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Journal of Volcanology and Geothermal Research 145 (2005) 68–80

Journal of volcanology and geothermal research

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Some consequences of volcanic edifice destruction for eruption conditions

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Received 29 July 2004; received in revised form 10 January 2005; accepted 17 January 2005

Abstract

Destruction of a volcanic edifice by landslides or phreatic explosions unloads the upper crust. Induced changes of stress field around, and of magmatic pressure within, a magma reservoir are investigated with an analytical model for the deformation of a liquid-filled cavity within an elastic half-space. Unloading affects the reservoir pressure, and hence the net result depends on how the liquid-filled reservoir responds to a change of remote stress. Magma compressibility is taken into account and may dampen changes of internal pressure in small volatile-rich reservoirs. The main consequence of edifice destruction is a decrease of magmatic pressure and stresses on the reservoir walls. In some cases, this may be responsible for dyke closure at the reservoir walls, which stops magma withdrawal and may prevent eruption. If an eruption does occur, edifice destruction affects the volume of magma erupted. Depending on edifice size, magma reservoir size and depth, the erupted volume may be smaller or larger than that which would be erupted with no damage to the edifice. These results suggest that major phreatic explosions may prevent magmatic eruptions.
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Keywords: volcanic edifice destruction; magmatic reservoir pressure; phreatic eruptions

1. Introduction

On 18 May 1980, a large landslide removed a large part of Mount St Helen's volcanic edifice and generated a thick avalanche deposit over a large area (Moore and Albee, 1981; Voight et al., 1981; Voight, 2000). Since this spectacular event, massive scarps truncating edifices have been identified on many volcanoes (Siebert, 1984; Siebert et al., 1987; Voight, 2000). The volumes of the associated avalanche

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doi:10.1016/j.jvolgeores.2005.01.012

deposits may be as large as 10 km^3, showing that major changes of edifice load have occurred. In most volcanic systems, the magma reservoir is shallow enough to be sensitive to loading by an edifice and hence to changes of edifice size. In oceanic volcanoes, evidence for this is provided by changes of lava composition which systematically follow edifice destruction (Presley et al., 1997).

Edifice destruction can be due to a host of different phenomena: shallow magma intrusion (Voight and Elsworth, 1997; Donnadieu and Merle, 1998), slope instability and phreatic explosions, as at Bandai-san, Japan, in 1888 (Yamamoto et al., 1999). These different phenomena may occur at different stages of a volcanic cycle and are likely to modify eruption behaviour. Decreasing the size (and mass) of a volcanic edifice acts to reduce the magnitude of compressive stresses in the upper crust. One might be tempted to infer that this facilitates magma ascent and hence eruption, but we shall show that this may not be true, depending on the location and size of the reservoir. Unloading affects the reservoir pressure, and hence the net result depends on how the liquid-filled reservoir responds to a change of remote stress. The aims of our paper are to (1) characterize the effect of edifice destruction on a volcanic plumbing system and (2) define under which conditions magma ascent can be stopped. We first describe the theoretical deformation model and determine changes of reservoir overpressure induced by a reduction of edifice load. We study how this affects tensile stresses at the reservoir walls with implications for dyke closure. Finally, we evaluate how edifice destruction affects the volume of magma which gets erupted from the reservoir. We close the paper by a discussion of phreatic eruptions. Considerations on the effects of magma compressibility and volatile content are developed in Appendix A. Assumptions behind the theoretical model are reviewed and discussed in Appendix B.

2. Model description

A volcanic edifice represents a load for the underlying crust which generates departures from the lithostatic stress field in the upper crust. This non-lithostatic component decreases with depth at a rate which depends on edifice radius. Detailed results can be found for an axisymmetrical model without a reservoir in Pinel and Jaupart (2000) and for a 2-D model with a reservoir Pinel and Jaupart (2003, 2004). Along the vertical axis, the horizontal stress is compressive at the base of the load and decreases in magnitude with increasing crustal depth until it becomes tensile. Such stress perturbations become negligible at depths greater than three times the edifice radius. Most magmatic reservoirs are located at depths shallower than this and hence are affected by edifice loading.

In a liquid-filled reservoir, the internal pressure gradient is constrained to be hydrostatic. Thus, the reservoir pressure may be written as follows:

$$P(z) = \rho_{\mathrm{m}} g z + \Delta P \quad (1)$$

where ρ_{m} is the density of the magmatic mixture in the reservoir (consisting of a liquid which may contain crystals and exsolved volatiles) and ΔP the magmatic overpressure. Changes of reservoir volume and overpressure are calculated using wall displacements and the equation of state for the magmatic mixture. Hoop stresses at the reservoir walls determine when and where the reservoir fails in tension. The theoretical deformation model uses the edifice load and internal reservoir overpressure as boundary conditions and allows predictions of all variables, including displacements and hoop stresses at the walls. Details, assumptions and full equations can be found in Pinel and Jaupart (2003).

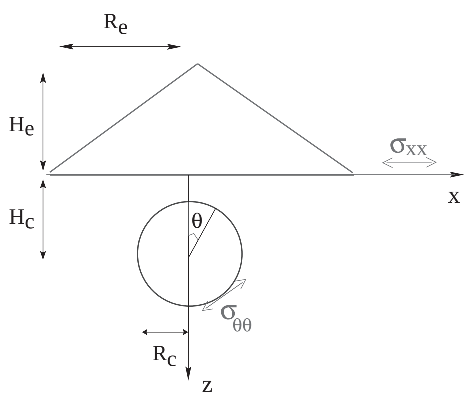

Fig. 1. Model geometry (described in Pinel and Jaupart (2003)). θ is the polar angle used for location along the reservoir walls. $\sigma_{\theta\theta}$ and σ_{xx} denote stresses normal to the reservoir walls and to Earth's surface respectively, used to investigate tensile failure in a volcanic system.

Table 1
Parameters and physical properties used in the calculation

Geometrical parameters		
Depth of the reservoir	H_c	
Radius of the reservoir	R_c	
Radius of the volcano	R_e	
Height of the volcano	H_e	
Physical properties		
Poisson's ratio	v	0.25
Rigidity (Pa)	G	1.125×10^9
Tensile strength (Pa)	T_s	2.10^7
Density of the magma (kg m^{-3})	ρ_m	2700
Bulk modulus of the magma (Pa)	K	10^{10}

Calculations are made in 2-D for a horizontal cylindrical cavity with circular cross-section in an elastic half-space. The state of reference is lithostatic with the reservoir lying at the level of neutral buoyancy such that magma and country rock have the same density. The upper crust is characterised by rigidity G and Poisson's ration v. The volcanic edifice has half-width R_e and height H_e. In the following, flank slope $H_e/R_e=0.6$, corresponding a stratovolcano. Chamber depth and radius are denoted by H_c and R_c respectively (Fig. 1). Model parameters and variables are listed in Table 1. Calculations are done for plane strain deformation such that the geometry of both the reservoir and the edifice is invariant along the y axis.

In the calculations below, the upper part of the edifice is removed and the implied changes of reservoir pressure are solved for (Fig. 2). For each stage (before and after destruction), we calculate the reservoir overpressure ΔP and the critical overpressure necessary for wall failure, ΔP_o. We have set the percentage of edifice volume (and mass) removed to 20%, which is in the upper range of values measured in the field (Siebert et al., 1995).

3. Pressure changes within the magma reservoir

3.1. Principle

Removal of part of the edifice produces an increase of reservoir volume ΔV_c and a change of internal pressure ΔP. Using the superposition principle, the calculation is made in two steps. ΔV_1 corresponds to zero internal pressure change (i.e., $\Delta P=0$) and to the prescribed change of surface load. ΔV_2 corresponds to pressure change ΔP and no change of surface load. For each case, the volume change is calculated from the radial displacement at the walls, u_r as follows:

$$\Delta V = 2R_c L \int_0^{\pi} u_r(\theta)\mathrm{d}\theta, \tag{2}$$

where L is the length of the reservoir in the third dimension. ΔV_1 is calculated with the formulae given in Pinel and Jaupart (2003). For ΔV_2, the solution can be expressed in compact analytical form (Jeffery, 1920):

$$\Delta V_2 = f\Delta P \tag{3}$$

where proportionality constant f is:

$$f=\pi\frac{R_c^2 L}{G\sqrt{H_c^2-R_c^2}}\left[2(1-v)H_c-(1-2v)\sqrt{H_c^2-R_c^2}\right] \tag{4}$$

The total volume change is $\Delta V_c=\Delta V_1+\Delta V_2$, which depends on the as yet unknown change of internal pressure ΔP. ΔV_c and ΔP are related through the equation of state for the magmatic mixture, which provides the third equation required for a solution.

For an incompressible magma, $\Delta V_c=0$, and hence:

$$\begin{aligned}\Delta P &= -\frac{1}{f}\Delta V_1 \\ &= \frac{2\sqrt{H_c^2-R_c^2}}{\pi R_c\left[2(1-v)H_c-(1-2v)\sqrt{H_c^2-R_c^2}\right]}G \\ &\quad\times\int_0^{\pi} u_r(\theta)\mathrm{d}\theta\end{aligned} \tag{5}$$

$u_r(\theta)$ is inversely proportional to rigidity G. Thus, in this equation, G cancels out. One important feature is

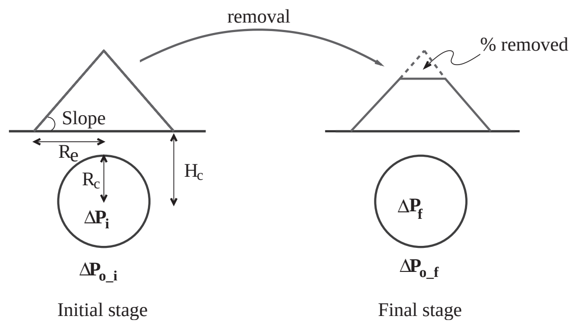

Fig. 2. Comparison between the initial stage and the final one, before and after a fraction of the volcanic edifice has been removed.

that it is not necessary to fully specify the initial state, as changes of internal pressure depend only on stress changes, and not on the total stress field.

Calculations for a compressible mixture are developed in Appendix A. We show that, for deep reservoirs, differences with the incompressible limit are small. Compressible effects become important for volatile-rich magma in small and shallow reservoirs.

3.2. Results

Removal of part of the edifice always causes a pressure drop within the reservoir (Fig. 3). For a given fraction of the edifice, which has been set at 20% here, the mass of material removed increases with edifice size, and hence so does the internal pressure drop. According to our calculations, one may expect changes of reservoir pressure which may be as large as $2.5 10^7$ Pa, which is comparable to threshold values for dyke initiation at the walls. One can surmise therefore that edifice destruction affects dyke behaviour in the vicinity of the reservoir.

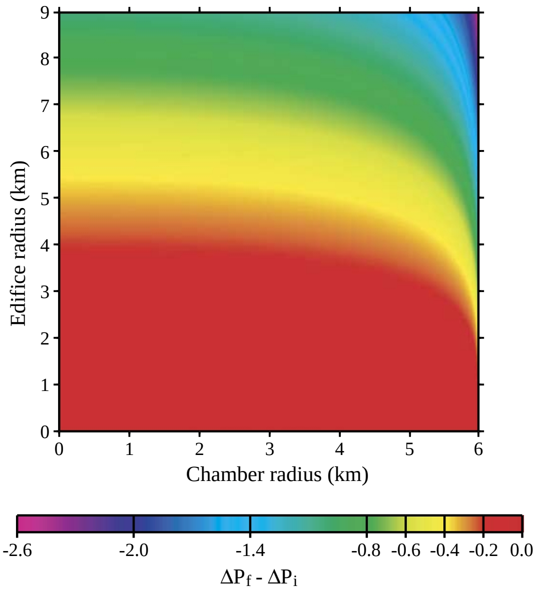

Fig. 3. Change of magma overpressure in a reservoir as a function of reservoir radius and edifice radius for a reservoir lying at a depth of 6 km. Calculations are made for the removal of 20% of a volcanic edifice with 30° slope. Magma is assumed to be incompressible. The pressure scale is set by the tensional strength of country rock, $T_s = 2.10^7$ Pa.

4. Dyke initiation and closure at the reservoir walls

4.1. Rupture criterion

Dykes are generated by rupture of the reservoir walls. In this paper, we use the classical tensile failure criterion such that the deviatory part of the hoop stress $\sigma_{\theta\theta}$ at the walls exceeds the tensile strength of wall rocks (Pinel and Jaupart, 2003). This failure criterion depends on the total stress field and cannot be specified using stress perturbations only. The failure criterion is written as follows:

$$\frac{\sigma_{\theta\theta} - \Delta P_o}{2} = -T_s \qquad (6)$$

where $\sigma_{\theta\theta}$ depends on magma overpressure inside the chamber ΔP_o and the edifice load at the free surface.

Without an edifice and when $R_c/H_c \to 0$, $\sigma_{\theta\theta} \to -\Delta P_o$, showing that the failure condition tends to that for an infinite medium, i.e. $\Delta P_o = T_s$ (Blake, 1981; Tait et al., 1989).

In some cases, edifice destruction leads to large deformations, in which case the elastic model may not be accurate. We have set the validity threshold for our calculations to be such that the change of reservoir volume remains smaller than 10%.

4.2. Changes of reservoir pressure through an eruption

ΔP_o is such that the reservoir walls fail, which allows dyke propagation away from the reservoir. With an eruption and magma withdrawal from the reservoir, the reservoir pressure decreases. The eruption ceases when the feeder dyke closes at the chamber walls (or soon afterwards).

Dyke closure at the chamber walls occurs if the chamber overpressure ΔP becomes lower than the local hoop stress $\sigma_{\theta\theta}$ (θ_r), where θ_r is the polar angle where dykes have breached the chamber walls (Fig. 1). Let us call ΔP_c the largest chamber overpressure which allows a dyke to remain open at the walls. By continuity, the magma pressure within the dyke at its entry point, at the chamber walls, is equal to the chamber pressure. Thus, the dyke can remain open if magma pressure exceeds the normal stress applied at its walls, which is equal to hoop stress $\sigma_{\theta\theta}$ (θ_r).

Therefore, the threshold overpressure for closure is such that:

$$\Delta P_{\mathrm{c}} = \sigma_{\theta\theta}(\theta_r). \tag{7}$$

For given edifice dimensions, ΔP_{c} may be written as a function of ΔP_{o}, the threshold overpressure for dyke initiation from Eq. (6):

$$\Delta P_{\mathrm{c}} = \Delta P_{\mathrm{o}} - \frac{2}{1+\phi(\theta_{\mathrm{r}})} T_{\mathrm{s}} \tag{8}$$

where function ϕ depends only on geometrical parameters:

$$\phi(\theta) = \frac{H_{\mathrm{c}}^2 + R_{\mathrm{c}}^2 - 2R_{\mathrm{c}}^2 \left(\dfrac{R_{\mathrm{c}} - H_{\mathrm{c}}\cos(\theta)}{H_{\mathrm{c}} - R_{\mathrm{c}}\cos(\theta)} \right)^2}{H_{\mathrm{c}}^2 - R_{\mathrm{c}}^2} \tag{9}$$

For small reservoirs such that $R_{\mathrm{c}} \ll H_{\mathrm{c}}$, $\phi \rightarrow 1$. For large edifices, as shown in Pinel and Jaupart, 2003, dykes are generated at the top of the chamber, such that $\theta_{\mathrm{r}}=0$, implying that $\phi=1$. In both cases, therefore, Eq. (8) for dyke closure can be simplified to:

$$\Delta P_{\mathrm{c}} = \Delta P_{\mathrm{o}} - T_{\mathrm{s}}. \tag{10}$$

For small edifices, this simplified expression is no longer valid. This case, such that the reservoir walls do not fail at the axis (Pinel and Jaupart, 2003), is of no practical importance because stress changes are very small.

Opening of feeder dykes at the reservoir walls is a necessary but not sufficient condition for eruption, which requires the dyke to remain open all the way to Earth's surface. For this, the internal magma pressure within the dyke must be larger than the normal stress applied at its walls at all depths. The limiting case separating successful dyke intrusion and failed eruption is that of a dyke which stalls just below Earth's surface. In such a static dyke, there is no viscous head loss in the vertical magma column, which allows calculations in the hydrostatic limit. For volatile-free magma, and assuming that country rock density does not change with depth, the internal dyke pressure is equal to ΔP at all depths. Pinel and Jaupart (2000, 2005) have studied under which conditions this overpressure is not sufficient to keep the dyke open at all depths. For magmas with dissolved volatiles, gas exsolution and expansion act to reduce the density of the magmatic mixture and the hydrostatic head in the dyke is smaller. As a consequence, the internal dyke overpressure is larger than ΔP at shallow levels, which favors dyke propagation to Earth's surface. Details can be found in Pinel and Jaupart (2000, 2004). The condition that the dyke remains open at all depths must therefore be made as a function of volatile content. If the initial dyke stalls at depth, replenishment into the reservoir continues, and the reservoir pressure increases until it is large enough to drive magma all the way to the surface. In practice, this extra overpressure is small and decreases with increasing volatile content, and will be neglected for simplicity purposes. We therefore use $\Delta P=\Delta P_{\mathrm{o}}$ as the condition for the beginning of eruption.

Most volcanic systems of interest involve an edifice of significant size, for which Eq. (10) holds. This equation has been obtained for a constant edifice size, i.e. the edifice is the same at the beginning of unrest and at the end of eruption. In this case, therefore, the total variation of reservoir pressure is constant from one eruption to the next.

5. The impact of edifice destruction on incipient eruptions

We next use these results and ideas to discuss how edifice destruction alters the behaviour of a volcanic system. Destruction of part of the edifice modifies the stress field around the reservoir, and hence may alter the course of an incipient eruption. We define two critical values of magma overpressure for tensile failure of reservoir walls before and after edifice destruction, called ΔP_{oi} and ΔP_{of} respectively. Our calculations show that the difference between these two overpressures can be positive or negative, i.e. edifice destruction can either increase of decrease the likelihood of an eruption (Figs. 4, 5 and 6). This can be understood using results given in Pinel and Jaupart (2003). An edifice on top of a shallow reservoir generates flexure of the roof region and hence tension at the top of the reservoir: in this case, edifice destruction acts to decrease tensile stresses in the roof region and hence to reduce the likelihood of an eruption. On the contrary, a large edifice puts the roof region in compression and its destruction enhances the likelihood of an eruption.

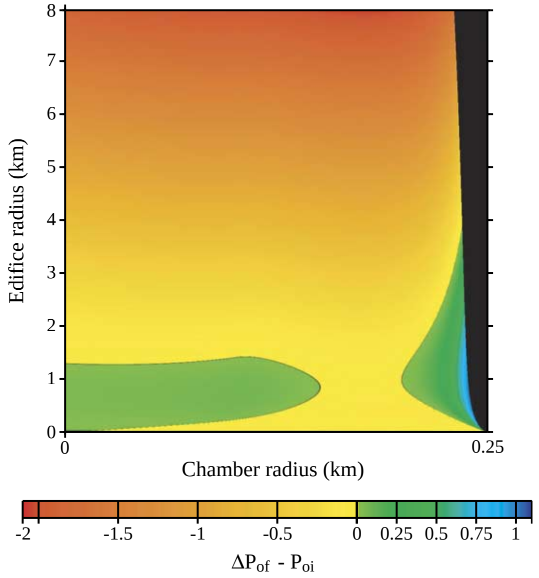

Fig. 4. Change of magma overpressure required for dyke initiation at the chamber walls as a function of edifice radius and chamber radius for a very shallow chamber (H_c=250 m). We consider the removal of 20% of an edifice with 30° slope. The pressure scale is T_s=2.10^7 Pa. The black domain is such that deformations are too large for the elastic model to be valid.

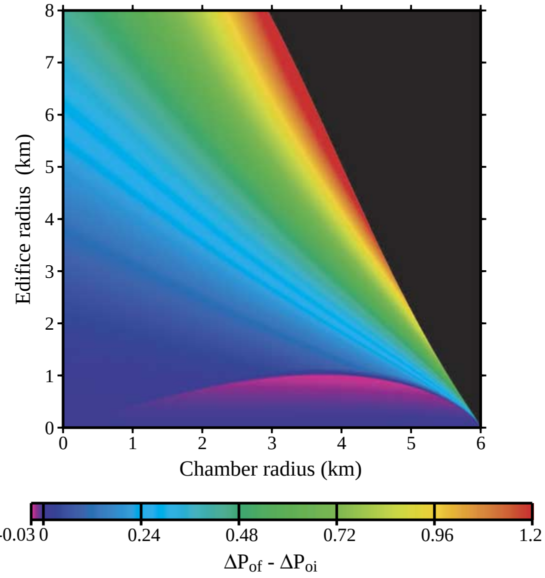

Fig. 6. Same as Fig. 4 for a deep chamber (H_c=6 km).

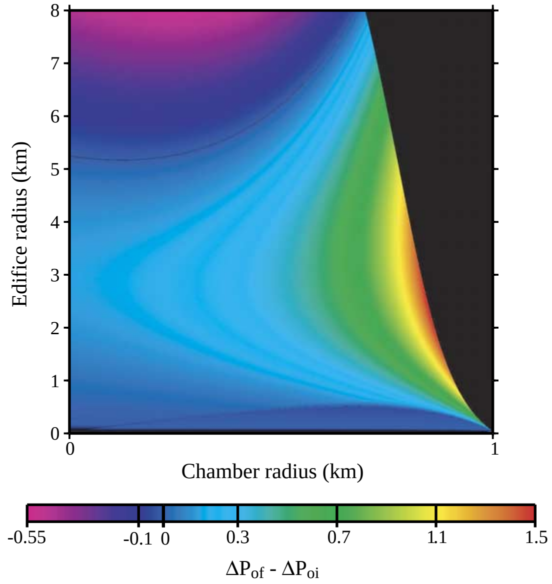

Fig. 5. Same as Fig. 4 for a shallow chamber (H_c=1 km).

Edifice destruction usually occurs when the volcanic system is in a phase of unrest and is indeed frequently followed by an eruption. For example, edifice breakdown is due to the intrusion of magma, as at Mount St Helens in 1980 (Moore and Albee, 1981; Hoblitt and Harmon, 1993; Voight and Elsworth, 1997). Thus, we set the initial condition to be such that the reservoir overpressure is at the failure threshold: dykes are about to propagate, or have just begun to propagate, away from the reservoir. This is written as $\Delta P = \Delta P_{oi}$. With no damage to the edifice, an eruption would follow and would last until pressure returns to the closure pressure ΔP_{ci}.

Edifice destruction modifies the stress field and hence the ensuing eruption sequence. When part of the edifice is removed, the reservoir pressure decreases and the state of stress along the reservoir walls changes. We wish to compare two different cases: one in which there is no edifice destruction, and the other in which the edifice gets destroyed sometime after dyke initiation. In the first case, the reservoir pressure lies within the range defined by ΔP_{oi} and ΔP_{ci}. With edifice destruction, the reservoir pressure starts from the same initial value ΔP_{oi} but decreases to a new final value ΔP_{cf}. In order to compare these two cases, it is useful to use the final edifice size. We define a "virtual" initial reservoir pressure ΔP_{of}

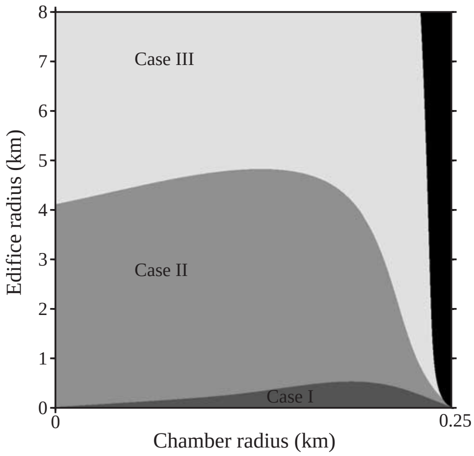

Fig. 7. Three different possibilities for a volcanic system after the partial destruction of the edifice by a major landslide or phreatic explosion. It is assumed that edifice destruction occurs at the end of volcanic quiescence, when chamber walls are ready to rupture or have just ruptured. Case I is such that the volume of magma erupted is larger compared to the volume which would have been erupted with no damage to the edifice. In Case II, the erupted volume is decreased. Case III is for no eruption. We consider the removal of 20% of an edifice with 30° slope above a reservoir filled with incompressible liquid. The chamber depth is 250 m. The elastic model may not be accurate for parameters in the black domain.

corresponding to the opening of dykes at the reservoir walls beneath the final edifice, i.e. beneath the partially destroyed edifice. As explained above, for fixed edifice dimensions, the difference between ΔP_{o} and ΔP_{c} does not depend on edifice size. Thus, $\Delta P_{\mathrm{oi}} - \Delta P_{\mathrm{ci}} = \Delta P_{\mathrm{of}} - \Delta P_{\mathrm{cf}}$. This result is not valid for small edifices, but such cases are rarely relevant.

We define different cases:

(1) Case I:

$$\Delta P_{\mathrm{oi}} > \Delta P_{\mathrm{of}}$$

This implies that $\Delta P_{\mathrm{oi}} - \Delta P_{\mathrm{cf}} > \Delta P_{\mathrm{of}} - \Delta P_{\mathrm{cf}}$, and hence that the total reservoir pressure variation from start to finish is larger than it would have been with no damage to the edifice. One consequence is an increase of the erupted volume.

(2) Case II:

$$\Delta P_{\mathrm{cf}} < \Delta P_{\mathrm{oi}} < \Delta P_{\mathrm{of}}$$

The reservoir walls have failed and feeder dykes have begun their ascent towards Earth's surface. The reservoir is overpressured with respect to the surroundings, and hence there is therefore a pressure differential driving magma ascent. Thus, even if the threshold overpressure for wall rupture has changed (i.e. $\Delta P_{\mathrm{oi}} < \Delta P_{\mathrm{of}}$), the starting dyke remains open and connected to the reservoir. However, the total pressure drop from start to finish is now smaller than it would have been without edifice destruction, implying a decrease of erupted volume.

(3) Case III:

$$\Delta P_{\mathrm{oi}} < \Delta P_{\mathrm{cf}}$$

This third possibility is somewhat unexpected, but highly interesting. The change of stress at the reservoir walls implies that feeder dykes get closed: the eruption ceases or does not occur at all. We return to this in a later section.

These three cases are displayed in Figs. 7, 8 and 9 for various values of edifice and reservoir sizes. The trend with increasing edifice radius is always the same. However, the relative importance of each regime is sensitive to reservoir depth. Case III, such that edifice destruction prevents a magmatic eruption, is most likely for intermediate reservoir depths, small reservoirs and large edifices. Case I, such that edifice destruction acts to increase the erupted volume, is rare and limited to early stages, when the edifice is small.

In a sense, these results illustrate the self-regulating effect of an edifice. As shown in Pinel and Jaupart (2003, 2004), edifice growth prevents the eruption of primitive magmas and hence favors reservoir forma-

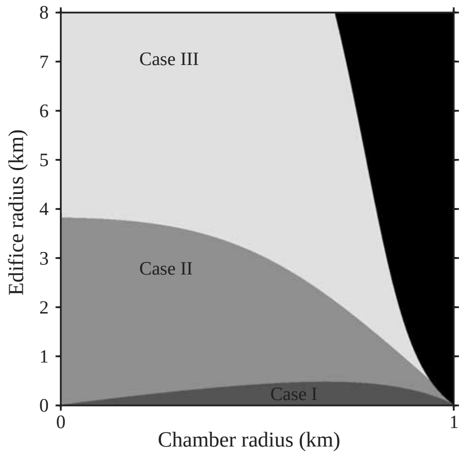

Fig. 8. Same as Fig. 7 for a 1 km-deep reservoir.

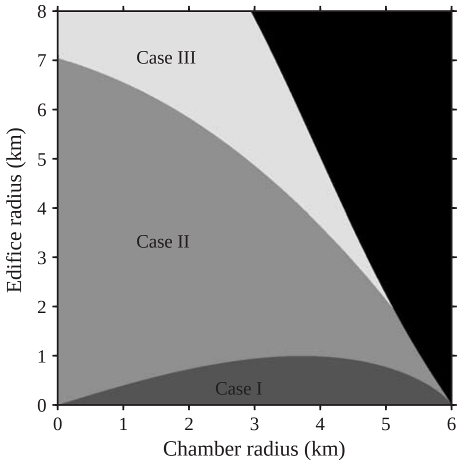

Fig. 9. Same as Fig. 7 for a 6 km-deep reservoir.

tion. Here, we further show that edifice destruction favors magma storage by decreasing the volume that gets erupted.

6. Other volcanological implications

Our calculations emphasize the changes that are brought to a volcanic plumbing system by edifice destruction. The effect of unloading Earth's surface on eruptive behaviour have already been illustrated in the completely different setting of deglaciation: Gudmundsson (1986); Jull and McKenzie (1996) have attributed a change in the rate of volcanism in Iceland to melting of ice sheets. We now discuss further implications of edifice destruction.

6.1. Reservoir size

Further developments of the theoretical ideas given here include a study of different reservoir shapes and dimensions. Comparison between changes of edifice size and eruption conditions yield information on the reservoir. Consider for example Mount St Helens, whose reservoir has been located at a depth of about 6 km (Barker and Malone, 1991; Moran, 1994). The edifice radius is about 5 km (Pike and Clow, 1981) and we know that the destruction due to the large landslide of May 18, 1980, did not stop the eruption. From Fig. 9, this implies that the radius of the Mount St Helens reservoir must be less than 3 km.

There have been some attempts to calculate chamber size from the mass of erupted products (Bower and Woods, 1997). Such calculations must be made with due consideration for edifice growth and destruction.

6.2. Caldera collapse

So far, we have focussed on rupture of the reservoir walls. However, rupture can also affect Earth's surface ($z=0$), which may be responsible for caldera collapse, as argued by Pinel and Jaupart (2005). If Earth's surface fails before the reservoir walls, our calculations may not be valid.

We have therefore calculated under which conditions Earth's surface fails in tension, using the same tensile failure criterion than for the reservoir walls. The relevant stress is the horizontal normal stress at the Earth's surface ($\sigma_{xx}|_{(z=0)}$, Fig. 1). Details can be found in Pinel and Jaupart (2005). With no reservoir, no tensile failure would occur at the Earth's surface because the edifice load generates compression. With no edifice, tensile failure occurs first at Earth's surface if $R_c/H_c > 1/\sqrt{2}$ (Pinel and Jaupart, 2005). Fig. 10 shows the parameter domain for which Earth's surface fails in tension before edifice destruction. This domain

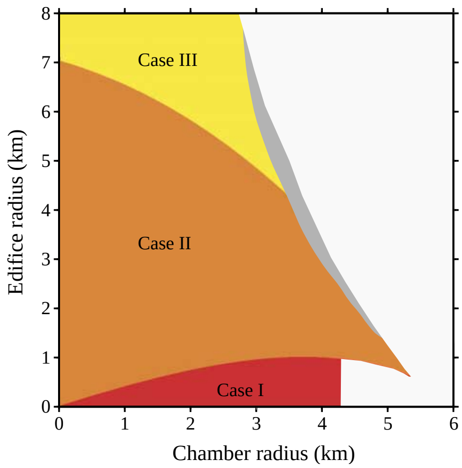

Fig. 10. Different evolutions for a volcanic system after edifice destruction. We consider the removal of 20% of an edifice with 30° slope. The chamber depth is 6 km. Magma is supposed to be incompressible. The pressure scale is $T_s = 2.10^7$ Pa. The white domain is such that failure occurs first at Earth's surface before edifice destruction. The grey domain correspond to the values for which failure occurs at the surface only after edifice destruction.

is such that the initial condition for eruption implies caldera collapse (Pinel and Jaupart, 2005). In this case, the present arguments are irrelevant. This domain extends the size of the domain where the elastic model is not really valid, as discussed above. Fig. 10 also shows a small parameter domain (the grey area) for which Earth's surface fails in tension after damage is done to the edifice. This raises the intriguing possibility that edifice destruction may lead to a caldera.

6.3. Changes of erupted lava composition

Changes of reservoir pressure affect the distribution of pressure within the volcanic plumbing system, with consequences for the rates of replenishment and withdrawal. We have shown that edifice destruction always induces a decrease of reservoir pressure. This may increase the rate of replenishment if reservoir and source had remained connected, or may trigger replenishment if the connection was closed. This may explain why, following edifice destruction, the eruption rate increases and the edifice rapidly grows back to its original size (Siebert et al., 1995). Furthermore, a reduced edifice load allows the eruption of denser magmas which otherwise would have stalled at shallow depth (Pinel and Jaupart, 2000). These two effects should lead to renewed eruption of primitive magmas.

The connection between edifice destruction and changes of erupted magma composition has already been discussed in Pinel and Jaupart (2000). For example, at Mount St Helens, the Pine Creek period (3000–2500 year BP) involved only dacitic lavas. This period ended with large landslides (Hausback and Swanson, 1990) and was followed by the Castle Creek period which saw basaltic and andesitic volcanism. The Castle Creek andesites are due to mixing between felsic dacite and calcalkaline basalt (Gardner et al., 1995). The early Kalama period following the *Wn* explosive eruption saw a similar pattern. In this case, the edifice load was reduced because a large crater was formed during the *Wn* eruption (Hopson and Melson, 1990). The combination of petrological studies of erupted products and reconstructions of the edifice should therefore improve our knowledge of magma reservoirs.

6.4. Phreatic eruptions and magmatic unrest

Phreatic eruptions remain poorly understood and yet can cause important damage. In his review of historical phreatic events, Barberi et al. (1992) found that a vast majority (115 out of 132) were not followed by a magmatic eruption. This has led many to conclude that magmas are not involved and that phreatic events have superficial causes. Nevertheless, the explanations that have been put forward, such as pressure release in a shallow aquifer due to unloading or infiltration of meteoric water into a magma reservoir, have not been tested quantitatively and are likely to remain so due to limited information on the behaviour of shallow volcanic environments. For want of a robust physical model, the basic line of reasoning seems to be the following: some phreatic explosions are followed by magmatic eruptions, and hence those that are not cannot be due to magmatic unrest. The present results identify a flaw in this reasoning.

Let us consider that pressure in the magma reservoir has reached breaking point and that a dyke has started to rise towards Earth's surface. This is likely to trigger phreatic events. One possibility is that this also triggers a landslide, which acts to close off the dyke at the chamber walls. Another possibility is that the phreatic event is so powerful that it leads to edifice destruction, as in 1888 at Bandai-san, Japan (Yamamoto et al., 1999). A final possibility is that, even without edifice destruction, phreatic events, because they vaporize and expel water in shallow aquifers, act to unload the upper crust with the same consequences as edifice destruction. In all these cases, phreatic events are due to magmatic unrest but modifications of the state of stress in the upper crust prevent eruption. Support for this idea is provided by the fact that precursor events to phreatic eruptions are often identical to those of magmatic eruptions (Barberi et al., 1992).

7. Conclusion

Partial destruction of a volcanic edifice by a landslide or a phreatic explosion always leads to a decrease of magmatic pressure inside the reservoir. As regards the composition of erupted products, important consequences may be: (1) renewed replenishment of the reservoir by primitive magma from a deeper source, (2) the reversal of the normal eruption sequence with the return to denser, and hence primitive, magma. Other important consequences are changes in the values of reservoir overpressure required for dyke initiation and dyke closure. Depending on edifice dimensions and reservoir size, edifice destruction may in fact prevent an impending eruption by shutting off dykes at the reservoir walls. In all cases, edifice destruction modifies eruption conditions and the volume of magma that can get erupted. Changes of erupted magma composition and eruption rates may well be due to modifications of the edifice itself, and may not reflect changes in the deep source of magma which feeds the reservoir.

Acknowledgments

The manuscript benefited greatly from the constructive and critical comments of two anonymous reviewers.

Appendix A. A compressible mixture in the reservoir

In the elastic model, the reservoir lies at the neutral buoyancy level, such that magma and country rock have the same density. Therefore no buoyancy force is applied to the reservoir walls. In this appendix, we calculate the effects of magma compressibility on changes of internal pressure and reservoir volume. In this case, by definition, the densities of magma and country rock do not remain equal at all times. Changes of density in the reservoir impact on the evolution of reservoir pressure and generate buoyancy forces. However, one may show that the latter are small and hence that, for most practical circumstances, one needs worry only about the effect on reservoir pressure.

For density change $\Delta\rho$, buoyancy forces induce stresses of magnitude $\approx \Delta\rho g R_c$ on the reservoir walls which can be neglected if:

$$\frac{\Delta\rho g R_c}{\Delta P} << 1 \tag{11}$$

For a mixture with bulk modulus K:

$$\frac{\Delta\rho}{\rho} = \frac{\Delta P}{K} \tag{12}$$

and condition (11) can be rewritten as:

$$\frac{\rho g R_c}{K} << 1 \tag{13}$$

For magma that is dry or undersaturated in volatiles, $K \approx 10$ GPa, which implies that condition (13) is met for all reasonable values of reservoir radius. For volatile-rich magma, density is sensitive to pressure. In this case, the equation of state is:

$$\frac{1}{\rho} = \frac{1}{\rho_m}\frac{1-x_o}{1-x} + \frac{1}{\rho_g}\frac{x_o - x}{1-x}, \tag{14}$$

where ρ_m and ρ_g are the densities of the magma and gas phases. x_o are the total mass fraction of volatiles in the mixture (dissolved+gas) and x the mass fraction dissolved in the melt respectively, such that $x_o - x$ is the mass fraction of gas in the mixture. Values for x_o and x are typically small (a few percent), and this equation simplifies to:

$$\frac{1}{\rho} \approx \frac{1}{\rho_m} + \frac{x_o - x}{\rho_g}, \tag{15}$$

For water vapor, we use the perfect gas approximation for the equation of state and take an empirical solubility law:

$$x = sP^{1/2} \tag{16}$$

with $s = 4.1 \times 10^{-6}$. As shown below, the effective bulk modulus of the melt–gas mixture is much smaller than that of the pure melt phase. Thus, we may neglect variations of melt density for simplicity purposes and find:

$$\frac{1}{\rho}\frac{d\rho}{dP} \approx \frac{\rho}{\rho_g}\left(\frac{x_o - x}{P} + \frac{dx}{dP}\right), \tag{17}$$

For $P = 200$ MPa and magma that is just saturated at that pressure, $K \approx 1$ GPa. For $R_c = 500$ m, $\rho g R_c / K \approx 0.01$. At $P = 100$ MPa, the same calculation leads to $K \approx 0.4$ GPa and $\rho g R_c / K \approx 0.03$. We therefore conclude that buoyancy changes have a small impact on the force balance at the reservoir walls. However, changes of magma density do affect the magnitude of pressure changes in the reservoir, as shown below.

Equation of state (12) specifies that:

$$\frac{\Delta V_c}{V_c} = -\frac{\Delta P}{K} \tag{18}$$

Using equations from Section 3:

$$\Delta P = \frac{-\Delta V_1}{f + \frac{V_c}{K}} \tag{19}$$

Comparison with the incompressible case (Eq. (5)) shows that ΔP decreases with decreasing bulk modulus, i.e. increasing compressibility. In the compressible case, ΔP depends on G, the rigidity of country rock. For a reservoir of given radius R_c and length L, its volume is affected by loading due to the edifice, such that:

$$V_c = \pi R_c^2 L + \Delta V_{ci} \tag{20}$$

where ΔV_{ci} is the change of volume with respect to the initial undeformed reservoir. In principle, therefore, the change of reservoir pressure now depends on the initial state.

Within the elastic approximation, $\Delta V_{ci} << \pi R_c^2 L$ and may be neglected with little error. Let us denote pressure changes in the incompressible and compressible cases by ΔP_∞ and $\Delta P(K)$ respectively. From the equations above:

$$\frac{\Delta P(K)}{\Delta P_\infty} = \frac{K\left[2(1-v)H_c - (1-2v)\sqrt{H_c^2 - R_c^2}\right]}{K\left[2(1-v)H_c - (1-2v)\sqrt{H_c^2 - R_c^2}\right] + G\sqrt{H_c^2 - R_c^2}} \tag{21}$$

This equation shows that compressibility effects increase with decreasing chamber radius. In one limit, as $R_c \to H_c$, corresponding to shallow reservoirs, $\Delta P(K)/\Delta P_\infty \to 1$. In the other limit, corresponding to small reservoirs, $R_c << H_c$, and $\Delta P(K)/\Delta P_\infty \approx K/(K+G)$.

For typical upper crustal rocks, which are fractured and permeable, $G \approx 1$ GPa. The above calculations show that values of K in volatile-saturated magma increase with increasing pressure. Magma compressibility has little influence on the evolution of reservoir pressure in deep reservoirs. For volatile-rich magma stored at shallow depth, Eq. (21) predicts that the reservoir pressure may be strongly affected if the reservoir is small. In this case, changes of the external stress field are dampened by magma expansion. Scandone and Giacomelli (2001) has investigated eruptions driven by a sudden decompression event with application to the May 18, 1980, eruption of Mount St Helens. The above calculations illustrate how country rock deformation and magma expansion within the reservoir complicate matters, such that the net reservoir pressure change can be much smaller than the load reduction.

For the sake of completeness, we account for changes of reservoir volume due to loading by the edifice and do not use the simplified solution (21). In this case, ratio $\Delta P(K)/\Delta P_\infty$ depends on edifice size. The solution depends on the initial state, which is such that chamber walls have just reached the failure threshold when the edifice gets destroyed. This initial condition sets the initial overpressure. For the sake of example, we consider dry or undersaturated magma with $K_o = 10$ GPa (Johnson et al., 2000; Gudmundsson, 1986; Spera, 2000). For $G = 1.125$ GPa, as shown in Fig. 11, one has:

$$0.9 < \frac{\Delta P(K)}{\Delta P_\infty} < 1. \tag{22}$$

The edifice size has a weak influence in the results, and Eq. (21) provides a good approximation.

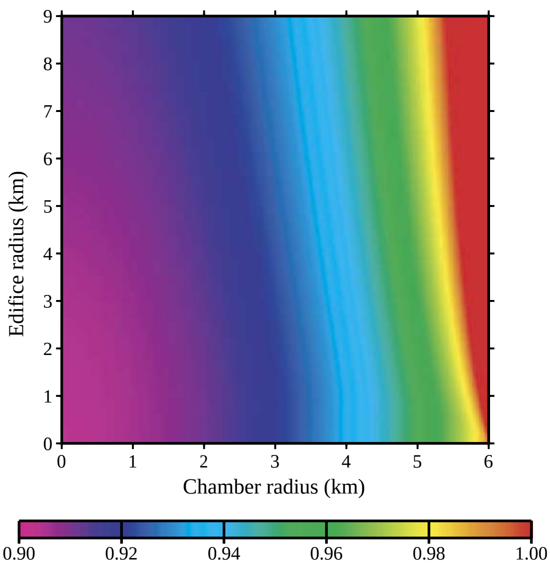

Fig. 11. Change of reservoir pressure for compressible magma, $\Delta P(K)/\Delta P_\infty$, as a function of edifice radius and chamber radius for a reservoir depth of 6 km, $K_o = 10$ GPa and $G = 1.125$ GPa. In the initial state, the reservoir overpressure is set to the threshold value for wall failure. The top 20% of a stratovolcano with 30° slopes have been removed.

Appendix B. Model assumptions

There is good field evidence for elastic (reversible) deformation on the short time-scale of an individual eruptive cycle. Thus, effects of reservoir pressure changes induced by a sudden unloading event can be adequately described by the present model. Transient elastic effects propagate at the speed of seismic waves, and hence can be treated as instantaneous for the magmatic system. Over larger time-scales, viscous relaxation may affect stresses due to edifice growth. Viscous relaxation time-scales have been estimated for the lithosphere as a whole, and are consistently larger than 5 Ma (Beaumont, 1981; Nunn and Sleep, 1984). Such estimates are vertical averages and hence provide lower bounds for the cold upper crust. On a smaller scale, rocks encasing the magma reservoir get heated up, but viscous behaviour is limited to a thin high-temperature halo. Elastic behaviour thus corresponds to an effective reservoir size which may be slightly larger than the true size.

Two issues involve the geometrical model set-up. The 2-D system studied here, the reservoir roof extends over a large horizontal distance in the direction normal to the (x, z) plane. For a reservoir elongated in the vertical direction, the roof would be able to sustain larger magmatic overpressures and flexure effects would be reduced. The threshold pressure for dyke initiation is thus probably underestimated by the present calculations. Another problem involves the change of edifice size, which rarely adopts a symmetric configuration. With a lop-sided edifice structure, the stress field at shallow depth departs from the predictions of the present model. However, due to the diffusive-like character of elastic equations, deviations from a symmetrical load get smoothed out at depth and the main effect is loading or unloading of a given mass.

A final issue is the fate of the destroyed parts of the edifice. In our calculations, the surface load is just decreased whereas in reality the same load gets redistributed over a larger area. We made calculations for which the mass removed is redistributed uniformly on a distance which is twice the edifice size. We found small differences with the results given here. Changes in chamber overpressure are slightly smaller, but the general trends as a function of edifice and chamber size are not affected. Boundaries between the domains for the three different scenario, shown in Figs. 7, 8 and 9, are almost the same. Obviously, this effect depends on the area over which the lost load gets redistributed. This additional parameter in turn depends on many variables, including the driving phenomenon for edifice destruction (i.e. landslide versus phreatic explosion). Such developments are outside the scope of this study.

We have assumed that an eruption may be prevented if feeder dykes get closed at the reservoir walls. In reality, if dykes have already propagated for some distance away from the reservoir, a finite amount of magma has been injected in country rock and may continue to rise. In this case, magma ascent cannot be driven by the reservoir overpressure and can only occur due to buoyancy. If magma is indeed buoyant, ascent can proceed if the crack is long enough (Rubin, 1995). The ability to reach Earth's surface further depends on the ascent velocity as magma may freeze if it does not rise fast enough (Spence and Turcotte, 1990).

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