Multiparametric quantification of volcanic hazards for eruption forecasting and communication

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2016

Winson, A. E. G. (2016). Multiparametric quantification of volcanic hazards for eruption forecasting and communication. Doctoral thesis, Nanyang Technological University, Singapore.

<http://hdl.handle.net/10356/66030>

<https://doi.org/10.32657/10356/66030>

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Asian School of the Environment

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Annie Elizabeth Grace Winson

A thesis submitted to the Nanyang Technological University in partial
fulfilment of the requirement for the degree of Doctor of Philosophy in
Earth Sciences

Asian School of the Environment

2016

Acknowledgments

I would like to thank my supervisors Chris Newhall and Fidel Costa, without whom this work would not have been possible. They have both been so wonderfully supportive over the course of my PhD research, dispensing advice, encouragement and humor as the situation called for. Our hours of discussion and debate have allowed me to explore this project and these themes with greater understanding of the nuances of the topics. They have also pushed me to think deeply and critically. It has been a privilege and a joy to learn from you both. Thanks also to my examiners John Pallister, Eliza Calder and Benoit Taisne for their thoughtful and thorough review of this work.

Special thanks to the Earth Observatory of Singapore and the Agency for Science, Technology and Research (A*STAR) for generously funding this work. I would like to thank Gordon Woo who met with me at NTU and in London and always gave me much to think about and usually many more questions to answer. Thank you to Christina Widiwijayanti, for her unfailing support throughout my PhD and specifically for all of the advice and assistance with the TITAN2D modeling. I am indebted to Thea Hincks for her kindness and patience in explaining Bayesian Belief Networks and helping me to understand how the plinian network should be designed. I am very grateful to Caroline Bouvet de Maisonneuve and to Patrick Whelley, for their willingness to discuss research roadblocks and to provide wise advice and counsel when it was needed. It would not have been possible to synthesize all the data in this study without the help of my multilingual colleagues; Corentin Caudron, Tarsilo Girona and Jay Wellik. Special thanks must go to Qiu Qiang for all of his patient assistance with writing code. I would further like to thank all of the members of the volcano group, for their advice and discussions. I would like to acknowledge and thank Susanna Jenkins for running the ash fall exceedance probability simulations. I am very grateful to both the Philippine Institute of Volcanology and Seismology (PHIVOLCS) and the Center for Volcanology and Geological Hazard Mitigation (CVGHM) for sharing their archives with me. I would specifically like to thank the observers at Mayon volcano, who always made me feel so welcome.

These years would have been very hard without my ‘Singapore Family’ who provided support and levity in all situations (Hunter, Chris, Jon, Nikki, Patrick, Beck, Chris, Manu, Kim, Amy and Ken). Finally I must express my deepest gratitude to my family for their unconditional encouragement and support throughout these years and always.

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Summary

When a volcano enters a period of unrest it is necessary to assess the threat an eruption will pose. This requires the integration of multiparametric data sets. I present two examples of how different data sets with different levels of uncertainty can be combined to better constrain volcanic hazards, and also how well this information is currently being communicated. Firstly, I develop a new database of plinian eruption characteristics and use it to inform a Bayesian Belief Network that calculates the probability of an eruption becoming plinian in style. I show that the presence of an excess gas phase is one of the key factors. Secondly, I integrate geologic field studies with pyroclastic flow and ash fall modeling to improve hazard assessment at Gede volcano, Indonesia. I use this new set of data to produce an event tree for future activity at Gede. Finally, I perform an analysis of the issuance of volcano alert levels worldwide. I find that current success rates of alert issuance are low and suggest that increasing monitoring networks to an optimal level is likely to improve this.

CHAPTER 1. SUMMARY

Introduction

Every year there are volcanoes around the world that change from a state of quiescence into unrest. This awakening of the volcano poses a challenge to the observers responsible for understanding its activity. Volcanologists must use their understanding of the monitoring data, the volcanic history and the dynamics of the magma system to forecast whether the volcano is likely to move towards eruption or if it will subside back to a quiescent state. The decisions that they make and the forecasts they present to civil defense authorities have wide-ranging implications for hazard and evacuation planning of local populations. These decisions may be based on multi-parametric data sets where available, or in other cases on very limited information but in both scenarios there is a large amount of uncertainty given the complexity of volcanic systems. Uncertainties in the causative processes of unrest can lead to uncertainties in forecasts, and it is still difficult to decide when signals of an impending eruption are clear enough to issue an evacuation of the surrounding area. The issuance of erroneous forecasts can undermine the scientists' credibility and damage the future effectiveness of warnings (Aspinall et al., 2003; Sparks, 2003; Selva et al., 2012; Neri et al., 2008). The need for proper estimations of hazards and forecasting is exacerbated by the growing global populations with increasingly more people living in volcanic areas (Small and Naumann, 2001). These areas are apt for agricultural intensification due to the fertile volcanic soils, are scenically beautiful and are often recreational areas, promoting tourism in these regions (Small and Naumann, 2001; Davis et al., 2006). This increasing exposure and vulnerability of society increases the demand for effective volcanic risk mitigation measures.

An assessment of volcanic risk must take into account: hazard factors (volcano type, volcanic history and potential eruption products), unrest factors (seismic, degassing and deformation), and exposure (population, aviation exposure and infrastructure) (Ewert, 2007). Accurate hazard assessment must also take into account periods of unrest that do not lead to eruption. This is because not all unrest is magmatic in character. In fact many volcanoes will exhibit periods of non-eruptive and non-magmatic unrest

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sometimes for prolonged periods of time (Rouwet et al., 2014). Over the last twenty years there has been a shift in the volcanological community to move from the informal interpretation of unrest factors, often based on a scientists personal experience, to a more formal structured approach. This has been achieved by the introduction of probability theory into eruption forecasting, either in an event tree structure or as a Bayesian Belief Network (Newhall and Hoblitt, 2002; Aspinall et al., 2003; Marzocchi and Woo, 2007; Marzocchi et al., 2008; Sobradelo and Martí, 2010; Marzocchi and Bebbington, 2012). These methods have been used to refine hazard preparedness and eruption forecasts.

Previous and on going works in the volcanological community indicate that effective eruption forecasting and hazard planning is facilitated by: 1) An understanding of the eruptive potential at the volcano in question. This maybe achieved through detailed geologic, stratigraphic and petrologic study of the eruptive history of the volcano. 2) The identification of areas that will be particularly vulnerable to the products of a volcanic eruption both locally (by pyroclastic flows, lava flows and lahars) and regionally (due to tephra fall). This will be facilitated by a combination of fieldwork studies and numerical simulations on the transport of volcanic materials over the earth's surface and in the atmosphere. 3) Geophysical, geochemical and remote monitoring of the volcano, preferably in real time. These data allow volcanologists to track changes at the volcano that may signal the onset of a new eruptive episode. 4) A volcanic alert level system that allows for civil defense authorities and local populations to be kept informed of the current level of volcanic unrest or activity. If used correctly these alerts should represent the perception of changing risk among the scientific observers responsible (De la Cruz-Reyna and Tilling, 2008)

This thesis consists of three chapters that address some of the issues raised above. The first chapter is focused on investigating the importance of knowing the previous history of the volcano and combining it with unrest data for Plinian eruptions, which are particularly hazardous and difficult to anticipate. I have constructed a database of information pertaining to plinian eruptions and have designed a Bayesian Belief Network that calculates the probability of an eruption becoming plinian in magnitude. By focusing on how we can assimilate information from geological field surveys, petrologic studies and sparse unrest data I suggest a structure for moving towards more effective eruption forecasting. In the second chapter, I integrate recent geologic field studies at Gede volcano (the closest active volcano to Jakarta, Indonesia) with volcanic hazard simulations of pyroclastic flows (using the TITAN2D software) and ash fall (using the ASHFALL software). These are synthesized to create an event tree for the progression from volcanic unrest to eruption in this location. I suggest that the hazard posed by this volcano, situated only 70km south of Jakarta has been

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previously underestimated and that there is a need for an increased awareness of the potential disruption even a small eruption of Gede could cause. The third chapter of the thesis is a global assessment of how well the volcano forecasts and unrest levels are communicated to local populations. To assess this, I test the efficacy of volcano alert levels to accurately represent the volcanic hazard. In this analysis I account for differences in volcanic systems, monitoring capacity and the local population. I found that for a large number of cases, there is still not a high success level and I suggest what could be done to improve the situation.

CHAPTER 2. INTRODUCTION

Can we forecast plinian eruptions? Insights from a new database and Bayesian analysis.

3.1 Abstract

Eruption forecasting as pertains to the onset of an eruption has vastly improved by the advancement of monitoring techniques. It is still, however, difficult to anticipate how large or explosive an eruption is likely to be. Moreover, we have very limited instrumental evidence of the precursory signs of very large plinian eruptions. At a volcano with a history of moderate-size events (VEI 3-4) interspersed with several $\text{VEI} \geq 6$ events, how do we prepare when the volcano enters a period of unrest? How can we most effectively use the history and any available unrest data to evaluate whether this event is more or less likely to be a large, explosive event? In this study we have compiled evidence of instrumental and non-instrumental data from 29 plinian and 13 sub-plinian historical eruptions to try to identify key parameters that characterize large explosive eruptions. We looked at three groups of evidence: the preconditions of the magmatic system (e.g. presence of an exsolved gas phase), additional conditions that may develop that will trigger, promote or sustain an eruption (fresh intrusions and magma mingling, viscosity and pre-eruptive water content); and finally unrest signals that may suggest that optimal conditions are developing (e.g. seismicity that suggests strong fluid pressurization and a fully established conduit). There have been very few plinian eruptions that have been monitored in real time. We therefore rely on detailed stratigraphic and petrologic work of historic eruptions, and identify anomalous unrest behavior from the few eruptions we have data for. We perform analysis to identify data trends and then use these in a Bayesian Belief Network. This has been designed to forecast the probability that the next eruption at a given volcano will be plinian. We test this network by performing a retrospective analysis of our database to see what the probability of the eruptions being plinian was prior to the climactic phase. We find that in the case of the largest eruptions we have studied, these probabilities were 4 times greater than *a priori* probabilities that are calculated from an analysis of the global catalogue. We find that the strongest control

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on whether an eruption will become plinian is the presence of a significant exsolved gas phase in the storage reservoir.

Annie Winson designed the project with Chris Newhall; AW collected all data, performed analysis and prepared database. Fidel Costa and CN advised on data interpretation. AW designed and constructed Bayesian Belief Network with advice from Thea Hincks. AW wrote the manuscript with FC and CN. All authors reviewed the manuscript.

CHAPTER 3. CAN WE FORECAST PLINIAN ERUPTIONS?

3.2 Introduction

There has been much progress in the field of Volcanology in the last decades and many parts of the complex volcanic system are now better defined and understood (Sparks, 2003). For example, we are now able to better interpret the signs that a given volcano is moving from a period of quiescence to unrest and from unrest to eruption. Although there will always be rapid onset eruptions at volcanoes (e.g. Calbuco eruption on 22nd April 2015, which began after only one hour of seismicity, (SERNAGEOMIN, 2015)), it is becoming less common for a volcano to erupt with no warning to the observers who are responsible for monitoring its activity (Winson et al., 2014). What remains very difficult to assess ahead of time, however, is the potential that the next eruption at any given volcano may be large and explosive. This is an important question to address because it has implications for evacuation and hazard management. If we are able to correctly interpret the information that would allow us to know that the eruption is going to be large, then we could establish larger perimeters of evacuation and have more chance of keeping local populations out of harm's way. Conversely if an eruption is going to be small then we need not uproot communities and move them away from their land and livelihoods.

Volcanoes are complex, multi-parametric systems and as such there are many different processes that can affect the size and explosivity of an eruption. For example, the amount of magma that is available, the amount of volatiles, and the geometry of the system all play a role in the style of a given eruption (Colucci et al., 2014). Moreover during ascent, the magma experiences crystallization and vesiculation that lead to rapid and transient changes in viscosity, and the exact combination of these parameters may change the outcome from effusive to explosive (Colucci et al., 2014). The combination of these can either encourage or inhibit eruption, and many of these occur within the volcano, where it is not possible to directly measure them. Thus, in addition to the potential number of variables involved, it is typical that these values are poorly constrained, and thus anticipating explosive events is an inherently uncertain exercise. Another limitation is that large eruptions occur infrequently, so our data sets are small. This can make finding significant correlation between parameters of unrest and volcanic datasets difficult.

Plinian eruptions are amongst the largest events that occur at volcanoes, characterized by eruption rates greater than 1×10^8 kg/s (Carey and Sigurdsson, 1989; Cioni et al., 2000; Pyle, 2000). They have the potential of being incredibly destructive. First described at Mount Vesuvius near Naples by the writer Pliny the Younger in 79 A.D., these kinds of eruptions became infamous for their destructiveness due to the decimation of the towns of Pompeii and Herculaneum (Sigurdsson et al., 1985; Carey and Sigurdsson, 1987). As there have been few of these events in human history, and even fewer that have been instrumentally monitored, there is an incomplete record of unrest prior to these large eruptions. However, these eruptions have been the subject of many detailed geologic and petrologic studies, and their results can be used to construct a database for determining key factors and *a priori* probabilities.

In this study we have compiled a new database of 42 large explosive eruptions that have occurred

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in historic times (29 plinian and 13 sub-plinian) from 35 volcanoes. We compare these events and assess various key parameters that might help forecasting plinian eruptions. Our aim is to better constrain the processes that increase the likelihood of a plinian eruption by a combined approach of compiling existing data from past events, identifying the key variables, and integrating such information in the formulation of Bayesian Networks. First we prepare a conceptual model that captures the main criteria we believe are necessary to generate a plinian eruption. Then we show how causal relationships between different factors can be explored through a Bayesian Belief Network (BBN). This statistical tool allows multiple variables that have a high uncertainty to be captured and combined using a combination of probability and graph theory. This permits us to calculate the probability of different outcomes, given a specific set of variables. This approach is different from the commonly used Event Trees (e.g. Newhall and Hoblitt, 2002; Marzocchi et al., 2006). An event tree is a branching graphical representation of scenarios, leading from a necessary prior event to increasingly specific outcomes, finally terminating in outcomes that represent future hazards or risk. Bayesian networks, by comparison are directed acyclic graphs that allow for criteria to be assessed independently, rather than in a hierarchical fashion (Jensen, 1996; Hincks, 2007). The nodes in Bayesian networks must also be mutually exclusive, unlike those in event trees (Marzocchi et al., 2006; Hincks et al., 2014).

Bayesian Networks were first developed as tools in designing computing systems, specifically artificial intelligence and in financial institutions to model uncertainty and risk (Walley, 1996; Cowell et al., 2007). They have also been used as a tool in medical decision-making (Jensen, 2001; Stassopoulou et al., 1998). In earth sciences, networks have been proposed to make estimates of flood threat, identify areas that are at risk from desertification following forest fires, and to assess catastrophe risk amongst other applications (Stassopoulou et al., 1998; Cowell et al., 2007; Li et al., 2010). More recently there have been networks developed for the field of volcanology. Sobradelo and Martí (2010) combined Bayesian statistics with an event tree to produce an estimate of the Teide-Pico Viejo volcanic hazard (on the Canary Islands), allowing them to account for sources of volcanic unrest that were not magmatic, as well as the effect of a different vent location. Hincks et al. (2014) developed a Bayesian belief network to retrospectively analyze the unrest at La Soufrière volcano, Guadeloupe 1975-1977. Using an expert judgment approach, they showed that there was a high probability of an intrusion during this unrest period, and a probability of an eruption occurring of approximately 0.4. Other BBNs have been developed to assess the potential for eruption progression in response to specific unrest criteria, such as during the recent crisis at Santorini (Aspinall and Woo, 2014). Bayesian networks were developed to allow for the combination of observed data and expert opinion (Cowell et al., 2007) and as such we believe that a Bayesian network will allow us to find causal links in our data that may otherwise not be clear. We have therefore designed a network to answer the following question: Given that an eruption is going to occur at a given volcano, what is the probability of that eruption being plinian?

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3.2.1 Data collection

We focused on collecting, analyzing and interpreting data from plinian eruptions that have been recorded during human history. We have included eruptions that were generated at stratovolcanoes. We do not include those occurred in large calderas, although in some cases calderas were produced by eruptions in our study (i.e. Santorini). To be included there needed to be available information on at least one of the following criteria:

1. Timeline of eruption progression: These are factors that are directly observable by any witnesses, such as the duration of unrest (both seismic and phreatic) prior to the eruption, and the onset of the main phase of the eruption, and any information about the column itself.
2. The volume of erupted products (Dense Rock Equivalent (DRE)).
3. A comprehensive previous eruptive history including the repose time since the last eruption greater than VEI 4.
4. Detailed petrologic information, such as bulk rock composition or the pre-eruptive H₂O and S contents (in melt inclusions as well as in the matrix glass).

The complete dataset can be found as an electronic attachment ('Plinian Database').

We have built detailed eruptive histories of the volcanoes in our study. These data have been collected from the Global Volcanism Program (2013) and from other published work (for a complete list of references, please see the electronic attachment ('Plinian Database') and the summary tables in this chapter).

In total we have studied 29 plinian and 13 sub-plinian eruptions (see Table 3.1). The sub-plinian eruptions have been included to allow for a comparison of eruption characteristics. They were chosen purely due to the amount of information available about them. By including them we are able to calculate the probability of a specific factor (i.e. repose time) increasing the likelihood of a plinian eruption, or else the likelihood of an eruption failing to become plinian.

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Table 3.1: Plinian and Sub-Plinian eruptions included in this study. All eruption dates are those recorded in Siebert et al. (2010)

Volcano	Eruption Type	Eruption Date
Tambora	Plinian	1815
Santorini	Plinian	1610 BCE
Changbaishan	Plinian	969
Katmai	Plinian	06/06/1912
Krakatau	Plinian	20/05/1883
Huaynaputina	Plinian	17/02/1600
Santa Maria	Plinian	24/10/1902
Pinatubo	Plinian	02/04/1991
Vesuvius	Plinian	24/10/0079
Cosigüina	Plinian	20/01/1835
Hudson	Plinian	08/08/1991
Öræfajökull	Plinian	15/06/1362
El Chichón	Plinian	28/03/1982
Bezymiany	Plinian	22/10/1955
Mount St Helens	Plinian	1480
Fuji	Plinian	16/12/1707
Hekla	Plinian	29/03/1947
Okatania	Plinian	10/06/1886
Colima	Plinian	17/01/1913
Vesuvius	Plinian	16/12/1631
Komagatake	Plinian	17/06/1929
Hekla	Plinian	1104
Mount St Helens	Plinian	27/03/1980
Asama	Plinian	09/05/1783
Agung	Plinian	18/02/1963
Usu	Plinian	16/08/1663
Mount St Helens	Plinian	1800
Askja	Plinian	01/01/1875
Mayon	Plinian	01/02/1814
Shiveluch	Sub-Plinian	12/11/1964
Augustine	Sub-Plinian	27/03/1986
Galunggung	Sub-Plinian	08/10/1822
Augustine	Sub-Plinian	22/01/1976
Soufrière St Vincent	Sub-Plinian	06/05/1902
Galunggung	Sub-Plinian	05/04/1982
Pelée	Sub-Plinian	23/04/1902

Continued on next page

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Table 3.1 – continued from previous page

Volcano	Eruption Type	Eruption Date
Fuego	Sub-Plinian	10/10/1974
Kelut	Sub-Plinian	19/05/1919
Merapi	Sub-Plinian	26/10/2010
Reventador	Sub-Plinian	03/11/2002
Merapi	Sub-Plinian	15/04/1872
Kelut	Sub-Plinian	10/02/1990

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3.2.2 Definition of eruption criteria

We define plinian eruptions as episodes that fulfill at least one of the following criteria:

1. The eruption column must have been more than 10km high and have been sustained for more than one hour.
2. The mass eruption rate of the paroxysmal episode of the eruption exceeds 1×10^8 kg/s (Pyle, 2000). We estimate the average climactic mass eruption rates for the eruptions in our database for which we have been able to determine the duration of the paroxysmal phase (Figure 3.1).
3. There was evidence in the literature of a well-sorted, thick, lapilli-rich fall deposit, consistent with a plinian column (Cioni, 2000).

Most Plinian eruptions have eruptive products of a DRE $\geq 0.1 \text{ km}^3$.

Conversely, the sub-plinian eruptions in our study met one or more the following criteria:

1. They are recorded as a VEI 4 by Siebert et al. (2010).
2. The eruptive volume is $> 0.01 \text{ km}^3$ DRE.
3. There was evidence in the literature of an explosive eruptive phase, with an eruptive column that reached more than 10km elevation. This column could have been established for less than an hour.
4. Finally, if an eruption was associated with dome building events and it is unclear whether the bulk of the eruptive volume recorded comes from these phases rather than from the explosive phase, we classify them as sub-plinian (e.g. Galunggung, 1982 eruption).

To calculate the Mass Eruption Rates we assume that the bulk of the DRE recorded was emitted during the explosive phase of the eruption. Using the volume, density of the magma and the duration over which the column is sustained we calculate the MER. These values are therefore an estimate and not absolute. Table 3.2 shows the values that were used for these calculations for each of the eruptions in our study. The density of magma is calculated using the CONFLOW program of Mastin and Ghiorso (2000). The petrologic information used to make these calculations can be seen in Appendix A, Table A.3. Throughout this study we use the average value of the DRE's seen in Table 3.2.

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Figure 3.1: Mass Eruption Rates (MERs) of the eruptions in our database, for which the duration the paroxysmal phase of the eruption is known. Most of the plinian eruptions in our study have MER's of $> 1 \times 10^8$ kg/s. Where this is not the case, the eruptions have fulfilled one of the other criteria. The only plinian eruption to fall below 0.1 km^3 DRE in our study is the basaltic eruption of Mount Mayon in 1814. This eruption was classified as plinian due to the findings of Mirabueno (2001) that there was lapilli-rich fall deposit associated with the eruption. If it is unclear whether the bulk of the eruptive volume is associated with a dome-building phase, rather than an explosive phase it is not included on this figure. Eruptions for which we have not been able to calculate magma densities are not included either.

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Table 3.2: Mass eruption rates of the eruptions in this study (kg/s). Values are calculated using DRE, Column Duration and Density. Where it has not been possible to ascertain that the bulk of the recorded DRE was released in the explosive phase of the eruption, we have marked the MER with * . These points have not been included in 3.1. MER's are calculated using average DRE's, Densities calculated using values from Table A.3 in Appendix A in CONFLOW (Mastin and Ghiorso, 2000). Column height is also shown for completeness.

Volcano	Eruption Type	Eruption Paroxysm	DRE (km ³)	Column Height (km)	Column Duration (hrs)	Density (kg/m ³)	Mass Eruption Rate (kg/s)	References
Tambora	Plinian	10/04/1815 19:00	41±4	43	24	2561.3	1.20E+09	Petroeschevsky, 1949 Kandlbauer and Sparks, 2014
Santorini	Plinian	1610 BCE	50.5±9.5	36	-	2533.9	-	Sigurdsson et al., 2006 Cadoux et al., 2014
Changbaishan	Plinian	969	24±5	25	-	2511.9	-	Horn and Schmincke, 2000
Katmai	Plinian	6/6/1912 13:00	13.5	25	60	3199.4	9.85E+06	Hildreth, 1983 Hildreth and Fierstein, 2012
Krakatau	Plinian	26/08/1883 14:00	12.5	26	-	2550.4	-	Simkin and Fiske, 1984 Mandeville et al., 1996
Huaynaputina	Plinian	19/02/1600 17:00	8.8	35	19	2527.31	3.30E+08	de Silva et al., 2000 Lavallée et al., 2006
Santa Maria	Plinian	25/10/1902 01:00	8.5	28	19±1	2545.3	-	Williams and Self, 1983 Andrews, 2014
Pinatubo	Plinian	15/06/1991 13:45	5.4±0.6	35	9	2530.7	4.10E+08	Self et al., 1996 Wolfe and Hoblitt, 1996 Self et al., 2004

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Table 3.2 – continued from previous page

Volcano	Eruption Type	Eruption Paroxysm	DRE (km ³)	Column Height (km)	Column Duration (hrs)	Density (kg/m ³)	Mass Eruption Rate (kg/s)	References
Vesuvius	Plinian	24/10/0079 13:00	4	32	12	2547.1	2.40E+08	Sigurdsson et al., 1985 Carey and Sigurdsson, 1987
Cosigüina	Plinian	21/01/1835 15:00	3±0.6	25	14	2555.4	1.50E+08	Self et al., 1989 Longpré et al., 2014a
Hudson	Plinian	12/8/91	2.7	18	-	2562.2	-	Scasso et al., 1994 Kratzmann et al., 2009
Öræfajökull	Plinian	15/06/1362	1.2	30	10±2	3228.3	-	Sharma et al., 2008
El Chichón	Plinian	4/4/82 01:35	1.1	31	2	2534.6	3.90E+08	Macías et al., 2003 Tilling, 2009
Bezmyiany	Plinian	30/03/1956 17:11:05	1	45	3.5	2532.4	2.00E+08	Gorshkov, 1959 Scaillet and Pichavant, 2003
Mount St Helens	Plinian	1480	1	-	-	2542.8	-	Scaillet and Pichavant, 2003
Fuji	Plinian	16/12/1707	0.7	23	5.5	2526.5	8.90E+07	Tsuya, 1955 Yoshimoto et al., 2004
Hekla	Plinian	29/03/1947 06:20	0.76	30	2	-	-	Einarsson et al., 1967 Thorarinsson, 1967
Okatania	Plinian	10/06/1886 03:30	0.7	11.5±1.5	5.5	2547.1	9.00E+07	Walker et al., 1984 Sable et al., 2006

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Table 3.2 – continued from previous page

Volcano	Eruption Type	Eruption Paroxysm	DRE (km ³)	Column Height (km)	Column Duration (hrs)	Density (kg/m ³)	Mass Eruption Rate (kg/s)	References
Colima	Plinian	20/01/1913	0.57	23	4.6	2530.6	8.70E+07	Saucedo et al., 2010
Vesuvius	Plinian	16/12/1631 07:30	0.55	28	14	2547.1	2.80E+07	Rolandi et al., 1993 Rosi et al., 1993
Komagatake	Plinian	17/06/1929 19:00	0.52	13	4.5	2553.9	8.20E+07	Kôzu, 1934 Jousset et al., 2000
Hekla	Plinian	1104	0.5	-	-	-	-	Scaillet and Pichavant, 2003
Mount St Helens	Plinian	27/03/1980 08:32	0.5	30	9	2541.8	3.90E+07	Sparks et al., 1986 Pallister et al., 1992
Asama	Plinian	04/08/1783 10:00	0.48	10	15	2530	2.20E+07	Aramaki, 1956 Yasui and Koyaguchi, 2004
Agung	Plinian	17/03/1963	0.4	24	3.5	2528	8.00E+07	Self and King, 1996 Self and Rampino, 2012
Usu	Plinian	16/8/1663	1	-	-	2529.5 2529.5	- -	Tomiya and Takahashi, 2005 Tomiya et al., 2010
Mount St Helens	Plinian	1800	0.25	-	-	2542.8	-	Scaillet and Pichavant, 2003
Askja	Plinian	28/03/1875 21:00	0.2	-	6.5	2528	2.20E+07	Sigurdsson and Sparks, 1981
Mayon	Plinian	01/02/1814	0.04	-	6	2523.8	4.70E+06	de Mata, 1814 Crossweller et al., 2012

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Table 3.2 – continued from previous page

Volcano	Eruption Type	Eruption Paroxysm	DRE (km ³)	Column Height (km)	Column Duration (hrs)	Density (kg/m ³)	Mass Eruption Rate (kg/s)	References
Mayon (cont)								Castillo and Newhall, 2004
Shiveluch	Sub-Plinian	12/11/1964 07:20 07:20	0.75	15	1	2540.2	5.3E+08*	Gorshkov and Dubik, 1970 Ui, 1983 Belousov, 1995
Augustine	Sub-Plinian	29/03/1986	0.5	12.2	2	2532.4	1.8E+08*	Holasek, R. E.; Rose, 1991 Tappen et al., 2009
Galunggung	Sub-Plinian	08/10/1822 15:00	0.5	-	-	2523.8	-	Katili and Sudradjat, 1984 Newhall et al., 1983
Augustine	Sub-Plinian	25/01/1976 14:57	0.39	10	-	2539.5	-	Lepel et al., 1978 Kamata et al., 1991 Tappen et al., 2009
Soufrière St Vincent	Sub-Plinian	07/05/1902 12:50	0.14	15	4.5	2540.8	2.20E+07	Anderson and Flett, 1902 Carey and Sigurdsson, 1978
Galunggung	Sub-Plinian	24/06/1982	0.1	20	-	2528.6	-	Newhall et al., 1983 De Hoog et al., 2001
Pelée	Sub-Plinian	08/05/1902 08:02	0.1	-	22	2533	3.20E+06	Anderson and Flett, 1902 Tanguy, 1994 Scaillet and Pichavant, 2003
Fuego	Sub-Plinian	17/10/1974 21:45	0.08	14	2.45	1949.7	1.80E+07	Rose et al., 1978 Scaillet and Pichavant, 2003

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Table 3.2 – continued from previous page

Volcano	Eruption Type	Eruption Paroxysm	DRE (km ³)	Column Height (km)	Column Duration (hrs)	Density (kg/m ³)	Mass Eruption Rate (kg/s)	References
Fuego (cont)								Rose et al., 2008
Kelut	Sub-Plinian	19/05/1919 02:00	0.076	-	4	-	-	Kemmerling, 1921 Badrudin, 1994 Bourdier et al., 1997 Crossweller et al., 2012
Merapi	Sub-Plinian	5/11/2010 00:02	0.06	17	5	2517.7	8.40E+06	Surono et al., 2012 Costa et al., 2013 Komorowski et al., 2013
Reventador	Sub-Plinian	3/11/2002 09:12	0.058	17	0.75	2523.8	5.40E+07	Hall et al., 2004 Johnson et al., 2006 Ridol et al., 2008 Samaniego et al., 2008
Merapi	Sub-Plinian	17/04/1872	0.05	9	-	2529.7	-	Hartmann, 1935 Surono et al., 2012
Kelut	Sub-Plinian	10/2/1990 12:35	0.043	10	4	-	-	Badrudin, 1994 Bourdier et al., 1997

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3.3 Conceptual model for magma reservoirs and unrest associated with Plinian eruptions

As a starting point for our analysis we designed a conceptual model aimed at capturing the main components of the volcanic system that we believe play a key role in producing plinian eruptions, and that are also relevant in anticipating them. Given the numerous processes that occur before and during many eruptions this is not straightforward. However, we propose that the following are necessary factors for addressing the probability of a plinian eruption:

1. **The pre-conditions of the magmatic reservoir for plinian eruption** We hypothesize that there are two key pre-conditions of the magmatic reservoir. The first is that there is a large volume of melt that has accumulated and is available to be erupted. Here the information about the past eruptive volumes and frequencies will be crucial. As we have discussed above, one of the key characteristics of a plinian eruption is a high mass eruption rate ($> 1 \times 10^8$ kg/s) that can be sustained for a long duration (more than 1 hour). For this to be possible there must be a large volume of material available to contribute to the eruption. The second factor is the existence of an exsolved volatile phase within the reservoir. When an eruption begins the depressurization of this pre-exsolved phase allows for the rapid upward propagation and fragmentation of the melt, causing a highly explosive eruption.
2. **Dynamic processes and physical properties in the magma reservoir prior to eruption** If there is an amount of melt primed for eruption then we need to assess the parameters that might make this eruption more or less likely. Although magma viscosity, density and other ranges of physical properties during the eruption are probably key in bringing the magma from the reservoir to the surface, these are difficult to constrain because they change with time. Instead, we use the pre-eruptive values. Another important part in transporting the magma is the creation of a conduit (usually in the form of dike injection directly prior to eruption), so that the reservoir is fully connected to the surface and not simply unlocking discrete portions at a time (Scandone et al., 2007), as this may also cause the ascending melt to slow. Finally, we also consider the potential triggering processes of eruptive events, such as magma mixing or regional earthquakes.
3. **What unrest processes may be observed to suggest that these optimal conditions are developing?** To assess this, we draw mainly from the monitoring data available from the eruption of Mount Pinatubo in 1991. This is the best-documented, normal plinian event (not triggered by a sector collapse as at Mount St. Helens), so our analysis must be guided by the activity that was seen at Pinatubo. Primarily we evaluate the importance of escalating unrest, including the seismicity and phreatic eruptions, as these may indicate the development of the conduit as well as the rising of fluids within the volcanic system.

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We also try to include data of the duration of unrest prior to the eruption, to determine if this will give any indication of the impending explosivity.

In the following sections we address in detail the data we have gathered for these different factors and the significance of each of the different variables.

3.3.1 Pre-conditions of the magma reservoir

It is not possible to directly observe whether there is a large body of melt accumulating in a reservoir below a volcano, and therefore we have to infer this from other information. A first order estimation of the accumulated magma volume can be done by considering the past eruptive history of a volcano, e.g. the pattern of frequency vs. size of the eruptions. Many of the volcanoes in our study exhibit a long repose time between eruptions. Assuming a volume-predictable model (e.g. Koyama and Yoshida, 1994; Marzocchi and Zaccarelli, 2006) with a constant flux of magma from depth, it is apparent that volcanoes that erupt very frequently are likely to have small eruptions, as there has been less time for the magma to accumulate in the reservoir. Conversely, the volcano with longer repose times would have amassed a larger volume of magma that can be erupted, following the so-called volume-predictable pattern. We use the eruptive frequency of a volcano and its previous repose time to infer the extent of the current magma accumulation in the reservoir. This approach has inherent uncertainties related to the completeness of the eruption record. As the record is more incomplete toward the past, it is possible that under-reporting of events may begin to skew the record to the larger eruptive episodes (Whelley et al., 2015). VEI 5 eruptions are likely to be observed in the volcanic record up to 1000 years after they occur, whereas a smaller eruption such as VEI 4 eruptions maybe lost after 210 years, VEI 2's after 110 years. Section 3.5.6 will discuss the affects of these uncertainties on our study.

Using the data from the Global Volcanism Program (2014) and from other published work (for a complete list of references, please see 'Plinian Database'), we calculate the following:

1. Average repose duration of the volcano. This is assessed as the average over the time period for which eruptions are recorded in Global Volcanism Program (2014) at each specific volcano.
2. The ratio between the number of eruptions of $VEI \geq 4$ and all eruptions of $VEI \geq 2$ from each volcano (this we call the Plinian Ratio). Again the time frame for these values reflects the Global Volcanism Program (2014).
3. The repose times prior to each give eruption in our database; and
4. The ratio of the current repose to the average repose duration at a given volcano (this we call the Repose Ratio).

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Moreover, the parameters above were calculated considering two eruptive histories: 1) All magmatic eruptions, for which we assume threshold of $VEI \geq 2$; and 2) Only eruptions with $VEI \geq 4$. We consider that the latter are large enough to deplete the mobile magma from the reservoir, and thus would reset the system for new magma to begin to accumulate.

Average repose duration

We calculated the average repose duration (ARD) of $VEI \geq 2$ and the ARD of $VEI \geq 4$ for the 35 volcanoes in our sample. Average repose duration is calculated in the following way:

$$ARD = t/ne,$$

where ne = number of eruptions, and t = interval of time (years). This correlates with the record in the Global Volcanism Program (2014). As such, it is different for each volcano.

Figure 3.2: A) Average repose duration of the volcanoes in our database, plotted against the volumes of specific eruptions in the study. The box highlighted is Krakatau. After the 1883 eruption of Krakatau the eruptive style of the volcano changed significantly, moving towards a more open vent regime. As a consequence, the ARD of this system is low compared to what it would have been prior to the 1883 eruption. B) Average repose duration for eruptions with $VEI \geq 4$. Tables show the percentage of volcanoes that fall within specific ARD's. These percentages are important later for calculating the conditional probabilities associated with our BBN

We show the correlation between the average repose duration of a system and relate it to the eruption that is in our database: i.e. we look for a relationship between these factors to see if we

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can draw conclusions about whether the volcano is behaving in a volume-predictable or time-predictable fashion. Do larger eruptions occur at volcanoes that generally have longer average repose duration? Values used to calculate the ARD can be found in Appendix A, Table A.1. We found a general positive correlation between the ARD of the volcano and the eruptive volume of specific eruptions when considering the ARD of magmatic eruptions ($VEI \geq 2$) (Figure 3.2). This suggests that as a first approximation the volcanoes are behaving as volume-predictable systems (e.g. Koyama and Yoshida, 1994), although if we consider only events with $VEI \geq 4$ there is no correlation.

The ARD for magmatic eruptions is longer at volcanoes that produced the plinian eruptions (61% of them have an ARD of more than 200 years), than at the volcanoes that produced the sub-plinian eruptions (only 33% have an ARD of more than 200 years). The volcanoes that produced the sub-Plinian eruptions all have ARDs for magmatic eruptions of less than 1000 years.

68% of the volcanoes that had plinian eruptions have ARD's of explosive eruptions that are longer than 1000 years. Conversely the volcanoes that produced the sub-plinian eruptions had overall, shorter ARD's for explosive eruptions. Only 44% of them are longer than 1000 years.

The Plinian Ratio

To further characterize the potential for large magma accumulation, we quantify the recurrence of plinian eruptions at the volcanoes in our study. In other words, we calculated the Plinian Ratio as the number of eruptions with $VEI \geq 4$ divided by the number of eruptions $VEI \geq 2$ (i.e. major explosive eruptions ($VEI \geq 4$) / all magmatic eruptions ($VEI \geq 2$)). This was done for all of the volcanoes in our database. We are trying to capture the difference in volcano type and its influence on the eruptive activity. For example, is the volcanic system one that produces mainly large, explosive eruptions, or is the large eruption we have in our database just one of many smaller ones of this volcano? As the groups of volcanoes that have had plinian and sub-plinian eruptions are mutually exclusive, we are able to relate the plinian ratio of a volcano to a specific eruption. We assess whether the plinian eruptions in our sample were dominantly produced by volcanoes that had higher Plinian Ratios than those that produced the sub-plinian eruptions. Please refer to Table A.1 for data of these calculations.

Volcanoes which had plinian eruptions, have a Plinian Ratio > 0.2 in 75% of the cases in our database. This means that at these volcanoes there was at least 1 plinian eruption for every 5 eruptions in their history. Of these volcanoes, 25% of them had a Plinian Ratio of at least 0.5 (1 plinian eruption for every 2 eruptions in the volcanic history). For the volcanoes with sub-plinian events we find that 55% of had a Plinian Ratio greater than 0.2. For these volcanoes, only 11% had a ratio greater than 0.5 (or 1 in 2 eruptions). Volcanoes that have had the larger eruptions in our database have a propensity towards larger eruptions, as we might expect. We believe that this is a significant finding of our analysis as it suggests that there is a greater potential for these

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Figure 3.3: The Plinian Ratio shows the ratio of eruptions at a given volcano that are plinian compared to all those in the history. Table shows the percentage of eruptions that fall into each Plinian Ratio group.

volcanoes to accumulate large amounts of melt.

The effect of repose time on magma accumulation

A volume-predictable model with constant supply of magma flux and volatiles into the base of the reservoir implies that the volume of the next eruption will depend on the current repose period. In general the size of the eruptions in our database increases in proportion to repose time prior to the eruption (Figure 3.4), with somewhat different relations if we consider the time since the last VEI ≥ 2 or since VEI ≥ 4 eruptions.

We found that 92% of the sub-plinian eruptions were preceded by a repose time of less than 100 years, if we consider the time since the last VEI ≥ 2 event. The only sub-plinian event in our database that occurred after a repose time of more than 100 years was the Galunggung eruption of 1882 (highlighted, Figure 3.4), and maybe an artifact of under-reporting of smaller eruptions that may have happened in the interim. Comparatively, we find that only 32% of the plinian

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Figure 3.4: A) Repose time between a given eruption and the last $VEI \geq 2$ eruption that occurred prior to it. B) Repose times between the eruption in our database and the last $VEI \geq 4$ eruption that occurred prior to it. In both figures the highlighted box shows the repose prior to the Galunggung 1822 eruption. Percentage Repose times for the different eruptions are shown in the tables below the graphs.

eruptions occurred after a magmatic repose time of less than 100 years, so that 68% of them have a repose > 100 years.

This relationship is also reflected in the correlation between the volume of the eruptions and the times since the last major explosive eruption ($VEI \geq 4$). About 76% of the plinian eruptions have a repose time to the last $VEI \geq 4$ greater than 500 years and 40% of them were greater than 1000 years. If we further limit this to plinian eruptions that produced more than 1 km^3 DRE, all of the eruptions in our study have repose times longer than 500 years. The sub-plinian repose times since the last $VEI \geq 4$ are generally shorter than those for the plinian eruptions. Only 25% of these eruptions were preceded by a repose time longer than 500 years, and for only 17% of those the time was more than 1000 years. An outlier in this group is the Galunggung eruption of 1822. The last magmatic episode recorded at this volcano was caldera-forming c.4000 years ago (Bronto, 1989; Global Volcanism Program, 2014). This event increases the average of the sub-plinian repose times for all of the sub-plinian eruptions from 26 years to 337 years. With the small sample size of the sub-plinian events it is hard to infer whether this event is anomalous. The other sub-plinian event that was preceded by a repose time of more than 1000 years is the eruption of Merapi in 1872. This similarly could reflect under-reporting. Many of the Indonesian volcanoes have more robust eruption history records after the settlement of the Dutch in the 1600's and this could be playing a role here. It also could be a product of the fact that deposits of $VEI 3$ or smaller eruptions can be lost from the geologic record, especially in the tropics. It is

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very likely that this is reflecting an incomplete geologic / eruption record.

The Repose Ratio

Another way to characterize the probability of $VEI \geq 4$ eruptions is to assess the ratio between the repose time since the last $VEI \geq 4$ eruption and the ARD of $VEI \geq 4$ of the volcano it occurred at. If we assume that the ARD $VEI \geq 4$ reflects the most likely time period between $VEI \geq 4$ eruptions at a given volcano, then we can assess how far through that period a volcano was when the $VEI \geq 4$ eruption occurred. Once a volcano has a repose time since the last $VEI \geq 4$ eruption that is longer than the ARD $VEI \geq 4$ eruption, we propose that the probability of an eruption of another $VEI \geq 4$ increases. This method means that if you are assessing the repose ratio for multiple eruptions at the same volcano, the ARD $VEI \geq 4$ will remain the same but the repose time since the last $VEI \geq 4$ will change. This will generate different repose ratios per eruptive event.

We find that 33% of the plinian eruptions in our database have a Repose Ratio >1 . Meaning that the time since the last $VEI \geq 4$ eruption at the volcano in question is more than the ARD of $VEI \geq 4$ for this volcano, and implies that given the volcano's eruption history, it could be considered as overdue for a large eruption. Therefore the probability of such an event increases. Comparatively we find that 17% of the sub-plinian eruptions can be classified as overdue. For the plinian eruptions, 48% have a Repose Ratio >0.7 , whereas only 17% of the sub-plinian eruptions have a Repose Ratio >0.7 . This finding suggests that the sub-plinian eruptions mostly occurred when the repose time was less than 0.7 that of the ARD of the volcanic system. We propose that the probability of a plinian eruption increases markedly once this 0.7 value is exceeded.

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Figure 3.5: The Repose Ratio is the ratio between the time since the last $VEI \geq 4$ eruption and the average repose duration of $VEI \geq 4$ at a given volcano. Ratios are calculated for the eruption record prior to the event in our database. Table shows the bins for the percentages of eruptions that occurred after <0.2, 0.2 - 0.7 and >0.7 Repose Ratios.

Exsolved volatiles

The primary driver for generating explosive eruptions is the presence of volatiles in magma, dissolved in the melt and/or as a separate volatile phase (Wallace et al, 1995 and 2015). As magma rises through the conduit, pressure decreases which causes a decrease in the solubility of volatiles, allowing the formation of new bubbles and/or expansion of an already existing volatile phase. The dominant volatiles in magmas are H_2O and CO_2 , followed by sulfur, chlorine and fluorine. We will focus on the effect of H_2O and sulfur. H_2O has large effects on magma crystallization, temperature, density and viscosity, all of which are important factors in the eruptibility of magmas (Wallace et al., 1995). H_2O and CO_2 are the most abundant gases and therefore are the principle drivers of explosive eruptions.

Although sulfur (S) has a lower abundance in melts than H_2O , the mass produced during an

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eruption is easily measured with remote sensing methods or can be estimated by other means (e.g. ice core records (Zielinski et al., 1994; Zielinski, 1995)). Studies of several recent large explosive eruptions have shown that the amount of sulfur measured or calculated by satellite or ice core data is much larger than the amount of sulfur that could have been exsolved from the erupted melt phase alone. The difference between the amount of SO_2 calculated petrologically (Devine et al., 1984) and the amount of SO_2 measured implies the existence of an exsolved volatile phase in the magma reservoir. That is, exceeding magma saturation and existing as a separate phase within the melt (e.g., the so-called excess gas; Gerlach et al., 1996; Wallace, 2001; Scaillet and Pichavant, 2003; Sharma et al., 2004; Shinohara, 2008).

Here we hypothesize that the presence of an exsolved volatile phase is key in producing plinian eruptions. If no exsolved phase exists at depth the bubbles have to nucleate, volatiles must diffuse into the bubbles and then the bubbles must expand. Even if the melt is saturated, the volatiles are not as available when the eruption begins as they are when an exsolved phase exists. If an exsolved phase exists at depth, the bubbles need only expand to drive the high-MER, sustained eruption.

The results presented here are for 16 of the plinian events, and 4 of the sub-plinian (48% of our entire dataset). These are the eruptions we were able to collect all of the relevant information to perform the calculations. Please see this data in Table 3.3. We find evidence for an ‘Excess gas’ phase in all of the plinian eruptions and 75% of the sub-plinian eruptions. 88% of the plinian eruptions release more than 1 MT of ‘Excess’ SO_2 during the main phase of the eruption, whereas none of the sub-plinian eruptions produce more than 1 MT of ‘Excess’ SO_2 . This is to be expected but highlights that there is a greater volume of volatiles present in the magma that produce the larger magnitude explosive eruptions in our study. To assess the amount of excess sulfur in these eruptions we calculate the difference between the amounts of SO_2 observed either directly from the eruption or inferred from ice core data (see Table 3.3).

About 55% of the plinian eruptions in our entire study have sufficient data available to calculate any excess sulfur, and all of these indeed had an excess gas phase. In 69% of the plinian eruptions that had an excess gas, it constitutes more than double of the petrologically derived estimate. The number of sub-plinian eruptions for which the excess gas phase can be calculated is only 30% of our initial dataset. Three of these eruptions show an excess gas phase. The Galunggung 1822 and Fuego 1974 eruptions released c.5% and c.12% respectively more SO_2 than inferred as the petrologic yield, only marginally significant at best. The Pelée 1902 eruption, however, released c.1450% more. The last sub-plinian eruption released less SO_2 than would have been suggested by the petrologic method (and therefore did not have an excess gas), releasing c.78% of the petrologic amount (Galunggung, 1982). The high number of plinian eruptions with excess gas suggests that it is very important in driving explosive eruptions. The low number of sub-plinian examples, however, makes it difficult to draw any robust conclusions about excess gas in these smaller events.

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Figure 3.6: A) Excess SO₂ emissions during the eruption (Mt) (data for 16 plinian and 3 sub-plinian eruptions). B) Shows the Excess SO₂ as the difference between the petrologically calculated and measured SO₂ / observed SO₂; data from the same eruptions are connected to each other (16 plinian, 3 sub-plinian). C) Excess Gas (Monitored-Observed SO₂) as related to the repose time since the last VEI ≥ 4 eruption (16 plinian and 3 sub-plinian). D) Weight percent of SO₂ as a product of the mass of magma involved in the eruption. The results presented here are for 16 of the plinian events, and only 3 of the sub-plinian (43% of our entire dataset).

A longer repose time prior to the eruption allows for greater quantities of melt to accumulate but also allows for greater amounts of volatiles to accumulate including sulfur and thus produce an exsolved gas. We found a positive relationship between the amount of SO₂ released and the repose time (Figure 3.6c). Finally we recalculated the SO₂ weight % of the total magma volume as a measure of the volatile per magma proportion, which could also be a critical indicator of the explosivity of the eruption. Surprisingly, we see a decreasing trend in the data set (Figure 3.6d), which suggest that smaller volume plinian eruptions might need a proportionally large amount of excess gas per magma volume to drive the final eruption towards a major, plinian regime. It is difficult to assess the uncertainty of these data (none of the S(MI), S(MG) or SO₂ data were provided with their errors), so it is possible that the figures for Galunggung 1822 and Fuego 1974 fall within the uncertainty of the data.

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Table 3.3: ‘Excess gas’ phase is inferred if there is a marked difference between the petrologically derived SO₂ and the SO₂ measured by other means, such as TOMS, DOAS, IC/OD data. Here we show the values for the Sulfur in the melt inclusions (S(MI)), the S in the matrix glass (S(MG)) and the Crystal content wt% that we used to calculate the petrologic sulfur. We use the same method as Wallace (2001) and Sharma et al. (2008). The ‘Excess’ SO₂ amount is simply the difference between this value and the measured SO₂. Finally we compute the ‘Excess’ SO₂ as a wt% of the original volume of the eruptive products. We use the average of the DRE seen in Table 3.2.

Volcano	Eruption Type	S(MI) (ppm)	S(MG) (ppm)	Crystal (wt%)	Petrologic SO ₂ (Mt)	Measured SO ₂ (MT)	Excess SO ₂ (MT)	Excess wt(%)	References
Tambora (1815)	Plinian	689±80	290 ±74	10	83.8	93±37	9.2	0.01	Self et al., 2004. Scaillet et al., 1998.
Santorini (1610 BCE)	Plinian	74	53	12.5	6.7	130.6	123.9	0.10	Scaillet and Pichavant, 2003 Cadoux et al., 2014
Changbaishan (969)	Plinian	455	205±40	3	9.0	26.2	17.2	0.03	Horn and Schmincke, 2000 Scaillet and Pichavant, 2003
Katmai (1912)	Plinian	65 *	19±5 *	1.5±0.5	0.6	20	19.4	0.04	Westrich et al., 1991 Hildreth, 1983
Krakatau (1883)	Plinian	200	100	11	7.0	36	29.0	0.09	Mandeville et al., 1996 Scaillet and Pichavant, 2003
Huaynaputina (1600)	Plinian	264±50	80	18.5	15.1	100	84.9	0.38	Costa, 2003 Scaillet and Pichavant, 2003
Santa Maria (1902)	Plinian	198	108	24.7	9.6	44	34.4	0.16	Scaillet and Pichavant, 2003

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*: Values are from the rhyolites in the Katmai eruptive sequence

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Table 3.3 – continued from previous page

Volcano	Eruption Type	S(MI) (ppm)	S(MG) (ppm)	Crystal (wt%)	Petrologic SO ₂ (Mt)	Measured SO ₂ (MT)	Excess SO ₂ (MT)	Excess wt(%)	References
Pinatubo (1991)	Plinian	66±40	30±10	47	4.5	17	12.5	0.09	Rutherford and Devine, 1996
Cosigüina (1835)	Plinian	580±120	130±40	22.5	15.5	26.8	11.3	0.15	Longpré et al., 2014a Longpré et al., 2014b
Öræfajökull (1362)	Plinian	77	50	2.3	0.05	1	1.0	0.02	Scaillet and Pichavant, 2003 Sharma et al., 2008
El Chichón (1982)	Plinian	200	150	24	0.67	7±2	6.3	0.23	Luhr and Logan, 2002
Bezymiany (1956)	Plinian	514	47	27	6.4	7.2	0.8	0.03	Scaillet and Pichavant, 2003
Mount St Helens (1480)	Plinian	71	38	27	0.45	13.06	12.6	0.50	Scaillet and Pichavant, 2003
Mount St Helens (1980)	Plinian	68	25	40	0.44	1	0.6	0.04	Bluth et al., 1993 Gerlach and McGee, 1994
Hekla (1104)	Plinian	71	10	5 ‡	0.08	20.2	20.1	1.61	Scaillet and Pichavant, 2003

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‡: Crystallinity is unknown so is assumed to be the same as the Hekla 1947 eruption.

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Table 3.3 – continued from previous page

Volcano	Eruption Type	S(MI) (ppm)	S(MG) (ppm)	Crystal (wt %)	Petrologic SO ₂ (Mt)	Measured SO ₂ (MT)	Excess SO ₂ (MT)	Excess wt(%)	References
Agung (1963)	Plinian	650	70	5.75	0.67	7	6.3	0.63	Self and King, 1996
Galunggung (1822)	Sub-Plinian	200	100	30 ◇	0.76	0.8	0.04	0.003	Bronto, 1989 Katili and Sudradjat, 1984
Pelée (1902)	Sub-Plinian	77	50	47	0.06	1	0.9	0.37	Scaillet and Pichavant, 2003
Fuego (1947)	Sub-Plinian	3450±1750	1300	32	2.8	3.2	0.4	0.17	Scaillet and Pichavant, 2003

◇: Crystallinity is unknown so is assumed to be the same as the Galunggung 1982 eruption.

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3.3.2 Dynamic processes and physical properties in the magma reservoir prior to eruption

Magmas are multiphase mixtures of crystals, liquid and gas (Pistone et al., 2013). As such it can be difficult to reconcile the competing role of different factors to assess the eruptibility of a melt, especially as these factors will all change over time and varying pressures (Takeuchi, 2011). However, the physical properties of magma whilst it is in the volcanic reservoir (as opposed to the pre-eruptive conditions previously discussed) also drastically affect the potential of volcanic eruptibility (Colucci et al., 2014). The changes in physical properties of the magma during its ascent vary on very short time scales and are affected by geometric factors (e.g. size of the conduit) and by many kinetic processes (bubble nucleation and expansion, crystal growth, viscous heating) and thus it is not possible to compile these values for the different events. Given this complexity we have instead looked at what can be gathered from the pre-eruptive conditions and processes that may make magma more likely to erupt more explosively and therefore allow for a greater magnitude of eruption. For example, if a melt is too viscous to move into dikes then it is likely to stall at depth, before reaching the surface. Thus there is a minimum viscosity necessary to erupt, since as magma is transported to the surface its viscosity will typically only increase. This sets scenarios for threshold values that may affect explosivity. We have also assessed factors that may lead to triggering of eruptive events, such as magma mixing and dynamic stress triggering of earthquakes.

Pre-eruptive water content

The control on the amount of pre-eruptive H_2O has on the size of the eruption has been explored. Ideally, one would like to calculate the amount of exsolved water in a similar manner to sulfur, but this is difficult to do since we also need to know various thermodynamic parameters to calculate the composition of the volatile phase (Burgisser et al., 2015). We don't see a significant correlation between measured higher pre-eruptive water contents and larger eruption volumes, although 80% of the eruptions greater than 10km^3 in DRE volume have values greater than 3.5 wt% H_2O (see Table 3.7. This lack of correlation is opposed to the results of numerical modeling by Neri et al. (1998) and Papale et al. (1998), who predicted a relationship between higher water contents and more explosive eruptions. They modeled that plinian eruptions would have water contents of c.6%. It is unclear if this is a product of our small dataset, but it does suggest that explosivity may not be controlled as strongly by water content as was previously supposed. It is unclear, however, if this is an artifact of our small dataset and/or leakage of water from melt inclusions, or a real lack of correlation. The amount of pre-eruptive water in the melt should also play a role in the pre-eruptive viscosity and thus the ability of magma to move into a conduit (see below).

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Figure 3.7: Pre-eruptive water content (weight %) as measured from the melt inclusions is shown for 88% of our sample. Where we have been unable to retrieve a value from the literature, we have use the plagioclase-liquid hygrometer (Lange et al., 2009) to calculate the pre-eruptive water (these values marked with a * in Table A.4 (Appendix A) and in this figure).

Viscosity

The viscosity of magma can determine whether a dike will propagate to the surface and erupt, rather than solidify enroute as a failed intrusion (Takeuchi, 2011; Moran et al., 2011). It is possible to calculate the viscosity of magmas that have been erupted in past event using petrological analysis of volcanic rocks and viscosity models. We used the model by Giordano et al. (2008) to calculate the melt viscosities and the Einstein - Roscoe equation to account for the phenocryst content of the magmas. This requires that we know the bulk rock composition, pre-eruptive water content and the temperature of the melt. If either the temperature or the pre-eruptive water content are unknown we calculate the value using the Lange et al. (2009) plagioclase hygrometer / thermometer, which requires that we know the An/Ab values for plagioclase, the bulk rock composition, pressure and either the temperature or water content. All values for these parameters were taken from the literature and can be seen (with references) in the attached 'Plinian Database'.

We were able to calculate the viscosities for 88% of our database. We see no clear correlation between the viscosity and the size of the eruptions, but there is perhaps a slight hint of positive correlation. High viscosity would not facilitate large eruption, but it might be that the presence of an excess gas in magma of the largest reservoirs ($\geq 10\text{km}^3$) causes the viscosity of these

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Figure 3.8: Viscosity and crystallinity compared to eruption volume. Where possible we have calculated the viscosity of the melts using Giordano et al. (2008). Weight percentage of pre-eruptive water or temperature and where unknown we calculated them using the Lange et al. (2009) plagioclase hygrometer / thermometer. These values are marked with a * in our database.

magmas to cluster around $\log_{10} 4$ and 5 Pa·s. Is this an optimal viscosity for larger eruptions, or just a coincidence? Our database shows an apparent decreasing trend of phenocryst content and increasing eruption volume, reflecting many voluminous, crystal-poor rhyolite eruptions. This does fit with numerical models that suggest that crystal poorer melts are associated with more larger magnitude eruptions, although these were also proposed to be water rich, which we do not find (Papale et al., 1998). This decreasing trend in crystallinity is certainly not a global one, as some very explosive eruptions also have products of high crystallinity (e.g. Fish Canyon tuff (Bachmann, 2002)).

Potential eruption triggers: Mafic magma intrusion and mingling

The triggering of silicic explosive eruptions by the introduction of a basic melt into a more silicic reservoir has been proposed by many studies (Anderson, 1976; Sparks and Sigurdsson, 1977; Pallister et al., 1992; Eichelberger, 1995; Murphy et al., 1998; Coombs et al., 2000). The injection of fresh melt into a crustal reservoir is thought to promote eruptions in at least two ways (Sparks and Sigurdsson, 1977; Eichelberger, 1995): (1) The higher temperature mafic melt transfers a large amount of heat to the reservoir; this reduces the viscosity and density at the base of the existing magma and would induce convection. This may lead to rapid overturning and vesiculation of the reservoir, which will become oversaturated with volatiles. (2) Volatile saturation and exsolution from the mafic magma either by the pressure decrease from the deep to shallow reservoir, or by rapid cooling and crystallization once within the cooler silicic magma, will lead to an increase in volume, decrease in density, and generate gas overpressure (Folch and

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Martí, 1998). It might also introduce more water into the magmatic system and therefore lower the viscosity. These processes should cause volumetric changes in the reservoir and thus increase the reservoir pressure leading to opening fractures that may ultimately facilitate the creation of a pathway towards the surface.

Although magma intrusion and mingling is considered to be a common trigger for explosive eruptions, the time frame between these two events is quite variable. For the eruption of Mt. Pinatubo (1991), Pallister et al. (1996) and Rutherford and Devine (1996) showed that, using disequilibrium arguments, there was intrusion of a basaltic melt into the dacitic magma within 4 days of the onset of lava dome extrusion. However, White (1996) linked 2 swarms of deep long period (DLP) earthquake events that occurred in late May (3 weeks before the paroxysmal phase of the eruption) to the injection of fluids at depth. These may reflect batches of basaltic melt being intruded into the roots of the plumbing system at that time, a process that probably took place in the preceding months. Other studies have reported ranges from months to hundreds of years between intrusion and mixing and explosive eruption (e.g. Zellmer, 2003; Costa and Chakraborty, 2004; Druitt et al., 2012) suggesting that multiple mafic injections before eruption prime a system before final triggering of an eruption.

We compiled evidence for magma mixing prior to the eruptions from the literature (Table 3.9). We focused on macroscopic compositional differences between the intruding and resident magmas. We consider the ‘resident’ magma as those present in storage and eventually, the most voluminous product. The ‘injected’ magma is introduced into the already established magma reservoir.

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Figure 3.9: Petrological or geochemical evidence for magma mingling / intrusion occurred before 12 plinian and 4 sub-plinian eruptions from our database. Red diamonds show the SiO₂ composition of resident magma and are connected to the red squares, which are the SiO₂ composition of the injected magma (both, bulk compositions). Sub-plinian magmas are represented similarly by the black diamonds and squares. Dashed lines show SiO₂ compositional boundaries.

Evidence for magma mingling is reported in 41% of plinian and 31% are sub-plinian events in our database. In all cases the injected magma has a lower bulk SiO₂ wt % than the resident magma. For the eruptions that we have studied, there tends to be a larger difference in the SiO₂ contents between the resident and the injected melts in the plinian eruption. The average difference in weight percent SiO₂ (resident minus intruding) for plinian eruptions is 12.5%, whereas this value is 8% for the sub-plinian melts. This suggests that larger compositional and thermal gradients between the resident and the injected magma occur for plinian events. This could lead to quenching of the injected magma, leading to rapid crystallization, volatile exsolution and the generation of larger overpressures, all of which will likely drive the eruption to become explosive at an early stage (Folch and Martí, 1998).

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Table 3.4: Data for evidence of magma mixing, including the bulk SiO₂ of inferred ‘Resident’ and ‘Injected’ Magma’s involved in the eruption. These are the values described in the literature for most silicic and most basic rocks analyzed from each of these eruptions, whether they be actively mingled or the deposits pass in phases from the first to the last. Where the deposits are mingled we assume that ‘Resident’ magma is that which makes up the larger proportion of the deposit.

Volcano	Eruption Type	Year	Resident Magma (SiO ₂ %)	Injected Magma (SiO ₂ %)	References
Santorini	Plinian	1610 BCE	70±0.5	58±4	Druitt, 2014
Katmai	Plinian	1912	77±0.6	58.6	Hildreth, 1983
Krakatau	Plinian	1883	74.4	61.6	Mandeville et al., 1996
Pinatubo	Plinian	1991	64	54.5	Pallister et al., 1992
Hudson	Plinian	1991	60	54	Kratzmann et al., 2009
Fuji	Plinian	1707	67	50	Watanabe et al., 2006
Colima	Plinian	1913	61.5	54.2	Saucedo et al., 2010; Crummy et al., 2014
Komagatake	Plinian	1929	60	54	Takeuchi and Nakamura, 2001
Asama	Plinian	1783	73	64	Yasui and Koyaguchi, 2004
Agung	Plinian	1963	57.55	53.8	Self and Rampino, 2012; Self and King, 1996
Usu	Plinian	1663	73.9	54.1	Tomiya and Takahashi, 2005
Askja	Plinian	1875	72.17	51.69	Sigurdsson and Sparks, 1981
Augustine	Sub-Plinian	1986	63	57	Roman et al., 2006
Augustine	Sub-Plinian	1976	64.2	58.6	Johnston, 1978; Johnson et al., 1996
Pelée	Sub-Plinian	1902	62	53	Pichavant et al., 2002
Reventador	Sub-Plinian	2002	66.8±0.5	58.5±0.3	Ridol et al., 2008

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Triggering by regional earthquakes

Regional earthquakes have also been proposed to trigger eruptions, specifically where magmas or volcanic systems are already primed for eruption. Linde and Sacks (1998) found an increase in the number of volcanic eruptions within 1-2 days of large, regional earthquakes. However, none of them were plinian or sub-plinian. It is not clear whether the triggering of an eruption is due to a change in the regional stress field following the earthquake or if it is a dynamic response to passage of seismic waves or even both. A change in the regional stress field could be due to a tectonic event, glacial unloading or sea level change (Manga and Brodsky, 2006). Static stress changes caused by slip on fault planes can affect flow pathways by unlocking pre-existing faults, allowing magma to rise more easily to shallow depths (Walter et al., 2007; Lupi and Miller, 2014). In other words, regional earthquakes could promote magma replenishment over succeeding months or years. Dynamic triggering by seismic waves can occur over a shorter range of timescales and a longer range of distances (Hill et al., 2002; Manga and Brodsky, 2006). This occurs due to loosening bubbles from walls, stimulating bubble nucleation, causing loose crystal mats to fall from a reservoir's roof and sink like a plume, displacing other magma upward, or by other mechanisms (Hill et al., 2002; Manga and Brodsky, 2006). It has been suggested that static stress changes caused by slip on fault planes will affect flow pathways. Thus allowing magma to rise more easily to shallow depths (Walter et al., 2007; Lupi and Miller, 2014). In other words, regional earthquakes could promote magma replenishment from depth. Whilst these relationships are still in need of further study it seems plausible that, under the optimal conditions, both static and dynamic stress changes associated with large earthquakes are capable of triggering unrest and potentially eruption (Hill et al., 2002; Hill and Prejean, 2015).

In our data set we only found four cases where large earthquakes appear to be both temporally and / or spatially linked to the eruption (that is within one year of the eruption and / or within 100km). At Mount Pinatubo (1991 eruption), there was a regional earthquake of magnitude c.7.8, 100km away from the volcano, 8 months before the onset of volcanic activity and 10 months before the paroxysmal eruption (Punongbayan et al., 1996). Another case is the Hiei earthquake (magnitude c.8.4) that occurred near the Kii peninsula, approximately 300km away from Fuji volcano on the 28th October 1707 (Chesley et al., 2012). This earthquake preceded the Fuji eruption by 49 days. At Santa Maria volcano (Guatemala) prior to the 1902 eruption there were 2 magnitude 8.3 regional events, the first on the 19th of April, and the second on the 23rd of September (189 and 32 days respectively before the paroxysmal eruption) (Williams and Self, 1983). Finally, the eruption of Vesuvius in 79 AD was preceded by an earthquake, which was believed to have occurred directly below the volcano on the 5th of February 62 AD, 17 years before the plinian eruption Scandone and Giacomelli, 2008.

Although the small number of cases precludes any conclusions, we find it instructive to incorporate additional observations to test the potential earthquake-eruption links. For the Mount Pinatubo 1991 and Fuji 1707 eruptions, the Repose Ratio of these eruptions is low (0.29 and

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0.42 respectively) meaning that the plinian event was not 'Due' when compared to the rest of the eruptive history of these volcanoes. In contrast, Repose Ratios for Santa Maria and Vesuvius were high (1.79 and 1.39, respectively) suggesting that those eruptions were overdue in both cases and that the system was therefore primed for eruption. In addition, all events show strong evidence for mafic magma intrusions, although they appear to have affected the magma reservoirs differently. The two eruptions with the low Repose Ratios demonstrate macroscopic mingling (Pallister et al., 1992; Watanabe et al., 2006). At Pinatubo (1991) there is evidence of recent injection (months) of more mafic melts and the first magma erupted was a hybrid (mixed) andesite that punched its way up through the resident but initially sluggish, crystal-rich dacite. At Fuji (1707) the deposits were dacitic in the early stages but the final products were basaltic in composition, also suggesting a mafic intrusion. In the case of Pinatubo 1991, the earthquake of 1990 is thought to have squeezed mafic magma upward (Bautista et al., 1996; Newhall et al., 2002). In the case of Fuji, the Hoei earthquake and 4-years-earlier Genroku earthquake are thought to have facilitated ascent of magma from 20km up to a reservoir at 8km depth, after which magma mixing took over (Chesley et al., 2012).

There is evidence of magma mixing at Santa Maria (E. Calder, personal communication, 2015). Williams and Self (1983) reported a component of slightly scoriaceous, dark gray, high-Al basalt scattered throughout the deposit near the volcano. They interpreted this to be from a juvenile intrusion of mafic magma into the main dacitic reservoir. It is possible that the two large earthquakes in the year prior to the Santa Maria 1902 eruption could have signified a recharge of mafic melt into the reservoir. Triggering the accelerated formation of bubbles, which changed the overpressure of the system and/or sinking of a plume of crystals and resulting convective overturn.

In contrast, the Vesuvius lacks macroscopic evidence of magma mixing. Deposits from this eruption does, however show mixing at a phenocryst level. The phenocrysts in the melt suggest that the phonolitic magma reservoir was re-charged multiple times in the century preceding the eruption (Morgan et al., 2006).

The connection between the eruption of Vesuvius in 79 AD and the earthquake of 62AD although well correlated spatially (the earthquake was reported to have occurred directly beneath the volcano) was separated temporally. The 17-year time lag appears to be too long for the eruption to have been a product of mechanical triggering. The earthquake seems be related to a large replenishment of the reservoir by a deeper magma (Morgan et al., 2006). The fact that the Repose Ratio was quite high and that there was still a pause of 17 years between the earthquake (and potential intrusion of new material) and eruption suggests that neither of these processes were the final triggers of the eruption.

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3.3.3 Diagnostic magmatic unrest.

To determine the probability of a plinian eruption it is critical to be able to make inferences about unrest as it unfolds. However, there is a dearth of monitored instrumental data for such large explosive events. We were able to record some factual accounts for most of the eruptions in our study, including when seismicity was noted to be higher than usual, and when ash plumes were first recorded. One of the few examples of monitored data from a plinian eruption is that of Pinatubo, 1991 and we'll base our inferences mainly upon this event. These data may not be representative of all plinian eruptions, but they do suggest parameters that might be important before such explosive eruptions. The initial statistics and data analysis will be improved with the addition of future events.

In addition, after a survey of the literature, the type of unrest data that we have compiled includes the time since the first recorded earthquakes and the first phreatic eruptions and the climactic event, the change in long period earthquakes, and the number and size of explosions before the climactic event.

Time of unrest before paroxysmal eruption

Passarelli and Brodsky (2012) found a positive correlation between the repose time of a volcanic system and the time between the onset of magmatic unrest and the onset of magmatic eruption. They define the onset of magmatic unrest as the point where juvenile magma is first noted in ash. We follow Passarelli and Brodsky (2012) and look for relationships between the repose time and DRE, and the seismic and phreatic run up times. The run up time is defined as the number of days between the onset of either anomalous seismicity or of ash plumes to the paroxysm of the eruptive episode. For most of our data it is unknown when in the eruptive sequence the first juvenile ash was erupted, and thus we have instead noted the first recorded instance of any ash as the beginning of the phreatic run up time. We also note that 'anomalous seismicity will usually refer to the first felt seismicity, as there is a lack of monitored data for these eruptions.

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Figure 3.10: A) Seismic run up time vs. DRE, B) Phreatic run up time vs. DRE, C) Seismic run up vs. time since the last eruption, D) Phreatic run up time vs. time since the last eruption. To see this data and associated references please see Appendix A, Table A.5

We find some weak correlations between the size of the eruption and the number of days of seismic or phreatic activity prior to the paroxysm of the eruption for all of the eruptions. Plinian eruptions have, on average, slightly shorter seismic lead times. 48% of the plinian eruptions occur after <10 days of seismic run up time, while for the sub-plinian eruptions this is 27%. The sub-plinian eruptions are also more likely to have seismic precursors that last longer than

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100 days (36.5% of eruptions) than the plinian eruptions (26% of eruptions). To a first order this suggests that the probability of a plinian eruption decreases after there has been 100 days of anomalous seismicity with no eruption. Apparently, once magma begins to ascend in the volcanic system, a volume capable of producing a plinian eruption is likely to accelerate quickly. As we have already discussed, there is a higher potential for an excess gas phase in plinian eruptions than in smaller eruptions. The availability of this gas could drive faster ascent, leading to an overall shorter seismic run up for these larger eruptions than we see in the sub-plinian eruptions.

Differences in the phreatic run up times are small, with 62% of plinian eruptions and 73% of sub-plinian eruptions having ≤ 10 days of phreatic activity prior to the main phase of the eruptions. We have no sub-plinian eruptions > 100 days of phreatic activity prior to main phase, whereas 14% of the plinian eruptions do. This means that we are seeing the opposite relationship to the seismic run up times, with > 100 days of phreatic activity increasing the probability of a plinian eruption. Larger (plinian) eruptions are often associated with large hydrothermal systems (i.e. Mount Pinatubo (Pallister et al., 1996)) that can start to produce phreatic ash eruptions even when magma is still deep and only starting to ascend. We note however, that it is difficult to draw firm conclusions, as the dataset for these criteria is small.

Passarelli and Brodsky (2012) proposed that if the repose time of the system is associated with magmatic recharge, there should be a linear relationship between the repose and the run up time. Magma recharge could begin almost immediately after the previous eruption, but it is mostly aseismic or without felt seismicity. We do see a hint of correlation between the seismic run up time and the repose time since the last magmatic eruption, but perhaps for a different reason. Volcanoes that have been quiescent for a longer time maybe well plugged and therefore more time is required to propagate new cracks to form a new conduit. For example for a repose time of $\geq 10,000$ days (27 years) and a seismic run up time of ≥ 10 days 75% of the eruptions that occurred were plinian and 25% were sub-plinian. We also see a positive relationship between the phreatic run up time and the time since the last eruption. For the same example as above ($\geq 10,000$ days repose and run up time of ≥ 10 days) 78% of the eruptions in our sample were plinian (22% sub-plinian).

The paucity of the data available makes it difficult to robustly link the size of eruptions, run up times an repose times, but our first order analysis of the available records (see Appendix A, Table A.5) suggests another tool for estimating probabilities of plinian eruptions.

Increasing shallow long-period earthquake swarms

It is challenging to isolate diagnostic unrest before plinian eruptions as we have a very small sample that has had any kind of instrumental monitoring. Whilst the Mount St Helens 1980 had monitoring prior to and during a plinian eruption, there was also bulging of the north flank. Because of the subsequent sector collapse and lateral blast, the normal progression toward a

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vertically directed eruption was foreshortened and we do not know what precursors might have occurred without the collapse and lateral blast. Other than the bulge (and the collapse and blast themselves), precursors to the plinian eruption of Mount St. Helens were unremarkable (personal communication C.Newhall). Similarly we are aware that there is some seismic information available for the eruption of El Chichon in 1982, but the station that recorded this data was part of a network monitoring a hydroelectric plant and was at a distance from the volcano (c.50km south), and is therefore not sufficient for our needs (Tilling, 2009). Thus, our ideas of ‘diagnostic unrest’ are largely predicated on the eruption of Mount Pinatubo in 1991, where there were several features of unrest that were sufficiently remarkable that, if noted during an unrest phase at another volcano, should be treated as indications of an increased potential for plinian eruption.

The first signs of unrest at Pinatubo were locally felt earthquakes beginning in mid-March 1991, culminating in a steam explosion on the 2nd of April (Wolfe and Hoblitt, 1996). At this time the Philippine Institute of Volcanology and Seismology (PHIVOLCS) installed seismometers and began to record 30-180 high frequency seismic events per day, which lasted into May (Wolfe and Hoblitt, 1996). Until the 1st of June these earthquakes were clustered in a zone between a depth 2 and 6km, approximately 5km to the northwest of the summit of Pinatubo (Cornelius and Voight, 1996; Harlow et al., 1996). These earthquakes were associated with a 10x increase in the amount of SO₂ flux (Cornelius and Voight, 1996; Wolfe and Hoblitt, 1996). During the first week of June a second cluster of earthquakes developed between the surface and a depth of 5km, this time only 1km to the north west of the summit (Wolfe and Hoblitt, 1996). This zone of seismicity was suggested to delineate a newly re-established conduit. At this time there was also a marked decrease in the SO₂ flux of the volcano (Daag et al., 1996; Wolfe and Hoblitt, 1996). This activity persisted until the 3rd of June when, at 19:39, the first small explosion was noted (Harlow et al., 1996; Wolfe and Hoblitt, 1996). In the days following there were several other small explosions (Harlow et al., 1996) and on the 8th of June a dome was confirmed to be growing in the crater (Wolfe and Hoblitt, 1996). From the 8th until the 12th of June, ash emissions increased (Wolfe and Hoblitt, 1996). There were also swarms of shallow earthquakes beneath the dome and episodes of harmonic tremor. The first major explosive eruption began at 08:51 on the 12th of June, producing an ash column that reached 19km (Cornelius and Voight, 1996; Hoblitt et al., 1996; Wolfe and Hoblitt, 1996). Over the next two days, three more explosions of similar size destroyed the dome that had formed (Wolfe and Hoblitt, 1996). The explosions accelerated on the 14th of June a sequence of 13 explosions began at 13:09, finally culminating on the 15th of June at 13:42 in the climactic phase of the eruption (Wolfe and Hoblitt, 1996). This paroxysm lasted for 9 hours and generated columns that reached 35km and produced an eruptive volume of 4.8-6.0km³ (DRE) (Wiesner et al., 2004). A full description of all parts of this eruption can be found in Newhall and Punongbayan (1996) .

Was any of this unrest at Pinatubo particularly indicative of an upcoming plinian eruption?

There were several phases of pre-eruptive seismicity (Harlow et al., 1996) in the run up to the climactic eruption on the 15th of June. We have plotted the RSAM of all seismicity (Fig:3.11).

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Figure 3.11: Normalized Cumulative RSAM recorded at Pinatubo before the climactic eruption. The data presented by the light blue segment of the graph is adapted from Harlow et al. (1996), and is the 4 hourly average RSAM. The data described by the darker blue segment of the graph is adapted from Cornelius and Voight, 1996); these are a 10-minute interval, cumulative RSAM recording. These data were taken from 2 different seismometers. * Shows the period dominated by LP swarms, the paroxysmal eruption is marked with an arrow.

The RSAM linearly increased from about 10th of May - 11th of June, with a fairly constant rate (with a slope of 0.02). From the 12th of June we see an acceleration in the RSAM, with the slope increasing 3 times to 0.06 between the 12th and 14th of June. During this period Harlow et al (1996) report that the seismicity was dominated by swarms of LP events that lasted from 1 - 2 hours and preceded explosions that produced ash columns to various heights. From the 14th of June until the paroxysmal eruption the RSAM further increased, the slope doubled to 0.13. The LP swarms that began 12th of June at 08:51 increased prior to the explosive eruptions that were occurring at the time. They have been attributed to pressurization of fluids within the magmatic system that were trapped behind temporary barriers that were subsequently destroyed during the eruption that followed (Cornelius and Voight, 1996; Harlow et al., 1996). As the eruption progressed from the 12th until the 15th the seismic spectral amplitude measurement (SSAM) records become increasingly saturated by these LP events (Power et al., 1996). Seismic energy release exceeded 1014 J/day. We are not aware of any other eruptions where shallow LP swarms such as these dominate the seismic record to such a degree. Future episodes of unrest that reflect these characteristics should be treated with great concern.

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The significance of the frequency and magnitude of pre-climatic eruptions

Pinatubo was restless for at least 74 days before the climatic VEI 6 eruption. During this period there were a series of explosions that lead to the production of ash columns to various heights. In the fully magmatic phase from June 12th to June 15th, the plumes reached altitudes of 24km on June 12th but then became progressively smaller and the eruptive episodes more frequent until the onset of the paroxysmal eruption. Hoblitt et al. (1996) suggested that the magma pressure reached a high level before the June 12th eruption. After this vent clearing, however, it was neither necessary nor possible for the pressure to build so high and so smaller eruptions resulted. The increasing frequency of these small eruptions suggests accelerating movement of magma up the conduit (Harlow et al., 1996; Hoblitt et al., 1996). Perhaps, accelerating magma ascent builds gas pressures ever more rapidly, and leaves correspondingly less time for magma remaining in the conduit to solidify, with less and less pressure needed to breach short-lived plugs (Harlow et al., 1996; Hoblitt et al., 1996). Hence, smaller, more frequent eruptions of decreasing intensity were observed in the run up to the paroxysmal event (Figure 3.12).

Scandone et al. (2007) suggested that the formation of a fully integrated conduit (with fractures connecting from reservoir all the way to the surface) is rare. Instead most eruptions can be explained by the ascent of isolated magma batches. Yet, a full integration of a conduit may be essential before a plinian eruption can develop. Several lines of evidence from the Pinatubo eruption suggest that this occurred during the months prior to the eruption: 1) seismicity recorded at the beginning of June highlighted a zone extending from the surface to a depth of 5km, and 2) the seismic 'groan' (Power et al., 1996) on the 12th of June, consisting of continuous VT earthquakes that lasted 1 hour and 25 minutes and was thought to represent the movement of magma and volatiles into a newly established system of cracks. The VTs were thought to be generated by brittle failure as these cracks opened, whilst lower frequency events represent the movement of fluids into these new spaces (Power et al., 1996). 3) The decreasing time between the eruptive episodes described above. All of these parameters combined point towards a conduit that by the time of the climactic eruption was fully connected, fulfilling another requirement for a large explosive eruption (Scandone et al., 2007).

In summary, to determine the likelihood of a plinian eruption at the volcanoes in our study we consider three discrete aspects of a volcano's behavior: the volcano's past record of plinian eruptions, the pre-eruptive and syn-eruptive dynamic processes that may help or hinder the development of a large explosive eruption, and finally the diagnostic unrest phenomena that may suggest an increased potential for a plinian eruption. In our analysis we have found some correlation in between the volcanic history, unrest parameters, and the size of the eruption. The most compelling links can be made between the size of an eruption and the presence of an excess gas phase, the repose ratio and the plinian ratio. Whilst the connections between eruption size and unrest are less persuasive we also feel that these are important.

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Figure 3.12: Ash plume height and frequency Pinatubo case study. Data are adapted from Hoblitt et al., 1996. Red lines show ash columns, the date that they were produced and the height that they reached. Dashed lines show ash plumes that were recorded but for which no height is known. Black arrow marks the paroxysmal eruption.

3.4 Bayesian Belief Networks and the calculation of the probability of a plinian eruption

Two key problems that we face in the forecasting of explosive eruptions are:

1. Volcanic systems are inherently multi-parametric. There are many processes that work together to impede or promote an eruption, most of which cannot be directly measured prior to an eruption.
2. There are few eruptions of this magnitude that have been observed and even fewer that have been monitored. Thus, we have to draw information from small sets of data; this limits our capacity for making robust estimates about the probability of eruption. To combine the different datasets we have described above, we have designed a Bayesian Belief Network that allows for calculating probabilities in complex and highly uncertain systems.

3.4.1 Introduction to Bayesian Belief Networks

A Bayesian Belief Network (BBN) allows one to combine expert opinions with observable data (Cowell et al., 2007) and then model a cause and effect relationship in the system. The basic method of Bayesian networks is to use conditional probabilities to treat uncertainties within a

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framework of a casual network (Jensen, 2001). With this tool we are able to assess the probability $[P(a)]$ of an event (a) given that it is conditioned by various factors. When building a BBN there are 3 types of information that must be reconciled (Jensen, 1996). The variables in the system, for example the presence of excess gas, form the ‘nodes’ of the network. These nodes are defined such that they have a finite set of mutually exclusive states and can either be directly observable or else inferred from the other nodes (Jensen, 1996; Cowell et al., 2007; Hincks et al., 2014). Once data are fed into the relevant node in the network, the information is propagated to all the other nodes in the system using ‘arcs of influence’. These show the direction of influence in the system and are defined by conditional dependencies, which are represented by a conditional probability table (Stassopoulou et al., 1998). Finally these nodes and arcs are connected in a directed acyclic graph (or network) that represents the casual structure of the hypothesis that is being tested.

Plinian Bayesian Belief Network

In designing our BBN for plinian events we have used the more robust variables from our data analysis outlined above. We use a static network that does not model time dependency of criteria; instead it is only evaluated at a specific time. However, it is possible to re-run this network multiple times as more data became available, allowing the ‘plinian forecast’ to be updated. When constructing the network, our central question is: “If we know that the volcano will erupt (e.g. the probability of eruption is already 1), what is the probability that the eruption will be plinian?” It is important to stress that this network is not designed to assess the probability of an eruption occurring in any specific time frame.

To design and compute the probability of a plinian eruption using BBN we have used GeNIe (Graphical Network Interface), the graphical interface of SMILE (Structural Modeling, Inference, and Learning Engine), developed by the Decision Systems Laboratory of the University of Pittsburgh. This software was chosen for its ease of usage and because it is open source. The structure and conditional probabilities of the BBN are based on the data and relationships we have discussed above. We have tested our network with data from some of the eruptions that we studied and try to quantify the range of values different inputs may have on the BBN.

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Figure 3.13: Plinian Bayesian Belief network. The green node represents the only 'hidden' node in our network, blue nodes represent 'observed' nodes and the pink node is our 'decision' node. The conditional probability table's of each of the nodes are also displayed (here, before values are added).

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3.4.2 Network structure

We have designed a minimalistic network that uses the smallest number of nodes and relations that we think are necessary to properly inform the forecasting of plinian eruptions. Other networks are possible. Our expectation is that as more data become available, the conditional and final probabilities will progressively become better constrained. The BBN (Figure 3.13) is controlled by the three main themes from the original data analysis:

1. Indicators of the eruption history of the amount of magma accumulating within the volcanic system.
2. The presence of an excess gas as a key to promoting the explosive nature of these eruptions (linked to the past eruption history and amount of time that magma has been accumulating).
3. Unrest that might be diagnostic of an impending plinian eruption.

Nodes of the Bayesian Belief Network

Nodes in a Bayesian belief network fall into three categories; **1)** ‘Hidden’ nodes, for factors that are not possible to directly measure - for example, we may wish to know whether or not magma is rising in the conduit. We cannot measure this directly but we can use other criteria such as the seismicity, amount of SO₂ in volcanic gases, inflation of the surface and the presence of juvenile material in phreatic explosions to infer that this is indeed the case. The hidden nodes are informed by **2)** ‘observable’ nodes which have measurable inputs. In the case of our example these are criteria such as petrology of the ash deposits and the amount of inflation, seismicity or SO₂ being measured. Thresholds for the nodes are pre-defined, which allows us to state that, given specific measurement, we expect the probability of our hidden node to increase or decrease.

A BBN can contain as many hidden and observable nodes as the designer wishes but all of these nodes will relate to the **3)** ‘Query’ node. This final node defines the question that we wish to answer. In the case of our network, our query node will address the question: What is the probability that the next eruption at volcano X will be plinian? To build our BBN we have chosen what believe are the most dominant criteria from our analysis. The definitions of the nodes are outlined below.

Node 1 - Eruptible magma: This node represents the probability that an impending eruption will be plinian when the Repose Ratio is either >0.7 (Due) or <0.7 (Not Due). We hypothesize a constant magma input into the magma reservoir, and if the time since the last $\text{VEI} \geq 4$ eruption is equal to or greater than the average repose duration of $\text{VEI} \geq 4$ eruptions at the volcano, there is a higher probability of the upcoming eruption being plinian. This is because we infer that there

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will be a large volume of eruptible magma available in the reservoir. Two states are defined; ‘Due’ (the value above is >0.7 , i.e. current repose is at least 70% of the average repose time of the system for $VEI \geq 4$) and ‘Not Due’ (if the values is <0.7). We selected the threshold at 0.7 for this node as it is the median value of the plinian data set.

Node 2 - Plinian Ratio: This node is conditional on the ‘Eruptible magma’ node. We assess the probability of the Plinian Ratio being ‘High’, ‘Moderate’ or ‘Low’ given that there is or is not an inferred large volume of magma at depth i.e. that the eruption is ‘Due’ or ‘Not Due’. We define 3 states; ‘High’ which is a value of greater than 0.5, meaning that at least 1 out of 2 eruptions in the history of the volcano are $VEI \geq 4$. ‘Moderate’ values fall between 0.5 and 0.2 (1 in 5 eruptions are $VEI \geq 4$), and values <0.2 is for ‘Low’.

Node 3 - Repose Time: This node is also conditional on the ‘Eruptible Magma’ node. It describes the probability of a plinian eruption occurring after high, moderate or low number of years, given that there is or is not an inferred large volume of magma and therefore that the eruption was due or not due. We use the repose time since the last $VEI \geq 4$ eruption occurred at the given volcano to inform this node. The three states are defined as: ‘High’ when >1000 years since the last $VEI \geq 4$ eruption, ‘Moderate’ >100 but <1000 years since the last $VEI \geq 4$ and ‘Low’ <100 years since last $VEI \geq 4$.

Node 4 - Excess volatiles: This node describes the probability of a plinian if an excess gas phase exists in the reservoir. The amount of ‘excess volatiles’ is likely related to the repose time and to how much eruptible magma there is in the volcanic storage system. As such we have made this node conditional on the eruptible magma node. As it is difficult to determine what should be considered a large amount of excess volatiles, we have made this node with only 2 states. Excess volatiles are assigned as ‘yes’ or ‘no’ depending on whether there is evidence of the presence or absence of an excess gas phase.

Node 5 - Diagnostic Unrest: This is the only hidden node in our network. It is defined as the probability of a plinian eruption occurring with or without some diagnostic unrest being recorded. There are few plinian eruptions in the global catalogue and is the number from recorded history is even smaller. Therefore, we cannot quantify the diagnostic value of monitored changes. This problem becomes less pronounced in more recent eruptions. An eruption of this size occurring without recorded unrest is less likely now that monitoring systems are advanced and can telemeter data from remote locations. Until there are more examples of well-monitored plinian eruptions, we allow this node to be informed, at face value without interpretation, by data from the observed nodes in the unrest section of the network.

Node 6 - Frequency and Intensity of explosive events: This node, like that for instrumental unrest, is defined due to observations from the Pinatubo eruption. This node estimates the probability of a plinian eruption being preceded by a phase of increasingly frequent and decreasingly intense explosive events. We define 2 states (‘Yes’ / ‘No’) for this node.

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Node 7 - Duration of seismic unrest: This node defines the probability of a plinian eruption given that seismic unrest (seismic run up) will occurred for <100 days or >100 days prior to a plinian eruption. These thresholds are defined by the analysis we performed of the plinian data in section 3.3.3. We expect eruptions to be preceded by seismicity but as with some of our historic eruptions it may not have been recorded at the time.

Node 8 - Duration of phreatic / phreatomagmatic unrest: Similarly to the previous node, we assess the number of days that phreatic and phreatomagmatic (phreatic run up) events were recorded prior to the paroxysm of the eruptions in our database. This node also has 2 states; More than 70 days of phreatic or phreatomagmatic events prior to the plinian paroxysm or less than 70 days. This value was also defined by the outcome of our analysis in section 3.3.3.

Node 9 - Shallow LP Seismicity: This node is influenced by the experience of scientists at Pinatubo. It refers to the probability of there being strong and accelerating, shallow LP seismicity prior to a plinian eruption. If there were evidence of accelerating, shallow LP seismicity such as that described earlier then we would expect a higher probability of a plinian eruption. The states of this node are ‘Yes’ and ‘No’.

Node 10 - Seismicity indicative of conduit connection: As previously discussed, Scandone et al. (2007) proposed that a fully connected conduit (that is a conduit that connects the magma chamber to the surface) would sharply increase the probability of highly explosive (plinian) eruptions. At this node, we assess the probability that an eruption will become plinian, given the appearance of seismicity all the way from the magma reservoir depths to the near surface. This node has 2 states: ‘Yes’ and ‘No’.

Node 11 - What is the probability of a Plinian eruption? This is the query node. The outcome of this node is the outcome of combining the rest of the conditional probabilities.

We note that we base the probabilities of nodes 6, 9, and 10 solely on unrest recorded at Pinatubo. As such we have assigned probabilities that will not make these nodes dominant in the calculation of the network. If these unrest phenomena are seen before other plinian eruptions in the future then these *a priori* probabilities can be changed to exert a greater affect on the network than they do currently.

3.5 Discussion

The data that were used to design this network were collected from over 300 documents, published between 1900 and 2014. It is difficult to assess the effects of scientific advancements and therefore the decrease of uncertainty in records from a dataset such as this. To test our proposed BBN, we chose to study four eruptions for case study that represent a range of degrees to which we know the values for the different nodes of the system.

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We use evidence from Tambora 1815, Pinatubo 1991, Merapi 2010 and Kelut 2014 eruptions. These eruptions were picked for different reasons. Pinatubo 1991 was the first monitored plinian eruption so much of this study has been guided by the data collected at Pinatubo. The eruption of Tambora in 1815 is one of the largest volcanic events in human history and precipitated the 1816 ‘year without summer’. Little is known of its precursors, but its products have been studied petrologically and its gas emissions are known from ice cores. We also consider two smaller eruptions in our catalogue for comparison. Firstly the eruption of Merapi 2010. Whilst this eruption was not plinian, it was anomalously large when compared to previous eruptions of Merapi. We also retrospectively assess the plinian potential of the most recent eruption of Kelut (2014). The fact that this was the fifth VEI 4 eruption in the past 100 hundred years (1919, 1951, 1966, 1990, 2014) makes Kelut an interesting case study.

An *a priori* probability of a plinian eruption (as defined by the criteria we have already outlined in 3.2.2) can be estimated as the percentage of all global eruptions that are $VEI \geq 4$ (Siebert et al, 2010). This calculation gives a probability of 8% for a $VEI \geq 4$ (636 eruptions of 7739 in the record) and sets the context for the calculations that follow.

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3.5.1 Case Study 1: Pinatubo

Prior to the eruption of Pinatubo in 1991 the volcano did not have an eruption $VEI \geq 4$ (or indeed any eruption) for c.500 years (Node 3: Moderate). This however, is not an unusually long repose time for a volcano when compared to the ARD $VEI \geq 4$ (which is c.1895 years, see Appendix A Table A.1, and it therefore would have been classified as ‘Not Due’ in the ‘Eruptible Magma’ node (Node 1: 0.28). The preliminary eruption history sketched out before the 1991 eruption was, however dominated by $VEI \geq 4$ eruptions with 4 out of 5 eruptions being of this magnitude (Node 2: High). Although we only know retrospectively, there was an excess volatile accumulation in the volcano prior to the eruption (Node 4: Yes). Seismicity was noted for 93 days before the eruption and latter there were swarms of shallow LP’s that caused a sharp escalation in RSAM (Node: $7 < 100$ days, Node 9: Yes). This seismicity was present from 5km depth to very shallow, indicating that the conduit was developing a full connection to the magma reservoir (Node 10: Yes). Ash and steam eruptions began and occurred intermittently for 75 days prior to the eruption paroxysm (Node 8: > 70 days). As the explosions became magmatic they increased in frequency but the amount of energy released decreased (Node 6: Yes). Using the BBN we estimate the probability of a plinian eruption at Pinatubo was 16%, double the *a priori* probability that can be calculated from the global distribution, and c.20 times the *a priori* probability of a $VEI \geq 6$ eruption.

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Table 3.5: Probability values for the eruptions in our case studies. Data from the plinian database are displayed in the first column for each eruption; the second column shows the resulting value this data is awarded in the BBN. Where we do not know the value for the criteria we will compute the probability of the network by varying node by every state available. This allows us to compute an average probability as well as an upper and lower boundary for this value. This is represented as the variance in this table.

Case Study	Pinatubo 1991		Tambora 1815		Merapi 2010		Kelut 2014	
	Data	Node State	Data	Node State	Data	Node State	Data	Node State
Eruptible Magma	0.29	Not Due	0.75	Due	0.05	Not Due	0.07	Not Due
Plinian Ratio	0.83	High	0.69	High	0.05	Low	0.21	Moderate
Repose Time	500	Moderate	815	Moderate	138	Moderate	24	Low
Excess Gas	Yes	Yes	Yes	Yes	Yes	Yes	Unknown	Yes/No
Conduit Seismicity	Yes	Yes	Unknown	Yes/No	Unknown	Yes/No	Unknown	Yes/No
LP Seismicity	Yes	Yes	Unknown	Yes/No	No	No	Unknown	Yes/No
Phreatic run-up	74	High	831	High	10	Low	1	Low
Seismic run-up	92	Low	831?	High / Low	5	Low	12	Low
Increasing Frequency and Decreasing Intensity of explosive events	Yes	Yes	Unknown	Yes/No	Unknown	Yes/No	Unknown	Yes/No
Average Probability of a Plinian Eruption	16		32		13		10	
Variance (+/-)	0		5		2		2	

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Figure 3.14: Example of BBN for the Pinatubo eruption. The values for all observable nodes are known and therefore are reflected as either 100% or 0%. The query node shows the final probability that the pending eruption would be plinian.

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3.5.2 Case Study 2: Tambora

Mount Tambora produced a VEI 7 eruption in 1815, with a volume of 41km^3 (DRE) (Kandlbauer and Sparks, 2014). At the time of eruption Tambora had a Repose Ratio (Node 1) of 0.75 and therefore could be considered as ‘Due’ for a large eruption. Large eruptions make up most of the volcanic history at Tambora and for that reason the Plinian ratio is 0.69 (Node 3: High). There are no eruptions recorded here prior to this event for c.800 years (Node 3: Moderate). This eruption had a large volume of excess volatiles ($51\text{--}58\text{Mt SO}_2$ (Self et al., 2004)). Node 5 therefore is set as ‘Yes’. Unrest began 3 years prior to the eruption. Passing ships reported seeing ash and steam clouds over the mountain and ‘rumbling’ sounds coming from the vicinity of the volcano (Stothers, 1984; Kandlbauer, 2013; Kandlbauer et al., 2013). We infer that the phreatic run up was therefore >800 days. It is unclear whether the ‘rumblings’ that were heard could have been earthquakes. We therefore assign Node 8 to be >70 days, but we test the network with a value of both <100 days and >100 days for Node 7. It has not been possible to assign values for Nodes 6, 9 and 10 as there is not enough information to make this judgment, and we have tested the network for both possible values of these nodes. Node 5 is informed by the other unrest nodes so it is not necessary for us to assign it a value. We attain an average probability for a plinian eruption at Tambora of $32\% \pm 5$. This high value shows that there was a great potential for a plinian eruption at Tambora prior to the eruption of 1815.

3.5.3 Case Study 3: Merapi 2010

The eruption of Mount Merapi in 2010, whilst not a plinian event, was quite large for this volcano. We find that the value for Node 1 is 0.5 and as such, we assigned the state ‘Not Due’, only 1 in 25 eruptions of Merapi are $\text{VEI} \geq 4$ (Node 2: Low) and the last explosive eruption of this size occurred 138 years ago (Node 5: Moderate). There is evidence of an excess gas phase (Node 4: Yes) involved in this eruption, although we have been unable to calculate the exact amount (Surono et al., 2012; Costa et al., 2013). Whilst seismicity and phreatic eruptions were recorded prior to the eruption, these were both for short amounts of time, Node 7 is therefore given state ‘ <100 days’ and Node 8 ‘ <70 days’. There were LP’s recorded at Merapi but it is unclear whether these swarms were dominating the seismic record of the eruption as we see at Pinatubo. We are not able to ascertain if the explosions prior to the main phase were increasing in frequency whilst decreasing in intensity. For this reason we have tested the network with both outcomes for these nodes. We have also allowed the value for Node 10 (conduit seismicity) to vary. We find that the probability of this eruption becoming plinian was $13\% \pm 2$, c1.5 times higher than the value we would have expected from the GVP *a priori* value.

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3.5.4 Case study 4: Kelut 2014

Mount Kelut is in our database with the sub-plinian eruptions in 1919 and 1990. Unrest began again on the 2nd of February 2014, when seismicity was noted to be at elevated levels. The paroxysm of this eruption occurred on the 14th of February 2014. The eruption occurred during the night and it is difficult to know exactly how long this phase lasted, but it has been estimated at c.6 hours (CVGHM web site), with a column of c.20km high. The eruption released 0.5km³ DRE of material, which would suggest a mass eruption rate of approximately 6×10^7 kg/s. We calculate the probability of a plinian eruption at Kelut, supposing that we were assessing the nodes of the network on the 13th of February 2014 at midday. We find that the value for Node 2 is 0.07 and assign the state 'Not Due', 1 in 5 eruptions of Kelut are $VEI \geq 4$ (Node 2: Moderate) and the last explosive eruption occurred 24 years ago (Node 3: Low). Prior to the eruption we would not have had any evidence about whether there was an excess gas phase in the reservoir. We therefore do not set a value for the excess gas node (Node 4), instead allowing it to be informed by the other nodes in the network. Seismicity was recorded prior to the eruption and on the 13th of February this would have been for duration of 11 days (Node 7 is <100 days). At this time there would have been no ash plumes observed, so we can set the value for Node 8 as <70 days. There were no explosions that fit the criteria of node 6, therefore we set it as 'No'. It is unclear whether Nodes 9 or 10 were observed so we not assigned a state to these nodes. With this evidence, we show that on the 13th of February at mid-day the probability of a plinian eruption would have been approximately $10\% \pm 2$. This value is slightly higher than that calculated for the global distribution.

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Figure 3.15: Example of the completed network for the highest probability outcome of the Kelut 2014 eruption. The values in the nodes are calculated by GeNIe based on the conditional probability tables that define them.

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3.5.5 Retrospective analysis of the probability of plinian eruption

Finally we calculate the plinian probability of all the events in our database. When data are not available to fully inform nodes both the highest and lowest probability are calculated (i.e. the value of the nodes are set at either 'High / 'Yes or 'Low / 'No and then different probabilities are calculated). We find that the probability of the larger eruptions ($>10\text{km}^3$) in the database becoming plinian reaches $\geq 30\%$, with the lowest probability being 25%. These probabilities are 3-4 times greater than that suggested by an assessment of the global distribution of eruptions. Probabilities estimated retrospectively for smaller events (down to sub-plinian) show a decreasing trend of eruption probability until they become steady for eruptions of around 0.1km^3 DRE where the probabilities of a $\text{VEI} \geq 4$ eruption fall between 5% and 9%. This is the probability we would expect from the global distribution, suggesting that for these eruptions this network is not providing any further predictive value.

Figure 3.16: Probability that eruptions in our database would be plinian in size. Black crosses represent the average value for sub-plinian eruptions, red crosses the average value for the plinian eruptions. The bars represent the maximum and minimum value of the probability, given that not all data is known to inform all nodes. The shaded area represents the a priori probability from the global distribution.

This distribution suggests that this network is able to correctly identify volcanoes that are more likely to have a plinian eruption. However, all of the volcanoes in this study have been evaluated retrospectively not whilst a crisis is occurring. It is possible to evaluate these probabilities ahead of time, even though it is not possible to know what some of the node values are. If the volcanoes

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have fairly robust eruption histories then nodes 1, 2 and 3 (all of which are based on the eruption history) could be assigned values prior to an eruption. Nodes 5 - 10 are all based on observable unrest factors so could also be set to specific values. It is not possible to assign a value to node 4, which accounts for whether there are excess volatiles associated with the eruption. Instead the user can either chose to leave the state free, therefore allowing the network to assign a state based on the conditional probability table associated with it. This will produce a probability that is most likely given the other data. It is also possible, however, to artificially impose a value on node 4 (either 'Yes or 'No). This would generate the highest and lowest possible probability of a plinian eruption, given the rest of the observations. Either of these methods could be used to estimate the possible probability of an eruption becoming plinian ahead of time. This second method was used to estimate the upper and lower bounds of the probabilities in Figure 3.16.

Some nodes in this network are more dominant then others. It is possible to perform a 'Sensitivity of Findings assessment for the network to understand the main controls on increasing the accuracy of evaluating the probability of a plinian eruption. This assessment essentially calculates how knowing the value of one node affects the probabilities at all other nodes. These results are then ranked from highest to lowest. In this network the nodes are ranked as follows; Excess Volatiles, Eruptible Magma, Unrest, Repose Time, Shallow LP seismicity, Seismicity indicative of conduit connection, Frequency and intensity of explosive events, Plinian Ratio, Duration of phreatic / phreatomagmatic unrest and Duration of seismic unrest. This suggests that knowing the state of the excess volatiles node will be the greatest control on the accuracy of the plinian probability calculation. Whilst it is not possible to know these values ahead of an eruption, aiming to improve the conditional probability tables that control these nodes is possible. It was difficult to gather data for these categories in our database, as the analysis has simply not been done in many cases. A concerted effort to evaluate sulfur content of matrix glass and melt inclusions for older eruptions and to tie more eruptions SO₂ release would allow for more accurate conditional probability tables here. We note that the accuracy of the entire network can be increased with more data and would expect that a network such as this should be continually revised as data becomes available. For nodes that can only be assessed retrospectively (i.e. 'Excess' Gas node), to apply this network to calculate the probability of a future event it would be necessary to evaluate the network twice, once with and once without 'Excess' Gas. This would give a maximum and minimum probability that could be used as a guide.

3.5.6 The effect of uncertainty on eruption forecasting

There can be high levels of uncertainty in data sets (Marzocchi and Bebbington, 2012). In this study particularly, where data have been taken from multiple sources published over a large time period, it is possible that uncertainties represent; under-reporting of eruptions, changes in analytic methods and an increases in the level of volcanic monitoring and understanding. For this

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reason when designing the BBN we have attempted not to over-constrain nodes. Here we discuss some of the most important sources of uncertainty in our data and show what the potential effects this could have.

Dense Rock Equivalent (DRE)

DRE (or Dense Rock Equivalent) relates to the volume of unvesiculated erupted magma. To assess this it is necessary to estimate the volume of tephra produced by an eruption (Pyle, 2000). This value is dependant on the preservation of poorly consolidated tephra deposits and as such there is some uncertainty associated with these values (Pyle, 2000; Crosweller et al., 2012). After deposition these deposits are subjected to erosional processes that may remove or thin them, causing anomalously thick deposits to form in topographic lows (Engwell et al., 2013). As a tephra volume estimates are derived from isopach maps this can introduce errors into the calculation of DREs. These errors are even more pronounced when deposits are carried off an island in the course of an eruption. The other source of error in estimating the DRE is the understanding the bulk density of the tephra. The densities of freshly emplaced tephra deposits can vary by a factor of 3 (e.g. from freshly fallen ash to welded ignimbrites) (Mason et al., 2004). To assess this completely it is necessary to sample all phases of the eruption and perhaps apply separate bulk densities for different phases. It has been suggested that volume estimates typically contain uncertainties of 10 - 15% (Bebbington, 2009), and it is likely that this is an effect on our data. Where they have been reported we have recorded the errors on DRE estimates (i.e. Tambora: 41 ± 4 (Kandlbauer and Sparks, 2014)), but in many cases this information is not available. As this is the case throughout this study we use the average value of the DRE's seen in Table 3.2.

Incomplete eruption histories

In this study we have studied the eruption histories of volcanoes to assess the potential for large explosive eruptions. It is well known that this information is often inhomogeneous, affected by incomplete observations that becomes more pronounced as we move further back in time (Bebbington, 2009). As several of the criteria we are concerned with in this study are dependent on a clear eruption history it is important to gauge the effect an incomplete eruption history may have on our results.

In the last 200 years there has been a clear increase globally in the number of eruptions recorded. This is related to the increase in global population and therefore the more effective recording of events (Bebbington, 2009; Siebert et al., 2010; Jenkins et al., 2012). Other studies have attempted to assess the completeness of eruption records on both a regional and global scale (Jenkins et al., 2012; Whelley et al., 2015). Here we use findings from Whelley et al. (2015), who analyzed apparent eruption frequencies and found that there was an exponential decay through time of

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eruptions of different sizes. They suggested that the incidence VEI 2 and 3 eruptions are well captured over the last 110 years, whereas for VEI 4 and 5 eruptions this is true for the last 210 and 1000 years respectively.

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Figure 3.17: We manipulate the eruption history for Mount Pinatubo to show the potential error of in calculating the Plinian Ratio. We add eruptions ($VEI \geq 2$ and $VEI \geq 4$) to the catalogue to demonstrate how this would affect the final outcome. The black star shows the Plinian Ratio that we record based on the eruption history reported by Siebert et al. (2010). A) The red diamonds show the effect of adding one VEI 2 eruption until the original value is doubled. The blue diamonds represent the same increase in $VEI \geq 4$ eruptions, the green diamonds show the affect of increasing both at the same time. B) Shows the affect of adding 1 VEI 5 : 10 VEI 2 eruptions to the record.

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Figure 3.17 shows the potential affect of under-reporting of eruptions at Mount Pinatubo. At this volcano there have been a total of 6 $VEI \geq 2$ and 5 $VEI \geq 4$ eruptions recorded over the last 9474 years (Siebert et al., 2010). This gives a Plinian Ratio of 0.83. If we vary the number of eruptions in this catalogue to reflect eruptions that could have been potentially missed we see the affect on the final result. In the BBN (Figure3.13) we have set a threshold value of 0.5 for an eruption to be considered as ‘Due’. In figure 3.17a we show that if an increasing number of $VEI \geq 4$ eruptions were under-reported then this Plinian Ratio would increase, meaning that the value of 0.83 could be considered a minimum for this eruption. If however, we assumed that the number of $VEI \geq 4$ eruptions was correctly reported and it is the smaller eruptions that have been under-reported (very possible given the higher potential for erosion of smaller deposits), then we see that the Plinian ratio drops. If we double the number of $VEI \geq 2$ eruptions, then we see a decrease in the Plinian Ratio to 0.42. This would remove Pinatubo from the High’ category in the BBN and place it in the moderate category this does not, however, change the plinian probability calculated by the BBN. It remains 16

Whelley et al. (2015) showed that a VEI 2 eruption may be lost from the record after 110 years, whereas a VEI 5 is more likely to be lost only after 1000 years. Using this relationship as a ratio, we assume that for every 1 VEI 5 eruption that has been lost from the record, 10 VEI 2 eruptions are also under-reported. Here we see that by doubling the number of $VEI \geq 4$ eruptions in the record (and also increasing therefore the number of $VEI \geq 2$) it is possible to decrease the Plinian Ratio from 0.83 to 0.17 (figure 3.17b). This would mean that the state for this node on our BBN would decrease from ‘High’ to ‘Low’. This, however, would still not change the final plinian probability.

Whilst it is not possible to know the full impact of under-reporting on this network we have tried to mitigate its effect. The Plinian Ratio, for example, has a much lower affect on the final probability then other criteria (such as the Excess Gas node) and this is why varying it doesn’t impact the final outcome for Pinatubo. It is important to note however, that this affect would be more extreme at volcanoes that had a low number of eruptions reported as missing even on from the record would have a large effect on the final statistics. This is one of the uncertainties we would attempt to better quantify through an expert elicitation process.

‘Excess Gas’ values

To calculate the ‘Excess Gas’ in an eruption it is necessary to have a value for; S(Melt inclusion), S(Matrix Glass), Crystallinity, DRE, Magma density and the amount of SO_2 released during the eruption. These values have been taken from many different sources (for references please see Table3.3), which of course makes the data inhomogeneous. Where possible we have reported the uncertainties for these data, as provided by the original study. Figure3.18 shows the errors that may be present in the ‘Excess’ SO_2 values.

We have shown that ‘Excess’ Gas is one of the controlling nodes on the BBN, however, we do

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Figure 3.18: Error bars show the upper and lower bounds on the ‘Excess’ SO₂ value, given the upper and lower values for the S(Melt inclusion), S(Matrix Glass), Crystallinity and DRE of the eruption.

not attempt to quantify how much ‘Excess’ SO₂ is likely to increase the probability of a plinian eruption. This is because it is difficult to assess how much the uncertainties on this data affect the final amount. Instead we suggest that the presence of an ‘Excess’ Gas phase is in itself enough to increase the probability of a plinian eruption. Figure 3.18 shows that only one of the plinian eruptions (Tambora) has error bars large enough to potentially mean that there was no excess gas phase present in this eruption. It is also possible that given these uncertainties that two of the sub-plinian eruptions wouldn’t have had ‘Excess’ Gas phases. Given this the uncertainties would not strongly alter the values used in the conditional probability tables for this node.

All of the data used in this study have been harvested from the literature and as such contain uncertainties that are difficult to quantify. In designed the BBN we have attempted to set thresholds that reflect the relationships we have described here, without overly interpreting the data and therefore disregarding these uncertainties. We aim to quantify these uncertainties more fully by conducting an expert elicitation, which could allow us to re-define some of the states in the current BBN.

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3.6 Conclusions

In looking for key controls on plinian eruptions we were hindered by the small sample size, the lack of instrumentally monitored data and an inability to sample or observe processes that are occurring within the volcano whilst the reservoir is developing. However, using our retrospective analysis of plinian eruptions we have isolated several parameters that appear to increase the probability of the next eruption at a volcano being plinian. Most important is the accumulation of a large volume of volatiles and thus readily, quickly eruptible melt. This allows for the generation of the large sustained eruptive columns, with high mass eruption rates that define plinian eruptions. In our study all the plinian eruptions for which we were able to calculate their SO_2 gas budget had an excess volatile phase in the reservoir at depth. In 60% of cases this volume, was at least double that we would have predicted by petrologic analysis. Whilst the data set is small, we think that this is a significant finding. It also suggests that mineral models that simulate eruptive processes should try to incorporate the role of excess gas in promoting explosive eruptions. As long known, it is possible to assess the potential of a plinian eruption by quantifying the eruptive history of a volcano. The probability of a plinian eruption is higher at volcanoes that have high (≥ 0.7) Repose Ratios and high Plinian Ratio (≥ 0.5 , 1 in 2 eruptions are plinian).

In general we found poor correlation between physical parameters that are considered critical such as viscosity, crystallinity and water content, and plinian eruptions. There is, however, evidence that plinian eruptions can be promoted by magma mingling. Specifically we find that larger compositional and temperature gradients between resident and injected magmas are linked to larger eruptions. Finally, we suggest unrest phenomena that may increase the potential for a plinian eruption; these are dominantly based on data from the only monitored plinian eruption in our sample, Pinatubo 1991.

To synthesize and quantify the implications of our database we develop a Bayesian Belief Network and performed a retrospective analysis of the probabilities of plinian eruptions. We show that the largest eruptions ($\geq 10\text{km}^3$ DRE) had probabilities of up to 32%, 4 times greater than the *a priori* probability of a $\text{VEI} \geq 4$ eruption that can be calculated from the global eruption catalogue. We also show that the sub-plinian eruptions in our study had probabilities of c. 9%, which is close to what the *a priori* probability would suggest. Further characterization of previous events and any new potential eruptions will increase the robustness of the conditional probabilities of the nodes and thus lead to the more realistic assessment of the probability of a plinian eruption.

Quantifying ash fall and pyroclastic flow hazard at Gede volcano, Indonesia

Gede volcano, Java, has erupted more than 26 times in the Holocene, with at least four moderate size explosive eruptions. Due to its proximity to Jakarta and Bandung it poses one of the most significant volcanic hazards in Indonesia. Between c.0.2 M to 38 M people live within a radius of 10 to 100 km of Gede's crater. Here we evaluate two of the most severe hazards from this volcano, ash fall and pyroclastic flows, by using a combination of numerical models and basic geological knowledge of past history of the volcano. We found that for eruptions VEI 3-5 there is $\geq 50\%$ probability of $> 1\text{mm}$ of ash affecting at least 5,000 and up to 200,000 people. Pyroclastic flows generated by Gede have the potential of travelling up to c.15km, directly into heavily populated areas. We also present an event tree that summarizes this information and which should be useful for preparation and raising awareness for the volcanic hazards related to this volcano

Author Contributions: Annie Winson designed the project, ran TITAN2D simulations, interpreted ASHFALL data, devised the Bayesian Event Tree and wrote the manuscript. Christina Widiwijayanti assisted with the TITAN2D simulations; Susanna Jenkins ran all ASHFALL simulations. Fidel Costa and Chris Newhall both provided insight into the geological history of Gede and advised on the construction of the BET. FC, CW, and CN all reviewed the manuscript.

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4.1 Introduction

The preparedness for the impact of volcano hazards on people and infrastructures critically depends on a good understanding of the past volcanic history and any potential unrest data. This is perhaps even more important for active volcanoes that have been dormant for a significant period of time. Any new reactivation may take everybody by surprise, for example, the recent reawakening of Sinabung volcano, in Sumatra, first in 2010 and then in 2013 occurred after >500 years of dormancy. After such a long period of dormancy there was very limited knowledge on its past history, virtually no instrumental data on seismicity, and the population and administration were accustomed to treating Sinabung as a mountain rather than as an active volcano. Thus, when the volcano reawakened there were many decisions that had to be taken with a very limited knowledge and in a highly uncertain context.

Gede volcano in Java (Figure 4.1) shares some similarities with Sinabung: it is heavily populated on its flanks, only 70km south of Jakarta, capital of Indonesia, and 60km to the west of Bandung. The eruptive history, geology and unrest of Gede were only approximately known until a recent collaborative project between EOS and CVGHM started in 2012 (e.g. Belousov et al., 2015; Krimer, 2015). The last recorded eruption at Gede was phreatic and occurred on the 13th of March 1957, and after this it fell into a largely quiet phase. Prior to this, the largest recorded eruption was in 1747, but it is simply recorded as ‘the most severe’. There are no historical records that pre-date this. In 1990, however, the first of 11 seismic swarms was recorded, which have continued for c.25 years. These have been of varying intensity and duration (a week to a month) and mainly comprise volcano tectonic events (Hidayat et al., 2012). These swarms suggest that magma may be intruding below the edifice of Gede, although the area is also quite active tectonically. There is a NE trending fault passing between Gede edifice and the neighboring Pangrango volcano (see geological map of Gede: CVGHM, 2013).

In this chapter we utilise the results of the recent geological and petrological studies and the unrest data from the CVGHM-EOS project to evaluate the main volcanic hazards that can be expected from Gede. Unrest data are given to describe the background of recent activity at Gede, and hazard analysis will be based on eruption history, stratigraphy and petrology. We evaluate the extent of ash fall and pyroclastic flows for a variety of scenarios. Volcanic ash has the most wide reaching impact on the communities that surround a volcano (Jenkins et al., 2012; Wilson et al., 2012). Thick ash falls (over 10cm of wet ash) can cause the onset of structural damage to buildings (Ewert and Harpel, 2004), but other critical infrastructure services can also be disrupted. Fine ash, for example, has been known to disrupt electricity networks (Wilson et al., 2012). Vehicles and air conditioning units are also susceptible to the effects of ash, as particles clog filters and abrade moving parts. Ash may also cause contamination to water supplies, either because it physically blocks reservoirs or else because the chemistry of the salts adhering to the ash particles can cause acidification of the water body (Wilson et al., 2010, 2012). The contamination of water sources has further implications for the stability of agriculture in the

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Figure 4.1: Location map of Gede volcano in the Sunda arc and the island of Java. Black triangles show location of active volcanoes according to GVP. Locations of Jakarta and Bandung are shown relative to Gede.

region around the volcano. Aviation can be particularly sensitive to the presence of volcanic ash (Prata and Tupper, 2009; Alexander, 2013).

Although lahars are also an important component of volcano-related hazards of Gede they will be evaluated in a future work. Moreover, despite the existence of a significant monitoring network including seismometers, GPS, and tiltmeters at Gede, there are not enough data to be used to construct an event tree that also incorporates such data. Notwithstanding, we hope that the results can be used for improving preparedness and making plans in anticipation on a future eruption of Gede, which even if small, would have major consequences for the large population and infrastructure that currently exist around the volcano.

The current hazard map available for Gede (Figure 4.2B), is based on a geological map that was produced by CVGHM in 1992 (Appendix C C.1). We base this study on the updated geologic descriptions from Belousov et al. (2015), these constrain the input parameters for the models we use and the generation of our Bayesian Event Tree.

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Figure 4.2: A) Map of the Gede / Pangrango area. The towns of Cibodas and Sukamahi are highlighted, which are important for the hazard assessment. Also noted is the location of the 'Istana Cipanas', the presidential palace. B) Shows the CVGHM hazard map from 2006. Official hazard zones III, II and I are shown, along with areas most likely to be affected by pyroclastic flows and lahars.

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4.2 Geological history and setting of Gede volcano

Gede is one of the 40 active volcanoes on Java, and is part of the Sunda volcanic arc (Figure 4.1). The arc is the result of the subduction of the Indo - Australian plate beneath the Eurasian plate at a rate of c.6-7cm yr⁻¹ (Demets et al., 1990; Handley et al., 2014). This is an area of high rates of volcanism (Handley et al., 2010). Gede rises c.2800m asl and the edifice volume is on the order of 100km³ (Belousov et al., 2015). Volcanism at Gede began in the mid - late Pleistocene with the construction of the stratovolcano 'Gumuruh'. This edifice makes up 95% of the current volume of Gede. This period was dominated by effusive eruptions of thick, blocky lava flows. The youngest emplaced deposits were formed > 43,500 yr BP. Gede then apparently entered a 30,000 - year period of repose (Belousov et al., 2015). This is a longer repose time than would usually be expected at subduction zone volcanoes. For this reason it is unclear whether the volcano was inactive or if the deposits can simply not be found today.

In the Holocene, Gede entered a period of renewed eruptive activity that was dominantly explosive. There are four major pyroclastic deposits that were formed between 10,000 and 1,000 years BP and these are one of the foci of our study.

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Figure 4.3: A) Geological map of the Gede / Pangrango area. This map is adapted from Belousov et al. (2015) and shows the units at Gede that are younger than 10,000 years old. These are the units we have referred to in this hazard evaluation.

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This study is concerned with four key deposits of Gede, described by Belousov et al. (2015);

1. The oldest Holocene deposits, dated by C_{14} , were formed 10,230-9,900 yrs BP. This pyroclastic flow deposit is composed of black scoriaceous basaltic andesite flows that are 25m thick at their maximum (Table 4.1). This deposit is described as massive, although there are four distinguishable flow sub units. They travelled 8 km from the crater, have a volume of $c.0.1\text{km}^3$. It is inferred that they were produced by an incandescent, gas rich and highly channelized pyroclastic flow which was produced by VEI 3-4 eruption (Belousov et al., 2015). Belousov et al. (2015) infer that this deposit was formed by a relatively explosive, voluminous and purely magmatic eruption, which created a crater c.900m wide. This sequence is informally referred to as 'The Black PF (these deposits can be seen in blue in Figure 4.3). This deposit is similar to those that were produced by the eruption of Mt Mayon in the Philippines in 1984 (Chris Newhall, personal communication).
2. These deposits were followed by a sequence of two pyroclastic flow units that are composed of ash and lapilli and are each 45m thick. This formation of this deposit is dated between 4050-3990 yrs BP. This deposit is characterized by banded scoria and frequent lithic blocks. It is poorly sorted and contains clastic material made up of ash and lapilli. They travelled a minimum 8km from the NNE of the summit of Gede and deposits are smaller than the previous flows, with a volume of 0.03km^3 . It is inferred that these deposits were produced by highly concentrated, valley channelised flows that were gas poor. Belousov et al. (2015) suggest that these were produced by a VEI 2-3 phreatomagmatic eruption (potentially vulcanian). Belousov et al. (2015) suggest that the crater that was formed by the earlier eruption was filled in this wet environment, forming a crater lake. The deposits formed by this eruption therefore are phreatomagmatic in character, with relatively low depositional temperatures and a high proportion of non-juvenile clasts. This deposit is informally named 'Banded Scoria PFs (these deposits can be seen in pink in Figure 4.3). The banded scoria in this deposit suggests a mingling between two magmas prior to eruption (andesite and basaltic andesite), which then was erupted into a wet environment. This type of eruption scenario is loosely comparable to the Taal 1965 eruption, although the Gede eruption would have been less explosive (Taal 1965 was a VEI 4 eruption) (Moore et al., 1966). It could also be compared to the Nevado del Ruiz eruption where mixed magma erupted through the volcanoes snow cap (Gourgaud and Thouret, 1990).
3. The following series of pyroclastic deposits were formed c.1200 years BP. This sequence of 4 flows consists of poorly sorted ash and lapilli deposits, in flow units 3-7m thick and with a volume of 0.15km^3 . They travelled in the same direction as the previous flows but reached a minimum of 12km from the summit. These flows were moderately hot, gas poor and highly concentrated when they are deposited in an eruption of approximately

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VEI 3. These deposits are noted to contain a significant proportion of non-juvenile, heterolithologic, dense angular lava clasts (Belousov et al., 2015). There is evidence that this eruption was also phreatomagmatic, potentially having interacted with the crater lake. This deposit is referred to as 'Lithic rich PFs (these deposits can be seen in green in Figure 4.3). We suggest that this eruption was comparable to the Merapi 1994 eruption (Abdurachman et al., 2000).

4. The deposits of the c.1000 yr BP eruption are purely magmatic in character, which supports the draining of any crater lake between 1200 BP and 1000BP. This pyroclastic flow is up to 12m thick and covers an area of 3km² and has a volume of 0.01km³. It is made up of moderately sorted lapilli and scoriaceous bombs. Most of the clasts are composed of juvenile black to dark gray, moderately vesicular basaltic andesite to andesite (Belousov et al., 2015). It is proposed that this deposit was formed by small incandescent avalanches erupted during a VEI 2-3 eruption at Gede (these deposits can be seen in yellow in Figure 4.3). The character of these deposits is similar to those created by the Merapi 2006 eruption, although from a single pulse rather than a series of events (Charbonnier and Gertisser, 2008).

None of these eruptions are correlated with any voluminous, coarse-grained deposits of tephra. This suggests that volcanic plumes from these eruptions were not buoyant. The one exception to this is a deposit on the flank of neighboring Pangrango volcano, which shows evidence of being deposited by inflated pyroclastic density currents during a Plinian eruptive phase. The ages of these deposits suggest that they are linked to the early stages of the 1200 BP eruption. These deposits appear to be the only evidence that there might have been an eruption at Gede having a plinian phase, although it is unclear why the deposits have not been found elsewhere.

It is important to note that all volumes quoted for these deposits were estimated, rather than calculated. This was due to the scarcity of outcrops in this region (A.Belousov, personal communication). As such it has also not been possible to estimate the volume that may have been lost due to erosion, which in this climate may have been significant. It is not possible to estimate the uncertainty associated with these volumes, it may be that they represent a minimum value for the eruption products.

The first recorded eruption after the 1000 yr BP deposit began in 1747. Since then there have been 23 eruptions from Gede, mostly small (Table 4.1). The volcano has not erupted since 1957.

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Table 4.1: Eruptive history of Gede volcano, compiled from Smithsonian Global volcanism program and Natuurk Tijdschrift von Nederlandsche Indie, and Belousov et al. (2015). 1909 present: Padang (1951), Taverne (1926) and Badan Geologi (2011).

Year	VEI	Description of activity
Pleistocene		Effusive volcanism, responsible for building the main stratocone 'Gumuruh'
>30, 000 YEARS OF REPOSE		
10,000 BP	3 - 4	Explosive, voluminous and purely magmatic eruption.
4,000 BP	2 - 3	Eruption was phreatomagmatic - suggesting the presence of a crater lake at this time
1,200 BP	3	Eruption was phreatomagmatic - suggesting the presence of a crater lake at this time
1100 - 1000 BP		Gravitational collapse of the crater wall.
1000 BP	2 - 3	Dry magmatic eruption
Repose		
1747	3	Most severe eruption
1761	2	
1832	3	29 th August 11:00 - 16:00 Ash-rain fall and 'thick smoke' observed
1840	3	12 th of November 03:00: Red flames rose from the crater amid rubblings and quakes. Heavy smoke continually rose from the crater and crashes heard, noise like 'a heavy cannonade'. The eruption went in a NW direction 'which moved down with such force that colossal trees were uprooted'. The flows were directed through ravines: Tji Kundul and Tji Walen. "if the eruption should, however, exceed all the previous ones in severity, then the regions between the ravines mentioned live in the danger zone"
		1 st December 05:00 - 05:30 Explosions
		2 nd December 08:30 Explosions
		3 rd December 18:00 Explosions
		11 th December 14:30 Eruption recorded to have been "more important than any of the previous ones"
1843	2	Eruption on the 28 th of July
1845	2	23 rd of July 11:30 - 15:00 Black smoke column arose out of Gede (maybe some tremors)
1847	2	17 th - 18 th of October: Light earthquake felt in the late morning. Eruption started in the night
1848	2	Eruption on the 8 th of May

continued ...

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Table 3.1, continued

Year	VEI	Description of activity
1852	2	28 th May: Reports from people who visited the crater on the 24 th indicate it was exhibiting 'no unusual manifestations'. After the eruption began is lasted 5 - 10 minutes. One report stated that 'a large amount of rocks from 2 - 12 feet in diameter and ash have been thrown out (of the volcano) during the eruption of Gede'
1853	3	14 th March 07:00 - 09:00 Gede ejected large columns of smoke and ash
1866	2	Eruption on the 18 th of September
1870	2	Eruption in August
		19 th of December 22:00 2 or 3 'booming crashes' were heard
1886		Underground rumbling almost the whole month of January. On the 29 th of January Gede was showing heightened activity by ejecting smoke columns.
	2	Eruption on the 10 th of June
		16 th of August 06:30 Ash fall noted in the surrounding areas.
		23 rd of August Decrease in flow of rivers from Gede was noted
1887		26 th of January Rumbling accompanied by expulsion of large smoke columns
		14 th of February at 14:08 An earthquake that lasted 5 - 6 seconds was noted. Gede was also rumbling.
		21 st of April at 07:30 a second earthquake occurred. Gede was also rumbling
		9 th of July Steam was escaping the conduit
	2	22 nd of October Eruption of Gede. At 10:45 Sounds 'like thunder' were followed by a rising smoke column (7000-8000 feet)
		Sounds like thunder, followed by a rising smoke column (7000-8000 feet)
1888	2	Eruption of Gede
		29 th of January 'Rumblings' were heard
		22 nd - 23 rd of March 'Rumbling' was heard
		28 th of May 05:30 - 07:30 Rumbling became continuous
1889	1	8 th May An earthquake occurred, Gede was strongly smoking, rumbling was heard

continued ...

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Table 3.1, continued

Year	VEI	Description of activity
1891	2	Eruption of Gede
		1 st of May 20:00 - Noise like “distant cannon shots” during the night. Smoke was illuminating the crater, the next morning a dense, grey cloud was seen over the summit. One week before this ‘smoke’ was constantly rising above the crater
		13 th and 14 th of May - More explosions, although smaller. “dull, weak sounds” accompanied by ash fall
1899		22 nd of November - A mass of smoke has been coming out of Gede’s crater
1900		7 th of January 16:10 - Earthquakes were recorded. Gede was expelling ‘smoke’ clouds.
		7 th of January 22:30 - 2 sharp, vertical shocks were recorded
		8 th of January at 15:00 - A sharp, vertical shock was recorded. Dark smoke rising above the crater
		15 th of January 01:30 - Gede was displaying a slight degree of heightened activity
1909	1	2 nd of May 17:00 - New fissures were reported, vapors escaping from them. Ash-rain fall (+30 mins). For 2 months before there were underground rumblings
1947	2	Between the 2 nd September 1947 and the 28 th of January 1948 - 14 eruptions of various strength were recorded
1948	2	15 th of November - Eruption of Gede
1956	2	28 th of April - Eruption of Gede
1957	2	13 th of March - Eruption of Gede

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The evidence from the Holocene suggests that Gede has the capacity for large, explosive eruptions but the historic events are limited to few, smaller magmatic eruptions interspersed with frequent small phreatic eruptions. Despite the lack of eruptive episodes since 1957, there is evidence of unrest at Gede. Seismic swarms have been recorded since 1990 and there have been 11 of these events since then. The largest swarm to date occurred in 2010 (Figure 4.4). This suggests that whilst Gede has not produced a large eruption in recent history, it is certainly not dormant. It is therefore important to understand the impacts of an eruption occurring with different scenarios of activity. Previous work by CVGHM lead to the publication of a hazard map in 2006, this analysis incorporated pyroclastic flow and ash fall hazard within a 7km radius of the summit (Figure 4.2). Now that detailed stratigraphic work has been completed and the volumes and ages of previous eruptions are better understood, we re-assess the hazard potential of Gede. We focus on the potential impact of pyroclastic flows and ash falls to evaluate if the 7km radius is appropriate. We complement the previous work with modeled results.

Figure 4.4: Cumulative seismic energy (erg) of Volcano Tectonic (VT) earthquakes recorded by Gunung Putri station (CVGHM), showing episodic seismic crisis observed from 2001-2014. The largest crisis, starting in December 2010 and ending in January 2011 is highlighted.

4.3 Population statistics around Gede volcano

We have calculated population statistics from ‘Gridded Population of the World data, obtained from the Centre for International Earth Science Information Network at Columbia University

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(for International Earth Science Information Network, 2015). These data are obtained with a proportional allocation gridding algorithm, which uses national and sub-national administrative units to assign estimates of the number of people per grid cell of 2.5 arc minutes (c.5km). The data are only from statistics of year 2000 and thus are minimum values for the population currently living in these areas. Investigation of the area around Gede volcano shows that there are at least c. 60 million people who live within 200km radius of the volcano (a further analysis of the population close to Gede is reported below).

Figure 4.5: Population numbers in South East Asia (a), Java (b) and around Gede. Circles around Gede volcano are for radii of 10, 20, 30 and 50km.

For hazards evaluation we choose radii of 200, 100, 50, 30, 20 and 10 km from the volcano, in line with previous studies (Small and Naumann, 2001; Ewert and Harpel, 2004; Ewert, 2007). These radii also broadly correlate to distances from the volcano that eruptive products can travel and pose specific hazards to local populations. For example, Newhall and Hoblitt (2002) show that a pyroclastic flow produced by a VEI 4-5 eruption has a low chance of travelling more than 30km distance from the vent (5%), but that at this distance tephra accumulation is likely to be more than 10cm (80% for a $VEI \geq 4$). In fact ash can affect infrastructure at substantial distances from the volcano (Small and Naumann, 2001). With this in mind we assess the potential population that could be exposed by the eruptive products of Gede at various distances.

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Table 4.2: Number of people who live within a; 10, 20, 30, 50, 100, 200km radius of Gede. Statistics taken from the ‘Gridded Population Data of the World and represent national administrative units. These figures are from the 2000 data set and therefore should be assumed to be minimums of the current population living around Gede.

Radius of contour (km)	Population
10	263,537
20	1,287,176
30	3,873,473
50	9,934,760
100	38,444,295
200	56,845,916

4.4 Approach to modeling and parameter values

4.4.1 Ash dispersal modeling

We calculated the probability of ash affecting the area around Gede by calculating the exceedance probabilities of 1mm, 10mm and 100mm thickness of ash on the ground (at a given location), from eruptions with VEIs 3 to 5. Ash dispersal modeling was carried out to determine the extent and potential thickness of ash fall from 3 different sized eruptions (VEI 3, VEI 4 and VEI 5). We used an ash advection diffusion model (ASHFALL; Hurst, 1994) to simulate these effects. The ash fall at a volcano is controlled by the eruption magnitude, character of ash particles and the vertical profile of wind speed and direction above the volcano at the time of eruption. The spatial distribution of ash needs to be calculated including the for given wind fields or else for averaged wind fields over a given period.

We have simulated 1000 eruptions of sizes equivalent to VEI 3, 4 and 5, and we use the following input parameters in ASHFALL software (Simulations were run at the University of Bristol by Susanna Jenkins, based on criteria provided from this work):

1. The Eruption volume: These are assumed by the VEI that we set (VEI 3, 4, 5). Random sampling is performed over a uniform distribution ranging between the upper and lower boundaries of the VEI.
2. Eruption Column height: The height of the eruption column is assumed to follow a relationship with eruption volume similar to that described by Carey and Sigurdsson (1989) for plinian eruptions. The production of higher columns will lead to ash particles remaining airborne for longer and therefore will lead to a larger dispersal in the direction of the dominant wind.
3. Ash settling velocities: These are extrapolated from particle size and density, which are modelled from a normal distribution. As with the eruption volume these sizes are

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uniformly sampled over a range of values (in this case between 2.5ϕ and 1ϕ , with an assumed density of 900 kg/m^3).

4. Wind conditions: In the model the wind conditions are compiled for a ten year period and then assess at heights equating to the mid-stratosphere, tropopause and the near surface (Figure 4.6). These are then randomly sampled to provide dominant wind directions at Gede. We assessed the seasonal wind patterns to explore if eruptions occurring at different times of year may produce different ash footprints. In this case we find that there is very little difference between the wind patterns for different seasons according to the National Centre for Environmental Protection (from 1997-2006) (NOAA and National Weather Service, 2014).

The model produces the exceedance probability for a grid of cells that are 1km^2 . This is the probability that the ash thickness in a given cell exceeds the thickness threshold given that an eruption of a specific size has taken place. We choose to calculate the probability of thickness exceeding 1mm, 10mm and 100mms.

These values were thought important for hazard assessment because of their impact on local infrastructure. For example 100mm of ash fall thickness may be enough to begin to collapse roofs, so it is necessary to know what area is prone to such accumulation. Finally we correlate the ash footprints generated by these eruptions with the population densities that we have calculated for the area. This allows us to estimate the number of people who could potentially be affected by the products of such eruptions from Gede.

4.4.2 Modeling pyroclastic flows

The stratigraphic history described by Belousov et al. (2015) shows that there are at least four pyroclastic flow units present to the NE of the summit crater of Gede. The summarized volumes and flow descriptions are (Table 4.3):

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Figure 4.6: Wind data used in ASHFALL modeling. Wind roses are a compilation for a ten year period for three different atmospheric levels (Mid-stratosphere, Tropopause and Near Surface). This figure was provided by Susanna Jenkins.

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Table 4.3: Main features of the four units described by Belousov et al. (2015). These information forms the basis of the following pyroclastic flow simulations.

Unit Name	Age (yrs BP)	Max thickness (m)	Run out distance from summit (km)	Volume of deposit (km)	VEI	Number of subunits
Black PF	10,230 - 9900	25	8	0.1	3 - 4	4
Banded Scoria PFs	4090 - 3910	10 (units: 4 - 5)	8	0.03	2 - 3	2
Lithic Rich PFs	1200	Each unit 3 - 7	12	0.15	3	4
Black lapili PF	1070 - 930	1 - 3		0.01	2 - 3	1

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As pyroclastic flows are highly destructive, the identification of such deposits calls for greater understanding on the controls of these phenomena. We use the details from past eruptions to simulate granular flows with TITAN2D software (Geophysical Mass Flow Group, 2007), a numerical mass flow model that can be used to simulate flows over digital elevation model (DEMs) of natural terrain. Combining the effect of present terrain the depth-averaged simulation allows for assessment of the impact and run out distances of either partial or total collapses of lava domes and eruptions columns.

This method has been proved useful in simulating the impact of mass flows at Mt. Taranaki (Procter et al., 2010), Cerro Machin (Murcia et al., 2010), Soufriere Hills, Montserrat (Ogburn et al., 2008; Widiwijayanti et al., 2009), El Misti (Constantinescu et al., 2011), and Merapi (Charbonnier and Gertisser, 2009), among others. This approach can either use the historical pyroclastic deposit coverage and genesis to simulate the impact of similar flows or else can be run for a broad range of eruptive scenarios. This allows for the forecasting of possible future events and for the development of hazard maps. To be effective, hazard forecasts must incorporate likely values for factors that will control future events, such as: the source location of the eruption, geometry and size of the material that is collapsing, the coulomb friction and the source mechanism (i.e. is it produced by a total/sustained dome collapse, sequence of partial dome collapses, or a column collapse event?). We have used the range of possible parameters from Table 4.3 to assign values for criteria such as: specific pile geometry, volume, starting coordinates, initial speed, direction of primary movement and internal and basal friction angles. We discuss the values input parameters below.

At Gede we have limited information about recent eruptive history and products, we therefore use the recent study by Belousov et al. (2015) as a guide for the input parameters. In this analysis we focus on two flows; the ‘Lithic Rich PF’ and the ‘Black Lapilli PF’. These units are the largest and the smallest described by Belousov et al. (2015), their footprints are also well constrained (Figure 4.3). The descriptions of these outcrops in Belousov et al. (2015), summarized in section 4.2 suggest that they could have been produced by dome collapse events.

To perform mobility test we calculate the L/H (where L is the run out distance and H is the total height of fall) value of the ‘Lithic Rich PF’. At Soufriere Hills volcano, Montserrat, Calder et al. (1999) found that different kinds of pyroclastic density currents showed contrasting mobilities that were indicated by the total height of the fall (H) and the run out distance (L) or the inundated area (A) and the volume ejected (V). L/H values for dome collapse events at Montserrat were found to be 4.5 ± 0.5 . For the ‘Lithic Rich PF’ we calculate a L/H value of 4, based on the assumed run out distance of 12km and the height of Gede (c.3000m). From this we assume that it appropriate to simulate this flow as a dome collapse event in TITAN2D. It has not been possible to calculate the L/H value for the ‘Black Lapilli PF’ as Belousov et al. (2015) do not quote a maximum run out distance. Instead we calculate the A/V value from the area that the flows covered and then volume of the deposit estimated by Belousov et al. (2015), this gives a value of 300. Compared to the values calculated for Montserrat by Calder et al. (1999), this

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suggests that the flow type similar to a derived flow. For this reason we model this flow as both a dome collapse and a column collapse event, comparing the final outcomes. In this analysis we aim to re-create some of the past collapse events, to inform potential hazard planning for future events.

It should be noted here that we have selected TITAN2D to perform these simulations as there is already a capacity for using this program at CVGHM.

Source location and initial dome geometry

We have produced a DEM using GRASS (Geographic Resources Analysis Support System, U.S. Army Corps of Engineers CERL), with a 20m resolution to represent the terrain. TITAN2D allows for re-sampling of the computational cell size using grid adaptation, which reduces the processing time required to simulate these flows. The initial lava dome represented by a pile of material in a paraboloid shape with assigned dimensions, and allow to fail due to gravity from the volcanoes summit. Its dimensions, location, and coulomb friction angles will control the direction of this failure as well as the run out distance of the flows produced (Williams et al., 2008). We have established piles that are of varying dimensions, controlled by the volumes of the deposits that were noted in Table 4.3. As the distribution of these deposits is dominantly to the NE and is controlled by the horseshoe topography of both the summit caldera and the Gumuruh caldera, we release the pile source in this direction, to allow the granular material flow to be controlled by gravity and the natural terrain. We have placed the source location, at the centre of the crater.

Most simulations that we have produced have been to model pyroclastic flows are from a dome collapse-type of event. Belousov et al. (2015) noted that the deposits were highly channelized, valley-concentrated flows that were not associated with voluminous coarse-grained deposits of fallout tephra around the volcano, so we model these events as domes collapses and vary other parameters to fit the deposits. Initial starting velocities are therefore 0 m/s.

Basal and internal frictional angles

TITAN2D uses two frictional parameters to account for how the particles in the flow are likely to interact with each other (internal friction angle) as well as with the ground surface (basal friction angle). The internal friction corresponds to the natural slope of the free surface that would form if a cylindrical pile of granular material was placed on a flat plane and allowed to collapse under its own weight (Murcia et al., 2010). Studies by Sheridan et al. (2005) showed that simulations are not sensitive to changes in this parameter. For consistency we select an internal friction angle of 25° for all simulations, following values used in simulations at Soufriere Hills Volcano (Widiwijayanti et al., 2009).

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The basal friction is used to estimate the resistance of a flow sliding due to its interaction with its underlying surface Procter et al. (2010). It corresponds to the minimum slope that the inclined surface must obtain before the material placed on it begins to slide from its static position (Murcia et al., 2010). The advancement of the flow and the distribution of any deposits is greatly controlled by this factor.

We initially use the ratio between the fall height (H) and the horizontal runout (L) or Heim coefficient to determine the best value for the basal friction parameter (Stinton et al., 2004). From this calculation we achieve a value of 0.25 for the ‘Lithic Rich PF’. Stinton et al. (2004) and Sheridan et al. (2005) use basal frictions of $c.12^\circ$ for Heim coefficient values such as these. Initially, therefore, we assign a value of 12° for this parameter. It has, however, been argued that the Heim coefficient is physically meaningless as the vertical fall and the horizontal runout distances are strongly affected by topography (Stinton et al., 2004). We therefore tune our simulations by varying the basal friction values around the 12° , up to 18° and down to 6° . These values reflect a survey of the literature to assess basal friction values from other volcanoes such as; Colima, El Chichon, Machin, Merapi, Mount Taranaki and others (Sheridan et al., 2005; Macías et al., 2008; Charbonnier and Gertisser, 2009; Murcia et al., 2010; Procter et al., 2010; Sulpizio et al., 2010; Capra et al., 2011). We compare the footprints of the unit mapped by Belousov et al. (2015) with the footprints of the simulated flows with different basal frictions, to assess which is the closest match.

The lithic rich PF was released as four pulses (inferred from the 4 sub-units reported by Belousov et al. (2015)). If pyroclastic material is being released in a sequence of dome collapses then this will have an effect on the basal friction. Each consecutive flow will decrease its basal friction, deposits have no time to compact meaning that the material is loose and therefore moves more easily. For this reason we have also generated flow sequences where the same volume is released with varying basal frictions (each successive flow decreasing by 2°). This allows us to assess the impact of successive pyroclastic flows over the course of an extended eruption, as is currently occurring at Sinabung, for example. In addition to the decrease in basal friction due to the deposition of fresh material, the terrain has been filled and basal surface smoothened by the previous deposits, successive dome collapse also generate increasingly mobile deposits. Dome collapse makes the following edifice increasingly unstable, leaving less time for the magma to degas before it fails. This generates flows that are increasingly dilute, and will therefore advance further (as is currently occurring at Sinabung) (C. Widiwijayanti, personal communication). We simulate this by assigning a lower basal friction to the flows. Conversely it appears that the ‘Black Lapilli PF’ was deposited in one event and therefore we assess different basal frictions simply to assess goodness of fit with the mapped deposits.

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Column collapse parameters

As the mechanism that generated the deposition of the ‘Black Lapilli PF’ is unclear, we model this as a column collapse event as well as a dome collapse event. We use the same source location as the dome collapse simulations, and allow for 1 flux source. We set the mean flux as $40 \text{ kg/m}^2\text{-s}$, with an active source duration of 70 seconds and a fall height of 1000m. The initial pile dimensions were set as 40m / 40m. The basal and internal frictions were set at 12° and 25° respectively. These parameters were informed by work by Charbonnier (2011) at Merapi.

To decrease the computation time TITAN2D uses an adaptive grid, which preferentially refines the simulation according to where the most active areas of the flows are. This means that the tail of the flow will be more coarsely modelled than the edges and the flow front. All simulations run for 10 minutes and stop at the edges when the deposit reaches $<0.1\text{m}$ thick.

Input parameters for dome collapse simulations can be seen in Table 4.4.

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Table 4.4: Input parameters used for TITAN2D simulations. Each volume is simulated with basal friction given. A sequence of simulations may also be used to represent a larger volume deposit, for example the four 37.5 million m³ flows also represent the 150 million m³ deposit. Coordinates of initial source locations are given in UTM. Bold numbers show simulations discussed in section 4.5.2, the other simulations' parameters are derived from the Belousov et al. (2015) unit descriptions. These values are used the construction of the Bayesian Event Tree (Figure4.6.1).

Type	Location	Volume (million m³)	Pile		Friction	
			Parameter		Parameter	
			Height (m)	Radius (m)	i	b
Dome Collapse	Center of Crater (719098/ 9249393)	150	200	490/490	25	12
		100	200	400/400		12
		37.5	200	250/250		9
						18
						16
						14
						12
						10
						8
						6
		30	200	225/225		12
						9
						6
						18
		10	200	125/125		16
						14
						12
						10
						8
						6
		1	100	60/60		12
						9
						6

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4.5 Results

4.5.1 Exceedance probability of ash fall from Gede volcano

We assess the exceedance probability of ash falling to a thickness of 1mm, 10mm and 100mm in the region surrounding Gede, given a specific eruption size (Figure 4.7). Given that wind direction patterns for most of the year are the same, for all eruption scenarios there is a higher probability of ash fall to the south west of Gede. We find that for the smallest eruption scenario (VEI 3) the possibility of having an ash fall of 100mm is low. The ash fall exceedance probability is <30% at a distance of 5km from the crater, the footprint of which is generally towards the SW. The total potential impacted area is only c.2300 km², which covers an area between our 20 and 30km contours (Figure 4.5).

There are 1.4 million people who live in this c.2300 km² area that could potentially experience 100mms of ash fall due to a VEI 3 eruption. For most, however these probabilities would be <10%. The area that would be impacted by 1mm of ash fall from an eruption of this size is much greater (c.100,000 km²). It is important to note that even with a VEI 3 eruption there is a low possibility (<5%) of ash fall in both Jakarta and Bandung (Figure 4.8A).

Footprints of ash fall exceedance probabilities are much larger for VEI 4 eruptions. The potential area affected by 100mm of ash fall from an eruption of this size is more than 10 times that of the VEI 3 eruption (27,000 km²; Figure 4.9C). For an eruption of this size, there is a possibility of ash thickness being 100mm in the western areas of Bandung, although due to the prevailing wind direction these values are <5%. The area that could potentially be affected by 10mm of ash fall from a VEI 4 eruption includes both Jakarta and Bandung (10% and 20% respectively). There are potentially 49 million people in this area that could be affected by this amount of ash fall (Figure 4.9B). The area affected by 1mm is 5 times larger then that affected by a VEI 3 eruption (500,000 km²) and could impact 65 million people.

The largest eruptive scenario we assessed was for a VEI 5 eruption (Figure 4.10). We find that the potential impacted areas cover both Jakarta and Bandung, and that the probabilities of ash fall in these locations are higher than for the previous simulations. The probability of 100mm ash in Bandung is c.20%, and about 10% in Jakarta. We also note that there is a very high probability that the area to the south west of Gede being covered with at least 10mm of ash. We would expect a very large area to be affected by 1mm of ash fall with this size of eruption. Approximately 1 million km² and 87 million people could be affected by an eruption of this size.

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Figure 4.7: Comparison of ash fall exceedance probability footprints for different sized eruptions (VEI 3, 4 and 5). Edge of the ash fall polygons represents the furthest most reach of potential ash fall. Simulations are for integrated wind patterns between 1997-2006. A) 1mm ash fall exceedance probability, B) 10 mm ash fall exceedance probability, C) 100mm ash fall exceedance probability.

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Figure 4.8: VEI3 ash fall exceedance footprints showing southern tip of Sumatra and W Java in UTM coordinates. Warmer colours indicate higher probability. We show the contour for exceedance probability of 50%. These areas are too small to show in the 100mm and 10mm figures. Simulations are for integrated wind patterns between 1997-2006. Jakarta and Bandung are marked to the north and east of Gede. A) 1mm ash fall exceedance probability, B) 10mm ash fall exceedance probability, C) 100mm ash fall exceedance probability.

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Figure 4.9: VEI 4 ash fall exceedance footprints. Warmer colours indicate higher probability. We show the contour for exceedance probability of 50%. Simulations are for integrated wind patterns between 1997-2006. Jakarta and Bandung are marked to the north and east of Gede. A) 1mm ash fall exceedance probability, B) 10mm ash fall exceedance probability, C) 100mm ash fall exceedance probability.

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Figure 4.10: VEI5 ash fall exceedance footprints. Warmer colours indicate higher probability. We show the contour for exceedance probability of 50%. Simulations are for integrated wind patterns between 1997-2006. Jakarta and Bandung are marked to the north and east of Gede. A) 1mm ash fall exceedance probability, B) 10mm ash fall exceedance probability, C) 100mm ash fall exceedance probability.

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4.5.2 Calculated run-out distances of pyroclastic flows

We used Titan 2D to simulate the potential extent of pyroclastic flows from a dome collapse at Gede volcano, using the volumes of deposits described from stratigraphic study (Belousov et al., 2015). We first aimed our simulations at reproducing the extent of the volcanic deposits from the known volumes, thickness, location, and stratigraphy (Figure 4.3). We explored the role of the initial conditions, the basal friction, and whether the deposit was emplaced in one or more events.

In general we found that the largest effect on these flows is the basal friction, which is also related with whether the deposit was emplaced in one or more events, since multiple events emplaced during short periods of time have much lower basal frictions due to lower surface roughness. We concentrated on the flows created by the largest, the ‘Lithic Rich PF, (150 million m³) and the smallest, the ‘Black Lapilli PF (10 million m³).

From our assessment of the L/H values and the character of the deposits we suggest that the ‘Lithic Rich PF (erupted about 1.2 kyr ago), was most likely created during a series of dome collapses. We know that this unit is made up of 4 flow deposits (Table 4.3), so we simulate four collapses of equal volume (37.5 million m³). The Heim coefficient suggests that a reasonable basal friction for these flows is 12°, we want to simulate the effect of decreasing friction due to repeated emplacement of flows, so we vary the basal friction and assess the impact on the run out distances. As there have been 4 flows, we are looking for a set of four simulations that will have a maximum run out distance similar to the footprint mapped by Belousov et al. (2015). We produce a series of simulations whose basal frictions vary from 18°- 6° to account for different eruption scenarios (Figure 4.11). The flows that would be produced from this would reach a maximum run out distance of 16.5km and would range in maximum thicknesses from c.42 m to c.10 m (see Table 4.5).

Figure 4.11 shows the affect of varying basal friction on the final run out distance of the deposit. We show that the best fit to the actual deposit appears to be a flow with a basal friction value of 6°. In this case we suggest that a progression of four flows with decreasing basal friction values of 12° (the value suggested by the Heim co-efficient) to 6° (the value that best fits the physical deposits) may best represent the eruption process for the deposition of the ‘Lithic Rich PF’ (Figure 4.12). We note that whilst it has been possible to match the maximum run out distances of the deposits for the ‘Lithic Rich PF’, these deposits are laterally more extensive then those mapped by Belousov et al. (2015).

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Figure 4.11: Simulated pyroclastic flow run out distances for 37.5M³, for basal frictions of 18° and 6°. These simulate a succession of four dome collapses that lead to the emplacement of a deposit of 150 million m³. This simulation represents the potential maximum run out distances of the 'Lithic Rich PF'. Co-ordinates are displayed in UTM. A) Basal friction 18°, B) Basal friction 16°, C) Basal friction 14°, D) Basal friction 12°, E) Basal friction 10°, F) Basal friction 8°. Main figure shows the closest match to the deposit mapped by Belousov et al. (2015).

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Figure 4.12: Compiled simulated pyroclastic flow run out distances for 37.5M³ compared to the distribution of the known 'Lithic Rich PF', for basal frictions of 12° and 6° that simulate a succession of four dome collapses that lead to the emplacement of a deposit of 150 million m³. This simulation represents the potential maximum run out distances of the 'Lithic Rich PF'. Co-ordinates are displayed in UTM.

We have now assessed the potential impacted areas for the largest unit described by Belousov et al. (2015). Next we show the results of simulations of the smallest deposit, the 'Black Lapilli PF'. It has not been possible to calculate the Heim coefficient for these deposits, as we do not have a value for the maximum run out distances. It has been possible to calculate the A/V (described by Calder et al., 1999), which suggests that the character of this deposit was more like a derived flow. We have modelled this deposit as both a dome collapse event and as a column collapse, to compare the extent of the resulting deposits.

Figure 4.13 shows the results from the simulations of the dome collapse event. We have followed the basal friction parameters that we used for the simulations of the 'Lithic Rich PF'. We find that unlike the 'Lithic Rich PF' deposits, a basal friction of 6° is too low for this scenario, with the simulated deposits vastly exceeding those that were mapped by Belousov et al. (2015). Conversely the simulations using higher basal frictions (18°) produce distributions that are too small to match those that have been mapped. The closest fit that we can produce from these dome collapse simulations is with a basal friction of 12°. This still does not exactly match the mapped deposit.

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Figure 4.13: Simulated pyroclastic flow run out distances for 10M³ from a dome collapse event, for basal frictions of A) 6°, B) 12° and C) 18°. This simulation represents the potential impacted areas of the 'Black Lapilli PF', if it were deposited by a dome collapse event. Co-ordinates are displayed in UTM.

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As the dome collapse simulations do not match the observed deposits, we also compare the results from a column collapse event (Figure 4.14). Here we show that at a basal friction of 12° (the best fit basal friction from the dome collapse events), the extent of deposits generated is greater than the observed unit. We suggest that this means that for the smaller deposits we are not able to accurately describe the emplacement mechanism and therefore are not modeling it correctly. We do note that the maximum run out distances for both the dome collapse and column collapse events are approximately the same (c.7km), when the basal friction angle is 12° , the main difference being that the column collapse event has a greater lateral extent.

Figure 4.14: Simulated pyroclastic flow run out distances for 10M^3 from a column collapse event, for basal frictions of 12° . This simulation represents the potential impacted areas of the 'Black Lapilli PF', if it were deposited by a column collapse event. Co-ordinates are displayed in UTM.

In general, we see that the main control on the maximum run out distance of the flows is the coulomb basal friction. This is clearly seen in the difference of flows produced by the simulation of the failure of 37.5 million m^3 as a succession of dome collapses (Figure 4.11). We have completed simulations for the other two deposits described by Belousov et al. (2015) - the 'Black PF' and the 'Banded Scoria PF'. For the purposes of this work we have assumed that these events more closely match the deposition characteristics of the 'Lithic Rich PF'. This is because both of these events have significant volumes (0.1km^3 and 0.03km^3 respectively) and yet are not associated with any widespread fall deposits on the flanks of the volcano, which maybe expected if they were produced by an eruptive column. As such we have modeled them as dome collapse events. The results of these simulations can be seen in Table 4.5.

The simulations of both the ash fall and pyroclastic flows show that there is indeed a large hazard posed by an eruption of Gede volcano. The pyroclastic flows are perhaps the most acute hazard.

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Table 4.5: Results from the simulations of dome collapses at Gede volcano. Maximum thicknesses and maximum velocities are calculated by TITAN2D at the end of each simulation.

Type	Volume (million m ³)	Basal Friction	Max run out (km)	Max thickness (m)	Deposit Area (m ²)
Dome Collapse	150	12	9	33	2.59×10^7
	100	12	8.5	30	3.10×10^7
		9	12	20	3.45×10^7
	37.5 (150)	18	3	42	7.71×10^6
		16	4	37	8.58×10^6
		14	6	32	1.12×10^7
		12	9	26	1.47×10^7
		10	11	21	1.52×10^7
		8	13	14	2.02×10^7
		6	16.5	11	1.69×10^7
	30 (100)	12	8.5	24	1.37×10^7
		9	12	16	1.41×10^7
		6	16	11	1.60×10^7
	10 (30)	18	3	29	2.67×10^6
		16	4	27	4.03×10^6
		14	5	27.5	4.32×10^6
		12	7	21	5.81×10^6
		9	7	16	5.81×10^6
		6	16.5	12	1.31×10^7
	1	12	4	18	1.36×10^6
		9	7	17	2.56×10^6
		6	10	16	3.13×10^6
Column Collapse	10	12	7	24	7.62×10^6

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We find that the minimum run out distance we could expect from a dome collapse of similar proportions to past events at Gede is 7km; this distance decreases to 4km for a significantly smaller event (1 million m³). It is important to note at this point that the limits of the nearest town to the NE of Gede, Cibodas, are reached at 5km from the NE of the volcano. This town is also well known, as one of the presidential palaces is located here, at c.8km from the summit of the volcano, to the NE. Gede therefore poses a significant hazard to the local area and may even disrupt the old highway that runs from Jakarta to Bandung, through Cibodas. We note that none of our deposits reach more than 16.5 km distance from the volcano; this is due to a topographic ridge close to Sukamahi, which is c.15km from the volcano, directly in line with the path of the deposits. The only area where the Sukamahi ridge is breached is where there is a break in the ridgeline, forming a topographic low. Along the rest of the ridge the pyroclastic deposits extend laterally and spread out. This final topographic control suggests that pyroclastic flows would be unlikely to extend further than c.16km, regardless of the size of the dome collapse.

4.6 Discussion

From our simulations we show that the current 5km hazard zone shown by the CVGHM map may underestimate the area that would be threatened by a dome forming eruption of Gede. We show pyroclastic flows can be expected to have greater run out distances and be laterally broader than anticipated, especially to the NE and that ash fall is likely to heavily impact the SW flank of the volcano. We use the information gained from this study to inform the following event tree with the aim of assessing the probability of specific eruptive scenarios. We hope to represent here the probability of different scale events affecting local and regional populations.

4.6.1 An event tree for Gede volcano

We follow methodology described in Newhall and Hoblitt (2002) and Newhall and Pallister (2014). We describe the conditional probabilities of each node of the event tree and the evidence that we use to inform the probabilities at each step. An event tree is a graphical representation that branches from a prior event to more specific events, until the final outcomes are reached.

- **P (1): Given that a potentially active volcano is in repose, what is the probability that unrest will begin?** As there is a history of seismic swarm at Gede we know that there is a potential for unrest here, the probability for this node will be 1.
- **P (2|1): Given that unrest is occurring, what is the probability that it is the product of a magmatic intrusion?** We know that there have been 11 periods of seismic swarm at Gede since 1990. The frequency of these events suggests that magma is being injected into the edifice. However, Gede also lies astride one or more active faults, so some of this

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seismicity may be of tectonic origin. Since we don't know the volcanic contribution, we use a rather arbitrary probability here of c.0.6 (following Newhall and Hoblitt, 2002).

- **P(3|2): Given that there has been a magmatic injection, what is the probability of a magmatic eruption?** Ideally here we would use the eruptive history of Gede and make an estimate about the number of unrest periods that don't lead to an eruption. Unfortunately at Gede it is not clear how many of the eruptions that have occurred since 1747 have a magmatic component (or are phreatic in character). In order to estimate a value for this node we assume that roughly 2 in 10 episodes of unrest lead to eruption at silicic volcanoes (Newhall and Hoblitt, 2002). Gede has mainly erupted andesitic magmas (Krimmer, 2015), which might be expected to have a higher 'success' rate at eruption, but as we know there have already been 11 unrest episodes without eruption, we use the 0.2 value.
- **P(4|3): Given that a magmatic eruption will occur, what is the probability of it being $VEI \leq 2$, $VEI = 3$ or $VEI \geq 4$?** In this case we use the frequency of events from the eruptive history to define the probability of eruptions of this magnitude occurring again. Once again, it is difficult to judge which eruptions in the Gede's history have been magmatic or not. For now we calculate the distribution assuming that all of the eruptions are the product of some magmatic intrusion. The values we calculate are: $VEI \leq 2$: $(19/26) = 0.75$, $VEI = 3$: $(6/26) = 0.25$, $VEI \geq 4$: $(1/26) = 0.05$.
- **P(5|4): Given that an eruption of a certain magnitude will occur, what is the probability of it generating specific eruptive phenomena?** Here we estimate the probability of an eruption of a specific magnitude generating an ash fall or a pyroclastic flow. We use the compilation of volcanic domes produced by Newhall and Melson (1983) to calculate probabilities of certain VEIs producing these products. We find that of the 160 dome eruptions, 140 were associated with normal explosive behavior, with vertically directed columns. We assume that these columns all could have produced a degree of ash fall around the volcano. There were 61 dome eruptions that were associated with pyroclastic flows and surges. Based on Ogburn et al. (2015): about 13% of 397 dome forming episodes had a large explosions or c.20% of the $VEI \geq 4$ eruptions compiled by GVP were associated with dome growth. Using this data we have assigned values to these nodes.
- **P(6|5): Given that the eruptive phenomena (ash fall / pyroclastic flows) will occur, what sector of the volcano will it affect?** We divide Gede into three sectors that we feel are impacted differently by the eruptive phenomena. This criterion is dominantly controlled by the topography of the crater. As we have stated the current configuration of the crater at Gede forms a horseshoe shape, open to the NE. The sector between N and SE ($0 - 119^\circ$) is therefore, most strongly impacted by the pyroclastic flows. A few of the larger volume dome collapses also generate some flows into the other sectors. We also

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note that the sector from the SW to the N (240 - 360°) is most heavily affected by the ash fall as the prevailing winds blow this way.

- **P(6|7): Given that sector 1, 2 or 3 will be affected, how far will the flows travel / how thick will the ash fall be?** We use the maximum run out distances of the flows from our simulations to assess the probabilities of these events. We assess the pyroclastic flow run out as either $\leq$ or $>$ than 7km. This distance would allow pyroclastic flows to reach the presidential palace and indeed the centre of Cibodas and therefore would present a significant hazard. The probabilities of thickness of ash fall in these regions are estimates based on the simulations.

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This event tree reflects our current understanding of the main hazards that are associated with Gede volcano. We show here that there is a significantly higher probability of Gede producing a VEI 2 eruption than a VEI 4 eruption. It seems, however, that any size eruption from Gede could produce pyroclastic flows that will travel to the NE of the crater for a distance of greater than 7km.

This tree does not include any description of lahar hazard, lava flows or sector collapse. Whilst there has been a sector collapse at Gede, the probability of this occurring again is small enough that we do not address it. Similarly, there have never been any lava flows from Gede that have reached populated areas and so we have not included them here. It will be necessary to assess the lahar hazard at Gede in the future. Whilst there is no crater lake or snowpack present at Gede, there is certainly a possibility of ash fall and heavy rainfall generating lahars after an eruption. Future work will address this limitation of our event tree.

It is also important to note that our TITAN2D simulations would be improved by a clearer interpretation of eruptive activity. This can only be achieved by further field work in this region, and should form part of future efforts. We have assumed that simulating dome collapse events is the best representation of most of our deposits, but this may not be the case given more information. Clarifying the mechanisms of this eruptive activity would allow for tuning of the models and therefore would inform a more precise hazard assessment. The current hazard assessment is focused on the northeast region of the flanks of Gede. This is because the current crater topography suggests that eruptions will preferentially affect this area. If there is a change to the summit topography, however, it is possible that other areas of the volcanoes flanks could be impacted. For this reason it would be beneficial to perform more detailed field campaigns in other sectors and augment these with TITAN2D simulations, to give a more complete hazard assessment of these areas.

Finally, we currently assess the 1, 10 and 100mm contours for ashfall exceedance probabilities for eruptions of VEI 3, VEI 4 and VEI 5. Further work would be improved by modeling a greater range of eruption scenarios and ashfall contours. This would in turn provide a more accurate input for the event tree.

4.7 Conclusions

We use the eruptive history of Gede to simulate the potential impacts of an eruption of this volcano on the surrounding local and regional area. The geologic history suggests that Gede has had a combination of magmatic and phreatic eruptions, the largest magmatic eruptions occurring in the Holocene. These four eruptions all produced channelized pyroclastic flows, and so we model the effect of dome collapses with different volumes (suggested by these historic flows) and mechanisms to anticipate key areas of volcanic hazard today. We find that the basal friction of the flow dominantly controls the maximum run out distances. By generating successive flows

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of 37.5 million m^3 (for example) we find that decreasing the basal friction from 12-6 doubles the maximum run out distance from 8-16.5km. This suggests that the mechanism by which the pyroclastic flows are released is also important. If the dome fails in the series of collapses then the basal friction will decrease, if however the dome fails in a single collapse it is unlikely for the flows to travel as far. We also suggest that whilst maximum run out distances are well captured by the observation of deposits, the lateral extent of these is potentially underestimated. We find that there is perhaps also a final topographic control of the larger flows, which mostly do not extend past the Sukamahi ridge (c.15km away). In this analysis we show that, in light of new geologic and stratigraphic data, an eruption could potentially hazardously affect a region of >7km.

We also assess the impact of ash fall on the area-surrounding volcano. Probabilities are highest to the SW of the volcano (controlled by dominant wind direction). The areas that have an >50% exceedance probability of ash fall of 1mm for VEI 3, 4 and 5 eruptions are estimated to have populations of 5,000, 70,000 and 200,000 respectively. Whilst the probabilities of ash fall in Jakarta and Bandung are lower, due to the prevailing wind direction, we still find that there is a 20% probability and a 10% from a VEI 5 eruption.

The 2006 Gede hazard map shows an area of higher hazard to the SW of the crater. We have not investigated this region, as we do not have detailed information about the deposits. Future studies will seek to further investigate; the affect of flows on this region, the impact of column collapse events and to simulate domes collapses whose primary failure direction is different to the NE failure we have simulated here.

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An analysis of the issuance of Volcano Alert Levels during volcanic crisis

Volcano Alert Levels (VALs) are a tool for volcanologists to quickly and simply inform local populations and government authorities of the level of volcanic unrest and eruption likelihood. Most VALs do not explicitly forecast volcanic activity but, in many instances they play an important role in informing decisions to define exclusion zones and in issuing evacuation alerts. We have performed an analysis on VALs from around the world (194 eruptions, 60 volcanoes) to assess how well VALs reflect the unrest before eruption and what other variables might be controlling them. We have also looked at VALs in cases where there was no eruption after an increase in alert level, and which we term ‘Unrest without eruption’ (UwE). We have analyzed our results in the context of eruption and volcano types, instrumentation levels, number of eruptions that a volcano has experienced, and the population living around the volcano. We found that globally, 19% of the VALs issued between 1990 and 2013 for events that ended with eruption accurately reflect the hazard at the volcano before eruption. This percentage increases to c.30% if we only consider eruptions with a Volcanic Explosivity Index (VEI) ≥ 3 . VALs of eruptions from closed-vent volcanoes are more appropriately issued than those from open-vents. These two observations likely reflect the longer and stronger unrest signals associated with large eruptions from closed vents. More appropriate VAL issuance is also found in volcanoes with monitoring networks that are moderately well equipped (3 - 4 seismometers, GPS and gas monitoring). There is a better correlation between VALs and eruptions when the volcanoes are situated in densely populated areas. We also see that over time (1990 to 2013) there has been an increase in the proportion of ‘UwE’ alerts to other alerts, suggesting increasing willingness to use VALs for one of their main purposes, giving alerts well before an eruption is certain. Counting all UwE alerts as appropriate, and adding these to the 19% of appropriate alerts before eruptions, we conclude that 55% of all VALs are appropriate. 19% ‘success’ before eruptions is alarmingly low; a 55% ‘success’ rate for all alerts (with or without eruption) is better, and improving over time, but still lower than it should be. We suggest that the low global accuracy of the issuance of VALs could be improved by having more monitoring networks equipped to a medium level, but also by using probabilistic hazard management during volcanic crisis.

Author Contributions: Annie Winson designed the project, compiled and analysed

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the data, and wrote the manuscript with Fidel Costa. Chris Newhall and Gordon Woo contributed ideas and interpretations of results. All read and approved the final manuscript.

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5.1 Introduction

Mitigation of volcano hazards relies on the correct identification of unrest signals (Marzocchi et al., 2012; Sparks et al., 2012), and thus good monitoring networks and complete knowledge of the system. This allows scientists to assess the probability of an unrest episode leading to an eruption or of a return to background levels (Newhall and Hoblitt, 2002; Marzocchi and Woo, 2009; Lindsay et al., 2010; Woo, 2011; Marzocchi et al., 2012; Sparks et al., 2012). It also requires an efficient means of communicating the information and interpretation of volcanic unrest to the government agencies for ensuring peoples' safety (De la Cruz-Reyna and Tilling, 2008; Fearnley et al., 2012). Volcano Alert Levels (VALs) are such a tool, using a simple color, letter or number code to report the level of unrest. Ideally the VAL should increase progressively before eruption or increase and then decrease where unrest does not proceed to eruption. In many cases, changes in VAL carry strong implications about the exclusion areas around a volcano and about evacuation and safe return of populations. However, decisions to change a VAL at any given volcano are fraught with uncertainty about the volcanic system itself, and also psychological and personal factors of the scientific decision-maker (e.g. Woo, 2014).

Given the importance of VALs for saving lives and property (Auker et al., 2013), we decided to analyze their issuance worldwide. We have compiled a dataset of VALs and eruptions dates and we investigate what parameters play a role in a more appropriate issuance of VALs. For example: are VALs issued in a timelier manner at volcanoes where there is a denser monitoring network? Are the unrest signals easier to interpret at some types of volcanoes than at others, leading to more appropriate VALs? Do more frequent eruptions at a volcano lead to more training of staff and therefore more appropriate VALs? We investigate how the issuance of VALs varies with the volcanic system (size and type of volcano) as well as with some technical (e.g. monitoring network) and social (e.g. population at risk) factors. Finally we also discuss the role of the scientific decision-maker on the appropriate issuance of VALs.

5.1.1 Past and current alert level schemes

The eruption of Kelut in 1919 led to the creation in Indonesia of a 'Volcano Watching Service', which was at the time part of the Department of Mines (Van Padang, 1983; Suryo and Clarke, 1985). The aim of the new service was to study how populations could be protected from volcanic eruptions and specifically considered volcano type, eruption history and volcanic unrest in order to warn and evacuate the population when necessary (Van Padang, 1983). Observation 'Posts' were established at Merapi from 1934 to warn of imminent volcanic threats (Van Padang, 1983). During the 1980s there was a growing interest in formalizing eruption warnings in the framework of VALs. Although never formally published, one of the first of these was developed by the Rabaul Volcano Observatory (RVO) in response to the unrest in the caldera during the early 1980's (RVO Mimeograph, 1983; United Nations Office of the Disaster Relief Coordinator,

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1985). When activity began at Mt St. Helens in 1980 it became apparent that the scheme in place (patterned after one used by the National Weather Service) did not capture rapid changes within the overall eruptive sequence. As an alternative, the USGS issued ‘factual statements’ - descriptions of current conditions and ‘extended outlook advisories’ that looked several weeks ahead (Newhall, 1984; Swanson et al., 1985). There was also a concerted effort to communicate risk to the people who went back to work in the areas decimated by the lateral blast in May 1980 (Newhall, 1984). These events raised the need for comprehensive VALs. The schemes that are in place today were designed in the late 1980’s and early 1990’s. Their creation was influenced by two significant events:

1. The encounter between a Boeing 747 bound for Anchorage International Airport and the ash cloud produced by the eruption of Redoubt (Guffanti and Miller, 2013). Although the crash was averted the event spurred the development of the aviation alert level scheme by the USGS scientists at the Alaskan Volcano Observatory (AVO) (Guffanti and Miller, 2013). The scheme was adopted by the Kamchatkan Volcanic Eruption Response Team (KVERT) in 1993 and was standardized across the United States in 2006 (Fearnley et al., 2012; Guffanti and Miller, 2013).
2. The eruption of Pinatubo in the Philippines in 1991 (Tayag et al., 1996). Escalating volcanic unrest preceded this eruption, and an alert level scheme was created and used to simplify communication between scientists and the local officials. The five - point scheme that was developed at Pinatubo was then extended to other volcanoes in the Philippines: Mayon, Taal, Kanlaon and Bulusan, though the VAL criteria developed by the Philippine Institute of Volcanology and Seismology (PHIVOLCS) are specific to individual volcanoes to take into account the different physical characteristics.

Further VAL schemes were developed globally through the 1990’s, and some have been simplified and standardized since then (Japan Meteorological Agency, 2009; Fearnley et al., 2012).

Most VALs are country-specific and further subdivided by the repose time (frequently active / reawakening). We looked at alert level information from 11 different countries. Their schemes can be quite different, but we found that they broadly contain several or all of the following points:

1. **Indicative Phenomena:** A description of the precursory activity currently occurring at the volcano, possibly including a description of seismic, geodetic or gas changes.
2. **Volcano Status:** A quick, one-line description of the overall activity at the volcano, e.g. ‘Minor eruption in progress’, ‘Hazardous eruption in progress’ or ‘Volcano is in typical, background, non-eruptive state’

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3. **Recommended Action:** Information on exclusion zones for a given specific activity as well as recommendations for evacuations.
4. **Time Scale:** Some indication of the potential time frame before an eruption, usually given in days, weeks or months.
5. **Number or Color Scale:** Either numbers or colors (or both) to communicate an escalating (or decreasing) volcanic hazard. When colors are used they are usually in the ‘traffic light’ scheme, where green is equated with normal conditions and red is the most dangerous level. Most numeric schemes link ‘0’ or ‘1’ to green and ‘4’ or ‘5’ to red, but some countries invert this, with ‘1’ being the highest level of unrest.

For all the volcanoes that we examined, the final level included the statement that ‘an eruption was imminent’ or that an ‘eruption was underway’.

5.2 Methodology

5.2.1 Data sources and treatment

Volcanological surveys issue VALs for local and national purposes. We retrieved these VALs from online archives of volcanological and geological surveys, the ‘Smithsonian/USGS Weekly Volcanic Activity Report’ and the Smithsonian Institution’s monthly ‘Bulletin of the Global Volcanism Network’, and internal hard-copy archives of some of the surveys in the study. We also drew supporting information from EM_DAT (Emergency Events Database) (CRED, 2008) and DesInventar (Disaster Information Management System) (UNISDR, 2011).

To determine which volcano observatories issue alert levels, we surveyed all of the volcanoes in the GVP database, recording the ones that had any weekly or monthly bulletins. We used reports of both volcano observatories and summarized monthly reports in the Smithsonian archives. We identified volcanoes that had activity after a given observatory began using VALs. Where we were unable to determine when a specific observatory began using alert levels, we took the date of the first record of any VAL at that volcano. This survey encompassed 133 volcanoes including 291 eruptions from 17 countries. However, many data sets were incomplete, so to ensure that the records were consistent with each other we only use cases where: (1) VALs had been established at the volcano before the eruption began, and (2) the records of VAL were continuous. We also excluded volcanoes that had no eruptions in the time frame of modern alert levels (e.g. since 1991) and volcanoes that were in continuous eruption for more than 10 years. After filtering by these criteria, our database includes 60 volcanoes and 194 eruptions from 11 countries: Chile, Colombia, Ecuador, Indonesia, Japan, New Zealand, Peru, Philippines, Russia, United States and Vanuatu. The eruptions occurred between 1th January 1990 and 31st December 2013. We

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note here that some countries with large numbers of volcanoes, such as Italy and Iceland, are not part of our study because either they don't publicly issue VALs or else we were not able to retrieve their data. Our complete dataset can be found in the electronic supplement.

5.2.2 Data classification

We made a timeline of the changes in VALs at each volcano and compared it with the start dates of eruptive episodes from the GVP. We plotted VAL data from 30 days prior to the onset of an eruption through to the end of the eruption. We assumed that any unrest within 30 days prior to eruption is related to the causes of the eruption. Using a longer time window (e.g. 60 days) does not change our results. We tracked how the VALs changed with time relative to the onset of eruption. Our main tenet is that a VAL is appropriate if it progressively increases before eruption, or if it increases then decreases when an intrusion of magma does not erupt.

We identified six possible VAL patterns (Figure 5.1): Category 1 has been termed 'Missed', where the onset of the eruption was not preceded by an increase in the VAL, and the VAL was not moved (and usually remained low) during the course of the eruption. In Category 2 ('Too late') the VAL was increased for the first time after the eruption began (days to weeks) but before the eruption ended. Category 3 is for 'Premature' alerts, where the VAL was increased but subsequently decreased to lower levels prior to the onset of eruption. Category 4 is called 'Almost' because the VAL was increased prior to the eruption, but not to the highest level that indicates 'increased potential of eruption' or 'high alert'. Often, the VAL was changed from the 'Normal' or 'Background' level to a higher level of unrest, but no further. Category 5 is called 'Timely' because the VAL was increased in a stepwise fashion and, at the point of eruption, was at an appropriately high level. Category 6 'UwE' (Unrest without Eruption) refers to cases in which the VAL was increased, remained elevated and then was lowered without an eruption occurring within the next 30 days. We consider Categories 4 ('Almost') and 5 ('Timely') to be the most appropriate issuance of alert levels in cases where unrest led to an eruption, and Category 6 'UwE' as appropriate when unrest did not lead to eruption. We have analysed reports from these episodes of unrest and found that the 'UwE' alerts recorded in this study are well correlated with the state of the volcano, and were thus appropriate. However, we have no way of counting the episodes of unrest that did not lead to an eruption that were not represented by an increase in the VAL. We therefore, cannot estimate the percentage of 'UwE' that were 'Missed' by the VAL systems.

We also note that the eruption start dates in the GVP often refer to the onset of phreatic precursors to magmatic eruptions. VALs, on the other hand, are more likely to track magmatic ascent. To test whether reference to phreatic or magmatic onsets would change our characterization of VALs we did an analysis for larger eruptions, all clearly magmatic ($VEI \geq 3$). Using the eruption reports we recorded the dates where magmatic activity can be assumed to have begun. This is defined as either the day that 1) Dome growth was recorded; 2) Pyroclastic flows were observed;

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Figure 5.1: Classification of the issuance of alert levels relative to eruptions. The VAL is plotted against time (number of days). The first 30 - day window shows the time before the eruption onset. The red line denotes the onset of the eruption, the plot continues until the end of the eruption (in this example at day 60).

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3) Eruption column reached 10km or greater, or 4) the date of the maximum eruption column height. We then assessed whether there was any change in how the event would have been ranked. We found this effect to be negligible (only 1 case out of 40 would have been classified differently).

5.2.3 Data sources for level of monitoring

We gathered information on the monitoring network at specific volcanoes from the records that are kept by World Organization of Volcano Observatories (WOVO) and by each observatory. To assess the level of monitoring we used a slightly modified scheme from that developed by Ewert et al. (2005), in the development of the National Volcano Early Warning System (NVEWS), where a numerical value was assigned to different degrees of volcano monitoring capacity. We focused on the seismic, deformation, gas and remote sensing capabilities of the network when we assessed the level of monitoring at the volcanoes in our study. Level 0 denotes ‘No In-Situ’ monitoring; eruption confirmation is usually by remote sensing. We have eight volcanoes with this level of monitoring. Volcanoes with Level 1 or ‘Limited’ have no real time monitoring data from ground-based sensors, but the volcano is within a regional seismic network. Eruption confirmation is still by remote sensing and there is a baseline inventory of satellite images available: five volcanoes from our sample have Level 1 monitoring. Level 2 network is considered ‘Basic’, and includes: 1 - 2 seismometers within 10 km of the volcano and any combination of: 3 continuous GPS or tiltmeters in the vicinity, repeated campaigns of gas measurements, and near-real-time images. There are 22 volcanoes that have this level of monitoring. Level 3 volcanoes have ‘Good’ monitoring, with 3 - 4 seismometers in the near field and any combination of: 6 seismometers within 20 km of the vent, least 6 continuous GPS in the vicinity and continuous or periodic gas monitoring. There are 21 volcanoes in our sample at this level. Level 4 or ‘Research Level Monitoring’ volcanoes include 12 - 20 stations within 20 km of the vent, sufficient continuous deformation stations to do detailed source inverse modeling and continuous gas monitoring. There are only four volcanoes in our study with this level of monitoring.

When we were in doubt about the level of instrumentation at a volcano we have awarded the higher ranking among the possible levels. Therefore our assessment can be seen as a measure of the maximum monitoring capability based on the information available. Because of the difficulty in accessing data about the changes in monitoring capability over time, we used the most up to date information. We have not been able to make an assessment of the staffing levels and capabilities at observatories, although we assume that there is loosely an increase in the number of staff and the level of training with an increasingly dense network.

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5.3 Results

From our sample of 194 eruptions we find that about 80% of VALs fall into the categories ‘Missed’ (Category 1; 76 eruptions, 39%) and ‘Too Late’ (Category 2; 76 eruptions, 39%), about 14% as ‘Almost’ (Category 4; 28 eruptions), and only about 5% can be classified as ‘Timely’ (Category 5; 10 eruptions; 5%) (Figure 5.2A). For the quantification of category ‘UwE’ we used the percentage of the number of ‘UwE’ from the total number of issued alerts (therefore excluding ‘Missed’). We found that overall 33% of alerts issued were ‘UwE’ (Figure 5.2B).

Figure 5.2: Overall results from the study. Pie chart (A) shows the relative frequencies of Categories 1 - 5. There is a large difference in the sample sizes (N). N values for categories are; Cat 1: 75, Cat 2: 76, Cat 3: 5, Cat 4: 28, and Cat 5: 10. Pie chart (B) shows the proportion of ‘UwE’ alerts as a percentage Categories 2 - 6, i.e. of total alerts less Category 1 (‘Missed’).

If we group the categories of ‘Timely’ and ‘Almost’ we find that only 19% of the alert levels were increased by any amount before the onset of eruption. We were surprised by this low percentage. Below we investigate what may be controlling the different categories by analyzing the results according to different parameters. First we look at the volcanic system: volcano type and eruption styles. Later we address the technical and societal factors that may be involved, including; the monitoring capacity, the effect of the number of eruptions within a system and the population at risk.

5.3.1 Volcano and eruption styles

We divided the eruptions according to the VEI (as assigned by the GVP) and found that in general there is an increase in the number of Category 5 (‘Timely’) VALs as the eruption size increases (Figure 5.3). For $VEI \leq 2$ there is a high percentage of ‘Missed’ and ‘Too Late’ VALs (42% and 39%, respectively) and of the 154 eruptions, we see only 3% of the VALs in the ‘Timely’ category for these small eruptions. As the eruption size increases to VEI 3 the VAL’s in the ‘Missed’ category decrease (26%), but those in the ‘Too Late’ increase (41%). There is also an increase in the percentage of ‘Timely’ (12%) and ‘Almost’ for eruptions of VEI’s ≤ 2 to VEI

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3. For $VEI \geq 4$ eruptions the percentages in the ‘Missed’, ‘Too Late’ and ‘Timely’ categories are about the same (33%), but the sample size ($n = 6$) is too small for statistical significance.

Figure 5.3: VAL results for eruptions with different VEIs. Percentages of data are classified by categories 1 (‘Missed’ in red), 2 (‘Too late’ in orange), 3 (‘Premature’, in yellow), 4 (‘Almost’, in light green) and 5 (‘Perfect’, in dark green).

We also looked at the character of the volcano to determine if this would have any effect on the issuance of VALs. For this, we simplified our classification into: (1) Open-vent volcanoes that produce many frequent, small eruptions. They are often continuously degassing and tend to have more effusive eruption products. This continuous release of gas seems to make large eruptions less likely, unless the system seals itself (Newhall, 2007), (2) Closed-vent volcanoes have less frequent eruptions and are commonly dormant for decades, centuries, or longer. About 53% of the volcanoes in our database were classified as open-vent, and produced 76% of the eruptions, whereas the remaining 47% are closed-vent, and produced 24% of all eruptions. We found a higher percentage of ‘Timely’ and ‘Almost’ VALs (10% and 19%, respectively) in closed-vent (Figure 5.4) than in open-vent systems (3% and 12%, respectively). Open-vent systems also show a higher number of ‘Missed’ events (43%) than closed-vent systems (22%). Finally the proportion of ‘UwE’ at open-vent volcanoes is 28% of the total number of alerts, but this proportion increases to 42% at closed-vent volcanoes.

5.3.2 Role of monitoring networks and equipment

We found that there is an increase of the ‘Timely’ and ‘Almost’ categories with the increase in the capacity of the monitoring network (Figure 5.5). Volcanoes with monitoring Levels 0 and 1 have a high number of ‘Missed’ and ‘Too Late’ VALs (88% and 80%, respectively for 13 total

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Figure 5.4: Proportion of VALs issues at Open and Closed Volcanoes. Percentages of data set that are classified as Categories 1 - 5 following the color code for Figure 5.2. Pie Charts show 'UwE' (Category 6, black) as a % of all alerts except 'Missed'. Data have been separated by volcano type

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events). For volcanoes with Levels 2 and 3, we find that there is an increase in the ‘Timely’ and ‘Almost’ VALs from 14 (Level 2) to 43% (Level 3). The value of the ‘Almost’ category for monitoring Level 4 still increases to about 50%, although given that we only have 4 volcanoes in this category, it is not clear if the increase is statistically significant. At volcanoes with low monitoring levels, the ‘UwE’ alerts are a very high percentage of total alerts issued (72% for Level 0 monitoring) and then broadly decrease as the monitoring level increases (33% for Level 4 monitoring).

Figure 5.5: Effect of the level of monitoring on the issuance of VAL's. Percentages of data set that are classified as Categories 1 - 5 following the color code for Figure 5.2. Pie Charts show ‘UwE’ (Category 6) as a % of all alerts except ‘Missed’.

5.3.3 Effect of eruption frequency on the issuance of VALs

To test whether an increased knowledge of the system (e.g. a given volcano) increases the reliability of the issuance of VALs, we looked for an increase in the number of ‘Timely’ and ‘Almost’ VAL categories with an increase in the number of eruptions at a given volcano. In our sample, 34 volcanoes had more than one eruption over the time period and this filter still includes 161 eruptions. We found a large spread in data and no relationship between the number of eruptions per volcano and the percentage of ‘Timely’ and ‘Almost’ VAL categories (Figure 5.6).

However, some pattern emerges if we consider frequency of eruption and level of monitoring.

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Figure 5.6: The effect of multiple eruptions on the issuance of appropriate VALs. Data are separated by Open (23 volcanoes, 135 eruptions) and Closed (10 volcanoes, 26 eruptions) vent volcanoes.

Volcanoes with frequent eruptions (mostly, open vent volcanoes) with the highest percentages of appropriate alert levels are those with monitoring levels 2 and 3 (Figure 5.7). The lowest percentages of appropriate VALs are for volcanoes that have fewer eruptions (mostly, closed-vent volcanoes), and a moderately high level of monitoring (level 2 or higher).

5.3.4 Effect of population on VALs

The purpose of VALs is primarily to inform local populations (and, in some cases, air traffic) of the potential for volcanic hazard. Local authorities also use them to inform decision making for evacuations. We checked to see if the size of population living around a given volcano affects the issuance of VALs. We used the population figures from the Smithsonian GVP, which was obtained from the LandScan 2008 data (Oak Ridge National Laboratory, 2008). We considered the population size within 30 km of the volcano, and normalized the data by the number of volcanoes within each population bin size (Figure 5.8A), to account for variable number of volcanoes per bin. We found that there are more ‘Timely’ alerts in the bin with the highest population, but in general there are similar proportions (within a few percentage points) for the ‘Timely’ and ‘Almost’ categories for all population sizes (Figure 5.8A), meaning that there is no significant correlation between appropriate VALs and population size. The proportion of ‘UwE’ varies between about 24 and 48% for all population sizes, without a clear change with population. As an additional measure to test the influence of the population on the decision to change VALs, we compared data from two countries with very different demographics. Country

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A) Closed Volcanoes

B) Open Volcanoes

Figure 5.7: Controls on the issuance of VALs due to changes in monitoring at volcanoes that have had multiple eruptions. Color shaded field's group volcanoes values with similar monitoring capacity. A) For 'Closed Vent' volcanoes: Blue area represents 3 volcanoes, 9 eruptions all have monitoring level 2. Green area represents 5 volcanoes, 15 eruptions 86% of which have a monitoring level of 2. B) For 'Open Vent' volcanoes: Red area represents 4 volcanoes with 48 eruptions, 75% of which have monitoring level 0, blue area represents 11 volcanoes, 53 eruptions 73.5% of which have monitoring level 2, green area represents 6 volcanoes, 16 eruptions 75% of which have level 3(+) monitoring.

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A has a large population living within 30 km of volcanoes, whereas Country B has much lower numbers of people (on average less than a 100) living close to its volcanic centers.

For Country A we found that for volcanoes with populations of less than 1 million people about 21% of the VALs fall into the ‘Timely’ or ‘Almost’ (Figure 5.8B), but this increases to 45% for volcanoes with more than 1 million people, so there is an improvement with population size. The highest proportion of ‘Missed’ events occurs within the population range of 1,000-10,000. The ‘UwE’ show a somewhat bi-modal distribution for Country A with peaks for the lowest (1,000-10,000) and the highest (>1,000,000) populations. In contrast, for Country B with sparsely populated volcanoes (less than 2,500 people living within a 30 km radius of the volcanoes) we see that even in the lowest populations (less than 1000) there are higher proportions ‘Timely’ and ‘Almost’ VALs (Figure 5.8C) compared to the highly populated Country A. But there is also an increase in appropriate VAL’s with population size. There is no dominant trend in the ‘UwE’ alerts that can be attributed to differences in population. The general decrease in ‘UwE’ alerts for Country B is more likely a factor of the small sample size, rather than a true trend.

5.3.5 Has the issuance of VALs improved over time?

Since 1990 there has been an increase in the number of countries and institutes that have adopted a form of VAL, from one in 1990 to 60 in 2013 (Figure 5.9). To be able to compare the results from different years we have normalized the data by the number of volcanoes. We find that there is quite a large variety from year to year and not a linear increase of any of the categories with time (Figure 5.10). Many ‘Missed’ and ‘Too Late’ VALs occurred prior to 2000 while from 2004 onwards most years have higher ‘Timely’ and ‘Almost’ categories (in general > 20%). To better test whether the VALs are improving with time, we performed a moving average analysis of the ‘Timely’ and ‘Almost’ categories (Figure 5.10) which shows a complex pattern but a general improvement after 2003.

We also compared the number of alerts and number of eruptions over time normalized by the number of volcanoes and some trends can be identified (Figure 5.11). We find that from 1996 to 2004 the normalized number of eruptions is higher than the number of alerts, which reflects that many eruptions occurred without any alert being issued. However, starting from 2005 the ratio of numbers of alerts to eruptions is similar or higher than one, so that on average at least one or more alert was issued per eruption, something that is also reflected in the decrease in ‘Missed’ events over this time period (Figure 5.10). Thus, in general we find that with time there is an increasing number of VALs issued per volcano and per eruption that could reflect an increase in the monitoring network or change in the scientific decision making process as discussed below.

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A) All Countries

B) Country A

C) Country B

Figure 5.8: Effect of Population on VALs. A) Distribution of VALs issued for all volcanoes in the study compared to the population living within 30 km of the volcano. B) Data for Country A - high average population within 30km. C) Data for Country B - low average population within 30km. All percentages are normalized by the number of volcanoes in each bin. Percentages of data are classified as Categories 1 - 5 following the color code of Figure 5.2. Pie charts show UWE (Category 6) as a % of all alerts except 'Missed.'

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Figure 5.9: Distribution of data in study over time. A) Change in cumulative number of volcanoes over time. B) Change in cumulative number of eruptions over time. C) Change in cumulative number of alerts over time. D) Change in cumulative number of 'UWE' over time.

Figure 5.10: Change in VAL issuance in episodes that ended with eruption and those of 'UWE' over time. Percentages of VALs issued are normalized by the number of volcanoes included in the study year to year. Percentages of data are classified as Categories 1 - 5 following the color code of Figure 5.2. Blue line is the % of 'UWE' and the Black line is the 3 year running average of the 'Timely' and 'Almost' categories.

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Figure 5.11: Total cumulative number of alerts issued (‘Timely’, ‘Almost’, ‘Premature’, ‘Too Late’ and ‘UwE’) and total cumulative number of eruptions in the study. Blue line is the ratio of the two data series. All data are normalized by the number of volcanoes included in the study year to year.

5.4 Discussion

We discuss how well VALs have reflected the state of the volcanoes, and two factors that may influence these results - the physical nature of the system, e.g. the size of the eruption or the type of volcano, the level of monitoring, or the number of outcomes (eruptions) from given volcano, and roles of population and scientific decision-making process.

5.4.1 How well do VALs reflect the state of a volcano?

We consider three more specific, component questions:

1. What percentage of eruptions was anticipated by appropriate alert levels? This is an eruption - centric evaluation of VALs. We sum Categories 4 (‘Almost’) and 5 (‘Timely’) and divide by the number of eruptions, getting 19%, a disconcertingly low number. Clearly, much improvement is needed.
2. What percentage of unrest was anticipated by appropriate alert levels? Unfortunately, we cannot answer this question, as there are an unknown number of episodes of unrest for which no VALs were issued. In our experience, there are many slight elevations of unrest that are not tagged with VALs so this ‘missing’ number might be significant.
3. What percentage of alert levels was appropriate to the status of the volcano? This is an alert-centric evaluation of VALs. We sum percentage of categories 4, 5, and 6, from the

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total sample of alerts. The result is 55% percent, higher than the result for eruption-centric evaluation but still rather low compared to what we expect of volcano observatories. Both the eruption-centric and this alert-centric 'success' numbers are kept low by the alarmingly high number of 'Missed' and 'Too Late' alerts.

5.4.2 Physical and technical parameters that affect issuance of VALs

One of the clearest trends that we have found is a general increase of 'Timely' and 'Almost' categories with VEI (Figure 5.3) which could be related to the difference in volcanic precursors and unrest signals for small versus large eruptions. VEI 1's include small ash puffs and phreatic explosions that may arise from hot gas and rock interacting with ground water (such as the eruption of Mt Mayon in the Philippines in May 2013), which have few or no discernible precursors and thus no basis for raising alert levels. The larger and longer episodes of unrest that are typically associated with VEI 3 or 4 eruptions allow for more progressive increase of VALs. Larger eruptions are better anticipated than smaller ones. One example is Merapi, which had much larger unrest signals and a much larger and explosive eruption in 2010 compared to those in 2006 and other past eruptions (Surono et al., 2012).

The more appropriate issuance of VALs at closed vents (Figure 5.4) is probably because magma needs to fracture rock or clear the conduit before it can reach the Earth's surface. This may also explain why 'UwE' are more prevalent at closed systems than at open systems because the longer repose periods would lead to a greater need for re-establishing pathways within the conduit and may cause multiple episodes of unrest over an extended period prior to eruptions. The signals at these volcanoes may therefore be stronger and a higher number of 'UwE' can be expected due to several intrusions or episodes of magma ascent before an eruption is triggered.

We find that VALs at open vent volcanoes (Figure 5.7b) are improved by a higher number of eruptions and opportunities for the observers to be 'trained' by the events, and/or by a better level of monitoring that allows for subtle precursors to be recorded. As closed vent volcanoes have fewer eruptions, there is less opportunity for observers to experience volcanic crisis. Instead they need to rely more heavily on monitoring networks and well-constrained eruptive histories (both from detailed stratigraphic analysis and the careful review of historical documents). The lack of experience in closed-vent volcanoes or in volcanoes that reawaken after a long period of dormancy (e.g. Sinabung volcano in 2013) could be mitigated by the use of a worldwide database of volcano monitoring signals (Widiwijayanti et al., 2013).

Another factor that significantly improves VALs is an increased level of monitoring (Figure 5.5). This finding could be anticipated as more data available to the volcanologist makes for better understanding of the system. However, the increase in the accuracy from Levels 0 to 1, or Level 1 to 2 is much less than from Level 2 to 3. This suggests that it is necessary to increase the

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monitoring network to at least 3-4 seismometers, and/or GPS and gas monitoring for an effective improvement of the appropriate issuance of the VALs. Such an increase is usually coupled with an increase in human resources, as more effort is invested in interpretation of the monitoring data. The results for monitoring Level 4 volcanoes are inconclusive because the sample size is only four volcanoes. It is also possible that an observatory with a monitoring level as dense as a Level 4 network would have a large volume of data from differing sources to reconcile before a decision could be made about the potential for eruption at a given time. The increase of the proportion of appropriate VALs and of 'UwE' with time (Figure 5.10) could be a reflection of an increase of the level of monitoring or understanding of the volcanic systems. Better monitoring enables detection of more subtle unrest signals that may not lead to immediate eruption but which could be related to magma intrusion at depth (e.g. Moran et al., 2011). Unfortunately it is not possible for us to quantify this relationship because we could not determine how the monitoring level at each volcano changed with time. Another explanation is that the increase in number of 'UwE' reflects a change in the decision-making process whereby there is less reluctance to issue an alert even if the certainty of the eruption is not higher than in the past (see below).

5.4.3 The influence of the decision-maker(s) on the issuance of VALs

So far we have analyzed the role of the physical parameters (volcano and eruption types) and technical factors (e.g. monitoring level) that may lead to more appropriate issuance of VALs. However, a change in VAL is ultimately a scientific decision-making exercise, usually involving a single individual or a group. It is difficult to identify a quantifiable measure to characterize the effect of the decision maker on the reliability of VALs, but it is probably very large and may explain why the parameters we have investigated do not fully explain the different percentages of the categories. Decisions in high-risk situations such a volcanic crisis may be affected by the individual risk aversion of the decision-maker and by personal interest (e.g., losing reputation) (Woo, 2014). To compensate for the potential bias of the scientist-decision-maker, several workers have advocated for a more robust methodology based on probabilistic hazard assessment and cost-benefit analysis (Marzocchi and Woo, 2007, 2009; Marzocchi and Bebbington, 2012; Woo, 2014).

Two factors from our data suggest a significant effect of the decision-making process on the appropriate issuance of VALs. One factor is the increase of 'Timely' and 'Almost' categories and another is the proportion of Alert Levels and eruptions with time (Figures 5.10 and 5.11). As noted above, these factors could reflect results from an improvement of the monitoring networks with time, but also include a realization that to improve VAL issuances it is necessary to be less conservative. Our data could indicate that decision are starting to be made from an optimal risk management perspective, rather than waiting for eruption probabilities to become very high (e.g., Woo, 2014). The other indication of influence from the decision making process is the

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relationship between the VALs, ‘UwE’, and the population size. The global population dataset shows only an increase of the ‘Timely’ category with increasing population density, whilst the ‘Almost VALs show no trend. However, we found that two different countries with different demographics both show an increase of appropriate VALs with increasing population suggesting that the decision makers are more risk averse when changing VALs depending on the population at risk. We find that there is a general increase of ‘Almost and ‘Timely VALs with increasing population but that this increase seems to occur at different country specific thresholds. Country B with its low population density shows an increase in ‘Almost and ‘Timely VALs at low population numbers (>100) (Figure 5.8B). In Country A, which has a higher population density, we see an increase at populations $>10,000$. This increase with population size suggests that although VAL schemes maybe designed simply to communicate the hazard, they are being implemented with risk to local populations in mind. In a specific example (Indonesia), CVGHM did not raise the VAL to the highest level at Sinabung in 2013 even when phreatomagmatic ash columns were reaching 7-10 km as there was little risk to the population on the ground. Instead, they waited until there was a high likelihood of a summit lava dome collapse, so that evacuations would be more commensurate with the risk (John Pallister, personal communication 2014).

The trends of ‘UwE’ alarms in Countries A and B may hint at the complexity of this decision-making process. In Country B (Figure 5.8C) the decrease in ‘UwE’ as populations increase may suggest that there is an effort to not issue alerts until there is a high level of certainty about impending eruptions. Country A, though, shows a more complex response (Figure 5.8B). It could be argued that the increase in ‘UwE’ alarms as populations increase (after 10,000 people) reflects intolerance to leaving populations at risk. It is interesting to note that the higher levels of ‘UwE’ seem to be coupled with higher levels of ‘Timely’ VALs.

5.5 Conclusions

The reliability of issuance of VALs depends on the type of volcano, size of the eruption and monitoring network. Large eruptions (VEI 3 or larger) from closed-vent volcanoes with reasonably instrumented monitoring networks (Level 3), and which have a large population living in their flanks are those for which VALs are issued most appropriately. Over time, we see modest improvements in the VALs that are issued, including an increasing number of ‘UwE’ alerts. Improvement is seen in both eruption-centric and alert-centric evaluations, and suggests improvement in the monitoring networks and also an increasing willingness to issue alerts even if unrest may not lead to eruption.

However, ‘improved’ does not mean ‘satisfactory’. Only 19% of the eruptions in our dataset were correctly anticipated by issuance of appropriate VALs. The low overall success in anticipating eruptions may reflect both the complexity of volcanic systems and the complex, multidisciplinary decision-making process of changing VALs during a volcanic crisis

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(Marzocchi et al., 2012; Sparks et al., 2012). We infer other remaining problems in using VAL schemes, such as an insufficient density of monitoring networks and/or reluctance to raise alerts because of concern that they may generate undue public concern, slow tourist arrivals, generate extra communication challenges for the scientists, etc.

Whilst we highlight a need for the improvement of monitoring networks in our findings, a conservative approach to changing alerts, or waiting for a clear cut evidence to become available seemingly leads to VALs being issued after the onset of eruptions and potentially leaving populations at risk. There are far more cases of ‘Missed’ and ‘Too Late’ than there are of ‘Premature’ alerts so many more that we suggest observatories need not worry too much about issuing ‘Premature alerts. We further suggest that to reduce the number of ‘Missed’ and ‘Too Late’ calls there is a need for less conservative decision-making raising alert levels freely - so that mitigating action can be taken even when there is still considerable scientific uncertainty. This is, indeed, one of the most important goals of VAL schemes. We invite countries to perform their own self-evaluation and weigh the cost of a higher number of alerts against the benefit of a higher accuracy in VAL issuances and to decide how to proceed accordingly with their own local populations. We also suggest that with the support of databases such as WOVOdat, the most effective way to increase the appropriate issuance of VALs could be to expand the implementation of the use of probabilistic hazards management and cost-benefit analysis so that decisions can be as balanced as possible.

Conclusions and Future Work

Forecasting the size and impact of volcanic eruptions is a complex, multi-parametric problem. Moreover most of the information and data used to make the forecast has a high degree of uncertainty. The communication of this information is also a complex matter: waiting to issue or increase a volcanic alert level until you are sure an eruption is going to occur can leave populations at risk, whereas issuing false alarms can cause a population to lose faith in the scientists abilities to successfully forecast activity. This could lead to costly and unnecessary evacuations and disruption to peoples lives and livelihoods. In my PhD I have assessed three different facets of this problem.

Firstly, I have constructed a database of the geological, magmatic and unrest information that is currently available for plinian eruptions that have occurred in historic times. I have used these data to calculate values for physical properties of the magma such as viscosities, magma density, excess gas and pre-eruptive water content. I have also found new criteria such as the apparent repose duration, the ‘Plinian Ratio’ and the ‘Repose Ratio’ to assess the effect of the volcanic history on the size of eruption. With this information I have developed a simple Bayesian Belief Network to calculate the probability of an eruption being plinian.

The key finding of this work is the importance of the accumulation of an excess volatile phase in generating large, explosive eruptions. This significant accumulation of a volatile phase, in excess of fluid saturation, whilst the magma is still in the storage reservoir appears to be instrumental in generating plinian eruptions. The presence of a large amount of free volatile phase in a readily available, quickly eruptible melt allows for the generation of the large sustained eruptive columns with high mass eruption rates that define plinian eruption. All of the plinian eruptions for which I was able to calculate this parameter, have an excess volatile phase.

In general I find poor correlation between physical parameters that are considered critical for high explosivity such as viscosity, crystallinity and water content and plinian eruptions. However, I found evidence that plinian eruptions can be promoted by magma mingling, with more explosive eruptions being linked to larger temperature and compositional gradients between resident and injected magmas. I also found poor correlation also, between run up times of seismic and phreatic activity and the size of the eruption.

By combining these data in a Bayesian Belief Network I show that the largest eruptions in our

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study ($\geq 10\text{km}^3$ DRE) have a probability of plinian eruption of up to 32%. This is four times greater than the *a priori* probability of a $\text{VEI} \geq 4$ eruption that can be calculated from the global eruption catalogue.

A limitation of the study is the small sample size that is currently available. This situation could be improved in three ways: 1) Adding more eruptions to increase the understanding of plinian precursors and confirm trends between different parameters. 2) A simple expert elicitation to better estimate the probabilities of the different nodes of the network. This would allow for the addition of any expert opinion in the network, and may change the structure that we have currently defined. An example of how this elicitation could be performed is attached as an appendix 'Elicitation Questionnaire'. This elicitation would potentially allow for the current states to be divided into smaller increments, making our network more accurate. 3) Use hidden multi-state markov model to define states and also the rate at which states can be switched. This allows for joint evaluation of the nodes and adds a dynamic component to the network; potentially it would be possible to compute timescales for hazard assessment from this approach.

The second part of my PhD has been to produce an assessment of the ash fall and pyroclastic flow hazard from Gede volcano in Indonesia. These data have been organized as an event tree that I hope it will aid to make more informal and accountable decisions in any upcoming volcanic crisis at Gede.

I show that the most important control on the simulated maximum run out distances for the deposits at Gede is the basal friction coefficient. The results of the simulations suggest that the mechanism by which the flows are generated are the controlling parameter on the maximum run out distances and therefore the number of people affected by such an event. If a dome fails in a series of collapses (and thus the basal friction is decreased) then the flows will travel further than if the dome collapses in a single event. It does not seem possible for flows to travel further than c.15km at Gede, due to the presence of a ridge at this distance.

Ash fall from a VEI 5 eruption at Gede has the potential to affect the cities of Jakarta and Bandung. However, due to the prevailing wind direction it is more likely that ash fall will affect areas to the SW of the volcano. For an eruption of VEI 3, 4, and 5 this could impact 5,000, 70,000 and 200,000 people respectively.

The hazard assessment at Gede could be more fully rounded out by an assessment of the lahar hazard from this volcano. It is possible to simulate potential lahars, following the TITAN2D runs I have already performed, using the program LAHARZ. It would also be beneficial simulate the affect of more column collapses, to supplement the existing simulations. This would allow for the assessment of different eruption scenarios from Gede.

Finally I have performed an evaluation of volcanic alert level (VAL) issuance at the global scale. The final results suggest a relative poor success rate for issuance of alert levels; I believe that in the consultation of monitoring data from analogous volcanoes, available from

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the WOVOdat, prior to eruptions would clarify some precursory phenomena. The expansion of probabilistic hazard management and cost benefit analysis would also help to generate more balanced decisions.

Whether or not a successful VAL is issued is dependent on: the type of volcano, size of the eruption and the monitoring network. The most appropriately issued alerts are from closed-vent volcanoes with reasonably instrumented monitoring networks, and which have a large population living on their flanks. The issuance of VALs has improved over time. These improvements are seen at both eruption-centric and alert-centric evaluations and suggest the improvement in monitoring networks and an increasing willingness to issue alerts, even when the outcomes are uncertain. The implementation of VALs worldwide could be improved by facilitating discussion between WOVO members about their experiences in volcanic crises; this could lead to the establishment of a best practices code for VAL issuance. With this in mind, I am a co-proponent of a newly established WOVO commission working group on volcano alert levels whose remit fits these goals.

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Appendix A: Tables of Plinian Eruption Criteria

APPENDIX A. TABLES OF PLINIAN ERUPTION CRITERIA

Table A.1: Data used to calculate $ARD\ VEI \geq 2$, $ARD\ VEI \geq 4$ and the Plinian Ratio. All values are taken from the Global Volcanism Program (2014), except for; Tambora (Kandlbauer, 2013), Katmai (Hildreth et al., 2003), Santorin (Druitt and Francaviglia, 1992) and Santa Maria (Rose, 1987) . To calculate the $ARD\ VEI \geq 4$ and Plinian Ratio it is necessary to have more then one $VEI \geq 4$ in the eruption history. If this is not the case then no value is calculated. These values are for each volcano.

Volcano	Eruption Type	Total $VEI \geq 2$	Total $VEI \geq 4$	Length of eruption record (yrs)	Average Repose Duration ($VEI \geq 2$)	Average Repose Duration ($VEI \geq 4$)	Plinian Ratio
Tambora	Plinian	16	11	12014	750.9	1092.2	0.69
Santorini	Plinian	13	3	3624	278.8	1208.0	0.23
Changbaishan	Plinian	10	3	4174	417.4	1391.3	0.30
Katmai	Plinian	2	2	10014	5007.0	5007.0	1
Krakatau	Plinian	45	2	1815	40.3	907.5	0.04
Huaynaputina	Plinian	2	1	9764	4882.0	9764.0	0.50
Santa Maria	Plinian	3	2	1112	370.7	556.0	0.67
Pinatubo	Plinian	6	5	9474	1579.0	1894.8	0.83
Vesuvius	Plinian	50	13	8954	179.1	688.8	0.26
Cosigüina	Plinian	5	1	514	102.8	514.0	0.20
Hudson	Plinian	15	3	10024	668.3	3341.3	0.20
Öræfajökull	Plinian	2	2	652	326.0	326.0	1
El Chichón	Plinian	13	5	8524	655.7	1704.8	0.38
Bezymiany	Plinian	58	4	9064	156.3	2266.0	0.07
MSH	Plinian	38	8	4354	114.6	544.3	0.21
Fuji	Plinian	57	5	10554	185.2	2110.8	0.09
Hekla	Plinian	64	15	7864	122.9	524.3	0.23

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APPENDIX A. TABLES OF PLINIAN ERUPTION CRITERIA

Table A.1 – continued from previous page

Volcano	Eruption Type	Total VEI ≥ 2	Total VEI ≥ 4	Length of eruption record (yrs)	Average Repose Duration (VEI ≥ 2)	Average Repose Duration (VEI ≥ 4)	Plinian Ratio
Okatania	Plinian	9	6	10064	1118.2	1677.3	0.67
Colima	Plinian	65	7	9704	149.3	1386.3	0.11
Komagatake	Plinian	15	6	6614	440.9	1102.3	0.40
Asama	Plinian	120	4	9214	76.8	2303.5	0.03
Agung	Plinian	16	13	5202	325.1	400.2	0.81
Usu	Plinian	13	4	8564	658.8	2141.0	0.31
Askja	Plinian	6	2	10924	1820.7	5462.0	0.33
Mayon	Plinian	43	2	5114	118.9	2557.0	0.05
Shiveluch	Sub - Plinian	102	43	10514	103.1	244.5	0.42
Augustine	Sub - Plinian	20	4	7434	371.7	1858.5	0.20
Galunggung	Sub - Plinian	5	3	4264	852.8	1421.3	0.60
Soufrière St Vincent	Sub - Plinian	18	2	4394	244.1	2197.0	0.11
Pelée	Sub - Plinian	53	21	10224	192.9	486.9	0.40
Fuego	Sub - Plinian	57	7	3594	63.1	513.4	0.12
Kelut	Sub - Plinian	34	7	2244	66.0	320.6	0.21
Merapi	Sub - Plinian	81	4	10794	133.3	2698.5	0.05
Reventador	Sub - Plinian	28	1	473	16.9	473.0	0.04

APPENDIX A. TABLES OF PLINIAN ERUPTION CRITERIA

Table A.2: Data for Repose time prior to calculate the Repose Ratio. All values are taken from the Global Volcanism Program (2014). To calculate the Repose Ratio it is necessary to have more than one $VEI \geq 4$ in the eruption history. If this is not the case then no value is calculated. These values are for each eruption.

Volcano	Eruption Type	Eruption Year	Time since last $VEI \geq 2$ (yrs)	Time since last $VEI \geq 4$ (yrs)	Average Repose Duration ($VEI \geq 4$)	Repose Ratio
Tambora	Plinian	1815	815	815	1092.2	0.75
Santorini	Plinian	1610 BCE	16390	16390	1208.0	13.6
Changbaishan	Plinian	969	789	1149	1391.3	0.83
Katmai	Plinian	1912	8000	8000	5007.0	1.6
Krakatau	Plinian	1883	199	1467	907.5	1.6
Huaynaputina	Plinian	1600	9350	9350	9764.0	0.96
Santa Maria	Plinian	1902	1000	1000	556.0	1.8
Pinatubo	Plinian	1991	541	541	1894.8	0.29
Vesuvius	Plinian	79	296	959	688.8	1.4
Cosigüina	Plinian	1835	126	Unknown	514.0	-
Hudson	Plinian	1991	20	3881	3341.3	1.2
Öræfajökull	Plinian	1362	Unknown	Unknown	326.0	-
El Chichon	Plinian	1982	622	622	1704.8	0.36
Bezymiany	Plinian	1955	1005	1005	2266.0	0.44
MSH	Plinian	1480	700	700	544.3	1.3
Fuji	Plinian	1707	7	907	2110.8	0.43
Hekla	Plinian	1947	34	102	524.3	0.19

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APPENDIX A. TABLES OF PLINIAN ERUPTION CRITERIA

Table A.2 – continued from previous page

Volcano	Eruption Type	Eruption Year Year	Time since last VEI $\geq$ 2 (yrs)	Time since last VEI $\geq$ 4	Average Repose Duration (VEI $\geq$ 4)	Repose Ratio
Okatania	Plinian	1886	576	576	1677.3	0.34
Colima	Plinian	1913	5	24	1386.3	0.017
Vesuvius	Plinian	1631	61	663	689.0	0.96
Komagatake	Plinian	1929	1	73	1102.3	0.066
Hekla	Plinian	1104	54	2204	524.0	4.2
MSH	Plinian	1980	123	180	544.0	0.33
Asama	Plinian	1783	6	675	2303.5	0.29
Agung	Plinian	1963	120	120	400.2	0.30
Usu	Plinian	1663	25	Unknown	2141.0	-
MSH	Plinian	1800	190	318	544.0	0.58
Askja	Plinian	1875	575	10785	5462.0	2.0
Mayon	Plinian	1814	14	None	2557.0	-
Shiveluch	Sub - Plinian	1964	20	110	244.5	0.45
Augustine	Sub - Plinian	1986	10	10	1858.5	0.005
Galunggung	Sub - Plinian	1822	4072	4072	1421.3	2.9
Augustine	Sub - Plinian	1976	5	93	1859.0	0.050
Soufrière St Vincent	Sub - Plinian	1902	88	90	2197.0	0.041
Galunggung	Sub - Plinian	1982	64	160	1421.0	0.11
Pelée	Sub - Plinian	1902	51	562	486.9	1.2

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APPENDIX A. TABLES OF PLINIAN ERUPTION CRITERIA

Table A.2 – continued from previous page

Volcano	Eruption Type	Eruption Year Year	Time since last VEI $\geq$ 2 (yrs)	Time since last VEI $\geq$ 4	Average Repose Duration (VEI $\geq$ 4)	Repose Ratio
Fuego	Sub - Plinian	1974	1	42	513.4	0.082
Kelut	Sub - Plinian	1919	18	93	320.6	0.29
Merapi	Sub - Plinian	2010	4	138	2698.5	0.051
Reventador	Sub - Plinian	2002	26	Unknown	473.0	-
Merapi	Sub - Plinian	1872	7	1852	2699.0	0.69
Kelut	Sub - Plinian	1990	23	24	321.0	0.07

APPENDIX A. TABLES OF PLINIAN ERUPTION CRITERIA

Table A.3: Data from whole-rock analyses for Plinian and Sub-Plinian eruptions, recorded from various references *Tambora: Self et al., 2004; Foden, 1986, Santorini: Cadoux et al., 2014; Michaud et al., 2000, Changbaishan: Horn and Schmincke, 2000; Zou et al., 2010, Katmai: Hildreth, 1983; Westrich et al., 1991; Hammer et al., 2002; Hildreth and Fierstein, 2012, Krakatau: Mandeville et al., 1996, Huaynaputina: Costa, 2003; Thoutet et al., 2002; Takeuchi, 2011, Santa Maria: Rose et al., 1977; Williams and Self, 1983; Andrews, 2014, Pinatubo: Gerlach and McGee, 1994; Pallister et al., 1996; Rutherford and Devine, 1996, Vesuvius 79: Marini et al., 1998; Cioni, 2000; Scaillet et al., 2008, Cosigüina: Longpré et al., 2014b, Hudson: Kratzmann et al., 2009, Öræfajökull: Sharma et al., 2008; Selbekk and Trønes, 2007, El Chichón: Luhr et al., 1984; Macías et al., 2003; Luhr and Logan, 2002, Bezmyaniy: Neill et al., 2010; Shcherbakov et al., 2013, Mount St Helens (1480): Gardner et al., 1995, Fuji: Miyaji et al., 2011; Sato et al., 1999, Hekla (1947): Thorarinsson, 1967, Okatania: Palais and Sigurdsson, 1989; Walker et al., 1984; Sable et al., 2009, Bruce Houghton, personal communication, Colima: Saucedo et al., 2010; Crumney et al., 2014; González et al., 2002, Vesuvius: Belkin et al., 1998; Rolandi et al., 1993; Belkin et al., 1998; Marini et al., 1998, Komagatake: Takeuchi and Nakamura, 2001; Okumura, 2011, Hekla (1104): Devine et al., 1984, Mount St Helens (1980): Rutherford et al., 1985; Blundy and Cashman, 2001, Asama: Sato et al., 1989; Okumura, 2011; Yasui and Koyaguchi, 2004, Agung: Self and King, 1996, Usu: Tomiya and Takahashi, 2005; Tomiya et al., 2010, Mount St Helens (1800): Gardner et al., 1995, Askja: Sparks et al., 1981; Sigurdsson and Sparks, 1981, Mayon: Castillo and Newhall, 2004, Chris Newhall, personal communication, Shiveluch: Gorshkov and Dubik, 1970; Siebert et al., 2010, Augustine (1986): Roman et al., 2006, Galunggung (1822): Bronto, 1989, Augustine (1976): Johnson et al., 1996; Roman et al., 2006, Soufrière St Vincent: Roobol and Smith, 1975; Heath et al., 1998, Galunggung (1982): De Hoog et al., 2001; Kelley et al., 2004, Pelée: Martel et al., 1998; Pichavant et al., 2002; Martel et al., 2000, Fuego: Roggensack, 2001; Rose et al., 1978, Kelut (1919): Stehn and Coert, 1929, Merapi (2010): Costa et al., 2013; Surono et al., 2012, Reventador: Samaniego et al., 2008, Merapi: Surono et al., 2012, McCallum et al., 2014, Kelut (1990): Bourdier et al., 1997*

Volcano	Eruption Type	Year	SiO ₂ (wt%)	TiO ₂ (wt%)	Al ₂ O ₃ (wt%)	Fe ₂ O ₃ (wt%)	FeOT (wt%)	MnO (wt%)	MgO (wt%)	CaO (wt%)	Na ₂ O (wt%)	K ₂ O (wt%)	P ₂ O ₅ (wt%)
Tambora	Plinian	1815	58.3	0.5	19.3	-	4.5	0.2	1.4	3.2	5.4	6.3	0.4
Santorini	Plinian	1610 BCE	70.6	0.5	14.6	-	3.1	0.1	0.7	2.4	5.1	3.0	0.1
Changbaishan	Plinian	969	75.7	0.2	10.1	-	4.0	0.1	0.0	0.2	4.6	4.4	0.0
Katmai *	Plinian	1912	77.4	0.2	12.4	0.4	0.9	0.1	0.1	0.9	4.3	3.2	0.0
Krakatau	Plinian	1883	70.6	0.6	14.1	-	2.7	0.1	0.7	2.3	5.3	2.3	-

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*:Note values for Katmai eruption come from 'Layer A' (Rhyolite)

APPENDIX A. TABLES OF PLINIAN ERUPTION CRITERIA

Table A.3 – continued from previous page

Volcano	Eruption Type	Year	SiO ₂ (wt%)	TiO ₂ (wt%)	Al ₂ O ₃ (wt%)	Fe ₂ O ₃ (wt%)	FeOT (wt%)	MnO (wt%)	MgO (wt%)	CaO (wt%)	Na ₂ O (wt%)	K ₂ O (wt%)	P ₂ O ₅ (wt%)
Huaynaputina	Plinian	1600	65.6	0.6	16.8	-	3.7	-	1.8	4.2	4.5	2.7	-
Santa Maria	Plinian	1902	63.5	0.5	16.9	2.3	2.6	0.1	0.2	5.0	4.9	1.6	0.2
Pinatubo	Plinian	1991	64.6	0.5	16.5	4.4	-	0.1	2.4	5.2	4.5	1.5	0.2
Vesuvius	Plinian	79	56.0	0.2	21.9	-	2.3	-	0.2	2.8	6.2	10.2	-
Cosigüina	Plinian	1835	57.8	0.7	18.4	-	7.0	0.2	2.2	7.5	3.9	1.3	0.2
Hudson	Plinian	1991	51.6	2.2	15.8	11.7	11.2	0.2	4.3	8.3	4.4	1.3	0.5
Öræfajökull	Plinian	1362	72.7	0.2	13.2	-	3.3	0.1	0.0	1.0	5.0	3.5	0.0
El Chichón	Plinian	1982	68.0	0.3	15.9	-	1.6	0.1	0.3	2.1	4.6	5.1	0.0
Bezymiany	Plinian	1955	54.3	0.6	17.8	6.6	-	0.1	2.7	6.6	3.5	1.3	0.2
St Helens	Plinian	1480	67.7	0.6	16.5	-	3.7	0.1	1.2	3.8	4.9	1.3	-
Fuji	Plinian	1707	63.1	1.0	16.1	-	6.3	0.1	2.2	5.6	3.6	2.0	0.3
Hekla	Plinian	1947	59.0	0.8	16.2	3.3	6.5	0.2	2.4	5.7	4.3	1.1	0.5
Okatania	Plinian	1886	52.1	0.8	19.8	-	7.5	0.2	3.6	11.8	3.2	0.6	0.2
Colima	Plinian	1913	58.6	0.8	17.7	6.6	6.1	0.1	4.0	6.8	4.5	1.2	0.2
Vesuvius	Plinian	1631	52.2	0.6	20.1	-	5.0	0.1	3.0	8.6	2.9	5.9	-
Komagatake	Plinian	1929	54.0	0.8	18.5	-	10.5	0.3	4.0	8.5	3.0	0.3	0.2
Hekla	Plinian	1104	71.2	0.2	13.7	-	3.0	0.1	0.1	1.8	4.7	2.2	-
St Helens	Plinian	1980	73.0	0.4	14.6	-	2.1	0.1	0.5	2.4	5.2	2.2	-
Asama	Plinian	1783	61.6	0.6	16.0	6.6	-	0.1	4.0	6.6	2.9	1.4	0.2
Agung	Plinian	1963	57.0	0.8	17.7	6.6	6.8	0.2	4.4	8.1	2.6	1.4	0.3
Usu	Plinian	1663	73.9	0.2	14.5	2.6	-	0.2	0.3	2.2	4.9	1.2	0.1

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APPENDIX A. TABLES OF PLINIAN ERUPTION CRITERIA

Table A.3 – continued from previous page

Volcano	Eruption Type	Year	SiO ₂ (wt%)	TiO ₂ (wt%)	Al ₂ O ₃ (wt%)	Fe ₂ O ₃ (wt%)	FeOT (wt%)	MnO (wt%)	MgO (wt%)	CaO (wt%)	Na ₂ O (wt%)	K ₂ O (wt%)	P ₂ O ₅ (wt%)
St Helens	Plinian	1800	64.0	0.7	17.7	-	4.7	0.1	1.8	5.0	4.6	1.3	0.0
Askja	Plinian	1875	72.2	0.7	12.6	-	4.1	0.1	0.9	2.6	4.2	2.2	0.3
Mayon	Plinian	1814	54.5	0.8	18.0	8.8	-	0.2	4.1	8.4	3.4	1.0	0.3
Shiveluch	Sub-Plinian	1964	58.8	0.9	16.2	3.7	2.8	0.1	5.0	6.8	4.4	1.6	-
Augustine	Sub-Plinian	1986	75.4	0.6	12.1	-	2.3	-	0.6	1.7	4.1	2.4	-
Galunggung	Sub-Plinian	1822	53.3	0.8	18.4	8.8	-	0.2	4.8	9.3	3.1	0.5	0.2
Augustine	Sub-Plinian	1976	73.7	0.3	14.3	-	2.4	0.1	0.8	3.0	4.2	1.3	0.1
Soufrière	Sub-Plinian	1902	53.5	0.9	18.8	9.5	-	0.2	3.5	9.9	3.1	0.5	0.1
St Vincent													
Galunggung	Sub-Plinian	1982	52.0	0.9	18.6	9.2	-	0.2	5.7	9.9	2.9	0.5	0.1
Pelé	Sub-Plinian	1902	76.7	0.2	12.6	-	2.2	0.0	0.3	2.2	3.8	1.9	-
Fuego	Sub-Plinian	1974	50.2	0.9	19.3	-	11.4	-	5.0	9.0	3.4	0.8	-
Kelut	Sub-Plinian	1919	55.3	0.9	17.7	3.8	4.9	0.1	3.4	8.9	3.2	1.1	0.2
Merapi	Sub-Plinian	2010	55.3	0.8	19.2	-	8.0	0.2	2.5	8.5	3.8	2.1	0.3
Reventador	Sub-Plinian	2002	59.4	0.7	17.5	6.5	-	0.1	2.8	5.9	4.4	2.2	0.4
Merapi	Sub-Plinian	1872	52.5	0.9	18.9	-	9.5	0.2	3.5	9.6	3.1	2.0	0.4
Kelut	Sub-Plinian	1990	55.2	0.6	18.7	8.7	-	0.2	3.5	9.1	3.1	0.8	-

APPENDIX A. TABLES OF PLINIAN ERUPTION CRITERIA

Table A.4: Magmatic Properties of Plinian and Sub-Plinian eruptions, including Temperature, Pre-eruptive water content and Phenocryst content reported from the literature. Viscosities calculated using Giordano et al. (2008).

Volcano	Eruption Type	Year	Temperature (°C)	Pre-eruptive H ₂ O (wt%)	Phenocryst (%)	Viscosity of Magma (log Pa·s)	References
Tambora	Plinian	1815	760	6	10	4.0	Foden, 1986 Self et al., 2004
Santorini	Plinian	1610 BCE	875±25	3.5±0.5	12.5±2.5	5.0	Michaud et al., 2000 Cadoux et al., 2014
Changbaishan	Plinian	969	740±40	5.2	3	5.2	Zou et al., 2010 Horn and Schmincke, 2000
Katmai	Plinian	1912	825±25	3.8	1.5±0.5	4.5	Hildreth, 1983 Westrich et al., 1991 Hammer et al., 2002 Hildreth and Fierstein, 2012
Krakatau	Plinian	1883	940±60	4	11±4	4.0	Mandeville et al., 1996
Huaynaputina	Plinian	1600	850	6	18.5±1.5	3.9	Thouret et al., 2002 Costa, 2003 Takeuchi, 2011
Santa Maria	Plinian	1902	845	4.75±0.25	24.7	4.5	Rose et al., 1977 Williams and Self, 1983 Andrews, 2014
Pinatubo	Plinian	1991	780±10	6±0.5	47	5.7	Gerlach and McGee, 1994

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APPENDIX A. TABLES OF PLINIAN ERUPTION CRITERIA

Table A.4 – continued from previous page

Volcano	Eruption Type	Year	Temperature (°C)	Pre-eruptive H ₂ O (wt%)	Phenocryst (%)	Viscosity of Magma (log Pa·s)	References
Pinatubo (<i>cont</i>)							Pallister et al., 1996
Vesuvius	Plinian	79	825±25	6.25±0.25	4±1	3.2	Rutherford and Devine, 1996 Marini et al., 1998 Cioni, 2000 Scaillet et al., 2008
Cosigiüina	Plinian	1835	985±25	3.3±1.7	22.5±7.5	3.3	Longpré et al., 2014b
Hudson	Plinian	1991	943±27	2±1	2	3.0	Kratzmann et al., 2009
Öræfajökull	Plinian	1362	828	1.±0.9	2.3±1.5	6.9	Selbekk and Trø nnes, 2007 Sharma et al., 2008
El Chichón	Plinian	1982	825±25	5.5±0.5	24	4.6	Luhr et al., 1984 Luhr and Logan, 2002 Macías et al., 2003
Bezymiany	Plinian	1956	910±20	5	27	2.9	Neill et al., 2010 Shcherbakov et al., 2013
Mount St Helens	Plinian	1480	847	5.2 *	27	4.6	Gardner et al., 1995
Fuji	Plinian	1707	900 *	3.5±0.5	25±5	4.4	Sato et al., 1999 Miyaji et al., 2011
Hekla	Plinian	1947	-	-	5	-	Thorarinsson, 1967
Okatania	Plinian	1886	1100	2.25±0.75	5	1.7	Walker et al., 1984 Palais and Sigurdsson, 1989

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APPENDIX A. TABLES OF PLINIAN ERUPTION CRITERIA

Table A.4 – continued from previous page

Volcano	Eruption Type	Year	Temperature (°C)	Pre-eruptive H ₂ O (wt%)	Phenocryst (%)	Viscosity of Magma (log Pa·s)	References
Okatania (<i>cont</i>)							Sable et al., 2009
Colima	Plinian	1913	977	5.7	32	3.0	Bruce Houghton, personal communication González et al., 2002 Saucedo et al., 2010 Crummy et al., 2014
Vesuvius	Plinian	1631	825±25	1.2±0.2	40	6.5	Rolandi et al., 1993 Belkin et al., 1998 Marini et al., 1998
Komagatake	Plinian	1929	1040	3.1	33±4	2.7	Takeuchi and Nakamura, 2001 Okumura, 2011
Hekla	Plinian	1104	-	-	-	-	Devine et al., 1984
Mount St Helens (<i>Cont</i>)	Plinian	1980	930±10	4.6±1.1	40	5.0	Rutherford et al., 1985 Blundy and Cashman, 2001
Asama	Plinian	1783	900 *	2.3±0.9	50	5.1	Sato et al., 1989 Yasui and Koyaguchi, 2004 Okumura, 2011
Agung	Plinian	1963	1070±30	4.3	5.75±1.75	1.9	Self and King, 1996
Usu	Plinian	1663	780	6.4	3.5±2.5	4.6	Tomiya and Takahashi, 2005 Tomiya et al., 2010

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APPENDIX A. TABLES OF PLINIAN ERUPTION CRITERIA

Table A.4 – continued from previous page

Volcano	Eruption Type	Year	Temperature (°C)	Pre-eruptive H ₂ O (wt%)	Phenocryst (%)	Viscosity of Magma (log Pa·s)	References
Mount St Helens	Plinian	1800	893	5 *	30.5	4.3	Gardner et al., 1995
Askja	Plinian	1875	1000±10	3	0.5	3.9	Sparks et al., 1981 Sigurdsson and Sparks, 1981
Mayon	Plinian	1814	975	2.5 *	20±5	3.1	Castillo and Newhall, 2004 Chris Newhall, personal communication
Shiveluch	Sub-Plinian	1964	1080	1.5	25±5	3.4	Gorshkov and Dubik, 1970 Siebert et al., 2010
Augustine	Sub-Plinian	1986	912±35	3.46	25	5.4	Roman et al., 2006
Galunggung	Sub-Plinian	1822	-	-	-	-	Bronto, 1989
Augustine	Sub-Plinian	1976	876	4±0.5	30	5.4	Johnson et al., 1996 Roman et al., 2006
Soufrière St Vincent	Sub-Plinian	1902	1060	5	50	3.3	Roobol and Smith, 1975 Heath et al., 1998 Pichavant and Macdonald, 2007
Galunggung	Sub-Plinian	1982	1050	2±0.5	30	2.8	De Hoog et al., 2001 Kelley et al., 2004
Pelée	Sub-Plinian	1902	887.5±12.5	5.8±0.5	47	5.7	Martel et al., 1998 Martel et al., 2000 Pichavant et al., 2002
Fuego	Sub-Plinian	1974	1050	4±0.8	32	2.1	Rose et al., 1978

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APPENDIX A. TABLES OF PLINIAN ERUPTION CRITERIA

Table A.4 – continued from previous page

Volcano	Eruption Type	Year	Temperature (°C)	Pre-eruptive H ₂ O (wt %)	Phenocryst (%)	Viscosity of Magma (log Pa·s)	References
							Roggensack, 2001
Kelut	Sub-Plinian	1919	-	-	-	-	Stehn and Coert, 1929
Merapi	Sub-Plinian	2010	950	5	30	3.1	Surono et al., 2012 Costa et al., 2013
Reventador	Sub-Plinian	2002	1000	5	27.5±2.5	2.9	Samaniego et al., 2008
Merapi	Sub-Plinian	1872	1010	4.75	46	3.3	Surono et al., 2012 McCallum et al., 2014
Kelut	Sub - Plinian	1990	1000	-	-	-	Bourdier et al., 1997

*: Values calculated using the Lange et al. (2009) hygrometer / thermometer.

Viscosity calculated using Giordano et al. (2008).

Calculations assume that Temperatures, Pre-eruptive water content and Phenocryst content are average values.

Viscosity calculate using Giordano et al. (2008).

APPENDIX A. TABLES OF PLINIAN ERUPTION CRITERIA

Table A.5: Unrest period prior to Plinian and Sub-Plinian eruption data. The first incidence of increased seismicity or ash / steam emission from the volcanoes in the study prior to the onset of eruption, as recorded in literature. The seismic and phreatic ‘run-up’ times are defined as the number of days between this onset and the eruption paroxysm.

Volcano	Eruption Type	Eruption Paroxysm	First recorded seismicity	Seismic run up (days)	First recorded ash emission	Phreatic run up (days)	References
Tambora	Plinian	10/04/1815 19:00	-	-	Jan 1812	816	Kandlbauer, 2013
Katmai	Plinian	06/06/1912 13:00	01/06/1912	5	05/06/1912	1	Martin, 1913 Hildreth and Fierstein, 2012
Krakatau	Plinian	20/05/1883 14:00	10/03/1882	170	13/04/1882	136	Simkin and Fiske, 1984
Huaynaputina	Plinian	19/02/1600 17:00	15/02/1600	5	19/02/1600 17:00	0.02	Barriga, 1951 de Silva et al., 2000
Santa Maria	Plinian	25/10/1902 01:00	18/01/1902	280	24/10/1902	1	Rose, 1972 Williams and Self, 1983
Pinatubo	Plinian	15/06/1991 13:45	15/03/1991	93	02/04/1991	75	Wolfe and Hoblitt, 1996
Vesuvius	Plinian	24/10/0079 13:00	21/10/0079	4	24/10/0079	0.17	Sigurdsson et al., 1985 Scandone et al., 2008
Cosigüina	Plinian	21/01/1835 1500	19/01/1835	2	19/01/1835	3	Caldclough, 1836 Williams, 1952
Hudson	Plinian	12/08/1991	08/08/1991	4	08/08/1991	4	Kratzmann et al., 2009

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APPENDIX A. TABLES OF PLINIAN ERUPTION CRITERIA

Table A.5 – continued from previous page

Volcano	Eruption Type	Eruption Paroxysm	First recorded seismicity	Seismic run up (days)	First recorded ash emission	Phreatic run up (days)	References
			14:00				Kratzmann et al., 2010
El Chichón	Plinian	04/04/1982 01:35	01/03/1980	764	29/03/1982	6	Tilling, 2009
Bezymiany	Plinian	30/03/1956 17:11	29/09/1955	183	22/10/1955 06:30	160	Gorshkov, 1959
Fuji	Plinian	16/12/1707	03/12/1707	13	15/12/1707	2	Tsuya, 1955
Hekla	Plinian	29/03/1947 06:20	25/03/1947	4	06:20 06:20	1	Einarrson et al., 1967 Thorarinsson, 1967
Okatania	Plinian	10/06/1886 03:30	10/06/1886 00:30	0.13	10/06/1886 01:30	0.08 0.08	Thomas, 1888 Sable et al., 2006
Colima	Plinian	20/01/1913	-	-	17/01/1913	3	Saucedo et al., 2010
Vesuvius	Plinian	16/12/1631 07:30	Septemembr 1631	93	16/12/1631 07:00	0.02	Rosi et al., 1993 Scandone et al., 2008
Komagatake	Plinian	17/06/1929 19:00	17/06/1929 00:23	1 1	17/06/1929 03:00	0.67 0.67	Tsuya et al., 1930 Kôzu, 1934
Mount St Helens	Plinian	18/05/1980 08:32	20/03/1980	59	27/03/1980	44	Moore and Rice, 1984
Asama	Plinian	04/08/1783 10:00	None recorded	-	09/05/1783	87	Aramaki, 1956
Agung	Plinian	17/03/1963	16/02/1963	29	19/02/1963	26	Zen, 1964

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APPENDIX A. TABLES OF PLINIAN ERUPTION CRITERIA

Table A.5 – continued from previous page

Volcano	Eruption Type	Eruption Paroxysm	First recorded seismicity	Seismic run up (days)	First recorded ash emission	Phreatic run up (days)	References
					03:00		Zen and Hadikusumo, 1964
Askja	Plinian	28/03/1875 21:00	18/12/1874	101	01/01/1875	87	Sparks et al., 1981
Mayon	Plinian	01/02/1814 08:00	31/01/1814	1	-	-	de Mata, 1814 Aragoneses, 1814
Shiveluch	Sub-Plinian	12/11/1964 07:20	24/01/1964	292	12/11/1964	0.01	Gorshkov and Dubik, 1970 Belousov, 1995
Augustine	Sub-Plinian	29/03/1986 09:15	July 1985	272	27/03/1986	2	Swanson and Kienle, 1988
Galunggung	Sub-Plinian	08/10/1822 15:00	08/10/1822 13:30	0.06	08/10/1822 13:30	0.06	Katili and Sudradjat, 1984
Augustine	Sub-Plinian	25/01/1976 14:57	October 1975	103	22/01/1976 17:59	3	Johnston, 1978
Soufrière St Vincent	Sub-Plinian	07/05/1902 12:50	Feb 1901	446	06/05/1902	1	Anderson and Flett, 1902
Galunggung	Sub-Plinian	24/06/1982	05/04/1982 00:00	80	05/04/1982	80	Newhall et al., 1983 Gerbe et al., 1992
Pelée	Sub-Plinian	08/05/1902 08:02	23/04/1902 08:00	15	24/04/1902 20:45	13	Roobol and Smith, 1975 Westercamp, 1987

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APPENDIX A. TABLES OF PLINIAN ERUPTION CRITERIA

Table A.5 – continued from previous page

Volcano	Eruption Type	Eruption Paroxysm	First recorded seismicity	Seismic run up (days)	First recorded ash emission	Phreatic run up (days)	References
							Tanguy, 1994
Fuego	Sub-Plinian	17/10/1974 21:45	-	-	10/10/1974 04:00	8	Rose et al., 1978 McNutt and Harlow, 1983
Merapi	Sub-Plinian	05/11/2010 00:02	31/10/2009	5	26/10/2010	10	Surono et al., 2012 Komorowski et al., 2013
Reventador	Sub-Plinian	03/11/2002 09:12	06/10/2002	28 28	03/11/2002 05:30	0.17	Hall et al., 2004 Samaniego et al., 2008
Merapi	Sub-Plinian	17/04/1872	15/04/1872	2	-	-	Innocenti et al., 2013
Kelut	Sub-Plinian	10/02/1990 12:35	18/11/1989	85	10/2/1990 11:40	0.04	Wirakusumah, 1992 Badrudin, 1994 Lesage and Surono, 1995 Bourdier et al., 1997

Appendix B: Elicitation Questionnaire

What is the probability that the next eruption occurring at a volcano will be Plinian?

Prior probability estimate of the potential for magma accumulation at depth.

1. What is the probability that there is a volume of magma present at depth, capable of sustaining a Plinian eruption?

Plinian eruptions caused by magma accumulation at depth

What is the probability of observing a history of Plinian eruptions at a volcano that is inferred to have accumulated a large volume of magma?

Where P_e = the number of Plinian eruptions in the eruptive history over the total number of eruptions in the eruptive history

- Low (P_0) = $P_e \leq 0.1$
- Moderate (P_1) = $0.1 < P_e < 0.6$
- High (P_2) = $P_e > 0.6$ days

2. What is the probability of a volcano having a low, moderate or high P_e given there is a large volume of magma present at depth?

3. What is the probability of a volcano having a low, moderate or high P_e given there isnt a large volume of magma present at depth?

Repose time

Where T_{repose} = the repose time since the last $VEI \geq 4$ eruption.

APPENDIX B. EXAMPLE EXPERT ELICITATION QUESTIONNAIRE

- Low (P_0) = $T_{repose} \leq 100$
- Moderate (P_1) = $100 < T_{repose} < 1000$
- High (P_2) = $T_{repose} > 1000$ days

4. What is the probability of T_{repose} being low, moderate or high, given there is a large volume of magma present at depth?

5. What is the probability of T_{repose} being low, moderate or high, given there isn't a large volume of magma present at depth?

Excess Gas

Where E_{gas} = the factor of difference between the expected (petrologically derived) and the observed SO₂ released in an eruption.

- Low (P_0) = $E_{gas} \leq 2$
- Moderate (P_1) = $2 < E_{gas} < 10$
- High (P_2) = $E_{gas} > 10$ times more

6. What is the probability of a Plinian eruption, given there is a low, moderate or high excess gas phase?

7. What is the probability of a Plinian eruption, given there isn't a low, moderate or high excess gas phase?

8. What is the probability of a low, moderate or high excess gas phase, given there is a large volume of magma present at depth?

9. What is the probability of a low, moderate or high excess gas phase, given there isn't a large volume of magma present at depth?

Unrest

10. What is the probability of a Plinian eruption, given that the volcano is in unrest?

11. What is the probability of a Plinian eruption, given that there is no volcanic unrest?

Frequency of ash plumes NB: The definition of unrest here means that these criteria cannot occur unless the volcano is in unrest.

12. Given that the volcano is in unrest, what is the probability of seeing increasingly frequent ash plumes, which have decreasing eruptive strength?

Ash run up Where T_{ash} = the number of days of ash eruptions prior to the paroxysmal eruptive phase:

APPENDIX B. EXAMPLE EXPERT ELICITATION QUESTIONNAIRE

- Low (P_0) = $T_{ash} \leq 1$ day
- Moderate (P_1) = $1 \text{ days} < T_{ash} < 70$ days
- High (P_2) = $T_{ash} > 70$ days

13. What is the probability of observing low, moderate or high values for T_{ash} , given that the volcano is demonstrating unrest?

Seismic run up Where T_{seis} = the number of days of elevated seismicity prior to the paroxysmal eruptive phase:

- Low (P_0) = $T_{seis} \leq 10$ day
- Moderate (P_1) = $10 \text{ days} < T_{seis} < 100$ days
- High (P_2) = $T_{seis} > 100$ days

14. What is the probability of observing low, moderate or high values for T_{seis} , given that the volcano is demonstrating unrest?

Conduit seismicity 15. Given that the volcano is in unrest, what is the probability of observing seismicity delineating a zone beneath the volcano, from the surface to the magma chamber?

Rapidly accelerating shallow LP earthquakes 16. Given that the volcano is in unrest, what is the probability of observing rapid acceleration of shallow LP earthquakes?

Geological Map of Gede

APPENDIX C. GEOLOGICAL MAP OF GEDE