 dissolution processes are more likely during infiltration and precipitation processes are more  
420 likely as water percolates out the base of a dome. In these models, the latter location is occupied  
421 by talus, but it could also include features such as brecciated zones at the base of shear  
422 lobes/flows (John et al. 2008); for example, the suites of hydrothermal alteration minerals in  
423 clay-rich lahars on Mount Rainier are often sourced from brecciated lavas and pyroclastic  
424 deposits with high primary permeabilities (Crandell 1971; Scott and Vallance 1995; John et al.  
425 2008). However, because water permeates through domes – which are composed of both  
426 fractured lavas and brecciated talus – relatively quickly, it is also possible that the water would  
427 not have the opportunity to form advanced alteration assemblages. This is consistent with  
428 observations at Santiaguito, where hot springs represent immature waters that have dissolved, but  
429 not equilibrated with, dome rock or formed minerals such as clays, and alteration on the dome  
430 surfaces is limited to the deposition of hydrous silica (Ball et al. 2013).

431 These results now provide a framework for investigations of specific mineral formation,  
432 for example, by combining temperature and flux data with aqueous geochemical data using a  
433 species model such as EQ3/6 (Wetzel et al. 2001). The simulation results suggest that alteration

434 minerals forming from dissolution processes (clays and kaolinite) are more likely to be found at  
435 the dome summit, while minerals resulting from precipitation processes (alunite) are more likely  
436 found at the base of the dome and talus around the slope break. Clays and alunite could  
437 contribute to shallow-seated collapses of the carapace and in the talus. Collapses triggered by  
438 talus erosion and dome undermining have been associated with intense rainfall (Calder et al.  
439 2005). It might be that increased fluid pressure in clay-bearing horizons contributes to increased  
440 instability associated with precipitation. In sufficient quantity, clays, alunite and silica minerals  
441 could also form impermeable layers, facilitating gas sealing (Elsworth et al. 2004), a mechanism  
442 that has been suggested for some deep-seated dome failures in active lava domes. Precipitation  
443 of vapor-phase cristobalite in domes has been shown to decrease porosity (and by inference, also  
444 permeability) of dome rock (Horwell et al. 2013). A similar effect could occur in low-  
445 temperature systems, such as modeled here, with other mineral precipitates (Figure 7); however,  
446 more-complex models accounting for factors such as fracture networks and high-permeability  
447 zones such as shear lobe boundaries would be necessary for a complete evaluation of this  
448 scenario.

449         The presence of hot, magmatic acidic gases rising from a magma source will accelerate  
450 alteration of the dome rock (Reyes 1990). However, Cox and Browne (1998) note that large-  
451 scale alteration of rock to smectite/montmorillonite is still possible even in neutral-to-alkaline pH  
452 systems. Additionally, while advanced argillic alteration contains a suite of weak sulfate  
453 minerals (such as alunite and jarosite) that require the presence of sulfur gases to form, clay  
454 minerals such as smectite also depend on the presence of water. Water vapor is the primary gas  
455 released in any volcanic eruption and meteoric water composes a significant percentage of the  
456 fluid available in any near-surface hydrothermal system (Goff & Janik, 2000) and it is reasonable

457 to assume that the degassing pathways followed by water vapor would be, at the shallow levels  
458 depicted in these models, essentially the same as those traveled by acidic gases released from a  
459 magmatic body. The behavior of water and water vapor are therefore useful both as first-order  
460 information and as proxies for interpretations about additional gases in lava dome hydrothermal  
461 systems.

462  
463 **Comparison to existing domes and volcanoes**

464         These simulations represent a first order approach into estimating the behavior of liquid  
465 water and water vapor in the interior of cooling lava domes. Physically validating this  
466 assessment of alteration is somewhat difficult, since there is currently little in the way of direct  
467 or remotely sensed field data about the hydrothermal systems of lava domes. Physical mapping  
468 of dissected domes (Duffield et al. 1995; Riggs and Carrasco-Nunez 2004) is generally limited to  
469 structural features or eruptive facies and neglects information about alteration mineral  
470 assemblages or the location of hydrothermal flow paths or fumaroles. Remote sensing is  
471 necessarily limited to surface materials, and while debris-avalanches may sample the interior of  
472 domes, the percentages and types of altered material involved in volcanic collapses have not  
473 been recorded in a systematic or detailed manner in most studies (Dufresne 2009 and personal  
474 communication). Voight et al. (2002) were able to trace hydrothermally altered material in the  
475 1996 collapse of the Soufriere Hills lava dome to the margins of the dome and buttressing crater  
476 wall, suggesting that permeability contrasts did indeed come into play (the altered material  
477 included pyroclastic deposits of brecciated lava). However in large collapses the deposits may  
478 consist of entire lava domes, making it difficult, if not impossible, to reconstruct internal  
479 structures and zones of alteration. As a result, the best available data about undisturbed dome  
480 interiors comes from geophysical investigations.

481           Nicollin et al. (2006) completed an electrical tomographic study of La Soufriere of  
482 Guadeloupe volcano and lava dome and created cross-sections of electrical resistivity indicating  
483 likely areas of hydrothermal alteration (altered materials tend to be less resistive/more  
484 conductive). They determined that there was a large area of low resistivity located at the base of  
485 the dome, which they interpreted as a hydrothermally altered zone; in addition, a medium  
486 resistivity layer on the western flank of the lava dome was found at depth below a layer of lower  
487 resistivity. Other zones of low resistivity were interpreted as cross-cutting faults which provided  
488 pathways for liquids and gases that promoted alteration. The authors interpreted the basal layer  
489 as related to the collapse of a highly-altered summit of the volcano (producing the crater in  
490 which the dome formed), but mentioned that the western low-resistivity zone could represent  
491 massive unaltered areas of the lava dome overlaid by a layer of thick better-drained scoriaceous  
492 or altered material from the dome carapace or brecciated units formed during the formation of  
493 flow lobes. This would be consistent with the RAI interpretations reached in this model, where  
494 alteration is likely to be concentrated at the interface between higher and lower permeability  
495 regions (higher permeabilities allow the passage of more liquid and vapor, which are essential to  
496 alteration processes). A recently published study by Brothelande et al. (2014) expands on this,  
497 indicating that hydrothermal ascending flows in the volcano are limited to the dome and its  
498 immediate proximity, confined in a collapse structure surrounding the dome (last modified in  
499 1530 by a collapse followed by an eruption).

500           Finn et al. (2007) and Finn and Deszcz-Pan (2007) conducted helicopter magnetic and  
501 electromagnetic surveys of Mounts Adams and Baker in order to determine the three-  
502 dimensional geometry of altered and saturated regions within those Cascade stratovolcanoes.  
503 Hydrothermal alteration significantly reduces the magnetization in volcanic rocks, while the

504 presence of alteration minerals, or of water in the rock, reduces electrical resistivity; by  
505 combining these methods, the authors were able to discriminate areas of dry, fresh rock from  
506 saturated fresh or weakly altered rock and variably saturated intensely altered rock (Finn et al.  
507 2007; Finn and Deszcz-Pan 2011). On Mount Adams, intensely altered and saturated rock is  
508 found in the core of the volcano, but layers of fresh or weakly altered saturated rock also underlie  
509 portion of the volcano's slopes. The authors interpreted these as 'fresh porous breccias', similar  
510 to the kind of material found in talus units at lava domes (Finn et al. 2007). Alteration at Mount  
511 Baker follows a different pattern, being restricted to thinner layers beneath the summit crater and  
512 fumarole field on the volcano's northeast flank (Finn and Deszcz-pan 2011), but similar un- or  
513 weakly-altered saturated layers underlie its slopes. While these methods were applied on an  
514 entire stratovolcano and did not discern features on the scale of lava dome structures, ground-  
515 based application of magnetic and electromagnetic surveys could provide that information at the  
516 scale of a lava dome. Muon radiography (Tanaka et al. 2007; Lesparre et al. 2012) is an  
517 emerging geophysical method which shows promise in determining density contrasts in lava  
518 domes, but it would be necessary to determine if altered material showed significant density  
519 contrasts with unaltered material, and if it was distinguishable from density differences in  
520 dome/conduit/talus material in the first place. Currently, muon radiography studies have been  
521 able to locate conduits within and beneath domes, but are limited by the fact that the method  
522 must encompass the whole thickness of the dome and cannot take a "slice" from it as in  
523 resistivity studies.

524         In order to corroborate the results of this study, future field and geophysical  
525 investigations of both old and young domes would be useful, with attention paid to the location,  
526 degree and character of alteration.



527

## 528 **CONCLUSIONS**

529           This study applies a multi-phase porous flow model to determine the flow of water and  
530 heat in low-temperature cooling lava domes over 100-year timescales. A number of conclusions  
531 can be drawn from these first-order simulations of lava dome hydrothermal systems.

532           1. The alteration potential in these domes is controlled by the contrasts in material  
533 permeability and the heat sources driving hydrothermal flow, and is highest where permeability  
534 contrasts are greatest, particularly at the interface between the less permeable dome core and  
535 more permeable talus. This suggests that alteration mineral formation is most likely to occur at  
536 the boundaries of lava dome structures.

537           2. Areas of increased alteration likelihood are sustained longer in low-permeability  
538 domes, but are more intense in domes with higher permeabilities and persistent heat sources. A  
539 dome without a sustained heat input will cool on geologically short time periods, and even faster  
540 if its overall permeability is high, denying the opportunity to develop alteration. However, at the  
541 low temperatures and high infiltration rates in these simulations there is no evidence for  
542 convection of water in the domes and flow is dominated by gravity, precluding the possibility of  
543 long-lived hydrothermal circulation.

544           3. Potential for dissolution (clay mineral formation) is highest near the summit at the  
545 core/talus interface of the simulated domes, while the potential for mineral precipitation (alunite,  
546 silica formation) is highest at the base of the domes. If alteration forms weak minerals at the  
547 core/talus interface the area could source shallow-seated collapses of the carapace.

548           In combination with geophysical and field studies, numerical modeling can provide an  
549 important first step in elucidating the behavior of post-eruptive volcanic systems. Incorporating

550 the results of numerical models with limited ground-based data and remote sensing can  
551 strengthen the interpretations drawn from both, and provide valuable insight into dome evolution  
552 and hazards. Coupling these first-order flow models with mineral species models and  
553 representing more complex dome structures and different fluid chemistries could allow  
554 simulations such as these to be used to evaluate potential collapse mechanisms at specific domes.  
555 The results of the models presented here provide a framework for future investigations, including  
556 field, geochemical and geophysical, into the way post-eruptive lava domes are altered by  
557 hydrothermal activity.

558

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573 **REFERENCES CITED**

- 574 Aizawa K, Ogawa Y, Ishido T (2009) Groundwater flow and hydrothermal systems within  
575 volcanic edifices: Delineation by electric self-potential and magnetotellurics. *J Geophys Res*  
576 114:1–12. doi: 10.1029/2008JB005910
- 577 Anderson S., Arthur M., Asimow P., et al (1999) *Encyclopedia of Volcanoes*. 1442.
- 578 Ball JL, Calder ES, Hubbard BE, Bernstein ML (2013) An assessment of hydrothermal alteration  
579 in the Santiaguito lava dome complex, Guatemala: implications for dome collapse hazards.  
580 *Bull Volcanol* 75:676. doi: 10.1007/s00445-012-0676-z
- 581 Barclay J, Johnstone JE, Matthews AJ (2006) Meteorological monitoring of an active volcano:  
582 Implications for eruption prediction. *J Volcanol Geotherm Res* 150:339–358. doi:  
583 10.1016/j.jvolgeores.2005.07.020
- 584 Bedrosian PA, Unsworth MJ, Johnston MJS (2007) Hydrothermal circulation at Mount St.  
585 Helens determined by self-potential measurements. *J Volcanol Geotherm Res* 160:137–146.  
586 doi: 10.1016/j.jvolgeores.2006.09.003
- 587 Boudon G, Villemant B, Komorowski J, et al (1998) The hydrothermal system at Soufriere Hills  
588 Volcano, Montserrat (West Indies): Characterization and role in the on- going eruption.  
589 *Geophys Res Lett* 25:3693. doi: 10.1029/98GL00985
- 590 Brothelande E, Finizola A, Peltier A, et al (2014) Fluid circulation pattern inside La Soufrière  
591 volcano (Guadeloupe) inferred from combined electrical resistivity tomography, self-  
592 potential, soil temperature and diffuse degassing measurements. *J Volcanol Geotherm Res*  
593 288:105–122. doi: 10.1016/j.jvolgeores.2014.10.007
- 594 Calder ES, Cortés J a., Palma JL, Luckett R (2005) Probabilistic analysis of rockfall frequencies  
595 during an andesite lava dome eruption: The Soufrière Hills Volcano, Montserrat. *Geophys*  
596 *Res Lett* 32:1–4. doi: 10.1029/2005GL023594
- 597 Calder ES, Luckett R, Sparks RSJ, Voight B (2002) Mechanisms of lava dome instability and  
598 generation of rockfalls and pyroclastic flows at Soufriere Hills Volcano, Montserrat. *Geol*  
599 *Soc London, Mem* 21:173–190. doi: 10.1144/GSL.MEM.2002.021.01.08
- 600 Carrasco-Núñez G, Díaz-Castellón R, Siebert L, et al (2006) Multiple edifice-collapse events in  
601 the Eastern Mexican Volcanic Belt: The role of sloping substrate and implications for  
602 hazard assessment. *J Volcanol Geotherm Res* 158:151–176. doi:  
603 10.1016/j.jvolgeores.2006.04.025
- 604 Chaudhuri a., Rajaram H, Viswanathan H, et al (2009) Buoyant convection resulting from  
605 dissolution and permeability growth in vertical limestone fractures. *Geophys Res Lett*  
606 36:587–596. doi: 10.1029/2008GL036533

- 607 Costa a., Melnik O, Sparks RSJ (2007) Controls of conduit geometry and wallrock elasticity on  
608 lava dome eruptions. *Earth Planet Sci Lett* 260:137–151. doi: 10.1016/j.epsl.2007.05.024
- 609 Cox ME, Browne P (1998) Hydrothermal alteration mineralogy as an indicator of hydrology at  
610 the Ngawha geothermal field, New Zealand. *Geothermics* 27:259–270. doi: 10.1016/S0375-  
611 6505(97)10015-3
- 612 Crandell DR (1971) Postglacial Lahars From Mount Rainier Volcano, Washington. *US Geol*  
613 *Surv Prof Pap* 677 75.
- 614 Dash Z V. (2003) Validation Test Plan (VTP) Results for the FEHM Application Version 2.21.  
615 76. <http://fehmlanl.gov/pdfs/fehmvvr.pdf>
- 616 Dash Z V., Fitzgerald MF, Pollock F (2003) Validation Test Plan (VTP) for the FEHM  
617 Application Version 2.21. 20. <http://fehmlanl.gov/pdfs/fehmvvp.pdf>
- 618 Del Potro R, Hürlimann M (2009) The decrease in the shear strength of volcanic materials with  
619 argillic hydrothermal alteration, insights from the summit region of Teide stratovolcano,  
620 Tenerife. *Eng Geol* 104:135–143. doi: 10.1016/j.enggeo.2008.09.005
- 621 Devoli G, Cepeda J, Kerle N (2009) The 1998 Casita volcano flank failure revisited — New  
622 insights into geological setting and failure mechanisms. *Eng Geol* 105:65–83. doi:  
623 10.1016/j.enggeo.2008.12.006
- 624 Duffield BWA, Richter DH, Priest SS (1995) Physical volcanology of silicic lava domes as  
625 exemplified by the Taylor Creek Rhyolite, Catron and Sierra Counties, New Mexico.
- 626 Dufresne A (2009) Influence of runout path material on rock and debris avalanche mobility: field  
627 evidence and analogue modelling. 268.
- 628 Elsworth D, Voight B, Thompson G, Young SR (2004) Thermal-hydrologic mechanism for  
629 rainfall-triggered collapse of lava domes. *Geology* 32:969. doi: 10.1130/G20730.1
- 630 Fields R, Soni BK, Thompson JF, et al (1996) Geological Applications of Automatic Grid  
631 Generation Tools for Finite Elements Applied to Porous Flow Modeling From : Numerical  
632 Grid Generation in Computational Fluid Dynamics and From : Numerical Grid Generation  
633 in Computational Fluid Dynamics and From : Methods 1–9.
- 634 Fink JH, Griffiths RW (1998a) Morphology, eruption rates, and rheology of lava domes: Insights  
635 from laboratory models. *J Geophys Res* 103:527. doi: 10.1029/97JB02838
- 636 Fink JH, Griffiths RW (1998b) Morphology, eruption rates, and rheology of lava domes: Insights  
637 from laboratory models. *J Geophys Res* 103:527. doi: 10.1029/97JB02838
- 638 Fink JH, Pollard DD (1983) Structural evidence for dikes beneath silicic domes, Medicine Lake  
639 Highland Volcano, California. *Geology*. doi: 10.1130/0091-7613(1983)11<458

- 640 Finn CA, Deszcz-Pan M, Anderson ED, John D a. (2007) Three-dimensional geophysical  
641 mapping of rock alteration and water content at Mount Adams, Washington: Implications  
642 for lahar hazards. *J Geophys Res* 112:1–21. doi: 10.1029/2006JB004783
- 643 Finn CA, Deszcz-Pan M (2011) Helicopter magnetic and electromagnetic surveys at Mounts  
644 Adams, Baker and Rainier, Washington: Implications for debris flow hazards and volcano  
645 hydrology. *Soc Explor Geophys Glob Meet Abstr* 15:3 pp. doi: 10.1190/1.3659065
- 646 Friant A Le, Boudon G, Komorowski JC, et al (2006) Potential flank-collapse of Soufriere  
647 volcano, Guadeloupe, Lesser Antilles: Numerical simulation and hazards. *Nat Hazards*  
648 39:381–393. doi: 10.1007/s11069-005-6128-8
- 649 Giggenbach W (1992) SEG Distinguished Lecture: Magma Degassing and Mineral Deposition in  
650 Hydrothermal Systems along Convergent Plate Boundaries. *Econ. Geol.*
- 651 Hale AJ (2008) Lava dome growth and evolution with an independently deformable talus.  
652 *Geophys J Int* 174:391–417. doi: 10.1111/j.1365-246X.2008.03806.x
- 653 Hale AJ, Calder ES, Loughlin SC, et al (2009a) Modelling the lava dome extruded at Soufriere  
654 Hills Volcano, Montserrat, August 2005-May 2006; Part I: Dome shape and internal  
655 structure. *J Volcanol Geotherm Res* 187:69–84. doi: 10.1016/j.jvolgeores.2009.08.014
- 656 Hale AJ, Calder ES, Wadge G, et al (2009b) Modelling the lava dome extruded at Soufriere Hills  
657 Volcano, Montserrat, August 2005-May 2006; Part II: Rockfall activity and talus  
658 deformation. *J Volcanol Geotherm Res* 187:53–68. doi: 10.1016/j.jvolgeores.2009.08.023
- 659 Henley RW, Ellis AJ (1983) Geothermal systems ancient and modern: a geochemical review.  
660 *Earth-Science Rev.* doi: 10.1016/0012-8252(83)90075-2
- 661 Horwell CJ, Williamson BJ, Llewellyn EW, et al (2013) The nature and formation of cristobalite  
662 at the Soufrière Hills volcano, Montserrat: Implications for the petrology and stability of  
663 silicic lava domes. *Bull Volcanol* 75:1–19. doi: 10.1007/s00445-013-0696-3
- 664 Hurwitz S (2003) Groundwater flow, heat transport, and water table position within volcanic  
665 edifices: Implications for volcanic processes in the Cascade Range. *J Geophys Res* 108:1–  
666 19. doi: 10.1029/2003JB002565
- 667 Hutnak M, Fisher a. T, Zühlsdorf L, et al (2006) Hydrothermal recharge and discharge guided by  
668 basement outcrops on 0.7-3.6 Ma seafloor east of the Juan de Fuca Ridge: Observations and  
669 numerical models. *Geochemistry, Geophys Geosystems.* doi: 10.1029/2006GC001242
- 670 Ingebritsen SE, Geiger S, Hurwitz S, Driesner T (2010) Numerical simulation of magmatic  
671 hydrothermal systems. *Rev Geophys* 48:1–33. doi: 10.1029/2009RG000287

- 672 John DA, Sisson TW, Breit GN, et al (2008) Characteristics, extent and origin of hydrothermal  
673 alteration at Mount Rainier Volcano, Cascades Arc, USA: Implications for debris-flow  
674 hazards and mineral deposits. *J Volcanol Geotherm Res* 175:289–314.
- 675 Keating GN (2005) The role of water in cooling ignimbrites. *J Volcanol Geotherm Res* 142:145–  
676 171. doi: 10.1016/j.jvolgeores.2004.10.019
- 677 Kerle N (2002) Volume estimation of the 1998 flank collapse at Casita volcano, Nicaragua: A  
678 comparison of photogrammetric and conventional techniques. *Earth Surf Process*  
679 *Landforms* 27:759–772. doi: 10.1002/esp.351
- 680 Khaleel R (1989) Scale dependence of continuum models for fractured basalts. *Water Resour*  
681 *Res* 25:1847. doi: 10.1029/WR025i008p01847
- 682 Lesparre N, Gibert D, Marteau J, et al (2012) Density muon radiography of La Soufriere of  
683 Guadeloupe volcano: Comparison with geological, electrical resistivity and gravity data.  
684 *Geophys J Int* 190:1008–1019. doi: 10.1111/j.1365-246X.2012.05546.x
- 685 McGuire W (2003) Volcano instability and lateral collapse. I:33–45.
- 686 Miller TA, Vessilinov VV, Stauffer PH, et al (2007) Integration of geologic frameworks in  
687 meshing and setup of computational hydrogeologic models, Pajarito Plateau, New Mexico.  
688 *New Mex. Geol. Soc. Guid. Book, 58th F. Conf. Geol. Jemez Mt. Reg. III.*
- 689 Nakada S, Shimizu H, Ohta K (1999) Overview of the 1990–1995 eruption at Unzen Volcano. *J*  
690 *Volcanol Geotherm Res* 89:1–22. doi: 10.1016/S0377-0273(98)00118-8
- 691 Nicollin F, Gibert D, Beauducel F, et al (2006) Electrical tomography of La Soufrière of  
692 Guadeloupe Volcano: Field experiments, 1D inversion and qualitative interpretation. *Earth*  
693 *Planet Sci Lett* 244:709–724. doi: 10.1016/j.epsl.2006.02.020
- 694 Opfergelt S, Delmelle P, Boivin P, Delvaux B (2006) The 1998 debris avalanche at Casita  
695 volcano, Nicaragua: Investigation of the role of hydrothermal smectite in promoting slope  
696 instability. *Geophys Res Lett* 33:4. doi: L15305 10.1029/2006gl026661
- 697 Raffensperger JP, Vlassopoulos D (1999) The potential for free and mixed convection in  
698 sedimentary basins. *Hydrogeol J* 7:505–520. doi: 10.1007/s100400050224
- 699 Reid ME, Sisson TW, Brien DL (2002a) Volcano collapse promoted by hydrothermal alteration  
700 and edifice shape, Mount Rainier, Washington. *Geology* 29:779–782. doi: 10.1130/0091-  
701 7613(2001)029<0779:VCPBHA>2.0.CO;2
- 702 Reid ME, Sisson TW, Brien DL (2002b) Volcano collapse promoted by hydrothermal alteration  
703 and edifice shape, Mount Rainier, Washington. *Geology* 29:779–782. doi: 10.1130/0091-  
704 7613(2001)029<0779:VCPBHA>2.0.CO;2

- 705 Reyes AG (1990) Petrology of Philippine geothermal systems and the application of alteration  
706 mineralogy to their assessment. *J Volcanol Geotherm Res.* doi: 10.1016/0377-  
707 0273(90)90057-M
- 708 Riggs N, Carrasco-Nunez G (2004) Evolution of a complex isolated dome system, Cerro Pizarro,  
709 central México. *Bull Volcanol* 66:322–335. doi: 10.1007/s00445-003-0313-y
- 710 Salaün A, Villemant B, Gérard M, et al (2011) Hydrothermal alteration in andesitic volcanoes:  
711 Trace element redistribution in active and ancient hydrothermal systems of Guadeloupe  
712 (Lesser Antilles). *J Geochemical Explor* 111:59–83. doi: 10.1016/j.gexplo.2011.06.004
- 713 Scott KM, Vallance JW (1995) Debris flow, debris avalanche, and flood hazards at and  
714 downstream from Mount Rainier, Washington. *Hydrol Investig Atlas* 9 (2 sheets).
- 715 Scott KM, Vallance JW, Kerle N, et al (2005) Catastrophic precipitation-triggered lahar at Casita  
716 volcano, Nicaragua: Occurrence, bulking and transformation. *Earth Surf Process Landforms*  
717 30:59–79. doi: 10.1002/esp.1127
- 718 Sheridan MF, Bonnard C, Careno R, et al (1999) Report on the 30 October 1998 Rock Fall /  
719 Avalanche and Breakout Flow of Casita Volcano, Nicaragua, Triggered by Hurricane  
720 Mitch. *Landslide News* 1202–1204.
- 721 Siebert L (2002) Landslides resulting from structural failure of volcanoes. *Catastrophic*  
722 *landslides Eff Occur Mech* 15:209–235. doi: 10.1130/REG15-p209
- 723 Smithsonian Institution (1991) Unzen. *Bull. Glob. Volcanism Netw.* 16:
- 724 Sparks RSJ, Barclay J, Calder ES, et al (2002) Generation of a debris avalanche and violent  
725 pyroclastic density current on 26 December (Boxing Day) 1997 at Soufriere Hills Volcano,  
726 Montserrat. *Geol Soc London, Mem* 21:409–434. doi: 10.1144/GSL.MEM.2002.021.01.18
- 727 Tanaka HKM, Nakano T, Takahashi S, et al (2007) Imaging the conduit size of the dome with  
728 cosmic-ray muons: The structure beneath Showa-Shinzan Lava Dome, Japan. *Geophys Res*  
729 *Lett* 34:L22311. doi: 10.1029/2007GL031389
- 730 Voight B, Elsworth D (1997) Failure of volcano slopes. *Géotechnique.* doi:  
731 10.1680/geot.1997.47.1.1
- 732 Voight B, Elsworth D (2000) Instability and collapse of hazardous gas-pressurized lava domes.  
733 *Geophys Res Lett* 27:1–4.
- 734 Voight B, Komorowski J, Norton GE, et al (2002) The 26 December (Boxing Day) 1997 sector  
735 collapse and debris avalanche at Soufriere Hills Volcano, Montserrat. *Geol Soc London,*  
736 *Mem* 21:363–407. doi: 10.1144/GSL.MEM.2002.021.01.17

737 Wadge G, Ryan G, Calder ES (2009) Clastic and core lava components of a silicic lava dome.  
738 *Geology* 37:551–554. doi: 10.1130/G25747A.1

739 Walker JA, Templeton S, Cameron BI (2006) The chemistry of spring waters and fumarolic  
740 gases encircling Santa Maria Volcano, Guatemala. *Geol Soc Am Spec Pap* 412:59. doi:  
741 10.1130/2006.2412(04)

742 Walter TR, Ratdomopurbo A, Aisyah N, et al (2013) Dome growth and coulée spreading  
743 controlled by surface morphology, as determined by pixel offsets in photographs of the  
744 2006 Merapi eruption. *J Volcanol Geotherm Res* 261:121–129. doi:  
745 10.1016/j.jvolgeores.2013.02.004

746 Watts RB, Herd R a., Sparks RSJ, Young SR (2002) Growth patterns and emplacement of the  
747 andesitic lava dome at Soufriere Hills Volcano, Montserrat. *Geol Soc London, Mem*  
748 21:115–152. doi: 10.1144/GSL.MEM.2002.021.01.06

749 Wetzel LR, Raffensperger JP, Shock EL (2001) Predictions of hydrothermal alteration within  
750 near-ridge oceanic crust from coordinated geochemical and fluid flow models. *J Volcanol*  
751 *Geotherm Res* 110:319–342. doi: 10.1016/S0377-0273(01)00215-3

752 Zyvoloski G (2007) FEHM: A control volume finite element code for simulating subsurface  
753 multi-phase multi-fluid heat and mass transfer. Los Alamos Unclassif. Rep. LA-UR-07-  
754 3359

755 Zyvoloski GA, Robinson BA, Dash Z V., Trease LL (1999) Models and methods summary for  
756 the FEHM application. Los Alamos Natl. Laboratory Publ. SC-194

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765 **FIGURE CAPTIONS**

766 **Fig. 1** Basic equations that combine to form the full governing partial differential equations in  
767 FEHM, as reported in Zyvoloski et al. (1999). For a full derivation and comments, see the  
768 supplemental material.

769

770 **Fig. 2** Schematic of the two modeling grids for (A) crater-confined lava domes (e.g. Santiaguito)  
771 and (B) perched lava domes (not confined by a crater, e.g. Unzen).

772

773 **Fig. 3** Temperature outputs for the crater-confined (left) and perched (right) dome geometries  
774 after 1 year for intermediate permeabilities and a conduit heat source. (Figures illustrate a subset  
775 of the full modeling domain above 1500 m elevation).

776

777 **Fig. 4** Fluid flux outputs for crater-confined and perched dome geometries at 1, 10, 50, and 100  
778 year intervals for intermediate permeabilities and a conduit heat source. Blues indicate higher  
779 fluxes and greens lower fluxes. Streamlines indicate flow direction only and are arbitrarily  
780 spaced.

781

782 **Fig. 5** Vapor flux outputs for crater-confined and perched dome geometries at 1, 10, 50, and 100  
783 year intervals for intermediate permeabilities and a conduit heat source. Higher fluxes are  
784 indicated by reds and areas of zero vapor flux by white.

785

786 **Fig. 6** RAI patterns for crater-confined and perched domes at 1, 10, 50, and 100 year intervals.  
787 RAI magnitude scales differ between the two dome geometries, but warm colors indicate  
788 positive RAIs (dissolution) and cool colors indicate negative RAIs (precipitation) in both.

789

790 **Fig. 7** Summary of dome alteration and potential collapse loci based on RAI patterns. Alteration  
791 mineral formation is most likely to occur at the talus/core interface early in the lifetime of the  
792 dome, and depending on the mineral species involved could either strengthen or weaken the  
793 dome and/or promote internal gas pressurization.

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**Table 1.** Summary of ranges for material properties derived from literature review.

Independent variable	Dome range	Talus range	Notes	References
Porosity ( $\phi$ )	0.1-0.5	0.1-0.6	Talus layers are likely to be more porous than core (porosity increases with increasing fragmentation, void space)	(García et al. 1989; Ingebritsen and Hayba 1994; Alt-Epping et al. 2001; Barmin et al. 2002; Hurwitz 2003; Keating 2005; Bartetzko et al. 2006; Flint et al. 2006; Scheu et al. 2006; Bernard et al. 2007; Mueller et al. 2008; Watanabe et al. 2008; Ikeda et al. 2008; Aizawa et al. 2009; Hicks et al. 2009; Wicks et al. 2011)
Permeability ( $m^2$ )	$1.0 \times 10^{-19}$ to $1.0 \times 10^{-12}$	$1.0 \times 10^{-17}$ to $1.0 \times 10^{-10}$	Talus layers are likely to be much more permeable (connected to porosity; see previous) due to higher porosity, fractured/rubbly state	(Sammel et al. 1988; Sekioka 1988; Ingebritsen and Hayba 1994; Barmin et al. 2002; Keating 2005; Bartetzko et al. 2006; Bernard et al. 2007; Mueller et al. 2008; Watanabe et al. 2008; Aizawa et al. 2009; Hicks et al. 2009; Platz et al. 2012)
Density ( $kg/m^3$ )	593-2890	--	Dome and talus are assumed to be composed of the same material	(García et al. 1989; Smith et al. 2001; Hurwitz 2003; Keating 2005; Scheu et al. 2006; Bernard et al. 2007; Watanabe et al. 2008; Ikeda et al. 2008; Hicks et al. 2009)
Thermal conductivity (W/m K)	0.537-3.430	0.35	Value for most volcanic rocks is ~2	(Sekioka 1988; García et al. 1989; Ingebritsen and Hayba 1994; Hurwitz 2003; Keating 2005; Hicks et al. 2009)
Specific heat (J/kg K)	730-1557	2036	Dome and talus are assumed to be composed of the same material	(Keating 2005; Hicks et al. 2009; Platz et al. 2012)

**Table 2.** *Constant simulation parameters and boundary conditions*

<b>Parameter</b>	<b>Value</b>
Porosity of Zones 2 and 3	0.3
Porosity of Zones 1 and 5	0.2
Porosity of Zone 4	0.1
Initial temperature of Zone 1	150 °C
Initial temperature of Zone 2	30 °C
Initial temperature of Zone 3	30 °C
Initial temperature of Zone 4	200 °C
Initial temperature of Zone 5	70 °C
Permeability of Zone 4	$1 \times 10^{-16} \text{ m}^2$
Thermal conductivity of all units	$2.0 \text{ W m}^{-1} \text{ K}^{-1}$
Rock density of all units	$2600 \text{ kg m}^{-3}$
Specific heat	1000
Atmospheric temperature (Zone 6)	25 °C
Atmospheric pressure (Zone 6)	0.1 MPa
Precipitation rate (Zone 6)	1700 mm/yr

**Table 3.** Yearly precipitation rates for different currently active, or recently active, lava domes.

<b>Dome/ Location</b>	<b>Min rate (mm/yr)</b>	<b>Max rate (mm/yr)</b>	<b>Average (mm/yr)</b>	<b>References</b>
Soufrière Hills Montserrat, W.I.	1250	2000	1625	(Barclay et al. 2006; Hemmings et al. 2015)
Soufrière of Guadeloupe Guadeloupe, WI	??	10000	10000	(Le Friant et al. 2004)
Merapi Central Java, Indonesia	2000	4500	3250	(Lavigne et al. 2000)
Casita Nicaragua			1250	(Velázquez and Gómez-Sal 2007)
Unzen Japan	2000	2600	2300	(Ogawa et al. 2007)
Galeras Columbia			1200	Meteorologia Aeronautica (Instituto de Hidrologia, Meteorologia y Estudios Ambientales) di Columbia ( <a href="http://www.meteoaeronautica.gov.co/">http://www.meteoaeronautica.gov.co/</a> )
Santiaguito Guatemala	1800	4000	2900	(Lopez 2004)

**Table 4.** *Individual simulation parameters*

<b>Run</b>	<b>Geometry</b>	<b>Permeability Dome Talus (Zone 2)</b>	<b>Permeability Dome Core (Zone 1)</b>	<b>Permeability Slope Talus (Zone 3)</b>	<b>Permeability Substrate (Zone 5)</b>	<b>Conduit Heat</b>
1	Crater- confined	$1 \times 10^{-12} \text{ m}^2$	$1 \times 10^{-13} \text{ m}^2$	$1 \times 10^{-12} \text{ m}^2$	$1 \times 10^{-13} \text{ m}^2$	Initial heat only
2		$1 \times 10^{-13} \text{ m}^2$	$1 \times 10^{-14} \text{ m}^2$	$1 \times 10^{-13} \text{ m}^2$	$1 \times 10^{-14} \text{ m}^2$	
3		$1 \times 10^{-14} \text{ m}^2$	$1 \times 10^{-15} \text{ m}^2$	$1 \times 10^{-14} \text{ m}^2$	$1 \times 10^{-15} \text{ m}^2$	
4	Crater- confined	$1 \times 10^{-12} \text{ m}^2$	$1 \times 10^{-13} \text{ m}^2$	$1 \times 10^{-12} \text{ m}^2$	$1 \times 10^{-13} \text{ m}^2$	200°C sustained
5		$1 \times 10^{-13} \text{ m}^2$	$1 \times 10^{-14} \text{ m}^2$	$1 \times 10^{-13} \text{ m}^2$	$1 \times 10^{-14} \text{ m}^2$	
6		$1 \times 10^{-14} \text{ m}^2$	$1 \times 10^{-15} \text{ m}^2$	$1 \times 10^{-14} \text{ m}^2$	$1 \times 10^{-15} \text{ m}^2$	
7	Perched	$1 \times 10^{-12} \text{ m}^2$	$1 \times 10^{-13} \text{ m}^2$	$1 \times 10^{-12} \text{ m}^2$	$1 \times 10^{-13} \text{ m}^2$	Initial heat only
8		$1 \times 10^{-13} \text{ m}^2$	$1 \times 10^{-14} \text{ m}^2$	$1 \times 10^{-13} \text{ m}^2$	$1 \times 10^{-14} \text{ m}^2$	
9		$1 \times 10^{-14} \text{ m}^2$	$1 \times 10^{-15} \text{ m}^2$	$1 \times 10^{-14} \text{ m}^2$	$1 \times 10^{-15} \text{ m}^2$	
10	Perched	$1 \times 10^{-12} \text{ m}^2$	$1 \times 10^{-13} \text{ m}^2$	$1 \times 10^{-12} \text{ m}^2$	$1 \times 10^{-13} \text{ m}^2$	200°C sustained
11		$1 \times 10^{-13} \text{ m}^2$	$1 \times 10^{-14} \text{ m}^2$	$1 \times 10^{-13} \text{ m}^2$	$1 \times 10^{-14} \text{ m}^2$	
12		$1 \times 10^{-14} \text{ m}^2$	$1 \times 10^{-15} \text{ m}^2$	$1 \times 10^{-14} \text{ m}^2$	$1 \times 10^{-15} \text{ m}^2$	

Figure 1

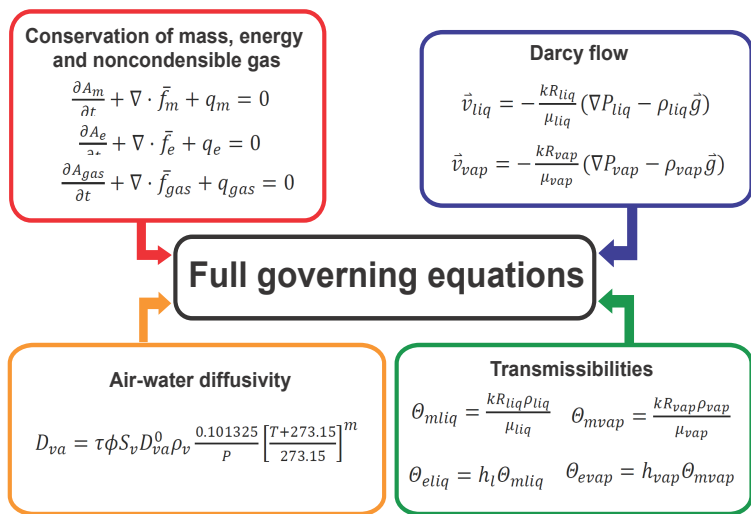


Figure 2

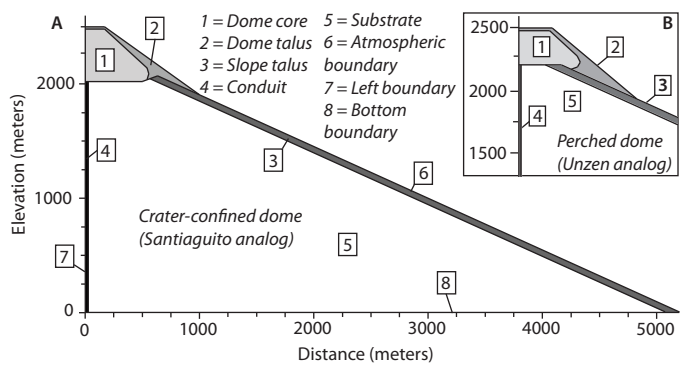




Figure 3

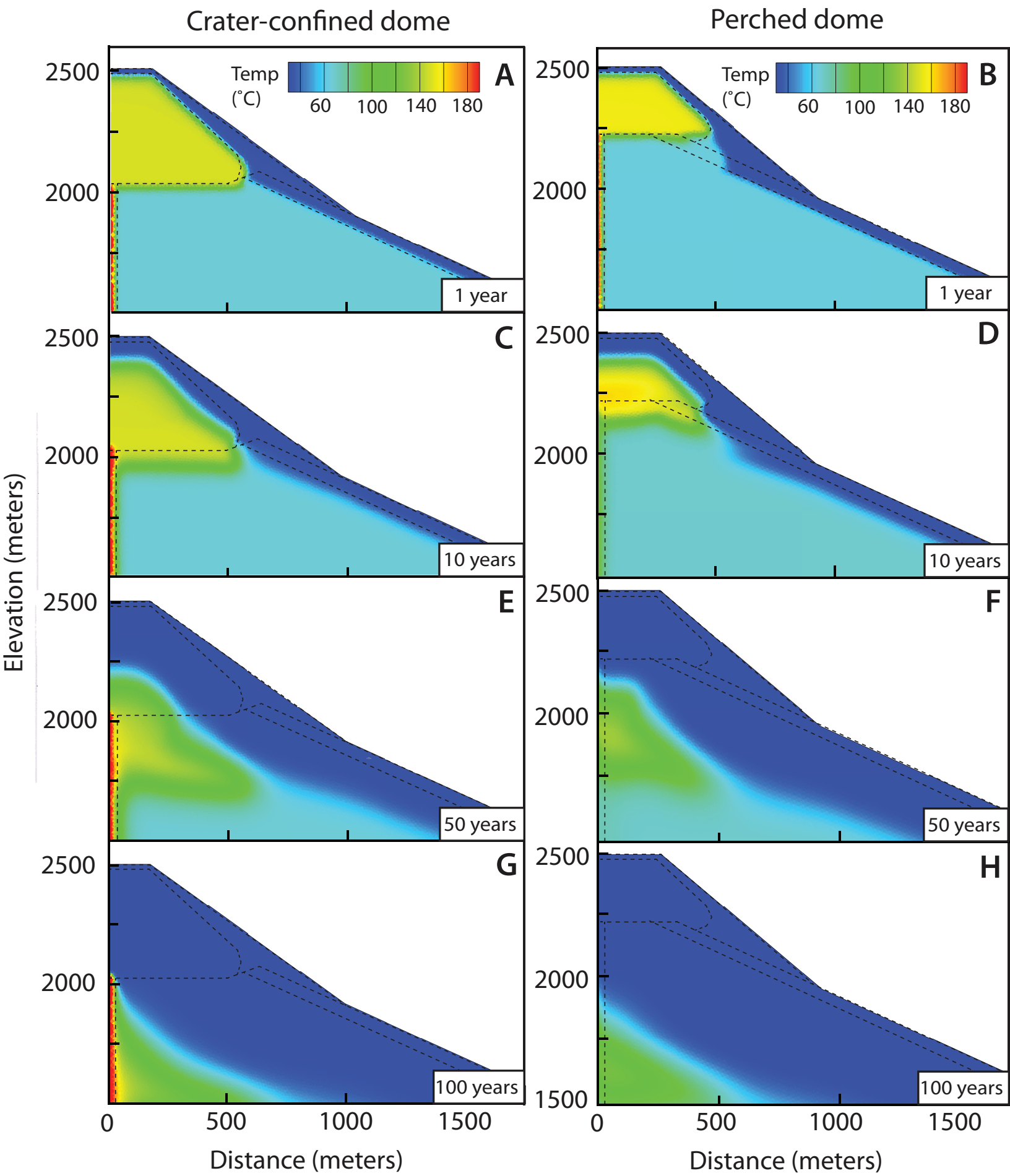


Figure 4

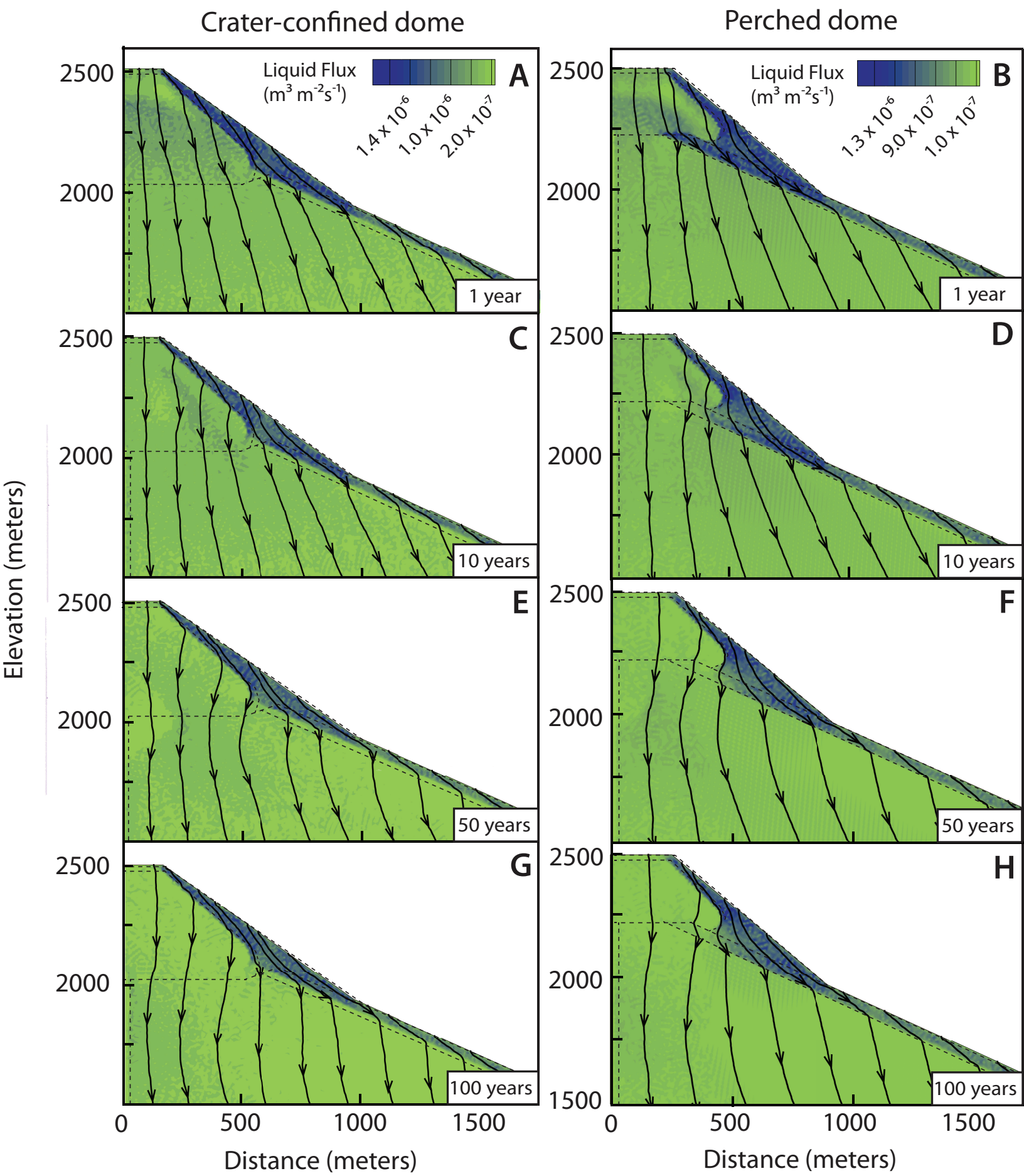


Figure 5

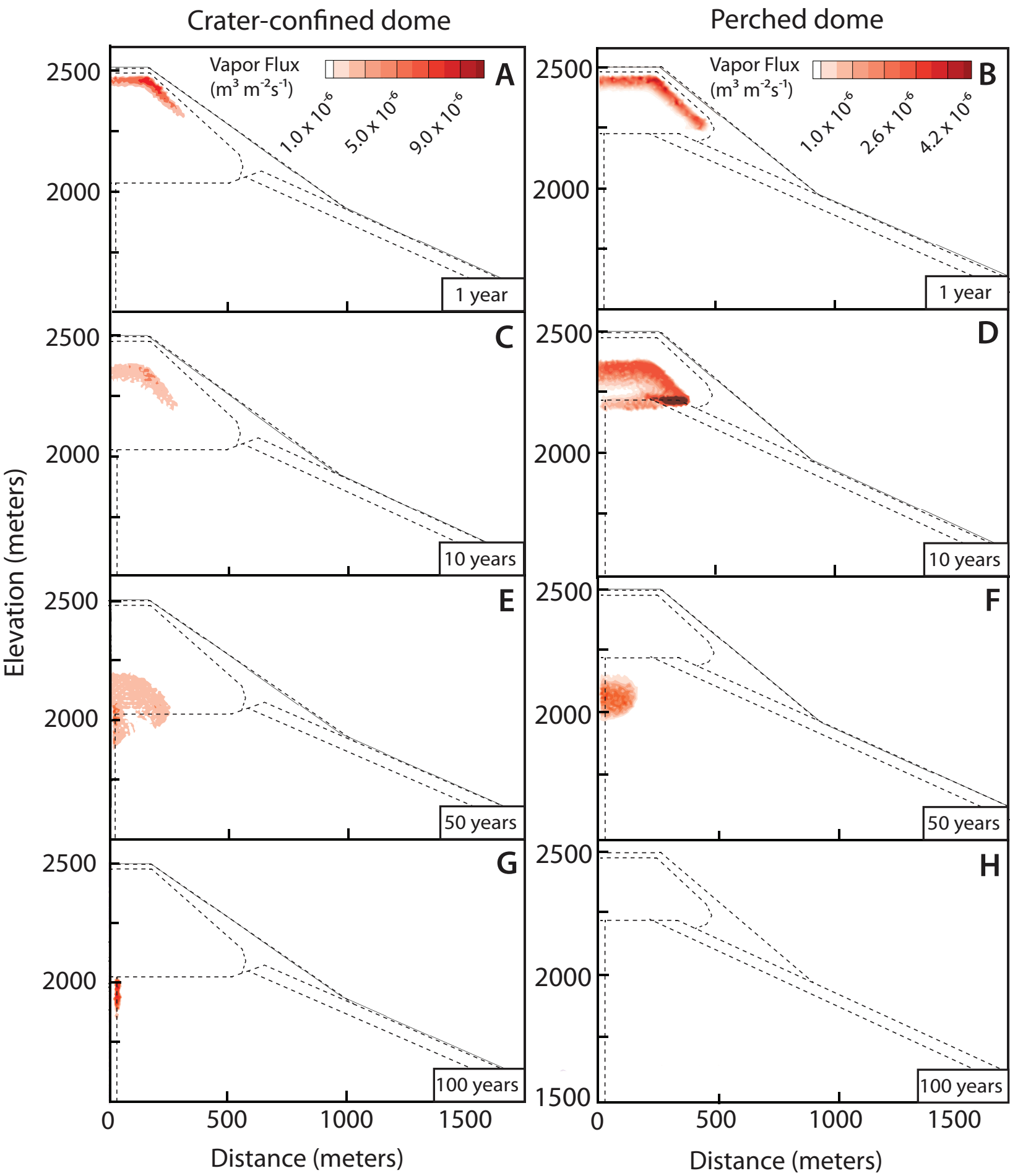


Figure 6

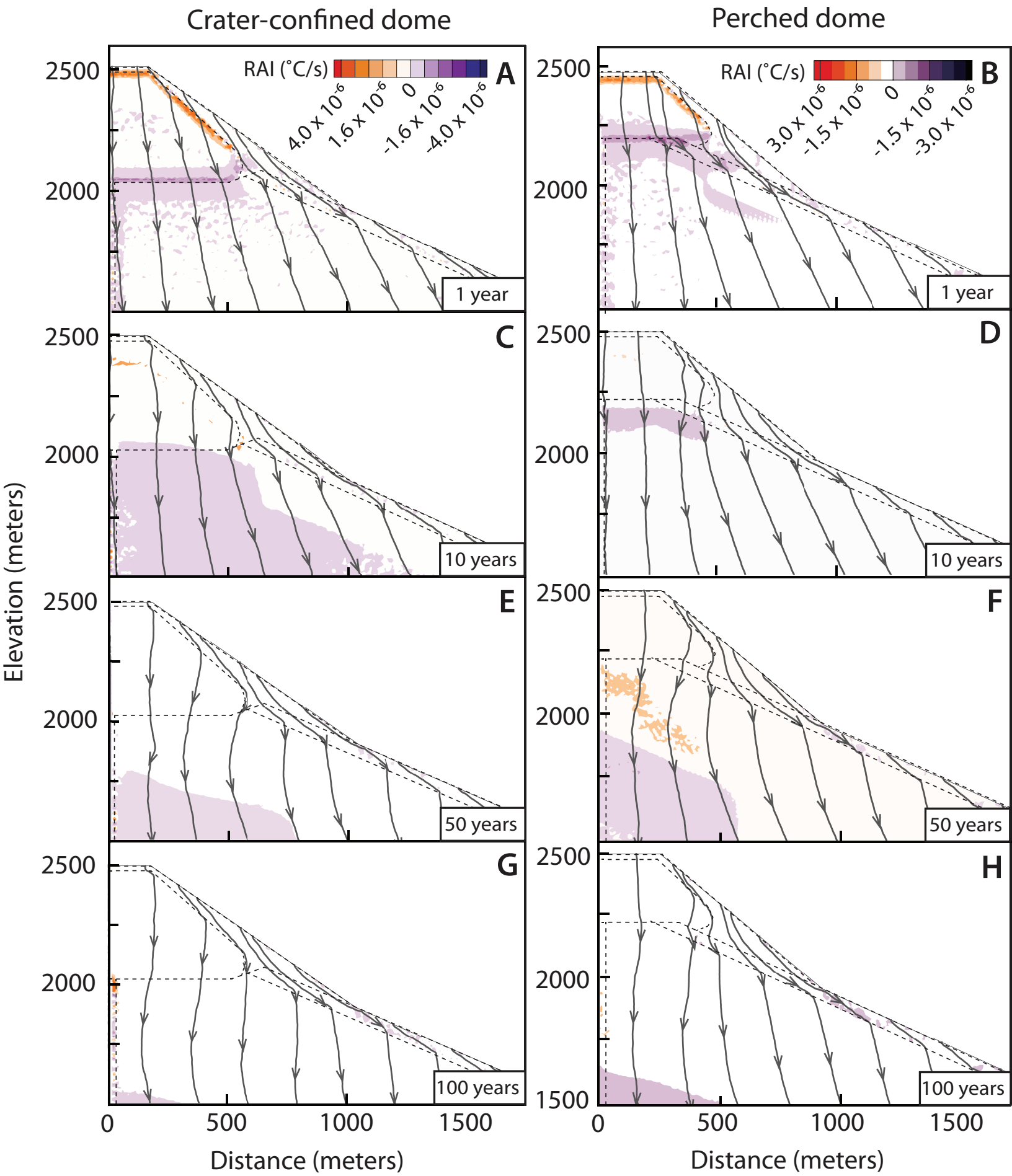
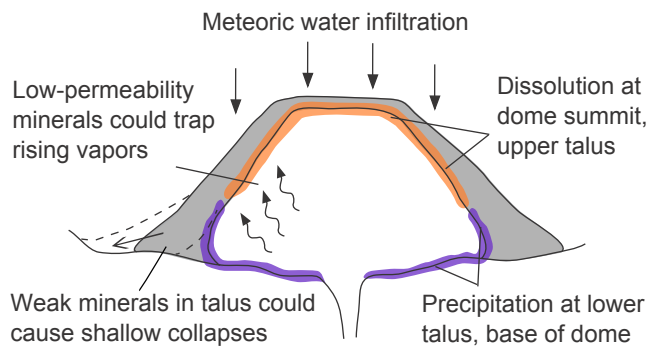


Figure 7



1 **Governing equations**

2           In FEHM, the governing partial differential equations for mass and heat transfer are  
3 discretized into a system of nonlinear algebraic equations, which are then solved using the  
4 Newton-Raphson iteration method (a way to find successively better approximations to the roots  
5 or zeroes of a real-valued function from an initial guess). In FEHM (as opposed to some other  
6 multi-physics computer codes like TOUGH2 (Pruess et al. 2012), the Newton-Raphson  
7 derivatives of the thermodynamic functions with respect to pressure and temperature are formed  
8 analytically rather than numerically in order to achieve faster convergence of the nonlinear  
9 system of equations (Zyvoloski 2007). FEHM equations of state are nonlinear because the  
10 porosity, permeability, density, enthalpy and viscosity are strong functions of pressure and  
11 temperature; in addition, relative permeabilities and capillary pressure can also be strong  
12 functions of saturation, which varies significantly with temperature and pressure. Pressure and  
13 temperature dependent behavior of density, enthalpy and viscosity are represented by rational  
14 polynomials derived from National Bureau of Standards (NBS) steam table data (Zyvoloski et al.  
15 1991).

16           All variables referenced in these equations are defined in Table 5. A more detailed  
17 breakdown of the equation derivation may be found in Zyvoloski et al. (1999) and Keating (  
18 2000).

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23 **Table 5.** Variables used in FEHM governing equations

$A_{e,m,gas}$	Energy/mass/noncondensable gas accumulation terms ( $\text{kg m}^{-1} \text{s}^{-2}$ )/ ( $\text{kg m}^{-3}$ ) / ( $\text{kg m}^{-3}$ )	$R$	Relative permeability
$C_p$	Specific heat ( $\text{m}^2 \text{s}^{-2} \text{ }^\circ\text{C}^{-1}$ )	$S$	Saturation
$f_{e,m,gas}$	flux vectors for energy/mass/noncondensable gas equation ( $\text{kg s}^{-3}$ )/ ( $\text{kg m}^{-2} \text{s}^{-1}$ )/ ( $\text{kg m}^{-2} \text{s}^{-1}$ )	$T$	Temperature
$g$	gravitational acceleration ( $\text{m s}^{-2}$ )	$t$	time
$h$	enthalpy ( $\text{m}^2 \text{s}^{-2}$ )	$v$	Velocity ( $\text{m s}^{-1}$ )
$k$	intrinsic rock permeability ( $\text{m}^2$ )	$\phi$	Porosity
$K$	Thermal conductivity ( $\text{kg m T}^{-1} \text{s}^{-3}$ )	$\eta_{vap,liq}$	Mass fraction of vapor/liquid
$P$	Pressure ( $\text{kg m}^{-1} \text{s}^{-2}$ )	$\mu$	Viscosity ( $\text{kg m}^{-1} \text{s}^{-1}$ )
$q_e$	Energy/mass/noncondensable gas source term ( $\text{kg m}^{-1} \text{s}^{-3}$ )/ ( $\text{kg m}^{-3} \text{s}^{-1}$ )/ ( $\text{kg m}^{-3} \text{s}^{-1}$ )	$\rho$	Density
<b>Subscripts:</b> <i>gas</i> = noncondensable gas, <i>vap</i> = vapor, <i>liq</i> = liquid, <i>f</i> = fracture, <i>r</i> = rock			

24

25 Conservation of mass for water is

26 
$$\frac{\partial A_m}{\partial t} + \bar{\nabla} \cdot \bar{f}_m + q_m = 0 \quad (1)$$

27 where the mass per unit volume  $A_m$  (a mass accumulation term) is a function of the porosity, the  
 28 saturation and the mass fraction of each phase:

29 
$$A_m = \phi(S_{vap}\rho_{vap}(1 - \eta_{vap}) + S_{liq}\rho_{liq}(1 - \eta_{liq})) \quad (2)$$

30 the mass flux is

31 
$$\bar{f}_m = (1 - \eta_{vap})\rho_{vap}\bar{v}_{vap} + (1 - \eta_{liq})\rho_{liq}\bar{v}_{liq} \quad (3)$$

32 and sources/sinks of mass are contained in the term  $q_m$ . The total of all water flux through the  
 33 system, water accumulated in the system, and sources/sinks of water are assumed to equal zero.

34 Conservation of fluid-rock energy is

35 
$$\frac{\partial A_e}{\partial t} + \bar{\nabla} \cdot \bar{f}_e + q_e = 0 \quad (4)$$

36 where the energy per unit volume (energy accumulation term)  $A_e$  is a combination of energy  
 37 transferred by the rock itself, energy transferred by water vapor and energy transferred by liquid  
 38 water:

$$39 \quad A_e = (1 - \phi)\rho_f C_{pr}T + \phi(S_{vap}\rho_{vap}C_{pvap}T + S_{liq}\rho_{liq}C_{pliq}T) \quad (5)$$

40 The energy flux,  $\bar{f}_e$ , can be stated either as a sum of the products of density, specific enthalpy  
 41 and velocity of the vapor and liquid, or the product of the thermal conductivity and temperature  
 42 gradients:

$$43 \quad \bar{f}_e = \rho_{vap}h_{vap}v_{vap} + \rho_{liq}h_{liq}v_{liq} = K\bar{\nabla}T \quad (6)$$

44 The conservation of noncondensable gas is described by

$$45 \quad \frac{\partial A_{gas}}{\partial t} + \bar{\nabla} \cdot \bar{f}_{gas} + q_{gas} = 0 \quad (7)$$

46 where the accumulation term  $A_{gas}$  is

$$47 \quad A_{gas} = \phi(\eta_{vap}S_{vap}\rho_{vap} + \eta_{liq}S_{liq}\rho_{liq}) \quad (8)$$

48 the gas flux is

$$49 \quad \bar{f}_{gas} = \eta_{vap}\rho_{vap}\bar{v}_{vap} + \eta_{liq}\rho_{liq}\bar{v}_{liq} \quad (9)$$

50 and the source and sink term  $q_{gas}$  is

$$51 \quad q_{gas} = \eta_{vap}q_{vap} + \eta_{liq}q_{liq} \quad (10)$$

52 It is assumed that Darcy's law applies to the movement of each phase; in this formulation, the  
 53 hydraulic conductivity term is replaced by the quotient of the intrinsic and relative permeabilities  
 54 divided by the viscosity of the phase, and the hydraulic head term is replaced with the difference  
 55 between the pressure gradient and lithostatic gradient:

$$56 \quad \bar{v}_{vap} = -\frac{kR_{vap}}{\mu_{vap}}(\bar{\nabla}P_{vap} - \rho_{vap}\bar{g}) \quad (11)$$

$$57 \quad \bar{v}_{liq} = -\frac{kR_{liq}}{\mu_{liq}}(\bar{\nabla}P_{liq} - \rho_{liq}\bar{g}) \quad (12)$$



58 By combining equations 1 – 10 with Darcy's law (11 & 12), the full governing equations are  
 59 derived for mass,

$$60 \quad -\bar{\nabla} \cdot \left( (1 - \eta_{vap}) \Theta_{mvap} \bar{\nabla} P_{vap} \right) - \bar{\nabla} \cdot \left( (1 - \eta_{liq}) \Theta_{mliq} \bar{\nabla} P_{liq} \right) + q_m +$$

$$61 \quad \frac{\partial}{\partial z} g \left( (1 - \eta_{vap}) \Theta_{mvap} \rho_{vap} + (1 - \eta_{liq}) \Theta_{mliq} \rho_{liq} \right) + \frac{\partial A_m}{\partial t} = 0 \quad (13)$$

62 Energy,

$$63 \quad -\bar{\nabla} \cdot \left( \Theta_{evap} \bar{\nabla} P_{vap} \right) - \bar{\nabla} \cdot \left( \Theta_{eliq} \bar{\nabla} P_{liq} \right) + q_e + \frac{\partial}{\partial z} g \left( \Theta_{evap} \rho_{vap} + \Theta_{eliq} \rho_{liq} \right) + \frac{\partial A_{ngas}}{\partial t} = 0$$

$$64 \quad (14)$$

65 and noncondensable gas:

$$66 \quad -\bar{\nabla} \cdot \left( \eta_{vap} \Theta_{mvap} \bar{\nabla} P_{vap} \right) - \bar{\nabla} \cdot \left( \eta_{liq} \Theta_{mliq} \bar{\nabla} P_{liq} \right) + \bar{\nabla} \cdot \left( D_{va} \bar{\nabla} \eta_{vap} \right) + q_{ngas} +$$

$$67 \quad \frac{\partial}{\partial z} g \left( \eta_{vap} \Theta_{mvap} \rho_{vap} + \eta_{liq} \Theta_{mliq} \rho_{liq} \right) + \frac{\partial A_{ngas}}{\partial t} = 0 \quad (15)$$

68 Transmissibilities, a measure of how easily a given layer transmits water, are represented by  $\Theta$  in  
 69 the equations and are given by

$$70 \quad \Theta_{mvap} = \frac{kR_{vap} \rho_{vap}}{\mu_{vap}} \quad (16)$$

$$71 \quad \Theta_{mliq} = \frac{kR_{liq} \rho_{liq}}{\mu_{liq}} \quad (17)$$

$$72 \quad \Theta_{evap} = h_{vap} \Theta_{mvap} \quad (18)$$

$$73 \quad \Theta_{eliq} = h_{liq} \Theta_{mliq} \quad (19)$$

74 The air/water diffusivity term  $D_{va}$  is given by

$$75 \quad D_{va} = \tau \phi S_{vap} D_{va}^0 \rho_{vap} \frac{0.101325}{P} \left[ \frac{T+273.15}{273.15} \right]^m \quad (20)$$

76 where  $\tau$  is the tortuosity factor (for diffusion in porous media, a measure of how circuitous a  
 77 typical flow path is through a medium),  $D$  (standard conditions) = 2.4e-5 m<sup>2</sup>/s and  $m$  = 2.334  
 78 (Zyvoloski et al. 1997). Transmissibilities are given by

79 
$$\theta_{mvap} = \frac{kR_{vap}\rho_{vap}}{\mu_{vap}} \quad (21)$$

80 
$$\theta_{mliq} = \frac{kR_{liq}\rho_{liq}}{\mu_{liq}} \quad (22)$$

81 
$$\theta_{evap} = h_{vap}\theta_{mvap} \quad (23)$$

82 
$$\theta_{eliq} = h_{liq}\theta_{mliq} \quad (24)$$

83 The nonisothermal multiphase transport model in FEHM is formed by equation 13, 14 and 15,  
 84 setting  $n = 0$  in Equation 13 and leaving out Equation 15 in the pure-water model.

85

86 **Relative permeability**

87 In cases where the system being modeled in FEHM does not remain fully saturated,  
 88 FEHM is capable of calculating relative permeability using either a linear, Corey-Brooks  
 89 (Brooks and Corey 1964), or van Genuchten formulation (van Genuchten 1980). The linear  
 90 formulation used in this study depends only on the residual liquid and vapor saturations and was  
 91 chosen for simplicity. Linear functions for relative permeabilities of liquid (the wetting fluid, in  
 92 this case water) and vapor (the ‘nonwetting fluid’) are defined by the equations

93 
$$R_l = \begin{cases} 0, & S_l \leq S_{lr} \\ \frac{S_l - S_{lr}}{S_{lmax} - S_{lr}} & S_{lr} < S_l < S_{lmax} \\ 1, & S_l \geq S_{lmax} \end{cases} \quad (27)$$

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$$R_v = \begin{cases} 0, & S_v \leq S_{vr} \\ \frac{S_v - S_{vr}}{S_{vmax} - S_{vr}} & S_{vr} < S_v < S_{vmax} \\ 1, & S_v \geq S_{vmax} \end{cases} \quad (28)$$

95 where  $S_l$  and  $S_v$  are liquid and vapor saturations,  $S_{lr}$  and  $S_{vr}$  are the residual liquid and vapor  
 96 saturations, and  $S_{lmax}$  and  $S_{vmax}$  are the maximum liquid and vapor saturations.

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99 **Table 6. Material properties of andesite and dacite lavas and pyroclastic deposits**

Reference	Location	Sample size/type	Rock type	Density (kg/m <sup>3</sup> )	Porosity (φ)	Permeability (m <sup>2</sup> )	Thermal conductivity (W/mK)	Specific heat (J/kgK)
(Aizawa et al. 2009)	Conceptual	Unknown	Sealing zone		0.01	3.00E-16		
		Unknown	Basement		0.05	3.00E-16		
		Unknown	Hydrothermal zone		0.1	3.00E-15		
		Unknown	Fractured rock surrounding hydrothermal zone		0.3	1.00E-13		
		Unknown	Conduit		0.1	1.00E-12		
		Unknown	Fresh volcanic rock		0.3	1E-13 - 2E-14		
(Bartetzko et al. 2006)	Undersea basin, PACMANUS field	Drill core	Dacite		0.22	1.25E-17		
		Drill core	Dacite		0.32	2.23E-17		
		Drill core	Dacite		0.24	4.48E-17		
		Drill core	Dacite		0.3	1.04E-16		
		Drill core	Dacite		0.24	4.46E-16		
		Drill core	Dacite		0.38	7.59E-16		
		Drill core	Dacite		0.21	1.50E-15		
		Drill core	Dacite		0.43	2.00E-15		
		Drill core	Dacite		0.01	1.17E-14		
		Drill core	Dacite		0.16			
		Drill core	Dacite		0.17			
		Drill core	Dacite		0.17			
		Drill core	Dacite		0.2			
		Drill core	Dacite		0.21			
		Drill core	Dacite		0.22			
(Bernard et al. 2007)	Mount Pelée	Hand sample	MB502	2700	0.038	1.00E-16		
		Hand sample	LPP Calebasse	2850	0.031	1.00E-15		
		Hand sample	LPP Plume	2890	0.035	1.00E-15		
		Hand sample	andesite, B&A flows	2780	0.099	1.00E-15		
		Hand sample	MB501	2740	0.102	1.00E-15		
		Hand sample	LPP Macouba	2760	0.134	1.00E-15		
		Hand sample	andesite dome	2720	0.098	2.00E-15		
		Hand sample	andesite, nuee ardente flows	2690	0.119	2.90E-15		
		Hand sample	andesite, indurated block and ash flows	2720	0.103	3.00E-15		
		Hand sample	andesite, indurated block and ash flows	2690	0.146	3.00E-15		
		Hand sample	andesite, indurated block and ash flows	2720	0.111	3.90E-15		
		Hand sample	andesite dome	2740	0.121	3.90E-15		
		Hand sample	andesite dome	2740	0.154	3.90E-15		

Hand sample	andesite, nuee ardente flows	2670	0.157	3.90E-15
Hand sample	andesite, indurated block and ash flows	2670	0.353	3.90E-15
Hand sample	andesite, indurated block and ash flows	2770	0.145	4.90E-15
Hand sample	andesite, nuee ardente flows	2670	0.169	5.90E-15
Hand sample	andesite, indurated block and ash flows	2710	0.132	8.90E-15
Hand sample	andesite, nuee ardente flows	2670	0.19	1.09E-14
Hand sample	andesite, nuee ardente flows	2670	0.099	1.88E-14
Hand sample	andesite, nuee ardente flows	2690	0.158	3.85E-14
Hand sample	andesite, indurated block and ash flows	2720	0.149	5.53E-14
Hand sample	andesite, nuee ardente flows		0.312	5.92E-14
Hand sample	andesite, nuee ardente flows	2670	0.174	1.09E-13
Hand sample	andesite, ash-and-pumice flows	2650	0.574	1.44E-13
Hand sample	andesite, nuee ardente flows	2700	0.197	1.67E-13
Hand sample	andesite, nuee ardente flows	2690	0.146	2.04E-13
Hand sample	andesite, ash-and-pumice flows	2630	0.547	2.07E-13
Hand sample	andesite, indurated block and ash flows	2620	0.213	2.27E-13
Hand sample	andesite, indurated block and ash flows	2760	0.24	2.54E-13
Hand sample	andesite, ash-and-pumice flows	2650	0.536	4.25E-13
Hand sample	andesite, indurated block and ash flows	2620	0.232	5.54E-13
Hand sample	andesite, indurated block and ash flows	2630	0.321	5.70E-13
Hand sample	andesite, nuee ardente flows	2670	0.415	6.70E-13
Hand sample	andesite, scoria flows	2810	0.28	1.03E-12
Hand sample	andesite, ash-and-pumice flows	2670	0.574	1.18E-12
Hand sample	andesite, nuee ardente flows	2700	0.304	1.18E-12
Hand sample	andesite, nuee ardente flows	2670	0.428	1.45E-12
Hand sample	andesite, scoria flows	2860	0.363	1.54E-12
Hand sample	andesite, nuee ardente flows	2690	0.251	1.58E-12
Hand sample	andesite, indurated block and ash flows	2680	0.305	1.77E-12
Hand sample	andesite, ash-and-pumice flows	2670	0.578	2.05E-12
Hand sample	andesite, nuee ardente flows	2680	0.274	2.91E-12
Hand sample	andesite, indurated block and ash flows	2710	0.289	5.06E-12
Hand sample	andesite, scoria flows	2840	0.352	5.69E-12
Hand sample	andesite, ash-and-pumice flows	2650	0.617	7.66E-12
Hand sample	andesite, block and ash flows	2710	0.328	1.02E-11
Hand sample	andesite, block and ash flows	2670	0.408	3.44E-11

		Hand sample	andesite, ash-and-pumice flows	2670	0.585			
(Flint et al. 2006)	Yucca Mountain	Drill core	Pyroclastic unit	1490	0.341			
		Drill core	Pyroclastic unit	1600	0.322			
(García et al. 1989)	Los Azufres	Drill core	Andesite	2053	0.02		1.05	
		Drill core	Andesite	2737	0.24		2.34	
(Hicks et al. 2009)	Soufriere Hills	Hand sample	Andesite	2600	0.2	1.00E-10		
(Hurwitz et al. 2003)	Cascades	Conceptual	Conduit (numerical model)	2500	0.15		2	
		Conceptual	Upper unit (numerical model)	2500			2	
		Conceptual	Basal unit	2500	0.01		2.5	
(Ikeda et al. 2008)	Unzen	Drill core	Volcanic breccia	2400	0.15			
		Drill core	Lava dike	2500	0.15			
		Drill core	Lava dike	2600	0.15			
		Drill core	Volcanic breccia	2500	0.3			
(Ingebritsen et al. 1994)	Cascades	Conceptual	Lava flows and domes younger than 2.3 Ma		0.15	1.00E-14	1.55	
		Conceptual	Lava flows and minor pyroclastic rocks from 4-8 Ma		0.1	5.00E-16	1.55	
		Conceptual	Lava flows from 8 to 17 Ma		0.05	1.00E-16	1.65	
		Conceptual	Chiefly volcanic and volcanoclastic strata from 18-25 Ma		0.05	1.00E-17	2	
		Conceptual	Quartz-bearing ash flow tuff		0.02	2.50E-14	2	
(Keating 2005)	Mount St. Helens	Hand sample	Dome rock	2200	0.6	2.10E-14	0.9	1557
		Hand sample	Pyroclastic flow deposits (substrate)	2200	0.25	1.00E-15	0.35	2036
		Hand sample	Pyroclastic flow deposits	2200	0.62	4.66E-14	0.35	2036
(Mueller et al. 2008)	Unzen	Hand sample	Dacite, breadcrust bomb		0.367	8.90E-14		
		Hand sample	Dacite, breadcrust bomb		0.475	1.47E-13		
		Hand sample	Dacite, dome rock		0.343	9.99E-13		
		Hand sample	Dacite, dome rock		0.349	3.41E-12		
		Hand sample	Dacite, dome rock		0.412	4.50E-12		
(Platz et al. 2012)	Mt. Taranaki	Cores from clasts	andesite dome	2555		6.80E-13	14.47	918
(Reid 2004)	Cascades volcanoes	"typical of volcanic rocks"	Unknown	2650			2	1000
(Sammel et al. 1988)	Newberry	Conceptual	Fill			1E-14 - 5E-12		
		Conceptual	Dikes and pipe			5.00E-15		
		Conceptual	Flow 1			1E-15 - 1E-13		
		Conceptual	Flow 2			1E-16 - 1E-14		
		Conceptual	Flow 3			1E-17 - 1E-15		
		Conceptual	Flow 4			5E-16 - 1E-13		
		Conceptual	Flow 5			1E-18 - 1E-16		
		Conceptual	Magma			1.00E-18		

(Scheu et al. 2006)	Unzen	Hand sample	Dacite	2490	0.041		
		Hand sample	Dacite	2420	0.073		
		Hand sample	Dacite	2280	0.12		
		Hand sample	Dacite	2180	0.163		
		Hand sample	Dacite	2100	0.193		
		Hand sample	Dacite	1930	0.259		
(Sekioka 1988)	Japanese geothermal fields	Calculated	Unknown			1.10E-12	2.04
		Calculated	Unknown			3.50E-13	2.33
		Calculated	Unknown			1.90E-12	3.43
		Calculated	Unknown			2.10E-12	2.87
		Calculated	Unknown			2.00E-12	1.49
		Calculated	Unknown			5.00E-14	0.93
		Calculated	Unknown			1.70E-12	1.65
(Smith et al. 2001)	Mount St. Helens	Cores from spine samples	dacite	2460	0.08		
		Cores from spine samples	dacite	2390	0.095		
		Cores from spine samples	dacite	2350	0.103		
		Cores from spine samples	dacite	2260	0.121		
		Cores from spine samples	dacite	2040	0.197		
(Watanabe et al. 2008)	Unzen	Drill core	C1-5-7, Brecciated dacite	2410	0.08	1.00E-19	
		Drill core	C14, Dacite dike	2500	0.16	1.00E-19	
		Drill core	C12, Volcanic breccia	2400	0.02	1.00E-17	
		Drill core	C14-2, Dacite dike	2560	0.04	1.00E-17	
		Drill core	C13, Dacite dike	2570	0.08	1.00E-17	

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- 105 Aizawa K, Ogawa Y, Ishido T (2009) Groundwater flow and hydrothermal systems within  
106 volcanic edifices: Delineation by electric self-potential and magnetotellurics. *J Geophys Res*  
107 114:1–12. doi: 10.1029/2008JB005910
- 108 Bartetzko A, Klitzsch N, Iturrino G, et al (2006) Electrical properties of hydrothermally altered  
109 dacite from the PACMANUS hydrothermal field (ODP Leg 193). *J Volcanol Geotherm Res*  
110 152:109–120. doi: 10.1016/j.jvolgeores.2005.10.002
- 111 Bernard ML, Zamora M, Géraud Y, Boudon G (2007) Transport properties of pyroclastic rocks  
112 from Montagne Pelée volcano (Martinique, Lesser Antilles). *J Geophys Res Solid Earth*  
113 112:1–16. doi: 10.1029/2006JB004385
- 114 Brooks RH, Corey a T (1964) Hydraulic properties of porous media. *Hydrol Pap Fort Collins CO*  
115 *Colo State Univ* 3:27 pgs.
- 116 Flint LE, Buesch DC, Flint AL (2006) Characterization of Unsaturated Zone Hydrogeologic  
117 Units using Matrix Properties and Depositional History in a Complex Volcanic  
118 Environment. *Vadose Zo J* 5:480. doi: 10.2136/vzj2004.0180
- 119 García A, Contreras E, Viggiano JC (1989) Establishment of an empirical correlation for  
120 estimating the thermal conductivity of igneous rocks. *Int J Thermophys.* doi:  
121 10.1007/BF00503174
- 122 Hicks PD, Matthews AJ, Cooker MJ (2009) Thermal structure of a gas-permeable lava dome and  
123 timescale separation in its response to perturbation. *J Geophys Res* 114:B07201. doi:  
124 10.1029/2008JB006198
- 125 Hurwitz S, Kipp KL, Ingebritsen SE, Reid ME (2003) Groundwater flow, heat transport, and  
126 water table position within volcanic edifices: Implications for volcanic processes in the  
127 Cascade Range. *J Geophys Res* 108:1–19. doi: 10.1029/2003JB002565
- 128 Ikeda R, Kajiwara T, Omura K, Hickman S (2008) Physical rock properties in and around a  
129 conduit zone by well-logging in the Unzen Scientific Drilling Project, Japan. *J Volcanol*  
130 *Geotherm Res* 175:13–19. doi: 10.1016/j.jvolgeores.2008.03.036
- 131 Ingebritsen SE, Mariner RH, Sherrod DR (1994) Hydrothermal systems of the Cascade Range,  
132 north-central Oregon.
- 133 Keating GN (2000) Multiphase thermal modeling in volcanic and contact metamorphic terranes.  
134 242.
- 135 Keating GN (2005) The role of water in cooling ignimbrites. *J Volcanol Geotherm Res* 142:145–  
136 171. doi: 10.1016/j.jvolgeores.2004.10.019
- 137 Mueller S, Scheu B, Spieler O, Dingwell DB (2008) Permeability control on magma  
138 fragmentation. *Geology.* doi: 10.1130/G24605A.1

- 139 Platz T, Cronin SJ, Procter JN, et al (2012) Non-explosive, dome-forming eruptions at Mt.  
140 Taranaki, New Zealand. *Geomorphology* 136:15–30. doi: 10.1016/j.geomorph.2011.06.016
- 141 Pruess K, Oldenburg C, Moridis G (2012) TOUGH2 User's Guide, Version 2. LBNL-  
142 43134:197.
- 143 Reid ME (2004) Massive collapse of volcano edifices triggered by hydrothermal pressurization.  
144 *Geology* 32:373–376. doi: 10.1130/G20300.1
- 145 Sammel EA, Ingebritsen SE, Mariner RH (1988) The hydrothermal system at Newberry  
146 Volcano, Oregon. *J Geophys Res* 93:10,149–10,162.
- 147 Scheu B, Spieler O, Dingwell DB (2006) Dynamics of explosive volcanism at Unzen volcano: an  
148 experimental contribution. *Bull Volcanol* 69:175–187. doi: 10.1007/s00445-006-0066-5
- 149 Sekioka M (1988) Tentative estimate of bulk permeability of basement rocks from heat  
150 discharges in a geothermal field. *J Volcanol Geotherm Res.* doi: 10.1016/0377-  
151 0273(88)90006-6
- 152 Smith J V, Miyake Y, Oikawa T (2001) Interpretation of porosity in dacite lava domes as  
153 ductile-brittle failure textures. *J Volcanol Geotherm Res* 112:25–35. doi: 10.1016/S0377-  
154 0273(01)00232-3
- 155 Van Genuchten MT (1980) A Closed-form Equation for Predicting the Hydraulic Conductivity  
156 of Unsaturated Soils1. *Soil Sci Soc Am J* 44:892. doi:  
157 10.2136/sssaj1980.03615995004400050002x
- 158 Watanabe T, Shimizu Y, Noguchi S, Nakada S (2008) Permeability measurements on rock  
159 samples from Unzen Scientific Drilling Project Drill Hole 4 (USDP-4). *J Volcanol*  
160 *Geotherm Res* 175:82–90. doi: 10.1016/j.jvolgeores.2008.03.021
- 161 Zyvoloski G (2007) FEHM: A control volume finite element code for simulating subsurface  
162 multi-phase multi-fluid heat and mass transfer. Los Alamos Unclassif. Rep. LA-UR-07-  
163 3359
- 164 Zyvoloski G, Dash Z, Kelkar S (1991) FEHMN 1.0 : Finite Element Heat and Mass Transfer  
165 Code. Los Alamos Natl. Lab. Publ. LA-12062-MS
- 166 Zyvoloski GA, Robinson BA, Dash Z V., Trease LL (1999) Models and methods summary for  
167 the FEHM application. Los Alamos Natl. Laboratory Publ. SC-194
- 168