

Deformation regime and long-term precursors to eruption at large calderas:

Rabaul, Papua New Guinea.

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5

6 ABSTRACT

7 Eruptions at large calderas are normally preceded by variable rates of unrest that continue for
8 decades or more. A classic example is the 1994 eruption of Rabaul caldera, in Papua New
9 Guinea, which began after 23 years of surface uplift and volcano-tectonic (VT) seismicity at
10 rates that changed unevenly with time by an order of magnitude. Although the VT event rate
11 and uplift rate peaked in 1983-1985, eruptions only began a decade later and followed just 27
12 hours of anomalous changes in precursory signal. Here we argue that the entire 23 years of
13 unrest belongs to a single sequence of damage accumulation in the crust and that, in 1991-
14 1992, the crust's response to applied stress changed from quasi-elastic (elastic deformation
15 with minor fault movement) to inelastic (deformation predominantly by fault movement
16 alone). The change in behaviour yields limiting trends in the variation of VT event rate with
17 deformation and can be quantified with a mean-field model for an elastic crust that contains a
18 dispersed population of small faults. The results show that identifying the deformation regime
19 for elastic-brittle crust provides new criteria for using precursory time series to evaluate the
20 potential for eruption. They suggest that, in the quasi-elastic regime, short-term increases in
21 rates of deformation and VT events are unreliable indicators of an imminent eruption, but
22 that, in the inelastic regime, the precursory rates may follow hyperbolic increases with time
23 and offer the promise of developing forecasts of eruption as much as months beforehand.

24

25

26 *Key words:* Large Caldera; Rabaul; Eruption Precursors; Eruption Forecasts; Volcano-
27 Tectonic Seismicity; Ground Deformation.

28

29 **1. Introduction**

30 Large calderas, with surface areas of 100 km² or more, are among the most populated active
31 volcanoes on Earth. At least 138 have records of historical unrest (Newhall and Dzurisin,
32 1988) and examples that have provoked recent emergencies include Rabaul in Papua New
33 Guinea (McKee et al., 1984; Nairn et al., 1995), Campi Flegrei in Italy (Barberi et al., 1984),
34 Long Valley in the USA (Hill et al., 2002), and Santorini in Greece (Parks et al., 2012).
35 During such emergencies, elevated unrest continues for ~0.1-1 years and is characterised by
36 caldera-wide uplift and volcano-tectonic (VT) events within the caldera to depths of
37 kilometres. Most episodes do not culminate in eruption and their activity has been attributed
38 to a combination of the intrusion of magma at depths of about 5 km or less and increased rates
39 of fluid circulation in near-surface hydrothermal systems (McKee et al., 1984; Battaglia and
40 Vasco, 2006; Geyer and Gottsmann, 2010; Woo and Kilburn, 2010; Bodnar et al., 2011; Parks
41 et al., 2012; Acocella et al., 2015).

42 Even though short-term emergencies tend to be evaluated independently as regards the
43 probability of eruption (McKee et al., 2004; Hill, 2006), it has long been recognised that they
44 may belong to longer-term unrest that will trigger an eruption only when a cumulative
45 threshold has been exceeded (Newhall and Dzurisin, 1988; De Natale et al., 2006; Hill, 2006;
46 Acocella et al., 2015). However, the connection between long-term unrest and eruption
47 potential has been described only qualitatively. Using data from Rabaul caldera, we argue (1)
48 that VT and deformation precursors at large calderas may belong to a unified sequence that
49 can be quantified over decadal timescales, and (2) that conditions for eruption are determined
50 by the transition from a quasi-elastic to inelastic response of the crust to applied stress. The
51 results confirm that, on their own, short-term changes in rates of unrest are unreliable guides
52 to the potential for eruption. They also identify new practical procedures for evaluating a
53 caldera's approach to eruption.

54

55 **2. Long-term unrest at Rabaul caldera.**

56 Among large calderas, Rabaul has a unique modern record of precursory unrest that lasted for
57 more than two decades, between 1971 and 1994, and included a non-eruptive emergency
58 during 1983-1985 before eruptions began on 19 September 1994. The caldera lies on the
59 northeastern coast of New Britain Island in Papua New Guinea (Fig. 1). At least five episodes
60 of collapse have occurred since the Late Pleistocene across an area about 14 by 9 km across,
61 most of which is now submerged beneath Blanche Bay and opens eastward into the Bismarck
62 Sea (Nairn et al., 1995). The most recent caldera collapse occurred 1,400 years ago (Nairn et
63 al., 1995). It formed an elliptical structure, about 10 by 6.5 km across and aligned
64 approximately North-South, since when andesitic-dacitic eruptions have occurred around its
65 margin at Tavurvur, Vulcan, Rabalanakaia and Sulphur Creek (Fig. 1; Nairn et al., 1995;
66 Wood et al., 1995; Johnson et al., 2010). The historical record dates back to the 18th Century
67 and consists of at least six events, in 1767, 1791, 1850, 1878, 1937-43 and 1994-Present
68 (McKee et al., 1985; Blong and McKee, 1994; Johnson et al., 2010). During the last three
69 eruptions, activity started almost simultaneously at Vulcan and Tavurvur, on opposite sides of
70 the caldera (Fig. 1), with the later stages becoming restricted to Tavurvur (Blong and McKee,
71 1994).

72 Unrest before the 1994 eruptions was recognized in late 1971 after 54 years of quiescence
73 (McKee et al., 1984; Mori et al., 1989). Six months after two tectonic earthquakes of local
74 magnitude (M_L) 8.0 had occurred in the Solomon Sea (Everingham, 1975), volcano-tectonic
75 (VT) earthquakes began to be detected at mean rates greater than the previously typical values
76 of 50-100 events per month (McKee et al., 1985). They were located at depths of about 4 km
77 or less and associated with normal and subsidiary reverse fault displacements (Mori and
78 McKee, 1987; Mori et al., 1989; Nairn et al., 1995; Jones and Stewart, 1997; Johnson et al.,
79 2010). Their epicentres were concentrated within the topographic expression of the 1,400 BP
80 caldera (Fig. 2), indicating a narrower zone of ring faults at shallow depth (Mori et al., 1989).

81 Approaching 2×10^5 VT events were recorded throughout unrest (Fig. 3). Most had local
82 magnitudes between about 0.5 and 2.0, although their relative frequency could not be
83 determined reliably for the full 23 years of unrest (Mori et al., 1989; Johnson et al., 2010).
84 Among the $\sim 10^4$ VT events recorded during the 1983-85 volcano-seismic crisis, the maximum
85 local magnitude was 5.1 and fewer than one percent had local magnitudes greater than 3.0
86 (Mori et al., 1989). For unspecified completeness magnitudes, estimates of the seismic *b*-
87 value during the crisis lie between 0.8 and 1.1 (Mori et al., 1989) and so, assuming a value of
88 1, fewer than 10% are expected to have had local magnitudes greater than 2.0.

89 Between 1971 and 1994, the VT event rate recorded by the same seismic network
90 (Itikarai, I., pers. comm.) showed mean values of 100-200 per month, rising and falling over
91 intervals of years around peak values from about 600 per month in 1973 to almost 2,000 per
92 month in 1993 (Fig. 3; McKee et al., 1985; Itikarai, 2008; Johnson et al., 2010). This trend
93 was interrupted between September 1983 and July 1985 by a surge in monthly numbers that
94 peaked at almost 14,000 (Fig. 3).

95 From 1973, the changes in VT event rate were accompanied by uplift along a deformation
96 route that ran northwards from Matupit Island (Fig. 3; McKee et al., 1984; Saunders, 2001).
97 The uplift decreased with distance from a maximum value measured on the south coast of the
98 island. Although uplift was recorded over about 6 km, some 80% occurred within 2.5 km of
99 the location of peak movement and is consistent with deformation over a pressure source at a
100 depth of about 2 km (McKee et al., 1984). At Matupit Island itself, the peak rate of uplift
101 varied from about 0.02 to 0.4 m per year, with the largest rates occurring during the early
102 stages of unrest and the 1983-85 peak in VT event rate (Fig. 3; McKee et al., 1984; Blong and
103 McKee, 1994; Saunders, 2001; Johnson et al., 2010).

104

105

106 **3. Precursors to eruption**

107 The VT and deformation trends show three remarkable features (Fig. 3). First, their annual
108 rates with time varied by at least one order of magnitude. Second, although both precursors
109 showed high rates during the 1983-85 crisis, their mean rates did not vary in direct proportion
110 for most of the unrest (Fig. 3). Third, although some 40% of the total uplift and number of VT
111 events occurred during 1983-85, this crisis did not culminate in volcanic activity; in contrast,
112 the eventual eruption on 19 September 1994 began without significant changes in the rate of
113 either precursor until 27 hours beforehand, when rates of seismicity increased to peak values
114 of about two felt events per minute and the ground was uplifted by several metres at least near
115 Vulcan and, also, Matupit Island, 2.5 km west of Tavurvur (Global Volcanism Program,
116 1994a, b; Blong and McKee, 1995); explosive activity finally commenced at 06.05 (local
117 time) at Vulcan and 07.17 at Tavurvur (Global Volcanism Program, 1994a). The first feature
118 shows that unrest at Rabaul occurred under varying rates of deformation and fault movement;
119 the second shows that deformation and VT precursors do not necessarily show the same types
120 of variation with time; and the third shows that, on their own, accelerations in precursory
121 signals with time may not be sufficient for assessing the potential for eruption.

122 Even when deformation and VT seismicity show contrasting behaviour with time, they
123 are expected to show preferred variations against each other, because both are responses of
124 the crust to changes in applied stress (Kilburn, 2012). In the case of Rabaul, Fig. 4 shows the
125 variation of VT events with uplift for the entire period of unrest before the 1994 eruption.
126 Following previous studies, uplift is represented by measurements from Matupit Island, where
127 the maximum amount of movement was recorded (McKee et al., 1984, 1989; Blong and
128 McKee, 1995; Johnson et al., 2010), although geodetic models suggest that the measured
129 value may have been about 85% of the potential absolute maximum towards the centre of the
130 caldera offshore (McKee et al., 1984).

131 The data show coherent trends between VT event number and uplift (Fig. 4).
132 Qualitatively, the number of VT events tends to accelerate with deformation for uplifts less
133 than c. 1.9 m (from 1973 to the end of 1988), followed by a transition to a linear trend for
134 uplifts larger than 2 m (from at least December 1991 to September 1994). The accelerating
135 phase may belong to a single trend or be composed of a non-linear increase for uplifts less
136 than c. 1.2 m and a quasi-linear increase for uplifts between 1.2 and 1.9 m, at a mean rate
137 faster than that for the uplift beyond 1.9 m. On its own, therefore, simple curve fitting cannot
138 provide a unique interpretation of the VT-deformation trends. However, as described in the
139 next section, a preferred interpretation does naturally emerge when additional constraints
140 from elastic-brittle deformation are taken into account.

141

142 **4. Regimes of deformation**

143 *4.1. Limiting regimes of deformation*

144 The emergence of different trends in Fig. 4 suggests an evolution in conditions controlling
145 Rabaul's deformation. The number of VT events and uplift are proxies, respectively, for
146 inelastic and total deformation of the crust. The field trend can thus be interpreted in terms of
147 the variation with deformation in the contribution from inelastic behaviour and this, in turn,
148 can be represented on a conventional stress-strain diagram for an elastic-brittle crust (Fig. 5).
149 When a differential stress is applied (*e.g.*, from overpressure in a magma body), the crust
150 ideally first responds elastically, so that the stress is supported by straining unbroken rock.
151 Above a critical differential stress, a proportion of the stress is lost inelastically by the onset
152 of fault slip (Fig. 5). The inelastic proportion increases with stress at an accelerating rate until
153 it becomes the sole mode of response (Fig. 5). The rates of stress increase and loss eventually
154 balance each other, so that the mean differential stress remains unchanged (Fig. 5). At this
155 stage, the total rate of deformation is determined by the rate of faulting, favouring a linear
156 increase between measures of deformation and seismicity. The field trends at Rabaul (Fig. 4)

157 are qualitatively similar to the evolution in deformation regime; the conditions for elastic-
 158 brittle behaviour may therefore yield also a quantitative description of the caldera's unrest.

159 Rabaul's VT events have typical magnitudes of less than 2 and record the movements of
 160 faults ~0.01-0.1 km across, much smaller than the dimensions of ~km for the deforming crust.
 161 The crust's bulk behaviour can thus be approximated to that of an elastic medium containing
 162 a dispersed population of relatively small faults. The rate of faulting is governed by slow
 163 crack growth around fault tips. Until a critical rate is achieved, continued cracking requires an
 164 increase in applied stress, as observed for quasi-elastic deformation (Kilburn, 2012); at faster
 165 rates, crack growth becomes a self-accelerating process even when the applied stress remains
 166 constant, as expected in the inelastic regime (Kilburn, 2003). Bulk deformation can therefore
 167 be described by a mean-field model for a crust with a population of small faults that slip at
 168 rates determined by slow crack growth (Kilburn, 2003, 2012).

169

170 4.2. Quasi-elastic deformation

171 For the quasi-elastic regime, the mean-field, slow-crack model yields an exponential increase
 172 in the mean inelastic deformation, ε_{in} , with total mean deformation, ε_t (Kilburn, 2012):

173

$$174 \quad \frac{d\varepsilon_{in}}{d\varepsilon_t} = \left(\frac{d\varepsilon_{in}}{d\varepsilon_t} \right)_{st} \exp \left(\frac{\varepsilon_t - \varepsilon_{QE,m}}{\varepsilon_{ch}} \right) \quad (1)$$

175

176 where $\varepsilon_{QE,m}$ is the total mean deformation at the end of quasi-elastic behaviour, when $d\varepsilon_{in}/d\varepsilon_t$
 177 $= (d\varepsilon_{in}/d\varepsilon_t)_{st}$, and ε_{ch} is a characteristic deformation. Setting the total number of VT events ΣN
 178 $\propto \varepsilon_{in}$ and uplift $h \propto \varepsilon_t$, Eq. (1) becomes:

179

$$180 \quad \frac{dN}{dh} = \left(\frac{dN}{dh} \right)_0 \exp \left(\frac{h}{h_{ch}} \right) \quad (2)$$

181

182 where $(dN/dh)_0 = (dN/dh)_{st} \exp(-h/h_{ch})$ is the value of dN/dh at the onset of quasi-
183 elastic behaviour, for which h is set at 0. (Note that some uplift is expected to have occurred
184 before the onset of the exponential trend, so that $h = 0$ in Eq. (2) corresponds to finite initial
185 uplift in the field). Integrating Eq. (2) with uplift then leads to

186

$$187 \quad \Sigma N = (\Sigma N)_0 \exp\left(\frac{h}{h_{ch}}\right) \quad (3)$$

188

189 where $(\Sigma N)_0$ denotes the number of VT events detected at $h = 0$.

190 Eq. (3) is consistent with observations at Rabaul for uplifts less than 1.9 m, yielding best-
191 fit values of $(\Sigma N)_0$ and h_{ch} of 4,120 events and 0.53 m (Fig. 4). The agreement supports the
192 interpretation that a single VT-deformation trend characterised this phase and reflects early
193 deformation of the caldera's crust in the quasi-elastic regime.

194 The model further provides an upper limit on the quasi-elastic value for h/h_{ch} . The total
195 quasi-elastic strain increases approximately in proportion with stress. The assumption that $h \propto$
196 ε_i , therefore leads to $h/h_{ch} \approx S_d/S_{ch}$, the ratio of applied differential stress to a characteristic
197 stress, defined below. The maximum quasi-elastic value for h/h_{ch} then represents S_F/S_{ch} ,
198 where S_F is the differential failure stress of the crust appropriate to the particular conditions of
199 loading (*e.g.*, in compression or extension). S_{ch} is the stress potentially available to deform
200 bonds as a result of fluctuations in atomic configuration (Kilburn, 2012). Its value depends on
201 the mode of deformation. For shear failure in compression, it is equivalent to the maximum
202 value $S^* = (3kT\phi + P_c - P_p)/3$, where T is absolute temperature (K), P_c and P_p are the
203 confining and pore-fluid pressures, k is the Boltzmann constant (1.381×10^{23} J molecule⁻¹ K⁻¹)
204 and ϕ is the number of molecules per unit volume, which in turn depends on rock chemistry
205 and density (Kilburn, 2012). Failure in tension, however, is limited only by the smaller stress

206 necessary to pull bonds apart, equivalent to the tensile strength σ_T . For extension, therefore,
207 $S_{ch} = \sigma_T < S^*$ and is the preferred characteristic stress for a crust stretching during uplift.

208 Different modes of failure are associated with specific ranges for the ratio S_F/σ_T (Shaw,
209 1980). In an extending crust, $S_F/\sigma_T \leq 4$ for failure in tension, but lies between 4 and 5.5 for
210 failure in combined tension and shear; in comparison, $S_F/\sigma_T > 5.5$ for failure in compression
211 (Shaw, 1980). At Rabaul, a maximum quasi-elastic uplift of 1.9 m was recorded at Matupit
212 Island (Fig. 4), with a potential maximum of 2.2 m near the presumed centre of uplift
213 offshore. Given $h_{ch} = 0.53$ m, therefore, the maximum value for h/h_{ch} yields the range 3.6-4.1
214 for S_F/σ_T , further supporting the interpretation of quasi-elastic crustal extension for the
215 exponential VT-uplift trend.

216

217 *4.3. Inelastic deformation*

218 In the inelastic regime, changes in total deformation are determined only by fault movement,
219 so that measures of total and inelastic deformation increase in proportion to each other. The
220 substitute parameters uplift and VT event number also increase together to yield a linear
221 trend:

222

$$223 \quad \Sigma N - (\Sigma N)_{i,0} = (dN/dh)_i (h - h_{i,0}) \quad (4)$$

224

225 where the gradient $(dN/dh)_i$ is the change in VT event number per unit uplift and the subscript
226 “ $i,0$ ” denotes values at the start of the inelastic regime. Such a trend well describes the VT-
227 deformation behaviour at Rabaul for uplifts greater than 2 m and yields a best-fit gradient of
228 6.23×10^4 events per metre (Fig. 4). The change from an exponential to a linear trend is thus
229 consistent with the transition from quasi-elastic to inelastic deformation of Rabaul’s crust.

230 Under the constant applied stress of the inelastic regime, the mean field model shows that
 231 the approach to bulk failure is characterised by mean rates of inelastic and total deformation
 232 that accelerate hyperbolically with time (Kilburn, 2003), equivalent to a linear decrease in the
 233 inverse event rate:

234

$$235 \left(\frac{d\varepsilon_{in}}{dt} \right)^{-1} = \left(\frac{d\varepsilon_t}{dt} \right)^{-1} = \left(\frac{d\varepsilon_t}{dt} \right)_1^{-1} \left[1 - \left(\frac{t-t_1}{\tau} \right) \right] \quad (5)$$

236

237 where $(d\varepsilon_t/dt)_1$ ($= (d\varepsilon_{in}/dt)_1$) denotes the mean deformation rate at the start of the hyperbolic
 238 trend, when time $t = t_1$. The timescale τ defines the duration of the trend, which depends on
 239 the failure strength of the crust and the rate of stress concentration around fault tips (Kilburn,
 240 2003).

241 The corresponding trends for changes in the mean rates of deformation and VT number
 242 will show the same form as Eq. (5). For example, the inverse deformation rate is given by:

243

$$244 \left(\frac{dh}{dt} \right)^{-1} = \left(\frac{dh}{dt} \right)_1^{-1} \left[1 - \left(\frac{t-t_1}{\tau} \right) \right] \quad (6)$$

245

246 which has a linear gradient of $-1/(\tau (dh/dt)_1)$.

247 The inelastic trend had become established by the beginning of 1992, when uplift at
 248 Matupit Island had reached 2 m (Fig. 4). Fig. 6 shows the associated inverse mean rate of
 249 uplift, derived from interpolations of the uplift data in Blong and McKee (1995) and Johnson
 250 et al. (2010) until the last date of published measurements in March 1994. The mean inverse
 251 rates rapidly settle to a trend that for at least 680 days is well-approximated by a linear
 252 decrease with time following a best-fit gradient of 14.4 days m^{-1} . The trend thus supports the

253 interpretation that, by the start of 1992, Rabaul's crust had entered an inelastic regime of
254 deformation controlled by self-accelerating crack growth.

255 Manipulating Eq. (6), the hyperbolic increase in contemporaneous VT event rate is:

256

$$257 \quad \left(\frac{dN}{dt}\right) = \left(\frac{dN}{dh}\right)\left(\frac{dh}{dt}\right) = \left(\frac{dN}{dh}\right)\left(\frac{dh}{dt}\right)_1 \left[1 - \left(\frac{t-t_1}{\tau}\right)\right]^{-1} \quad (7)$$

258

259 where the rate of change of VT event number with uplift, dN/dh , is assumed to be constant at
260 6.23×10^4 events m^{-1} , given by the mean gradient of the inelastic trend line in Fig. 4.

261 Fig. 6 compares the observed and calculated variations in VT event number with time.
262 The two trends agree well during the start of the inelastic regime, but begin to diverge after
263 about 11 months. At this time, the rate of increase in event rate with time changes from
264 hyperbolic to approximately constant, yielding values as much as 3% greater than expected
265 from the calculated trend (Fig. 6). Although the difference in values is small enough to give
266 an approximate proportionality between ΣN and h in Fig. 4, the difference suggests a
267 systematic deviation from the expected trend.

268 Relaxing the assumption that dN/dh is constant, Eq. (7) shows that a linear increase in VT
269 number could be produced if dN/dh decreased with time to counterbalance the acceleration in
270 dh/dt . A decrease in dN/dh is equivalent to an increase in the amount of uplift per fault
271 movement. Two mechanisms favouring such a condition are the coalescence of faults and the
272 opening of discontinuities much larger than the faults triggering VT events. In the first case,
273 the coalescence of faults would yield an increase in the average magnitude of VT events, so
274 reducing the number of events associated with a given amount of uplift (Meredith et al.,
275 1990). In the second case, upward bending of the crust around a magma body (Pollard and
276 Johnson, 1973) would favour the progressive opening of pre-existing, sub-vertical
277 discontinuities in the ring-fault system, as well as an upward migration in the depth at which

278 opening can occur (Fig. 7). Slip in the crust around zones of opening would not necessarily
279 require any coalescence of faults and so not lead to a significant change in the average
280 magnitude of VT events. At the same time, opening of the discontinuities themselves need not
281 generate significant seismicity, because tensile stresses normal to the plane of a discontinuity
282 do not favour slip along its length (Troise et al., 1997).

283 A decrease in dN/dh would produce total numbers of events smaller than those expected
284 from the calculated trend and, by itself, cannot account for the observed increase above the
285 expected values. Larger numbers of VT events could be obtained by an increase in the total
286 number of faults being activated around the caldera rim. Coalescence would favour a decrease
287 in fault number (Smith and Kilburn, 2010), whereas growth of a large discontinuity would
288 favour an increase in the number of smaller faults activated in its vicinity. Qualitatively,
289 therefore, the deviation of VT event numbers from the idealised inelastic trend may reflect the
290 onset of significant stretching and upward opening of the ring-fault system around the rim of
291 the caldera. An opening fault would in addition provide an initial pathway from which magma
292 escaping from the shallow reservoir could force its way to the surface (Fig. 7). Indeed, the
293 propagation of magmatic fractures could account for the increases in local seismicity and
294 uplift observed during the 27 hours before the start of the 1994 eruptions (Global Volcanism
295 Program, 1994b; Blong and McKee, 1995).

296 Saunders (2001) has also proposed that Rabaul's deformation was controlled by the
297 ascent of magma through caldera ring faults. Magma was considered to have been forcefully
298 injected into faults throughout unrest, generating an overpressure in the resulting dykes that
299 induced lateral compression and associated uplift of the caldera floor. Such a mechanism is
300 distinct from the current interpretation, in which overpressure in a magma reservoir favours
301 surface uplift and inward bending of the crust near the reservoir until the ring faults open
302 sufficiently to favour magma ascent (Fig. 7).

303

304 5. Discussion

305 5.1. The shallow magmatic system at Rabaul caldera.

306 The 1994 eruptions at Rabaul were preceded by a progressive extension of the caldera's crust
307 from at least 1971. During 1991-1992, the style of deformation evolved from quasi-elastic to
308 inelastic, as shown by the following: (1) the VT event number increased at first exponentially
309 and then linearly with uplift (Fig. 4), (2) the transition between trends occurred when the
310 inferred ratio of applied stress to tensile strength was approximately 4, and (3) the final
311 inelastic behaviour was characterised by a linear decrease with time in the inverse mean rate
312 of uplift, equivalent to a hyperbolic increase in the mean rate of deformation (Fig. 6).

313 The onset of bulk failure is expected to have occurred at or near the margin of a magma
314 body, either through the equivalent of hydraulic failure of the body itself, or by the opening of
315 a major discontinuity at the caldera's boundary. In either case, the recorded VT events reflect
316 slip along small faults in the crust surrounding the magma body or major discontinuity.

317 In the case of hydraulic failure, the maximum compressive stress (σ_1) at the margin of the
318 magma body is the sum $P_{e,0} + \Delta P_m$ of the effective confining pressure (the confining pressure
319 – pore-fluid pressure) at the start of deformation and the overpressure imposed by the magma
320 body (Gudmundsson, 2011a). Opening of a ring fault may instead occur as the crust of the
321 uplifting caldera is pulled inwards and away from its periphery (Fig. 7), in which case $\sigma_1 \approx$
322 $P_{e,0}$. For crustal failure in extension, therefore, it is expected that $\sigma_1 \geq P_{e,0}$ which, assuming
323 initially lithostatic conditions, is given by $\rho_{cr}gz$ and $\rho_{cr}(1 - (\rho_f/\rho_{cr}))gz$ for dry and fluid-
324 saturated crust, where ρ_{cr} and ρ_f are the mean density of the crust and pore fluid, g is gravity
325 and z is the depth at which failure begins. The corresponding differential stress, $S_d \approx \sigma_1 + \sigma_T$,
326 which gives $S_d/\sigma_T = (\sigma_1/\sigma_T) + 1$. Since inelastic deformation occurs under the differential
327 stress established at the end of the quasi-elastic regime, the maximum value of S_d/σ_T is 3.6-
328 4.1, so that σ_1 lies between $2.6\sigma_T$ and $3.1\sigma_T$ or, for a notional tensile strength of 10 MPa, in

329 the range 26-31 MPa. For a crust with an assumed mean density of 2,200 kg m⁻³ and an
330 aqueous pore-fluid with a density of 1,000 kg m⁻³, pressures of 26-31 MPa correspond to
331 maximum depths of c. 1.2 km and 2.6 km for dry and water-saturated conditions. Given that
332 most of the caldera lies under water (Fig. 1), the preferred solution is for a fluid-saturated
333 crust with a pressure source at depths of less than 2.6 km, consistent with the geodetic
334 modelling of McKee et al. (1984) and Geyer and Gottsmann (2010).

335 Remarkably, the types of deformation regime are revealed by the numbers of VT events
336 and uplift as simple measures of seismicity and deformation. No filtering has been applied,
337 for example, to select VT events within a prescribed range of completeness magnitudes.
338 Throughout unrest, therefore, the proxy measures appear to have remained in the same
339 relative proportion to the amounts of inelastic and elastic deformation and imply that, at
340 Rabaul: (1) the bulk mechanical behaviour of the crust can be approximated to that of a
341 medium containing a dispersed population of relatively small discontinuities; (2) the
342 magnitude-frequency distribution of VT events remained approximately constant; and (3) the
343 essential geometry of deformation also remained approximately constant.

344

345 *5.2. Applications to large calderas in general.*

346 The evolution from quasi-elastic to inelastic deformation is expected to be common during
347 long-term unrest at large calderas. Conditions for recognising the evolution were particularly
348 favourable at Rabaul because uplift continued throughout unrest to provide a simple measure
349 of increasing deformation. Other large calderas may instead show alternating episodes of
350 uplift and subsidence over years to decades, such as at Campi Flegrei in southern Italy (Piochi
351 et al., 2013) and Yellowstone in Wyoming, U.S.A. (Dzurisin, 2007). Such short-term
352 oscillations must be removed from the total amount of ground movement before the elastic-
353 brittle model can be applied to the sustained, long-term changes in crustal deformation.

354 Short-term oscillations have been attributed to changes in pore-fluid pressure in shallow
355 hydrothermal systems heated by underlying magma or magmatic gases (Fournier, 1989;
356 Troiano et al., 2011). Changes in pore-fluid pressure alter the isotropic confining pressure on
357 rock (Secor, 1965) and, hence, the total volume of the crust on which a differential stress is
358 being applied. They allow ground movement to occur without altering the magnitude of either
359 the applied differential stress or the difference between between total and elastic deformation.
360 Ideally, therefore, increases and decreases in pore-fluid pressure can induce ground uplift and
361 subsidence without changing the applied differential stress or triggering significant additional
362 seismicity (notice that the ideal conditions do not refer to fluids becoming pressurized within
363 existing fractures, which will instead promote fracture growth and attendant seismicity).

364 Increased pore pressure, however, also reduces the differential stress at which bulk failure
365 occurs (Secor, 1965) and this reduction may trigger additional fracturing. Hence the idealised
366 case of aseismic ground movement following a change in pore pressure implicitly assumes
367 that induced changes in failure stress can be neglected. Such pre-failure behaviour can be
368 recognised on a VT-deformation diagram by abrupt changes in deformation without a change
369 in the total number of VT events (Fig. 8). Once recognised, the movement due to changes in
370 pore-fluid pressure can be removed before applying the elastic-brittle model to the remaining,
371 sustained deformation caused by changes in differential stress (Fig. 8).

372 Deviations from model conditions will also occur if the crust develops a rheology outside
373 elastic-brittle conditions. For example, Di Luccio et al. (2015) have suggested that long-term
374 exposure to high temperatures may allow the emergence of plastic deformation, which would
375 favour surface movement under a constant differential stress with fewer VT events than
376 expected from the elastic-brittle model. The significance of such processes will vary among
377 calderas and so needs to be evaluated on a case-by-case basis. By virtue of its simplicity, the
378 elastic-brittle sequence provides a reference model with which to compare observed patterns

379 of unrest and, when significant deviations from the model are observed, to provide a starting
380 point for identifying the physical causes of the deviations.

381

382 *5.3 Forecasting intra-caldera eruptions after long repose intervals.*

383 The elastic-brittle trends show how the regime of crustal deformation evolves during the
384 approach to bulk failure, independent of the time required for stress and strain to be
385 accumulated. Although eruptions after extended repose are normally preceded by bulk failure,
386 this does not mean that all episodes of bulk failure are followed by eruptions (Bell and
387 Kilburn, 2011). Failure may produce a major discontinuity in the crust that does not intersect
388 a magma body; even when magma does enter a new discontinuity, structures in the crust may
389 favour the emplacement of a sill, rather than a dyke (Gudmundsson, 2011b; Woo and Kilburn,
390 2010), and, in the second case, magmatic pressure gradients may be too small to allow the
391 fracture to reach the surface (Gudmundsson, 2002). Precursory sequences may thus culminate
392 in a seismic swarm or intrusion, instead of an eruption. However, given that an eruption might
393 occur, the analysis of Rabaul's precursory sequence has identified new procedures for
394 evaluating the possibility of such an outcome.

395 First, the potential for eruption depends on the proportion of inelastic behaviour during
396 deformation. In addition to monitoring changes in seismic and geodetic precursors with time,
397 therefore, it is important to follow changes in VT event number with a measure of
398 deformation to determine the regime of crustal deformation.

399 Second, an eruption is most likely to occur after the transition to inelastic behaviour. The
400 transition can be recognised by a change from an exponential to a linear trend in the variation
401 of VT event number with deformation; additional support for a physical change can be
402 obtained from the maximum value of the deformation ratio (in this case h/h_{ch}) at the end of
403 quasi-elastic behaviour, which for deformation in tension is not expected to exceed 4.

404 Third, when inelastic behaviour has been identified, an accelerating approach to bulk
405 failure can be tested by checking for linear decreases in the inverse mean rates of deformation
406 or VT event number with time, following the classic method of Voight (1988). The preferred
407 time for eruption occurs when the inverse mean event rate becomes zero, because this is
408 equivalent to an approach to infinite rates and is interpreted to mark a catastrophic change in
409 the crust, such as bulk failure. At Rabaul, however, simple extrapolation of the inverse-rate of
410 deformation between January 1992 and March 1994 (Fig. 6) would have indicated a preferred
411 time for eruption at the beginning of June 1994, three months before it actually occurred.
412 Thus, although the linear trend indicates deformation in the inelastic regime, additional
413 analyses may be necessary before it can be utilised to make a specific forecast of an eruption
414 (Bell et al., 2013; Boué et al., 2015). Nevertheless, a sustained linear decrease in the inverse
415 mean rate of deformation may justify a progressive raising of alert levels in the vicinity of a
416 caldera.

417 Finally, the long-term evolution of the precursory sequence confirms that short-term
418 accelerations in individual precursors are not reliable indicators of an imminent eruption.
419 Instead, they reflect short-duration increases in the rate at which the crust is evolving along
420 the much longer term trend required before an eruption can occur. Thus the 1983-85 crisis at
421 Rabaul represents a two-year acceleration along a trend destined to last for 23 years (Fig. 4).
422 The acceleration is associated with a rapid pressure increase in the magmatic system (McKee
423 et al., 1984), perhaps caused by the injection of magma into a shallow reservoir following a
424 regional earthquake of Magnitude 7.6 in March 1983, 200 km east of the caldera (Acocella et
425 al., 2015).

426 The long precursory sequence contrasts with observations at stratovolcanoes, for which
427 precursory trends tend to evolve over a year or less (Voight, 1988; De La Cruz-Reyna and
428 Reyes-Davila, 2001; Kilburn, 2003; Bell and Kilburn, 2011). Short-term accelerations of
429 precursory signal with time thus have different implications for eruption potential at

430 stratovolcanoes and at large calderas, so that experience from one cannot be translated to the
431 other without first establishing whether a volcano is deforming in the quasi-elastic or inelastic
432 regime. Time-dependent protocols for emergency responses at large calderas are therefore
433 expected to be significantly different from those at stratovolcanoes.

434

435 **6. Conclusions**

436 The 1994 eruptions at Rabaul were preceded by a single sequence of unrest that continued for
437 23 years, during which crustal deformation evolved from the quasi-elastic to inelastic
438 regimes. Upward bending of the crust favoured the opening of ring faults that provided
439 weakened zones for magma to reach the surface. The change in regime was shown by a
440 change from an exponential to a linear increase in the VT event number against uplift, which
441 indirectly measured changes in inelastic against total deformation. Shorter-term accelerations
442 with time in precursory signals provided unreliable indications of the potential for eruption:
443 rapid changes in 1983-85 were not followed by eruption, whereas unremarkable changes in
444 1994 culminated in simultaneous eruptions at Tavurvur and Vulcan, on different sides of the
445 caldera. Additional measurements that identify the regime of deformation will improve
446 assessments of eruption potential and may reduce the possibilities of false alarms or failed
447 forecasts.

448 Rabaul's 1971-1994 sequence provided a particularly clear example of the change from
449 quasi-elastic to inelastic behaviour, because unrest was characterised by persistent uplift. At
450 other calderas, uplift may be interrupted by episodes of subsidence, owing to the effects of
451 local processes such as changes in pore-fluid pressures in geothermal systems. In such cases,
452 the effects of local processes must be filtered out before patterns of surface movement and
453 seismicity can be used to identify the regime of deformation.

454

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584 **Figure Captions**

585

586 **Figure 1.** Map of Rabaul caldera, showing the outlines of the Blanche Bay complex of nested
587 calderas (*large black dashes*) and of the most recent collapse in 1400 BP (*small, grey dashes*).
588 The 1971-1994 unrest ended with simultaneous eruptions from Tavurvur and Vulcan (*black*
589 *circles*) and provoked evacuation of Rabaul Town. Previous eruptions (*white circles*) have
590 occurred within and outside the Blanche Bay complex. (Modified from Johnson et al. (2010).)

591

592 **Figure 2.** The distribution of local seismicity recorded during 1971-1992. The VT events
593 occurred at depths of less than 4 km and their epicentres (*black dots*) appear to mark an inner
594 zone of faults associated with the 1400 BP caldera (*grey dashes*). Compared with events at
595 depths of 2 km or less (*left*), the distribution becomes extended towards the south among
596 events deeper than 2 km (*right*). Tavurvur and Vulcan (*black circles*) lie outside the outer
597 margin of the more shallow VT events. See Fig. 1 for names of additional locations.
598 (Modified from Jones and Stewart (1997) and Saunders (2001).)

599

600 **Figure 3.** During the 1971-1994 unrest, the monthly VT event rate (*columns*) and uplift at
601 Matupit Island (*line*) both varied irregularly with time (*left*). The 1983-1985 seismo-volcanic
602 crisis was characterised by increases in rate by at least one order of magnitude, but decayed
603 without an eruption. The eruptions in 1994 (*star*) were not preceded by anomalous behaviour
604 until 27 hours beforehand and the onset of a seismic swarm that raised the monthly number of
605 VT events to about 3,700 and caused uplift near Vulcan and Tavurvur by several metres (*not*
606 *shown*). The amount of long-term uplift decayed along the measuring line northward from
607 Matupit Island (*top right*). *REF* shows the reference base station and *A* and *B* the end stations
608 used to illustrate the pattern of uplift during 1971-1984 (*bottom right*). The uplift has been

609 normalised against the uplift recorded at Station *B* on Matupit Island. (Modified from (*left*)
610 Johnson et al. (2010), (*top right*) Saunders (2001) and (*bottom right*) McKee et al. (1984);
611 note that the times in the original Fig. 46 of Johnson et al. (2010) should be moved earlier by
612 six months.)

613

614 **Figure 4.** The variation of VT event number (ΣN) against uplift (h) at Matupit Island shows a
615 change from an exponential to linear trend at an uplift of about 1.9 m. The best-fit relations
616 (for h in m) are $\Sigma N = 4120 \exp(h/0.53)$ and $\Sigma N = 2.79 \times 10^4 + (6.23 \times 10^4)h$, with
617 corresponding r^2 regression coefficients of 0.98 and 0.99. The change in trends are consistent
618 with an evolution from the quasi-elastic to inelastic regimes of deformation (Fig. 5). The
619 shaded area shows conditions during the 1983-1985 seismo-volcanic crisis (Fig. 3).

620

621 **Figure 5.** As the applied differential stress increases (*left*), the deformation of an elastic-
622 brittle material (*black curve*) deviates by an increasing amount (*horizontal dashed arrows*)
623 from ideal elastic behaviour (*dashed line*). The deviation is caused by fracturing and slip
624 along discontinuities. The rate of stress increase is eventually balanced by the rate of stress
625 loss by slip (*circle*), after which deformation continues only by slip and the mean stress
626 remains constant at its failure value S_F . The change in regime is recorded in the field (*right*)
627 by a change from an accelerating to a constant rate of change in VT event number with total
628 deformation. The vertical bars illustrate conditions for elastic (*white*), quasi-elastic (*grey*) and
629 inelastic (*black*) behaviour. The continuous arrows show the direction of evolution with time
630 under an increasing stress. The expected field trends are observed during the 1971-1994
631 unrest at Rabaul (Fig. 4).

632

633 **Figure 6.** The inverse mean rate of uplift $\Delta t/\Delta h$ decays linearly with time t (*top*) during the
634 1992-1994 inelastic deformation at Rabaul (Fig. 4). The best-fit trend is described by $\Delta t/\Delta h =$
635 $12784 - 14.38 t$ ($r^2 = 0.99$; h in m and t in days from 01 January 1992); the bars show $\pm 10\%$
636 limits about the mean value. For ideal inelastic behaviour, the deformation trend yields an
637 acceleration in contemporaneous VT event number with time (*bottom, black curve*). Observed
638 VT event numbers follow the expected trend until Day 320, after which they instead follow a
639 constant rate of increase (*dashed line*, fit using only the shaded circles). The deviation may be
640 caused by the progressive opening of ring faults (Fig. 7). The vertical bars illustrate a 3%
641 increase above the VT numbers calculated from the deformation trend.

642

643 **Figure 7.** Bending of crust (*light grey*) around a pressurizing magma body (*dark grey*)
644 favours the inward opening of weak zones (*white*) within a caldera's system of ring faults
645 (*dashed lines*). As bending proceeds (*top to bottom*), the zone of opening extends towards the
646 surface and, when the walls of the magma body are breached, favours the ascent of magma
647 and its eruption at the surface. The opening migrated slowly upwards during the 1992-1994
648 interval of inelastic deformation, but was only utilized by ascending magma about 1-2 days
649 before eruption.

650

651 **Figure 8.** (*Left*) Changes in the pore-fluid pressure of hydrothermal systems may produce
652 transient variations in ground movement (*dot-dashed line*) in addition to the sustained
653 deformation caused by the pressurization or intrusion of an underlying magma body (*dashed*
654 *line*). The transient movements must be removed from the total observed deformation (*black*
655 *line*) before the elastic-brittle model can be applied. The example shows uplift accompanied
656 by seismicity, followed by aseismic subsidence as the thermal pulse decays and the pore-fluid
657 pressure in the hydrothermal system decreases to its pre-intrusion level. (*Centre and right*)

658 The later-stage decrease in pore–fluid pressure is revealed on stress-strain (S_d - ε_t) and total
659 seismicity-strain (ΣN - ε_t) diagrams by a decrease in strain under constant differential stress and
660 without significant seismicity (*Positions 3 to 4*), here illustrated for deformation in the quasi-
661 elastic regime. Removing the transient movement reveals the elastic-brittle trends associated
662 with changes in differential stress alone (*dashed curves from Positions 1 to 2 to 4*).

Figure 1

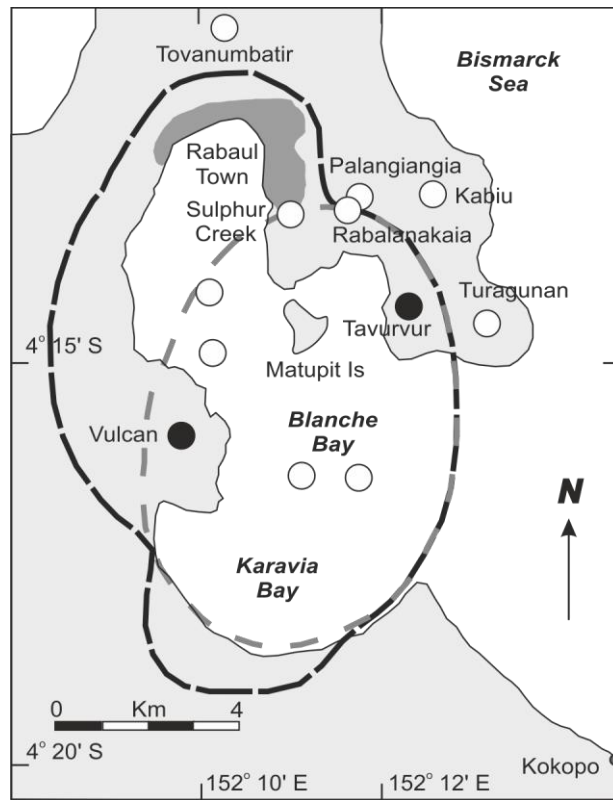


Figure 2

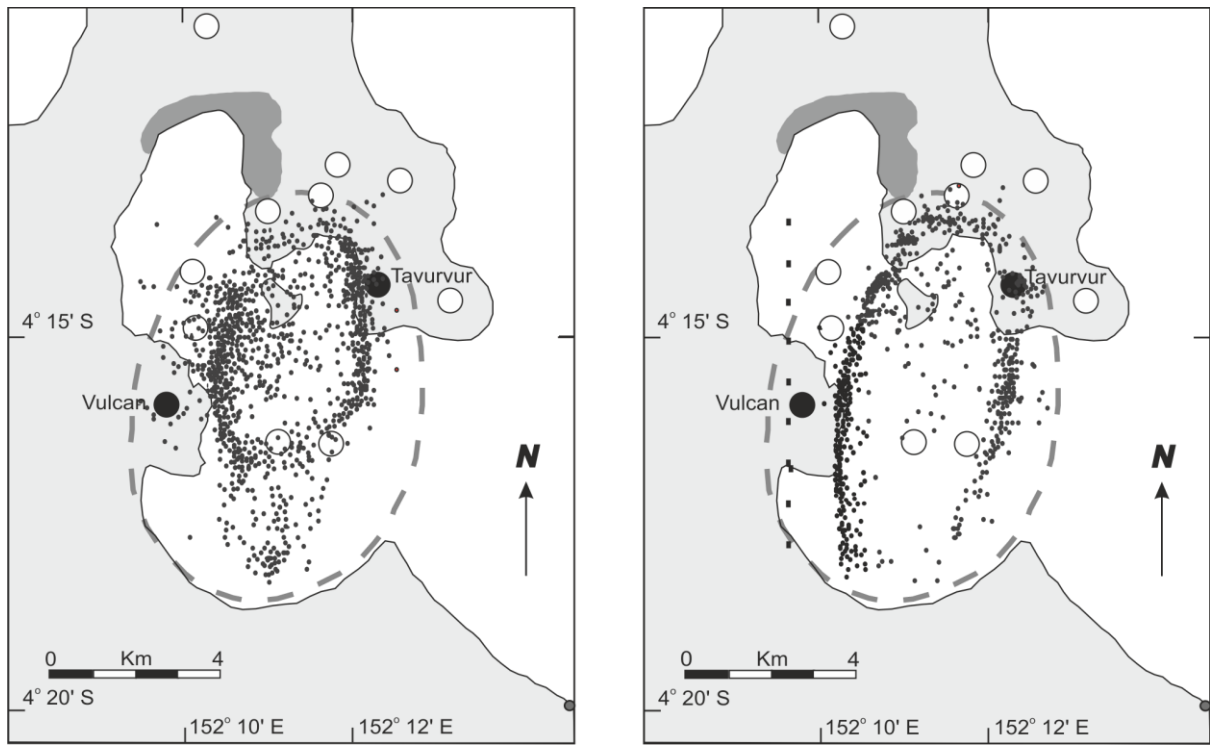


Figure 3

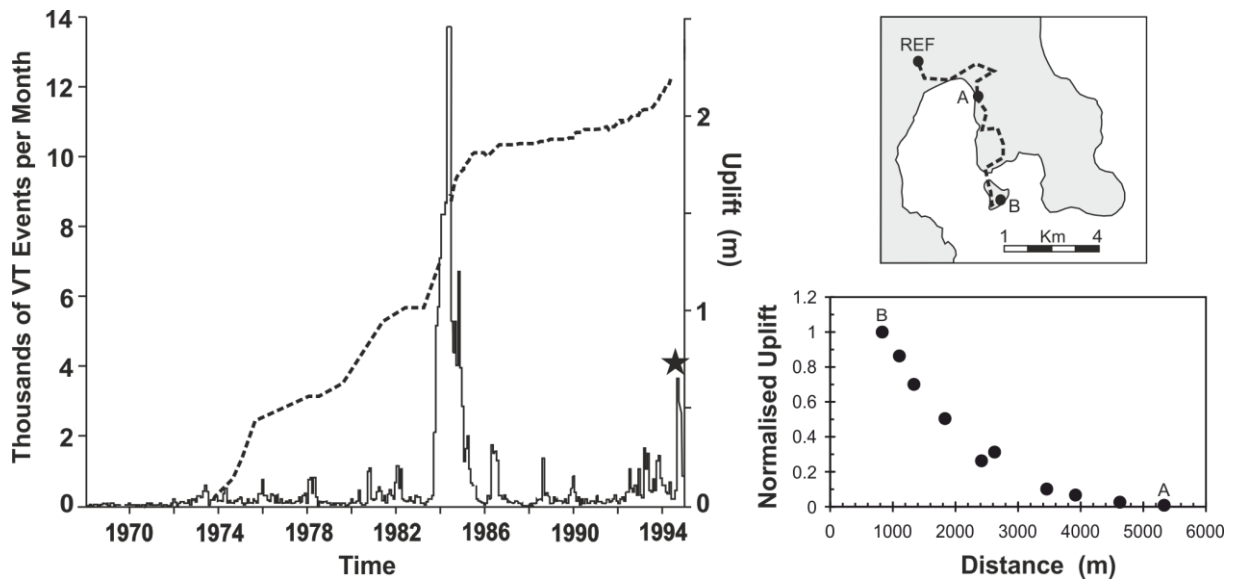


Figure 4

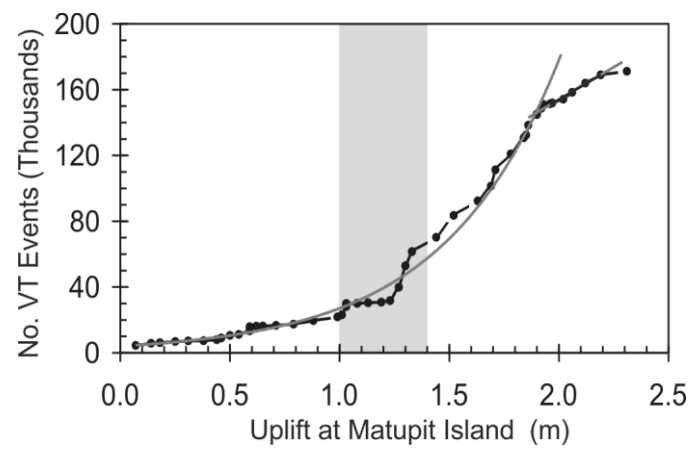


Figure 5

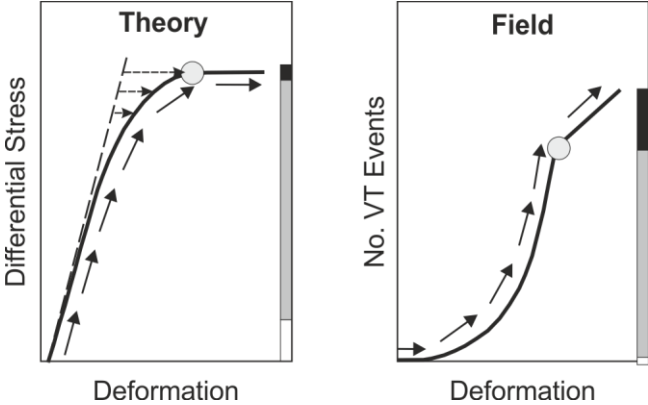


Figure 6

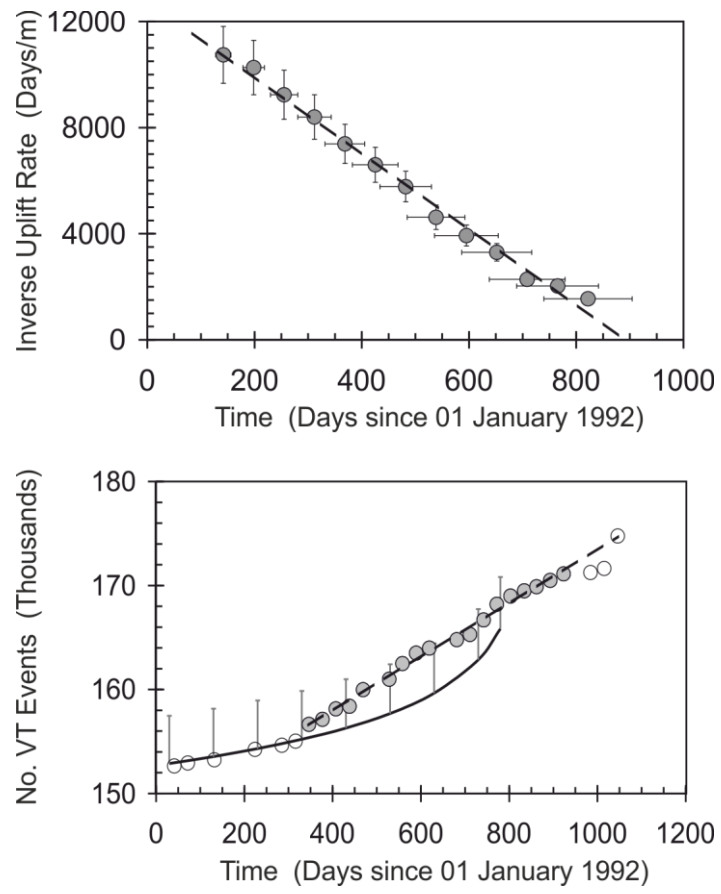


Figure 7

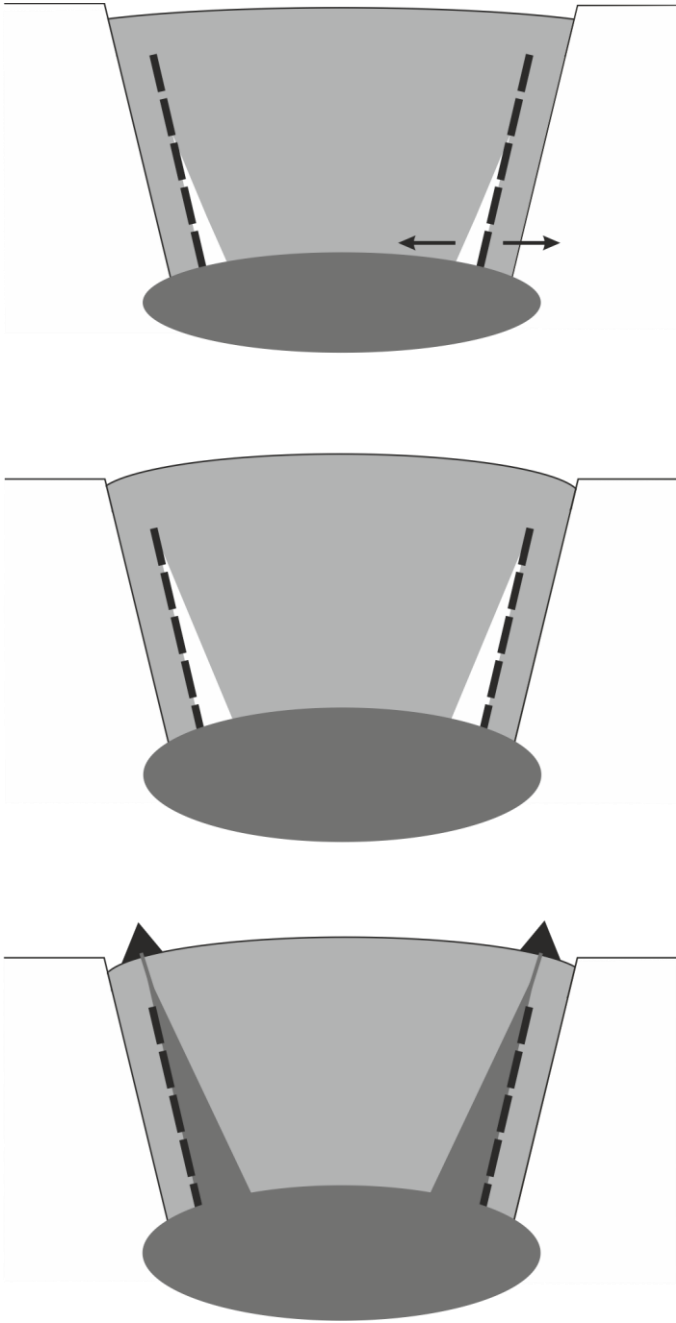


Figure 8

