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during four historic caldera-forming events

Magma dynamics and collapse mechanisms

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16 Abstract

Four historic caldera-forming events were studied to understand the relationship of 17 magma withdrawal processes and caldera subsidence mechanisms. Two calderas are 18 19 silicic (Katmai 1912, Pinatubo 1991) and two are basaltic (Fernandina 1968, Miyakejima 2000). All events have sufficient geophysical, geologic, and petrologic data with which to 20 21 examine and model magma withdrawal and caldera collapse. The data reveal that the 22 magmas erupted at Katmai and Pinatubo were in a bubbly state in the reservoir immediately before and during caldera collapse. The bubbly magma allowed for its 23 efficient extraction from the reservoir, causing significant underpressures to develop 24 rapidly, particularly in the case of Katmai where the erupted rhyolite was voluminous, 25 nearly aphyric, and very low viscosity. The rapidly developing underpressures at Katmai 26 and Pinatubo caused sudden en masse caldera collapse halfway through the climactic 27 eruptions, thereby liberating large amounts of seismic energy. At Fernandina and 28 Miyakejima, by contrast, caldera collapse was initiated early and continued for an 29

extended period of time from weeks to months, consisting of a series of discrete
subsidence events manifested by large earthquakes at Fernandina and by very long period
(VLP) signals at Miyakejima. Systematic changes in earthquake magnitudes and
quiescent intervals at both volcanoes reveal changes in friction as collapse took place
during extended time intervals.

35

36 **1.** Introduction

Our understanding of calderas has increased significantly in recent years. This 37 38 improvement is the result of new experimental and theoretical research which has complemented detailed field studies of calderas. By marrying these different approaches, 39 we have gained important insight into the surface and subsurface workings of caldera 40 41 systems. Experiments by *Roche et al.* [2000] have shown, for example, that the aspect ratio of the roof (thickness/width) plays an important role for the style of caldera 42 collapse, with piston-style behavior at low aspect ratios and more piecemeal collapse at 43 larger aspect ratios, often accompanied by subsurface stoping. Roche and Druitt [2001] 44 developed a failure criterion for piston collapse which is related to the aspect ratio of the 45 roof. Experiments by *Kennedy et al.* [2004] showed that many calderas are polygonal in 46 nature rather than circular, due in part to crustal heterogeneities. Modelling by Folch and 47 Martí [1998] and Martí et al. [2000] examined pressure variations during caldera-48 forming eruptions, while Legros et al. [2000] studied the effect of changing conduit 49 geometry upon such large-scale eruptions. Kumagai et al. [2001] developed a pumping 50 model of caldera subsidence, whereby changing pressures in the reservoir controlled the 51 52 style and amount of subsidence.

53 Despite these impressive advances, there are a number of important unresolved questions regarding caldera formation. Perhaps most importantly, the timing of collapse 54 is not well known for real calderas. Does subsidence begin at an early stage during the 55 eruption, or does it commence at a later stage? Does subsidence occur in a continuous or 56 incremental fashion throughout the eruption, or does it occur suddenly and en masse at 57 some specific time? The aspect ratio of the roof may play an important role in the timing, 58 with small values favoring early and/or incremental collapse, and large values resulting in 59 late-stage en masse subsidence. The occurrence of lithic lag breccias interbedded with 60 61 ignimbrite deposits can help shed light on the timing of collapse [e.g., Druitt and Bacon, 1986]. Nevertheless, our understanding of this problem is still rudimentary. A second 62 related issue is the relationship between magma withdrawal and caldera collapse. This 63 64 question is important, since it bears upon the amount of underpressure, due to withdrawal of magma, which develops in the magma reservoir before the onset of subsidence. Third, 65 the issue of collapse dynamics is still unresolved. Does subsidence occur as a response to 66 magma withdrawal from the reservoir? Or does subsidence of the roof forcefully push 67 magma out of the reservoir, in the process repressurizing the system [Druitt and Sparks, 68 1984; Martí et al., 2000; Kumagai et al., 2001]? Finally, what is the effect of magma 69 rheology upon the magma extraction and subsidence processes? For example, magmas 70 71 that are poor in crystals may behave differently during their extraction compared to crystal-rich magmas. The extent to which a magma is compressible also may influence 72 the amount of underpressure that develops due to magma withdrawal. 73 74

The purpose of this report is to explore these issues by examining four historic caldera-forming events which are well-documented and have adequate geologic and 76 geophysical data with which to address these questions. Of the four, two are silicic eruptions (Katmai 1912, Pinatubo 1991) and two are mafic (Fernandina 1968, 77 Miyakejima 2000), thus providing both compositional similarities and contrast. For each 78 example, the process of magma extraction is examined in relation to caldera collapse, as 79 manifested by seismicity which was recorded at the time. Additionally, the magma 80 rheology and dynamics are analyzed to study their effects upon underpressure generated 81 by magma extraction, as well as upon the subsequent subsidence. The goal is to shed 82 light on these and related issues by integrating observational, experimental, and 83 84 theoretical data related to caldera formation.

85

86 2. Chronology of Caldera Development

87 2.1. Katmai 1912

Much of the following summary of the 1912 Katmai eruption draws upon the 88 important papers by Abe [1992] and Hildreth and Fierstein [2000]. The climactic 89 eruption of Katmai lasted 60 hours and erupted 13.5 km³ of magma (dense rock 90 equivalent or DRE), frequently with simultaneous plinian and ignimbrite activity. The 91 eruptive materials consisted of nearly aphyric rhyolite (7-8 km³ DRE), dacite (4.5 km³). 92 and andesite (1 km³), both with 30-50% crystals. The rhyolite was erupted at the 93 beginning of the eruption and was predominant. The aspect ratio of the magma 94 reservoir's roof was approximately 2, based on a thickness of 4-5 km and an average 95 width of 2.3 km [Hildreth and Fierstein, 2000] (Table 1). 96 A critically important phase of the climactic eruption occurred at about 1000 UTC 97

on 7 June (midnight local time on 6 June, the time difference being 10 hours), about 11

hours after the onset of the eruption. To this point, a cumulative amount of ~ 8.7 km³ 99 DRE of magma had been erupted, representing 64% of the 13.5 km³ total volume, with 100 an average evacuation rate of $\sim 2.2 \times 10^5 \text{ m}^3 \text{ s}^{-1}$. This 8.7 km³ volume was nearly all 101 rhyolite; at this point in time, however, the relative proportions changed dramatically, 102 with rhyolite decreasing to 40-50% and dacite and andesite increasing to $\sim 40\%$ and 103 ~10%, respectively. Subsequently, the evacuation rate declined abruptly to ~2.8 x 10^4 m³ 104 s^{-1} well into 8 June. For a period of 5-6 hours on 7 June, the relative proportions of the 105 three magma types fluctuated rapidly and substantially. 106

107 Seismicity during the climactic eruption was characterized by very high levels of liberated energy, mainly due to several large-magnitude earthquakes [Abe, 1992; Hildreth 108 and Fierstein, 2000]. The pattern of seismicity exhibited several large, rapid jumps when 109 110 plotted on a diagram of cumulative energy vs. time (Fig. 1a). Energies were calculated as $\log E = 1.96 M + 2.05$ where E is the energy in Joules and M the magnitude of the 111 earthquake. The first significant increase occurred at 0956 UTC on 7 June, with a M 6.5 112 earthquake coinciding with the change in magma compositions described above. This 113 event may indicate the initiation of caldera collapse, since the first lithic mud layer was 114 erupted from the volcano at this time, followed by a second mud layer approximately six 115 116 hours later [Hildreth and Fierstein, 2000]. The second major jump began with a M 7.0 earthquake at 0736 UTC on 8 June, followed by a M 6.8 event at 0848 UTC and a M 6.6 117 event at 1300 UTC [Abe, 1992] (Fig. 1a). During this 5¹/₂ hour period, approximately 118 60% of the total energy associated with the climactic eruption was released, likely 119 corresponding with the bulk of caldera subsidence [Hildreth and Fierstein, 2000]. 120

| 121 | We have redrawn Figure 1 as a non-dimensional diagram (Fig. 2) where time is |
|-----|---|
| 122 | normalized to the duration of the climactic eruption, and cumulative energy is normalized |
| 123 | to the total energy released during and immediately after the climactic eruption. At |
| 124 | Katmai, there is a clear mismatch between erupted magma and released seismic energy |
| 125 | during the climactic eruption, with significant amounts of magma being erupted well |
| 126 | before the bulk of the seismic energy release (Fig. 2a). For example, by the first M 6.5 |
| 127 | earthquake at 0956 UTC on 7 June, ~64% of the total magma volume had already been |
| 128 | erupted, while only \sim 4.5% of the cumulative seismic energy was released. At this stage, |
| 129 | the eruption was less than 20% complete. By the time of the M 7.0 earthquake at 0736 |
| 130 | UTC on 8 June, approximately halfway through the climactic eruption, 80% of the |
| 131 | magma had been erupted and 51% of the seismic energy released. |
| | |

A final point is that of the total released seismic energy associated with caldera collapse at Katmai, 74% was released during the 60-hour climactic eruption. Thus, there was additional seismic energy released after the climactic eruption, which may be related to structural adjustments and further collapse of the unstable edifice.

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137 **2.2 Pinatubo 1991**

The climactic phase of the 15 June 1991 Pinatubo eruption lasted approximately 8.8 hours, beginning at 0542 UTC (1342 local time, the time difference being 8 hours) [*Hoblitt et al.*, 1996]. The eruptive products consisted almost entirely of dacite, with 3.7-5.3 km³ of DRE magma erupted [*Scott et al.*, 1996]. According to these values, the mean magma discharge rate ranged from $1.2 \times 10^5 \text{ m}^3 \text{ s}^{-1}$ to $1.7 \times 10^5 \text{ m}^3 \text{ s}^{-1}$. *Koyaguchi and Ohno* [2001] have suggested that initial discharge rates may have reached as high as 3.6 x 144 $10^5 \text{ m}^3 \text{ s}^{-1}$, subsequently declining to ~1.2 x $10^5 \text{ m}^3 \text{ s}^{-1}$ during the course of the eruption. 145 The aspect ratio of the roof was ~2.4, based upon the observed caldera diameter of 2.5 146 km and depth to the top of the magma reservoir of ~6 km [*Mori et al.*, 1996a; *Rutherford* 147 *and Devine*, 1996] (Table 1). This is a maximum value, as the lateral extent of the roof 148 may have been greater at depth.

Significant seismicity began with a M 5.1 event at 0739 UTC on 15 June, two 149 hours after the beginning of the climactic eruption [Mori et al., 1996b]. The pattern of 150 cumulative seismic energy resembles that of Katmai, with large jumps occurring midway 151 through the eruption (Fig. 1b). At 1041 UTC, a M 5.5 event was recorded, and 34 152 minutes later at 1115 UTC, the largest earthquake of the sequence (M 5.7) was observed. 153 The bulk of caldera collapse may have occurred at the time of these large-magnitude 154 155 events [Scott et al., 1996], although there is some uncertainty as to whether the earthquakes were related directly to collapse or were tectonic in origin [Bautista et al., 156 1996]. Lithic breccias and lithic-rich pyroclastic flow deposits were observed to overlie 157 most of the pumiceous pyroclastic flow deposits [Scott et al., 1996], also suggesting that 158 the majority of collapse took place at a relatively late stage during the climactic eruption. 159 Similar to Katmai, there is a mismatch in timing between erupted magma and 160 released seismic energy (Fig. 2b). By the time of the M 5.5 earthquake at 1041 UTC, at 161 least 57% of the total magma volume of 3.7-5.3 km³ had been erupted, while only 14% of 162 cumulative energy was released. The value of 57% may in fact be appreciably larger if 163 discharge rates were high during early stages of the climactic eruption, as suggested by 164 Koyaguchi and Ohno [2001]. When the M 5.7 event occurred at 1115 UTC, at least 63% 165

of the magma had been erupted, with 43% of the cumulative energy released. At thisstage, the climactic eruption was nearly two thirds complete.

Compared to Katmai, proportionally more seismic energy was released at Pinatubo after the climactic eruption stopped (51% for Pinatubo vs. 26% for Katmai) than during the eruption itself. Discharge rates at Pinatubo appear to be higher than at Katmai, while the climactic eruption was significantly shorter. As a result, structural adjustments after the climactic eruption may have been more significant at Pinatubo than at Katmai.

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174 **2.3 Fernandina 1968**

By contrast with Katmai and Pinatubo, caldera development at Fernandina in 175 1968 occurred over an extended period of 250-300 hours between 12-21 June [Simkin 176 177 and Howard, 1970; Filson et al., 1973]. The magma was basaltic in composition. Nearly all magma displacement occurred beneath the surface, with only a small amount erupted. 178 A caldera was already present when this collapse episode took place. During 12-21 June 179 180 1968, the caldera floor subsided a maximum of 350 m in the southeast sector. A preexisting tuff cone on the caldera floor was observed to be undisturbed after the collapse 181 event [Simkin and Howard, 1970; Filson et al., 1973]. These observations suggest 182 trapdoor-style piston subsidence. The new collapse volume at the surface was 2.0-2.4 183 km³. Based upon a 1 km-thick roof and a 3200 m equivalent diameter of the caldera floor, 184 the aspect ratio of the roof was ~ 0.3 (Table 1), although there is some uncertainty of the 185 roof thickness [Filson et al., 1973]. 186

187 The sequence of events began with an eruption accompanied by lava flows which 188 were observed on the eastern flank of the volcano on 21 May 1968. A possible explosion

| 189 | occurred on 8 June at 0220 UTC (2020 local time on 7 June, the time difference being 6 |
|-----|--|
| 190 | hours). On 11 June, a vapor cloud was observed after a M 3 earthquake at 1618 UTC. |
| 191 | This emission was followed by two large explosive eruptions at 2218 and 2308 UTC |
| 192 | which may have been partly phreatomagmatic [Simkin and Howard, 1970; Filson et al., |
| 193 | 1973]. After these events, significant amounts of lithic-rich tephra began to fall, an |
| 194 | occurrence which may have marked the initiation of caldera collapse [Simkin and |
| 195 | Howard, 1970]. Eruptive activity then diminished on 12 June, while seismic activity |
| 196 | increased, with a series of $M \ge 5$ earthquakes beginning at 2221 UTC [<i>Filson et al.</i> , |
| 197 | 1973]. These large earthquakes probably represented the early stages of caldera |
| 198 | subsidence. Mean minimum magma evacuation rates were $3.0-3.4 \times 10^3 \text{ m}^3 \text{ s}^{-1}$, based on |
| 199 | a caldera collapse volume of 2.2 x 10^9 m ³ and a duration of 182 hours from 12-20 June |
| 200 | for subsurface drainage of magma from the reservoir. These rates are several orders of |
| 201 | magnitude less than at Katmai and Pinatubo. |
| 202 | The seismicity associated with collapse at Fernandina was exceptional [Filson et |

al., 1973]. From 2221 UTC on 12 June to 0420 UTC on 15 June, there were a series of M 203 > 5 events which were regularly spaced at ~6 hours. Beginning at 0851 UTC on 15 June, 204 the large events decreased (a) in magnitude to M 4.7-4.9 and (b) in spacing to ~4-hour 205 intervals. Starting late on 16 June or early on 17 June, the seismic pattern changed yet 206 again, with more closely spaced events of generally smaller magnitudes (i.e., M 4.0-4.5). 207 At about 1200 UTC on 20 June, the number of earthquakes decreased rapidly. For 208 example, 89 events were recorded on 20 June, 13 on 23 June, and only 5 on 24 June 209 [Filson et al., 1973]. Recorded seismicity stopped on 27 July. 210

By contrast with Katmai and Pinatubo, the overall pattern of seismic energy released by Fernandina during caldera subsidence was quite regular, showing initially high rates of energy which gradually subsided with time to zero (Fig. 1c). The trend between 25-75 hours is charaterized by a series of abrupt jumps, which are the M 5 events occurring mainly on 13-14 June. The jumps then become smaller and eventually indiscernible on the diagram by 150 hours as earthquakes diminished in magnitude and increased in number.

218

219 **2.4**

Miyakejima 2000

Caldera development at Miyakejima from 8 July of 2000 was even more 220 protracted than at Fernandina, extending over a period of about 40 days. However, the 221 222 nature of magma evacuation at Miyakejima resembled that at Fernandina. The total volume of the caldera of $6 \times 10^8 \text{ m}^3$ [Geshi et al., 2002] is much larger than the volume of 223 ejecta from the eruption (9.3 x 10⁶ m³) [*Nakada et al.*, 2005], since magma was 224 225 displaced almost entirely in the subsurface, with very little material being erupted. The caldera was intermittently enlarged during July-August 2000, and its final size was 1.6 226 km in diameter and 450 m in depth [Geshi et al., 2002]. The depth of the magma 227 chamber related to caldera formation is not well-known, but seismic and geodetic surveys 228 suggest that a pressure source is located at a depth of 3-6 km [Kobayashi et al., 2003; 229 Irwan et al., 2003; Ueda et al., 2005]. Using 3-6 km as the thickness of the caldera block, 230 the aspect ratio of the roof is evaluated as $1.9 \sim 3.8$. 231 The magmatic activity began with a gigantic seismic swarm on 26 June around 232

233 0859 UTC (1759 local time on 26 June, the time difference being 9 hours) [Ueda et al.,

234 2005; *Uhira et al.*, 2005] caused by northwestward dike-forming intrusions of magma, followed by a small submarine eruption [Nakada et al., 2005; Kaneko et al., 2005]. The 235 earthquake swarm beneath the island died out by the next day, and activity shifted to a 236 comparatively quiescent stage. However, small volcanic earthquakes began to be 237 recorded beneath the summit at the beginning of July, leading to a summit eruption on 8 238 July at 0941 UTC, accompanied by significant subsidence at the surface. A void was 239 already present just beneath the summit several days before the first summit eruption, so 240 that the subsidence on 8 July is interpreted as collapse of a block into a shallow void 241 space [Furuya et al., 2003]. The caldera grew incrementally after the eruption [Geshi et 242 al., 2002], with infrequent explosive eruptions. The caldera had grown to about 1.5 km in 243 diameter by early August and was widened later by landslides off its steep walls [Nakada 244 et al., 2005]. Intermittent explosive eruptions that had occurred from 10 August 245 eventually led to the largest explosion on 18 August around 0900 UTC which was 246 vulcanian to subplinian in nature. After this eruption, significant caldera development 247 was not observed. The caldera size finally reached $\sim 0.6 \text{ km}^3$ in volume [Geshi et al., 248 2000]. Based on the duration between the first summit subsidence on 8 July and the 249 largest eruption on 18 August (984 hours), the mean magma evacuation rate is estimated 250 at $\sim 170 \text{ m}^3 \text{ s}^{-1}$. The rate is significantly less than the other three volcanoes. After the 251 largest eruption, the characteristic activity was strong degassing. By the end of August 252 and early September, the volcano began to emit substantial quantities of sulfur dioxide, at 253 times exceeding 100,000 metric tonnes SO₂ per day [Kazahaya et al., 2004]. 254 The seismic activity during caldera formation was characterized by very long-255

256 period (VLP) seismic pulses having a pulse width of about 50 seconds [Kikuchi et al.,

2001; Kumagai et al., 2001]. A series of VLP signals were observed once or twice a day 257 from 9 July at 1339 UTC to 17 August at 1907 UTC. This period corresponds exactly 258 with the observed duration of caldera development. The equivalent moment magnitudes 259 are 5.0 to 5.6, according to *Kikuchi et al.* [2001]. In addition to the VLP signals, other 260 earthquakes occurred during the caldera formation stage, e.g., series of earthquakes 261 262 which were observed as precursors to the VLP signal [Kobayashi et al., 2003], but the magnitude of these individual events was only 1 to 2. Thus, the total released seismic 263 energy during caldera development can be generally described by the VLP signals alone 264 (Fig. 3). The released seismic energy of each VLP event is calculated as $\log E_s = \log M_0$ -265 4.3, where the seismic energy E_s is in Joules and the seismic moment M_0 is in Nm. The 266 overall trend of released seismic energy is similar to that of Fernandina; the seismic 267 energy was regularly released at a nearly constant rate, unlike the pattern of Katmai and 268 Pinatubo. In early stages, the occurrence interval was approximately 12 h, but it became 269 substantially longer at the beginning of August (~500 h on Fig. 3), with magnitudes of 270 271 individual VLP events correspondingly increasing. This trend of occurrence is opposite to the case of Fernandina. The repetitious character of these VLP signals, as well as the 272 constant growth rate of the caldera, implies that the subsidence process at Miyakejima 273 was highly regular and systematic. *Kumagai et al.* [2001] have interpreted these signals 274 as a volumetric expansion of the magma reservoir which may be associated with slip of a 275 vertical piston in the conduit, followed by magma outflow and gradual depressurization. 276 Slow deflation was recorded by the tiltmeters on the island [e.g., Ukawa et al., 2000] 277 during the interval of VLP occurrence representing rapid short-term inflations of the 278 279 magma reservoir. Thus, these seismic events can describe a basic framework of magma

withdrawal on short timescales controlling caldera development, a process which is
superimposed upon deformation trends at longer timescales. This may be a common
feature for other volcanoes as well.

283

3.

Modelling the Subsidence Process

The high aspect ratios of the Katmai and Pinatubo caldera blocks illustrate the 284 presence of comparatively thick crust which likely provided "bridging" roof support for 285 the magma reservoirs beneath. Accordingly, Katmai and Pinatubo illustrate the mismatch 286 between early magma withdrawal and comparatively late caldera collapse. By contrast, 287 magma withdrawal and caldera subsidence at Fernandina and Miyakejima appear to have 288 taken place during short intervals and in a progressive manner. In the case of Fernandina, 289 the thin roof may not have provided much support or resistance before or during 290 291 subsidence. Miyakejima is a more complicated case which exhibited both a thin roof during initial caldera collapse [*Kikuchi et al.*, 2001], as well as a thicker crustal block 292 which subsided in a progressive fashion during the summer of 2000. These differences 293 294 pose interesting questions regarding the properties of the magmas, the nature of the magma withdrawal process, and the dynamics of caldera formation during these 295 eruptions. 296

Because the eruptions have been well studied, many parameters relating to the eruptions are well constrained, and the eruptions thus can be modeled with some confidence in terms of magma withdrawal and caldera development. *Kumagai et al.* [2001] have modeled caldera subsidence at Miyakejima in the middle of 2000. Their approach can be adapted for use at Katmai, Pinatubo, and Fernandina with certain caveats, in particular that caldera subsidence occurs as a piston-style process. By this, we

| 303 | mean that the roof of the magma reservoir subsided coherently as a single block. This |
|---|---|
| 304 | style of subsidence is illustrated schematically in Figure 4. We recognize that many |
| 305 | calderas have more complicated subsidence styles, such as piecemeal, downsagging, a |
| 306 | series of concentric downthrown blocks, etc. [Kennedy and Stix, 2003]. Nevertheless, |
| 307 | while this collapse style is likely an oversimplification, it is sufficient for the modelling |
| 308 | presented here, since Katmai and Pinatubo collapsed en masse and Fernandina and |
| 309 | Miyakejima as a series of discrete steps. Kumagai et al. [2001] show that the piston |
| 310 | begins to move downward after a time T: |
| 311 | |
| 312313314315 | $T = \frac{2(F_{\rm s} - F_{\rm d})}{p'A} \tag{1}$ |
| 316 | where $F_s - F_d$ are static and dynamic frictions, respectively, p' is the rate of pressure |
| 317 | decrease due to outflow of magma from the magma chamber, and A is the cross-sectional |
| 318 | area at the base of the piston (all symbols are explained in Table 2). Equation (1) can be |
| 319 | rewritten as |
| 320 | |
| 321 | $p'TA > 2(F_{\rm s} - F_{\rm d}) \tag{2}$ |
| 322 | |
| 323 | which shows that the piston will move downward when the pressure decrease $p'T$ |
| 324 | exceeds the rock friction or rock strength. Kumagai et al.'s [2001] model is essentially |
| 325 | one of repeatedly pumping the magma chamber with the piston: |
| 326 | |

 $p_1 = p_0 - p'T + p$ (3)

| 329 | where p_0 is the original pressure in the magma chamber, $p'T$ is the pressure decrease due |
|-----|---|
| 330 | to magma outflow, p is the pressure increase in the chamber once the caldera block |
| 331 | begins to move downward, and p_1 is the re-equilibrated pressure in the magma chamber. |
| 332 | The equation describes a stick-slip sense of movement of the caldera block, where stick is |
| 333 | associated with underpressure as magma is erupted or drained from the reservoir. After a |
| 334 | certain time, underpressure exceeds friction, and the block begins to slip. This movement |

may squeeze the magma, causing it to be repressured. Equation (3) can be written as

337
$$p_1 = p_0 + \kappa [(-\alpha T + A_z) / V_0]$$
(4)

where κ is the bulk modulus of the magma, α the evacuation rate of the magma, z the displacement of the block, and V_0 the initial volume of the magma chamber before caldera subsidence is initiated [Kumagai et al., 2001].

With respect to events at Katmai and Pinatubo, we are interested primarily in pressure variations of the magma reservoir before the bulk of collapse occurred. As observed in Figures 1-2 above, there was an extended period of magma withdrawal before the bulk of collapse was initiated approximately halfway through the climactic eruptions, as manifested by the sudden onset of large-scale seismicity. This period of magma withdrawal indicates that pressure was decreasing in the magma reservoirs at both volcanoes. The pressure decrease prior to caldera subsidence can be calculated as follows:

$$p'T = \frac{\kappa \alpha T}{V_0} \tag{5}$$

353 354

For Katmai and Pinatubo, we assume that the bulk of collapse occurred in a short 355 time interval about halfway through the climactic eruption. For Fernandina, collapse 356 occurred incrementally separated by quiescent intervals of six hours. The parameters T, α , 357 and V_0 are reasonably constrained at all three volcanoes (Table 1), while κ is not. 358 Therefore, we have plotted the pressure decrease against a range of bulk moduli in Figure 359 5. For Katmai and Pinatubo, bulk moduli range from $\sim 10^6$ Pa to 10^{10} Pa for pressure 360 decreases from 1 MPa to nearly 10 GPa, respectively. 361 362 Equations (2) and (5) are valid once the caldera collapse process is in a steadystate condition, i.e., the caldera block is alternately sticking and slipping during the 363 course of subsidence. At Katmai and Pinatubo, however, this condition may not be 364 365 strictly true, as the evacuating magma reservoir and increasing underpressure were developing before the bulk of subsidence was initiated. Here, the resistance to subsidence 366 was provided by the shear strength and fracture processes of the country rocks. In this 367 case, an independent approach for estimating underpressures is to use the failure criterion 368 method of *Roche and Druitt* [2001], which calculates the underpressure p'T_{crit} which is 369 required to exceed a critical shear stress upon the rocks above the magma reservoir: 370 371

(6)

372 $p'T_{\rm crit} \ge 4R\tau_{\rm c}$

where *R* is the aspect ratio of the caldera block as defined by its thickness *H* divided by its diameter *D*, and τ_c is the critical shear stress for failure. This shear stress can be solved as follows:

377

$$\tau_{c} = \tau_{o} + \mu \sigma_{n}$$

379

where τ_0 is the cohesion of the caldera block, μ is the coefficient of internal friction of the block (~0.6) [*Byerlee*, 1978], and σ_n is the mean stress normal to the plane of failure. This last parameter can be calculated as $\sigma_n = kp_{\text{lith}}$ where k is a constant (~0.6) [*Cornet and Valette*, 1984], p_{lith} is the lithostatic pressure $\rho_r g H/2$, ρ_r is the average density of the caldera block, and g is the acceleration due to gravity.

Using these equations gives underpressures of 166-205 MPa for Katmai and 265-385 312 MPa for Pinatubo for a range of cohesions from 0.1 MPa to 5 MPa [Roche and 386 Druitt, 2001]. These values may be overestimates, as shear strengths of rocks beneath 387 volcanoes are typically on the order of 1-100 MPa [Martí et al., 2000]. Nevertheless, 388 these calculations provide a useful upper limit of underpressures at these two volcanoes, 389 while our calculations from Equation (5) use a lower limit of 1 MPa. For these levels of 390 underpressure, therefore, bulk moduli for Katmai range from a low of 1.2×10^6 Pa to a 391 high of 2.0-2.5 x 10^8 Pa, while for Pinatubo the range is from 2.0 x 10^6 Pa to 5.2-6.1 x 392 10^8 Pa (Fig. 5). The bulk modulus for bubble-free magma is $\sim 10^{10}$ Pa, while that for 393 bubbly magma is typically 10⁷-10⁹ Pa [*Huppert and Woods*, 2002]. It is thus reasonable 394 to conclude that reservoir magmas at Katmai and Pinatubo were bubbly immediately 395 before and during caldera collapse. 396

(7)

| 397 | For Fernandina, the case is less clear, as pressure decreases range from ~ 0.3 MPa |
|-----|--|
| 398 | to ~290 MPa for bulk moduli of 10^7 - 10^{10} Pa (Fig. 5). This large range can be further |
| 399 | constrained by the approach of Roche and Druitt [2001], as done above for Katmai and |
| 400 | Pinatubo, and these calculations result in maximum critical underpressures of 3-9 MPa |
| 401 | for Fernandina using the same range of cohesions as above. The comparatively low |
| 402 | magma evacuation rates at Fernandina $(10^3 \text{ m}^3 \text{ s}^{-1} \text{ compared to } 10^4 10^5 \text{ m}^3 \text{ s}^{-1} \text{ at Katmai}$ |
| 403 | and Pinatubo) also imply relatively small amounts of underpressure. Thus, we suggest |
| 404 | that the basaltic magma in the reservoir beneath Fernandina also had a reduced bulk |
| 405 | modulus and appreciable vesicularity as it drained in the subsurface (Fig. 5). |
| 406 | It is important to assess the quality of the values we have used in these |
| 407 | underpressure calculations, in particular the parameters T , α , and V_0 (Table 1). In the case |
| 408 | of T , we are fairly confident of the data, since they are constrained (1) by the time |
| 409 | between the start of the climactic eruption and the sudden onset of large-scale seismicity |
| 410 | for Katmai and Pinatubo, or (2) by the periods of seismic quiet between major |
| 411 | earthquakes at Fernandina. For α , the values we have used for Katmai and Pinatubo |
| 412 | appear reasonable, as they are constrained by the volume of magma erupted during the |
| 413 | course of the climactic eruption divided by its duration. Further refinements based on |
| 414 | field studies have been possible for both volcanoes [Hildreth and Fierstein, 2000; |
| 415 | Koyaguchi and Ohno, 2001]. For Fernandina, the case is not so clear because very little |
| 416 | magma was erupted. Thus, we have used the caldera collapse volume at the surface |
| 417 | divided by the duration of the collapse events. This is probably a minimum rate, so we |
| 418 | have doubled this value to provide an upper limit. Lastly, the V_0 parameter is not |
| 419 | particularly well constrained. Thus we have simply used the volumes of erupted magma |

420 (Katmai, Pinatubo) or caldera collapse volume (Fernandina) to represent V_0 . These are clearly minimum values; however, larger values of V_0 would result in decreased 421 underpressures and bulk moduli (Fig. 5), reinforcing the argument that the magmas were 422 bubbly upon evacuation. A magma reservoir which empties itself entirely during a 423 climactic eruption is an interesting concept [Martí et al., 2000]. Such conditions might 424 promote its replenishment during and/or immediately after the eruption [e.g., Stix and 425 Gorton, 1993]. Another possibility is a larger magma reservoir in which the largely liquid 426 portion is erupted while the more crystallized portion is not. 427

428 In these calculations, the T parameter is of crucial importance, since it is directly related to the amount of underpressure that is generated before the roof fails and the bulk 429 of caldera collapse is initiated. It is thus essential to know if the T parameter is being used 430 431 appropriately and correctly in the calculations above. This can be verified by calculating the displacement of the subsiding caldera block as follows. If T is significantly longer 432 than the duration of piston displacement, as appears to be the case for Katmai and 433 Pinatubo and probably Fernandina as well (Figs. 1, 2), then the displacement of the 434 caldera block can be estimated thus: 435

436

437

 $z = \alpha T / A \tag{8}$

438

Results for Katmai and Pinatubo are shown in Figure 6 and Table 3. Since magma
evacuation rates likely varied during the course of the climactic eruptions, we have used a
range of values from the minimum and maximum estimates. Allowing for some

uncertainty in the magma evacuation rates, the agreement between calculated andobserved displacements is good (Table 3).

Equation (8) also can be used to examine the nature of subsidence at Fernandina. 444 In this case, T and A are known while α is not. Figure 7 shows the effects of evacuation 445 rate and time before subsidence on the amount of displacement of the piston. For 446 evacuation rates ranging from a minimum of 2970 m³ s⁻¹ to a maximum of 6730 m³ s⁻¹. 447 individual displacements are 8-18 m, respectively. Notably, the spacing between major 448 earthquakes changed on 15 June from 6 hours to 4 hours, with a corresponding decrease 449 in the magnitudes of the earthquakes. Inputting these intervals as values for T reveals that 450 individual displacements, as manifested by the earthquakes, decreased on 15 June from 451 8.0 m to 5.4 m, using a constant evacuation rate of 2970 $\text{m}^3 \text{ s}^{-1}$. In summary, it is certain 452 that (a) subsidence at Fernandina was incremental, in contrast to Katmai and Pinatubo, 453 and (b) the amount of displacement for individual subsidence events declined with time. 454 It is possible that both declining evacuation rates and time intervals between earthquakes 455 contributed to these decreasing displacements. Alternatively, the evacuation rate 456 remained approximately constant, and the smaller displacements were the result of 457 reduced time intervals between earthquakes. This issue is examined more fully in the next 458 section. 459

460

461 **4. Discussion**

The analysis above has relevance for the dynamics of magma extraction, the nature of caldera subsidence, the presence of bubbly magma stored in shallow magma reservoirs, and the progressive weakening of caldera faults. Furthermore, it is clear that in
 many cases these issues are inter-related. These points are discussed in turn below.

467

4.1 Early Extraction and Eruption of Silicic Magma at Katmai and Pinatubo

The bulk of caldera collapse appears to have occurred midway through the 468 climactic eruptions at Katmai and Pinatubo. Most of the magma was withdrawn and 469 erupted rapidly in the early stages of the eruptions, allowing significant underpressures to 470 quickly develop in the magma reservoirs. The early-erupted magma was highly 471 extractable, principally due to its low viscosity. In the case of rhyolite magma at Katmai, 472 for example, viscosity calculations using the method of *Hess and Dingwell* [1996] 473 indicate that viscosities ranged from a maximum of 1.4×10^5 Pa s at 4 wt. % H₂O and 474 805° C to a minimum of 7.1 x 10^3 Pa s at 7 wt. % H₂O and 850° C [Westrich et al., 1991; 475 Lowenstern, 1993; Cowee et al., 1999; Hildreth and Fierstein, 2000; Coombs and 476 Gardner, 2001]. As magma transited the conduit, flow was likely shearing at high strain 477 478 rates. Under these conditions, viscosities would be lowered even further due to shear thinning behavior induced by the presence of bubbles [Stein and Spera, 2002] and/or 479 viscous dissipation effects [Lavallée et al., 2007]. It is therefore possible that the 480 viscosity of rhyolite magma at Katmai approached values as low as 10^3 Pa s, which is 481 similar to the viscosity of dry basalt [Khitarov et al., 1976]. These low viscosities thus 482 allowed rapid extraction of the early-erupted magma from the reservoir, causing high 483 underpressures to develop as shown by Equation (5). High magma extraction rates also 484 may have been aided by large pressure gradients due to the free gas phase and by large 485 conduit dimensions. 486

Equation (5) also reveals that there may be a delicate balance between the bulk modulus and evacuation rate of the magma as a system is underpressured. In order to maintain reasonable underpressures (~100 MPa), bubbly magmas of low bulk modulus may be associated with high magma evacuation rates while bubble-poor magmas of high bulk modulus may be extracted at lower evacuation rates [*Huppert and Woods*, 2002].

At Katmai, a rheological interface was reached in the magma chamber at an early 492 stage in the eruption when significant amounts of mixed crystal-rich dacite and andesite 493 began to be erupted at ~1000 UTC on 7 June. Significant seismicity was initiated at this 494 point, while the volumetric eruption rate decreased by nearly an order of magnitude. 495 These changes likely are the result of the eruption's tapping a rheological boundary 496 between low-viscosity rhyolite above and high-viscosity dacite and andesite underneath. 497 498 Using the underpressure arguments from above, the bulk modulus also may have changed abruptly at this boundary from low values in rhyolite to high values in dacite and 499 andesite. The seismicity may indicate onset of caldera collapse; if so, subsidence of the 500 501 caldera block may have helped mix dacite and andesite.

Rheological boundaries in silicic magma chambers may be common occurrences, 502 serving to slow or stop the course of an eruption [Smith, 1979, Bacon and Druitt, 1988; 503 Scaillet et al., 1998; Hildreth, 2004]. At the 121 Ma Ossipee ring complex in New 504 Hampshire, Kennedy and Stix [2007] have shown that crystal-poor rhyolite was erupted, 505 resulting in caldera subsidence. After collapse, crystal-rich quartz syenites were intruded 506 into the ring dyke. The magma chamber configuration thus consisted of low-viscosity 507 rhyolite magma which was erupted from the top of the reservoir, underlain by high-508 509 viscosity quartz syenite crystal mush which was not extractable and not erupted.

In summary, magma rheology and rheological boundaries appear to strongly influence the course of caldera-forming eruptions. For such systems, the nature of magma extraction has profound implications for the timing and nature of caldera subsidence.

513

514 4.2 Magma Extraction, Caldera Subsidence, and Seismic Energy Release

Hildreth and Fierstein [2000] have shown that the seismic energy released at 515 Katmai was 2-3 orders of magnitude higher than that observed at Pinatubo and 516 Fernandina (and Miyakejima as well), attributing this difference to high-strength rocks, 517 lateral magma flow, and horizontally layered structure at Katmai. As noted above, a 518 comparatively large volume of magma was withdrawn and erupted at Katmai prior to the 519 bulk of collapse. Experiments by Roche et al. [2000] and Kennedy et al. [2004] have 520 521 shown that high aspect ratios of the caldera block result in late collapse after a significant volume of magma has been extracted from the reservoir. Such a situation allows large 522 underpressures to develop in the reservoir before collapse is initiated. The experiments 523 also reveal that subsidence occurs in a progressive fashion by stoping at low evacuation 524 rates. In the case of high evacuation rates, however, we theorize that collapse may occur 525 en masse after a significant proportion of magma has been withdrawn quickly from the 526 reservoir, the subsidence involving a large fault surface area and significant slip. This 527 style of collapse could be further facilitated by explosive enlargement of the ring fault or 528 faults. In such a situation, the high seismic moment is principally the result of a large 529 quantity of easily extractable magma which is erupted at a high rate from a deep 530 reservoir, thus generating significant underpressure in a short period of time. The 531 elevated evacuation rates result in a focusing of stress [Sunde et al., 2004], causing the 532

crust to fail catastrophically along major caldera faults. The ensuing caldera collapse
occurs abruptly, releasing large amounts of seismic energy. If this hypothesis is correct, it

implies that this style of collapse is closely linked to the depth of the magma reservoir,

the magma's rheology, and the extraction rate of the magma.

537

538 4.3 Bubbly Magma in Shallow Reservoirs

The bulk modulus of a substance is a measure of its incompressibility. It is a balance between an imposed pressure difference and the associated volume change of the material:

- 542
- 547 Where Δp is the pressure difference and ΔV the change in volume. The calculations above demonstrate that the Katmai and Pinatubo reservoir magmas were in a bubbly state 548 during their extraction prior to the bulk of caldera collapse. However, the calculations do 549 550 not reveal the physical state of the magmas before the initiation of magma withdrawal. They may have been undersaturated, saturated, or oversaturated in volatiles. Other lines 551 of evidence indicate that a pre-eruptive, separate fluid phase was present beneath 552 553 Pinatubo [Wallace and Gerlach, 1994] and possibly Katmai [Coombs and Gardner, 2001; Hammer et al., 2002]. It is thus reasonable to infer that these magmas may have 554 been partly vesicular before eruption, thereby aiding their extraction from the reservoir. 555 The occurrence of bubble-rich magma in shallow reservoirs has important 556 ramifications for eruptive processes. Such magma will have high buoyancy and low 557

558 viscosity, particularly so if the magma contains few crystals; as a result, it will be mobile and easily extracted and erupted from the reservoir. These properties can explain the high 559 eruption rates observed during the early stages of the Katmai and Pinatubo eruptions 560 [Hildreth and Fierstein, 2000; Koyaguchi and Ohno, 2001] and are consistent with 561 Huppert and Woods' [2002] model which shows enhanced eruption rates for magmas of 562 low bulk modulus. Under these conditions, the rapid withdrawal of bubbly magma 563 inevitably will lead to rapid fragmentation at deep levels in the conduit or even in the 564 reservoir itself, since the fragmentation threshold is a sensitive function of vesicularity 565 566 [*Spieler et al.*, 2004].

The presence of bubbly magma also will allow efficient syn-eruptive release of 567 volcanic gas [Wallace et al., 1995, 1999]. Silicic magmas, as exemplified by Katmai and 568 569 Pinatubo, have a melt phase which is depleted in volatile components such as sulfur, and a free gas phase which is correspondingly enriched in these components. During 570 eruptions, the volatiles will be released preferentially at an early stage, due to their 571 buoyancy and concentration in the upper levels of the magma reservoir. This early release 572 may be observable by remote sensing methods [Rose et al., 2000]. We hypothesize that 573 many basaltic magmas also contain free gas in the upper parts of their subsurface 574 plumbing. If so, these sulfur-rich systems may release large amounts of sulfur and other 575 gases at the beginning of an explosive eruption [e.g., Rose et al., 2003]. 576

577

4.4 Modification of Caldera Faults During Subsidence at Fernandina and
Miyakejima

580 By contrast with the sudden onset of caldera subsidence at Katmai and Pinatubo, collapse at Fernandina and Miyakejima was incremental and progressive. Fernandina and 581 Miyakejima provide contrasting insight during caldera subsidence. At Fernandina, the 582 patterns of seismic energy release can be grouped into discrete periods. For a given 583 period, the interval of time between large earthquakes remained essentially constant, as 584 did the magnitude of the earthquakes. The constant time interval implies that a critical 585 stress value was attained repeatedly; once this level was reached, it was relieved by 586 downward movement of the subsiding caldera block. The similar earthquake magnitudes 587 588 suggest that (a) the block repeatedly subsided the same distance, as calculated above, and (b) the stress level returned to a common baseline value. The stress then started to build 589 again during the next interval of time as magma continued to drain from the reservoir. 590 591 Notably at Fernandina, earthquake magnitudes and quiescent intervals both decreased abruptly after a period of time, best examplified at 0851 UTC on 15 June when 592 large earthquakes decreased in magnitude from M > 5 to M 4.7-4.9 and in spacing from 593 594 ~6-hour to ~4-hour intervals. The concurrent decreases in both parameters strongly suggest that they are physically linked. The declining time interval can be expressed by 595 Equation (1), which can be rewritten as 596

597

598
599

$$T = \frac{2V_0(F_s - F_d)}{\kappa \alpha A}$$
(10)
601

where $p' = \kappa \alpha / V_0$. Here the important parameters are V_0 , α, κ, and $F_s - F_d$, which are discussed in turn. As the caldera block subsided, the initial volume of the magma chamber V_0 was reduced; such a change should manifest itself by a continual decline in *T*. This pattern was not observed, however, instead showing discrete periods of constant Tfollowed by a decline, as described above. A change in the magma's evacuation rate α may also affect the value of T; a decrease of α would increase T, but the opposite behavior is observed. An increase of α would lower T and result in greater displacements of the subsiding caldera block and thus higher-magnitude earthquakes, but the opposite is observed.

Increased values of κ would lower T as shown by Equation (10), implying that the 611 magma became less bubbly with time [Filson et al., 1973]. This interesting possibility 612 suggests that the initially large subsidence events caused vesicular magma in the upper 613 parts of the reservoir to be compressed and pressurized reducing the amount of bubbles, 614 to a point where the caldera block stopped moving downward. Between subsidence 615 616 events, the reservoir drained, allowing magma beneath the caldera block to revesiculate to an extent where the block again subsided. With time, decreasing volatile contents in 617 the magma reduced vesiculation, hence produced less subsidence and lower-magnitude 618 619 earthquakes. The relationship between increasing bulk modulus on the one hand and reduced subsidence and smaller-magnitude seismicity on the other hand can be expressed 620 quantitatively thus: 621

622

$$\begin{array}{ccc}
623 \\
624 \\
625 \\
626 \\
\end{array} \qquad z = \frac{2V_0(F_s - F_d)}{\kappa A^2} \tag{11}$$

The shorter intervals between subsidence events may imply that evacuation ratesincreased over time.

| 629 | A second possibility is that $F_s - F_d$ declined during magma withdrawal and |
|-----|---|
| 630 | subsidence, causing T to decrease, as suggested originally by Filson et al. [1973] and |
| 631 | examined recently by Kobayashi et al. [2003] at Miyakejima. Essentially, a lower critical |
| 632 | stress threshold was reached sooner, and the block subsided. For a given evacuation rate, |
| 633 | the shortened time interval between subsidence events resulted in smaller amounts of |
| 634 | downward displacement of the caldera block. As a result, earthquake magnitudes also |
| 635 | decreased. It appears that the friction forces resisting downward movement of the caldera |
| 636 | block decreased with time in a stepwise fashion. This weakening process appears to have |
| 637 | occurred at discrete intervals rather than as a continuous process. |
| 638 | A final view is that the caldera faults were inward dipping. As magma drained |
| 639 | from the reservoir initially, the extent of collapse was comparatively large. As the magma |
| 640 | continued to drain, however, the ability of the caldera block to subside diminished |
| 641 | progressively due to the constricting faults. With time, the block became wedged between |
| 642 | the faults, and subsidence ceased. This hypothesis is difficult to evaluate, as we have no |
| 643 | direct evidence regarding the configuration of the caldera faults. |
| 644 | We can identify a similar temporal change of magnitude and time interval in the |
| 645 | individual VLP events of Miyakejima, which provides complementary information for |
| 646 | the understanding of frictional control. Kobayashi et al. [2003] have studied sequences of |
| 647 | small-amplitude earthquakes, especially in the events at the beginning of the VLP activity, |
| 648 | which were observed to occur about an hour before the appearance of a VLP signal. |
| 649 | Initially, time intervals between these precursor earthquakes exhibited progressive and |
| 650 | linear declines, while the earthquake amplitudes remained constant until near the end of a |
| 651 | sequence when the amplitudes declined precipitously. To explain these patterns, |

652 *Kobayashi et al.* [2003] use an asperity model where the caldera piston is strongly coupled to the country rock by means of "asperites". The repeated earthquakes of similar 653 amplitude represent the progressive breaking of individual asperites which maintain 654 constant strength. As asperites break, the stress level returns to a baseline which then 655 increases more rapidly with time, since there are progressively fewer asperites holding 656 the block stationary against the country rock. Eventually, all asperites are broken, the 657 critical stress level declines rapidly to the baseline value, and the block starts to move. 658 This view of an individual subsidence event at Miyakejima shares similarities with the 659 sequence of subsidence events at Fernandina. 660

When examining the overall activity at Miyakejima, however, the temporal trends 661 provide a contrasting case to Fernandina. At Day ~500 about halfway through the events 662 of 2000 (Fig. 3), the time interval between successive VLP events increased, as did the 663 magnitudes and energies of individual VLP signals. This interesting observation is 664 exactly opposite to that seen for Fernandina. The seismic energy of a VLP signal is 665 related to an enhanced internal pressure of the magma chamber Δp , which can be 666 rewritten as $2V_0(F_s - F_d)/A$. Assuming that the A and V_0 parameters remain constant, the 667 friction parameter $(F_s - F_d)$ controls the temporal features in magnitude and time interval 668 simultaneously, suggesting that the effective friction on the caldera faults may have 669 increased in early August 2000. If this interpretation is correct, then a longer time 670 interval was required to reach the point of failure. If the magma evacuation rate remained 671 constant, the longer time interval implies that a greater amount of magma was drained 672 from the reservoir than previously occurred. Hence, the caldera block was able to fall 673 further during an individual subsidence event, producing a large-magnitude VLP signal. 674

| 675 | In | summary, for both Fernandina and Miyakejima, friction acting on the conduit wall |
|-----|----|--|
| 676 | ma | y play a pivotal role for the process of incremental subsidence of the piston. |
| 677 | | |
| 678 | | |
| 679 | 5. | Conclusions |
| 680 | | The principal conclusions from this work are the following: |
| 681 | 1. | Caldera collapse at Katmai and Pinatubo occurred en masse at a comparatively late |
| 682 | | stage in the course of the eruption, while subsidence at Fernandina and Miyakejima |
| 683 | | proceeded incrementally. The en masse collapse was caused by the caldera block |
| 684 | | subsiding in response to magma evacuation. By contrast, the incremental collapse |
| 685 | | observed at the basaltic calderas may have resulted from the periodically subsiding |
| 686 | | block forcefully pushing magma out of the reservoir. |
| 687 | 2. | During its evacuation, the magma in the reservoirs beneath Katmai and Pinatubo |
| 688 | | possessed significant porosity and compressibility. Due to its compressibility and low |
| 689 | | viscosity, the rhyolite magma at Katmai was easily extracted and erupted, leading to |
| 690 | | significant underpressures prior to caldera subsidence. |
| 691 | 3. | At Katmai, changes in magma rheology appear to have played an important role in |
| 692 | | magma withdrawal and timing of collapse. The development of underpressure may be |
| 693 | | controlled by a delicate balance among certain parameters including the evacuation |
| 694 | | rate of the magma, its bulk modulus, and its crystal content. |
| 695 | 4. | Incremental collapse at Fernandina and Miyakejima reveals contrasting behavior |
| 696 | | which is probably related to modification of the caldera fault systems over the course |
| 697 | | of protracted subsidence. |
| | | |

5. The amount of extractable and eruptable magma present in the reservoir determines,
at least in part, the degree of underpressure in the reservoir, the evacuation rate of the
magma, and the timing and style of caldera collapse. Such magmas possess low
viscosities, small amounts of crystals, and substantial amounts of bubbles.

702

At Katmai and Pinatubo, magma discharge rates did not appear to increase when the 703 bulk of subsidence occurred, since most of the ignimbrites had already been erupted 704 before the major collapse episodes were initiated. Intracaldera ponding of ignimbrite does 705 706 not appear to be significant at either volcano. For all four volcanoes studied here, the subsiding caldera blocks appear to have behaved coherently, either as pistons or as en 707 masse subsidence. Aspect ratios of the caldera blocks for Katmai, Pinatubo, Miyakejima, 708 709 and possibly Fernandina as well, may exceed those of larger caldera systems, where roof blocks may resemble thin plates. For these large systems, collapse may occur more 710 incrementally and at an earlier stage due to the smaller aspect ratio. Incremental collapse 711 712 promotes progressive ponding of ignimbrite, increasing the lithostatic load of the subsiding block. Thus, incremental collapse may result in an eruption style quite different 713 from that associated with en masse collapse. 714

715

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896 **Figure captions**

Figure 1. Plot of cumulative energy released by large earthquakes during the (a) Katmai 898 1912, (b) Pinatubo 1991, and (c) Fernandina 1968 caldera-forming eruptions. The 899 steplike patterns of energy released at Katmai and Pinatubo contrast with the progressive 900 trend seen at Fernandina. Note the different time and energy scales for the three events. 901 Energies were calculated as $\log E = 1.96 M + 2.05$ where E is the energy in Joules and M 902 the magnitude of the earthquake. Souces of data as follows: Katmai, *Abe* [1992]; 903 904 Pinatubo, National Earthquake Information Center, http://earthquake.usgs.gov/regional/neic/; Fernandina, Filson et al. [1973]. 905 906 907 Figure 2. Cumulative energy plots for (a) Katmai 1912 and (b) Pinatubo 1991, in which the diagrams have been made non-dimensional. Time plotted on the x axis is normalized 908 relative to the duration of the climactic eruptions; for Katmai, the duration was 60 hours, 909 910 while for Pinatubo the duration was ~8.8 hours. Cumulative energy released by large earthquakes is normalized relative to the total amount of cumulative energy liberated 911 during and immediately after the climactic eruptions. 912 913 Figure 3. Plot of energies of VLP events, both individual and cumulative, recorded at 914 Miyakejima in July and August 2000. Using the relationship between the seismic energy 915 (Joules) and seismic moment (Nm) $\log E_s = \log M_0 - 4.3$, we can evaluate the released 916 seismic energy of each VLP event. In this calculation, we used the seismic moment of 39 917 VLP signals listed in *Kikuchi et al.* [2001] and obtained the values for M_0 from their 918

| 919 | Figure 9. Also shown are caldera volumes plotted as a function of time and shown as |
|-------------------|--|
| 920 | solid circles; data from Geshi et al. [2002]. |
| 921 | |
| 922 | Figure 4. Schematic diagram of piston subsidence (a) prior to caldera collapse (e.g., |
| 923 | Katmai, Pinatubo) and (b) after an increment of collapse (e.g., Fernandina, Miyakejima). |
| 924 | The aspect ratio is $R = H / D$. See Table 2 for symbology and text for equations. |
| 925 | |
| 926 | Figure 5. Magma underpressure plotted as a function of bulk modulus of the magma for |
| 927 | the Katmai, Pinatubo, and Fernandina eruptions. Typical rock shear strengths are also |
| 928 | shown. See text for discussion. |
| 929 | |
| 930 | Figure 6. Caldera subsidence plotted against evacuation rate. The amount of |
| 931 | displacement is calculated using Equation (8). A range of evacuation rates is used based |
| 932 | on published values of volumetric eruption rates. (a) Katmai 1912; (b) Pinatubo 1991. |
| 933 | |
| 934 | Figure 7. Caldera subsidence at Fernandina as a function of (a) evacuation rate of magma |
| 935 | and (b) the time interval before subsidence. The amount of displacement is calculated |
| 936 | using Equation (8). |
| 937 | |
| 938 | |
| 939 940 941 | |
| 942 943 044 | |
| 744 | |

| 945 | | | | | |
|-----|--|---------------------------|-------------------------|---------------------------|------------------------|
| 946 | | | | | |
| 947 | Table 1. Input values for mo | deling | | | |
| 948 | | | | | |
| 949 | | | | | |
| 950 | Parameter | Katmai | Pinatubo | Fernandina | Miyakejima |
| 951 | | | | | |
| 952 | Aspect ratio of caldera | 2.0 | 2.4 | 0.31 | 1.9 - 3.8 |
| 953 | block, <i>R</i> (dimensionless) | | | | |
| 954 | | 10 | 0 | 0 | 10 |
| 955 | Initial volume of magma | 1.3×10^{10} | $5.0 \ge 10^9$ | 2.2×10^9 | 3.7×10^{10} |
| 956 | reservoir, $V_{\rm o}$ (m ³) | | | | |
| 957 | | <i>(</i> | <i>,</i> | <i>,</i> | <i>,</i> |
| 958 | Cross-sectional area of base | $4.0 \ge 10^6$ | 4.9 x 10 ⁶ | 8.0 x 10 ⁶ | $2.0 \ge 10^6$ |
| 959 | of caldera block, $A(m^2)$ | | | | |
| 960 | | | | | |
| 961 | Duration of magma | 117,360 | 17,953 | 21,600 | 43,200 to |
| 962 | evacuation before caldera | | | | 86,400 |
| 963 | block begins to subside, T | | | | |
| 964 | (s) | | | | |
| 965 | | | - | 2 | |
| 966 | Magma evacuation | 2.8×10^4 to | 1.2×10^{5} to | 3.0×10^3 to | 4.5×10^{1} to |
| 967 | rate, α (m ³ s ⁻¹) | 2.2×10^{5} | 3.6×10^{5} | 6.7 x 10 ³ | $1.7 \ge 10^2$ |
| 968 | | | | | |
| 969 | | | | | |
| 970 | | | | | |
| 971 | Sources of data: Katmai, Hil | dreth and Fiers | <i>stein</i> [2000]; Pi | natubo, <i>Hoblitt</i> | t et al. [1996], |
| 972 | Koyaguchi and Ohno [2001], | , <i>Mori et al</i> . [19 | 96a, 1996b], S | <i>cott et al</i> . [1996 | 6]; Fernandina, |
| 973 | Simkin and Howard [1970], I | Filson et al. [19 | 73]; Miyakejir | na, <i>Kumagai et</i> | <i>al.</i> [2001], |
| 974 | Geshi et al. [2002], Kobayasi | hi et al. [2003], | <i>Irwan et al.</i> [2 | 003], Ueda et d | <i>ıl</i> . [2005]. |
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|------|-----------------|--|
| 992 | Table | 2. Symbols used in equations |
| 993 | | |
| 994 | | |
| 995 | | |
| 996 | | |
| 997 | Ε | energy of volcanotectonic earthquake (J) |
| 998 | E_s | energy of VLP event (J) |
| 999 | М | magnitude of volcanotectonic earthquake |
| 1000 | M_0 | seismic moment of VLP event (N m) |
| 1001 | Α | cross-sectional area at the base of caldera block (m ²) |
| 1002 | Η | thickness of caldera block (m) |
| 1003 | D | diameter of caldera block (m) |
| 1004 | R | aspect ratio of caldera block, H/D (dimensionless) |
| 1005 | Z. | displacement of caldera block (m) |
| 1006 | $V_{ m o}$ | initial volume of magma chamber prior to caldera subsidence (m ³) |
| 1007 | ΔV | change in volume (m ²) |
| 1008 | Т | time between initiation of magma chamber evacuation and downward movement |
| 1009 | | of caldera block (s) |
| 1010 | α | evacuation rate of magma (m ³ s ⁻¹) |
| 1011 | $F_{\rm s}$ | static friction (N) |
| 1012 | F_{d} | dynamic friction (N) |
| 1013 | p_{o} | original pressure in magma chamber (Pa) |
| 1014 | р | pressure increase in magma chamber from downward movement of caldera block |
| 1015 | 200 | (Pa) |
| 1016 | pT | underpressure in magma chamber (Pa) |
| 1017 | $pT_{\rm crit}$ | critical magma underpressure to cause failure of caldera root and initiate caldera |
| 1018 | , | collapse (Pa) $rate = 1$ $rate = 1$ |
| 1019 | p n | rate of pressure decrease in magma chamber (Pa's) |
| 1020 | p_1 | lithostatic prossure (Pa) |
| 1021 | Plith | prossure difference (Pa) |
| 1022 | Δp | hulls modulus of magma (Da) |
| 1023 | K T | oritical shear stress for failure of caldere block (Pa) |
| 1024 | ι_c | cohesion of caldera block (Pa) |
| 1025 | ι ₀ | mean stress normal to the plane of failure (Pa) |
| 1020 | U U | coefficient of internal friction of the caldera block (dimensionless) |
| 1027 | μ k | constant (~ 0.6) (dimensionless) |
| 1020 | л От | density of caldera block (kg m^{-3}) |
| 1029 | Pr Ø | acceleration due to gravity (m s ⁻²) |
| 1031 | 0 | |
| 1032 | | |
| 1033 | | |
| 1034 | | |
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|------|-----------------------------------|-------------------|---------------------|-----------------------|---------------------|
| 1038 | | | | | |
| 1039 | | | | | |
| 1040 | Table 3. Downward displa | cement of the c | caldera block as | a function of 1 | nagma |
| 1041 | evacuation rate. | | | | |
| 1042 | | | | | |
| 1043 | | | | | |
| 1044 | | Katr | nai | Pina | tubo |
| 1045 | Parameter | Minimum | Maximum | Minimum | Maximum |
| 1046 | | | | | |
| 1047 | | | - | - | - |
| 1048 | Observed evacuation | 2.8×10^4 | 2.2×10^{5} | 1.2×10^{5} | 3.6×10^{5} |
| 1049 | rate of magma $(m^3 s^{-1})$ | | | | |
| 1050 | | | | | |
| 1051 | Calculated displacement | 822 | 6484 | 440 | 1320 |
| 1052 | of the caldera block (m) | | | | |
| 1053 | | | | | |
| 1054 | Observed surface | 1200-1300 | | ~900 | |
| 1055 | displacement of the | | | | |
| 1056 | caldera block (m) | | | | |
| 1057 | | | | | |
| 1058 | | | | | |
| 1059 | | | F 6 0 0 0 1 | | 1.5.7 1 11 |
| 1060 | Sources of data: Katmai, <i>E</i> | lildreth and Fi | erstein [2000]; | Pinatubo, <i>Jone</i> | s and Newhall |
| 1061 | [1996], Koyaguchi and Ohi | 10 [2001], Scot | t et al. [1996]. | | |
| 1062 | | | | | |
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| 1064 | | | | | |
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| 1066 | | | | | |



Fig. 1



Fig. 2



Cumulative energy released (Joules)



Caldera subsidence is initiated when $p'T_{crit}A = 2(F_s - F_d)$



Underpressure





Fig. 6

Evacuation rate





