Caldera Unrest Management Sourcebook

S H Potter
B J Scott
G E Jolly

GNS Science Report 2012/12
July 2012

BIBLIOGRAPHIC REFERENCE

Potter, S. H.; Scott, B. J.; Jolly, G. E. 2012. Caldera Unrest Management Sourcebook, *GNS Science Report* 2012/12. 73 p.

S. H. Potter, GNS Science, Private Bag 2000, Taupo
B. J. Scott, GNS Science, Private Bag 2000, Taupo
G. E. Jolly, GNS Science, Private Bag 2000, Taupo

ISSN 1177-2425
ISBN 978-0-478-19892-8

CONTENTS

FIGURES

TABLES

KEYWORDS

Caldera unrest, volcano, Taupo, Okataina, Tarawera, Reporoa, Kapenga, Whakamaru, Mangakino, Rotorua, Ohakuri, Mayor Island, Raoul Island, Caldera Advisory Group.

EXECUTIVE SUMMARY

CALDERA UNREST MANAGEMENT IN NEW ZEALAND

The largest and most unpredictable of New Zealand's volcanoes are calderas, those which have erupted so explosively that the ground has collapsed to form large craters many kilometres across (such as Taupo and Rotorua volcanoes). These low frequency, high impact eruptions are preceded by geophysical and geochemical signals produced by the volcano as the magma forces its way through the ground, which can be interpreted by scientists to enable forecasting of the most likely future scenarios. The signals, collectively forming volcanic unrest episodes, occur far more frequently than there are eruptions. Volcanic unrest can be dangerous to nearby communities, even if there is no resulting eruption, as seen both in New Zealand and overseas. Caldera unrest can include damaging earthquakes, meters of ground deformation, hydrothermal explosions and poisonous gas emissions.

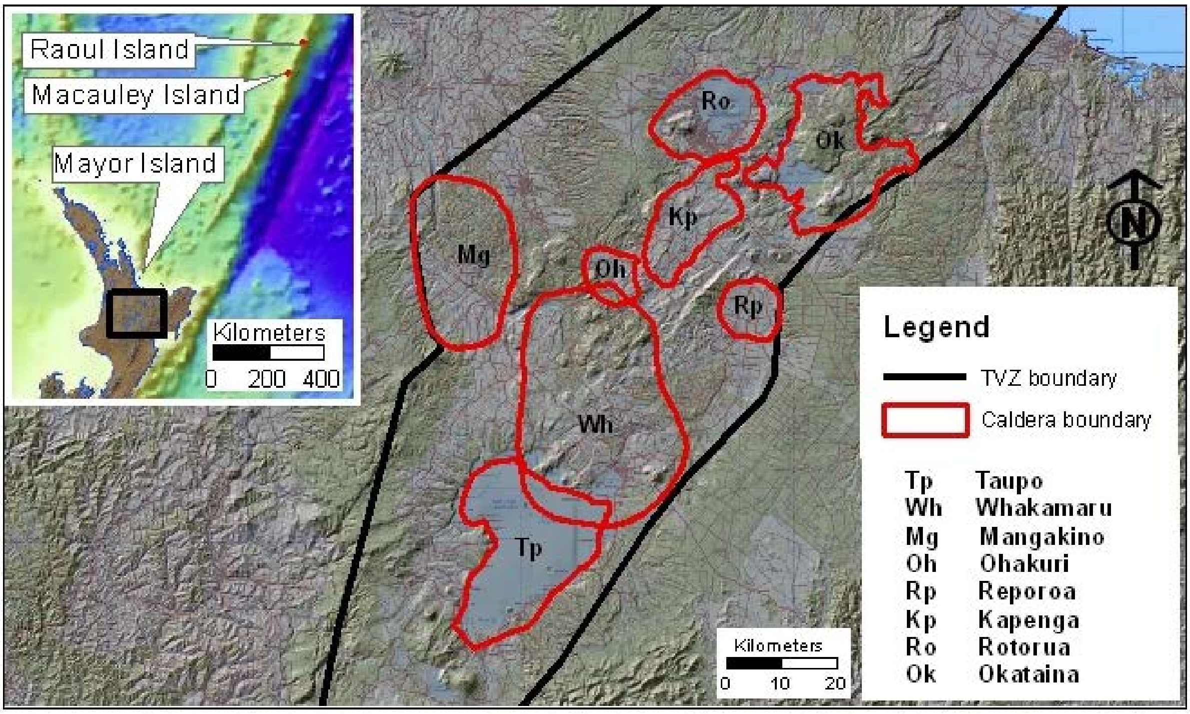

Figure ES1 Map of the eleven calderas in New Zealand.

New Zealand has eleven calderas (Figure ES1), many of which have shown frequent signs of unrest in the past 150 years. Two of these unrest episodes have resulted in eruptions (Tarawera (1886) and Raoul Island (2006)). Unrest has the potential to affect the local and national economies, the tourism industry, infrastructure of national importance, the psychological and physical health of the nearby residents and to undermine the trust between the community, media, emergency management officials and scientists. Caldera unrest episodes can last for days to decades, and must be carefully prepared for to avoid casualties and minimise the impact on society.

The calderas are monitored for signs of activity by GeoNet, an EQC-funded project run by GNS Science (Figure ES2).

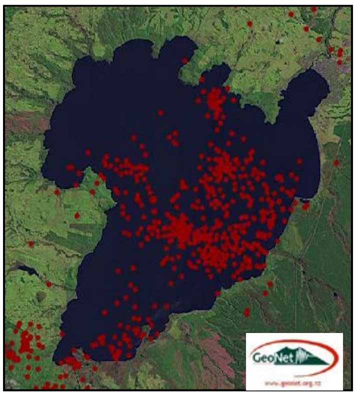

Figure ES2 An example of GeoNet monitoring data: earthquake epicentres at Taupo Caldera, April to September 2008.

Likely consequences of caldera unrest

Physical hazards

- *Ground shaking* from earthquakes (ranging from unnoticeable to damaging)
- *Ground deformation* (uplift and subsidence of millimetres to metres per day)
- *Gas poisoning* (potentially lethal in depressions)
- *Hydrothermal system changes* (including potentially large steam explosions)

Social effects

- *Psychosocial,* including public anxiety from months of earthquakes, and frustration and anger over economic impacts
- *Economic,* including likely impacts on tourism, local and national economies, insurance and investment industries

New Zealand caldera unrest episodes

During historical times New Zealand has experienced eruptions at calderas and numerous relatively small episodes of unrest[1]. We are yet to experience the magnitude of unrest seen internationally, in terms of episode duration and intensity of phenomena. Some examples are given in Table ES1.

Table ES1 Summary of caldera unrest episode examples in New Zealand and overseas.

Caldera name	Date of episode	Seismicity	Deformation	Hydrothermal/ other	Social impact & response	Eruption?
Okataina	1886	Felt seismicity started only about 1 hour before the eruption	No surface deformation is known	No unusual hydrothermal activity was noted. New features formed post eruption.	Unknown impact during unrest; 108 died in eruption	Yes (Tarawera)
Taupo	1895	M6 to 7.5 with 6 weeks of frequent aftershocks felt; liquefaction	Landslips; fissures; unknown if subsidence or uplift	0.6 m tsunami in lake; spring temperature changes	Chimneys collapsed; minor injuries; anxiety; self-evacuations	No
Taupo	1922	Thousands of earthquakes, max M6 over 10 months	Subsidence of 3.7 m at Whakaipo Bay; faulting; liquefaction	Changing hydrothermal activity at Mokai, Orakei Korako, Wairakei	Chimneys collapsed; tourism affected from misreporting; self-evacuations	No
Raoul Island	2006	5 days of earthquakes distant to volcano	No deformation recorded	No unusual hydrothermal changes	One fatality during eruption	Yes (hydro-thermal)
International examples of caldera unrest						
Campi Flegrei (Italy)	1982-1984	Hundreds of felt earthquakes, some large (<M4.2).	3.5 m uplift	Gas concentration increases	40,000 evacuated, damaged buildings	No
Long Valley (U.S.)	1979-1984	Swarms between 1982-4, 3 M6 quakes in 1 day	25 cm uplift in <6 months	No confirmed hydrothermal changes	Anger and frustration; economic & political impact	No

[1] Potter, S. H.; Scott, B. J.; Jolly, G. E. 2012. Caldera Unrest Management Sourcebook, *GNS Science Report* 2012/12. 74 p.

1.0 INTRODUCTION

Many of New Zealand's recently active volcanoes are situated near population centres, including our largest city, Auckland. There is a range in the type of volcanoes in New Zealand, including stratovolcanoes (see glossary at the back of this report for explanations of unfamiliar terms) such as Ruapehu, Tongariro/Ngauruhoe and Taranaki, volcanic fields such as Auckland, and calderas such as Okataina and Taupo (Wilson et al., 2004). Taupo Volcano is regarded as one of the most frequently active rhyolitic caldera systems in the world. The majority of the caldera volcanoes are in the Taupo Volcanic Zone (TVZ) which is located in the centre of the North Island of New Zealand (Figure 1).

The range of eruption styles between New Zealand's volcanoes and of the resulting landforms is largely due to the different chemistry of the magma. This also influences the frequency of eruptions, hazards and the extent of the area affected by an eruption. Rhyolitic magma has a high silica content compared to basaltic, andesitic or dacitic magma, it is viscous (doesn't flow easily) and can build up higher pressure before erupting. This is an influencing factor on why rhyolitic volcanoes don't erupt as often as less silicic (basaltic, andesitic, dacitic) volcanoes, but when they do erupt it can be in a very large, explosive way. Over the past 1.6 million years, at least 25 caldera-forming eruptions have occurred in the TVZ, most or all of which have caused widespread very dangerous pyroclastic (hot ash) flows (Wilson et al., 2009). New Zealand's calderas have erupted almost exclusively rhyolitic material (Wilson et al. 1984). Basaltic eruptions have also occurred in the TVZ, however they only make up <0.1% of the volume of deposits (Wilson et al., 1995). Andesitic and dacitic eruptions also occur in the TVZ, for example, Ruapehu, Tongariro and White Island erupt andesites, and Tauhara is formed from dacite, however dacite is rare. These eruptions tend to be smaller volume and impact a smaller area.

Part of the difficulty with understanding processes involved with rhyolitic eruptions is the very long gaps between eruptions (periods of quiescence), therefore very few rhyolitic eruptions have occurred worldwide during human existence. Of those that have, some have occurred in unpopulated areas. Therefore the precursors before rhyolitic eruptions have very rarely been witnessed, let alone recorded with modern monitoring equipment. Of the few witnessed eruptions at calderas which had previously been in quiescence, some erupted within only hours to days after the onset of noticeable unrest. However these volcanoes did not have monitoring networks such as in the TVZ, so it is likely a longer length of warning time will exist before eruptions here. Three of New Zealand's calderas have erupted post-settlement. At the Okataina Volcanic Centre, the Kaharoa eruption formed the summit domes of Mt Tarawera about 1314AD (Leonard et al., 2010) and Tarawera unusually erupted basalt in 1886. In the Kermadec Islands, five small eruptions have occurred at Raoul Island (1814, 1870, 1886, 1964 and 2006) as well as two potential eruptions at Macauley Island in 1825 and 1887. Over 110 fatalities have resulted from these eruptions, the vast majority from the 1886 Tarawera event.

This report summarises the current understanding of the eruption histories of New Zealand's 11 most recently active calderas – Raoul Island, Macauley Island, Mayor Island, Okataina, Rotorua, Kapenga, Reporoa, Ohakuri, Mangakino, Whakamaru and Taupo Calderas. Due to the large range of eruption sizes and styles from each of the calderas in the past, especially Taupo, it is extremely difficult to predict the size and style of the next eruption.

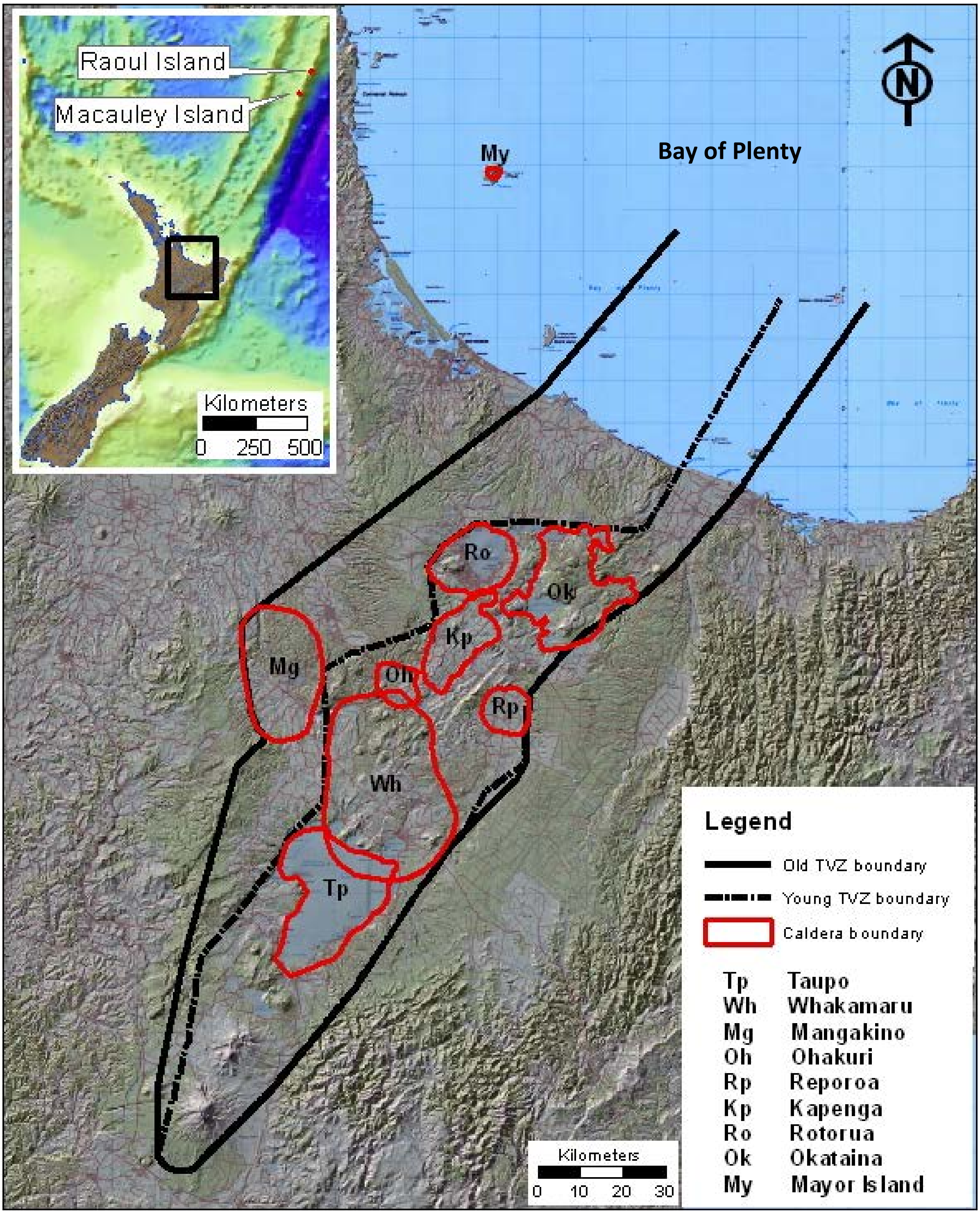

Figure 1 Map of the geomorphic boundaries of the eight calderas within the Taupo Volcanic Zone (TVZ). Three additional calderas lie outside the old TVZ boundary, Mayor Island in the Bay of Plenty and Raoul and Macauley Islands on the Kermadec Ridge, shown in the inset map. Based on Wilson et al. (2009) and Nairn (2002).

While eruptions at the calderas are relatively infrequent, volcanic unrest caused by magma and fluids moving underground and regional stress adjustment occurs more frequently. Most unrest episodes at calderas do not result in an eruption (Newhall & Dzurisin, 1988). Due to their frequency unrest episodes have been documented globally, including in New Zealand.

A summary of the known unrest episodes which have occurred in New Zealand is included in section 3 of this report for each caldera. Although unrest episodes are relatively frequent, they usually do not leave any trace in the geological record (except the occasional surface fault rupture, and large hydrothermal eruption deposits), so the knowledge is largely restricted to areas and times of human occupation. In New Zealand's case, this is reasonably limited, therefore to gain an understanding of what the unrest indicators may look like before future caldera eruptions in New Zealand, we must look overseas to countries with similar volcanoes and longer histories of settlement. This report provides descriptions of eruptions and unrest at rhyolitic calderas similar to New Zealand's including at Campi Flegrei (Italy), Long Valley (U.S.A.), Rabaul (Papua New Guinea), Chaitén (Chile), Aira (Japan), Taal (Indonesia), Novarupta (Alaska, U.S.A.) and Yellowstone Volcanic Centre (U.S.A.). Yellowstone is similar to the Taupo Volcanic Zone as both have had a similar discharge rate of magma in the past 2.2 million years and are a similar size, however the TVZ has had more frequent and smaller eruptions than Yellowstone (Houghton et al., 1995a).

Unrest phenomena may include seismicity, ground deformation and changes in the hydrothermal systems. These have the potential to be hazardous, damage buildings and infrastructure, and can result in psychosocial and economic impacts, all of which have occurred at Taupo Caldera in the past 160 years, as well as overseas (for example at Long Valley and Campi Flegrei Calderas). Unrest episodes can last for hours to decades. They need to be carefully managed by the CDEM sector, responding agencies, local and regional government, media, the public and scientists, even if there is no resulting eruption. Calderas with long periods of quiescence are particularly difficult to manage due to the public, media and public officials not fully recognising the hazards of the volcano, and the potential size and style range of any future eruption. There will be high levels of uncertainty for all groups, particularly as to the outcome of the unrest episode. This highlights the need for developing excellent pre-event interagency communication and cooperation as well as with the public and media.

In this report, we will discuss the physical, social and economic impacts of unrest and outline some of the implications for management of an unrest episode. We will not address in detail the eruption hazards from calderas. Volcanic eruption hazards, particularly for the Bay of Plenty region, have been outlined in Leonard et al., (2010).

1.1 What is a caldera?

A caldera is the depression in the ground formed by the withdrawal of underground magma (molten rock), causing the roof of the magma chamber to collapse. These depressions are usually formed during, but at a late stage in a coinciding large volcanic eruption from the caldera or a nearby vent. The natural bowl shape of these depressions can collect water, filling with lakes such as Lake Taupo and Lake Rotorua. Lakes within Okataina Caldera are broken up by lava domes formed in smaller eruption episodes well after the caldera-forming eruption itself.

Calderas can be created at a volcano with any type of magma, for example at basaltic volcanoes such as Kilauea (Hawaii) in 1750-1790; andesitic and dacitic stratocones such as Kuwae (Vanuatu) in ~1450 AD, Pinatubo (Philippines) in 1991 and Krakatau (Indonesia) in 1883; and rhyolitic volcanoes such as Taupo in 232AD (Lipman, 2000). The majority of New Zealand's calderas were formed during rhyolitic eruptions, which tend to form the largest calderas internationally.

The surface area of a caldera is many times larger than the individual vents that magma was erupted through. Small calderas (<5 km in diameter) can be formed during lava eruptions at andesitic and basaltic volcanoes, while calderas with a larger diameter (<75 km across) are usually formed during voluminous ignimbrite-forming (hot ash flow, called pyroclastic flow, deposits) eruptions (Lipman, 2000); generally ignimbrite-forming eruptions are rhyolitic. The size and geometry of calderas largely depends on the pre-existing host rock types, tectonic influences, magma chamber properties (such as size and shape) and the volume of material erupted (Lipman, 2000). Although calderas are usually formed from one or two very large eruptions, their magma system can also be the source of many smaller eruptions. This range in potential eruption size and style causes a large amount of uncertainty for decision makers during unrest.

The central TVZ can be viewed as one caldera complex akin to Yellowstone (e.g. Houghton et al., 1995 comparison). As such, unrest, hydrothermal eruptions, rhyolite magma eruptions and occasionally new calderas can occur outside of existing past caldera boundaries. This is very important to the interpretation of unrest, as the same uncertainties may exist as to the future outcome of unrest either inside or outside of known past calderas in the central TVZ.

1.2 What is caldera unrest?

Caldera unrest is simply volcanic unrest at a caldera volcano. Barberi and Carapezza (1996) define volcanic unrest as "the appearance on a dormant volcano of a multitude of anomalous phenomena indicative of possible eruptive reactivation (e.g. increased seismicity, ground uplift, physico-chemical changes in fumaroles and hot springs, increased heat flow, and changes in the gravimetric [and] magnetic…fields)". ‘Dormant’ generally means not-in-eruption, so in other words, volcanic unrest is signs that a sleeping volcano is starting to wake.

Volcanic unrest occurs when regional tectonic and/or volcanic processes cause magma (underground molten rock) and/or its fluids to interact with pre-existing rocks and sub-surface fluid. As the magma forces its way through the pre-existing rock, the rock can fracture causing earthquakes, and the ground surface may deform by millimetres to metres. Gas emitted by the magma (some of which can be hazardous) can be released at the ground surface, through the soil or at fumaroles (vents emitting gas and steam). Hydrothermal explosions, powered by steam, which can have been heated by the magma body, may occur at existing geothermal fields. Regional groundwater levels and spring temperatures can change due to alterations to the underground fluid systems, and the introduction of a new hot magma body.

If a monitoring or research programme exists, seismic, geodetic and other geophysical data like electromagnetic, magnetics and gravity may indicate the existence and size of an underground magma body.

1.3 Physical hazards during caldera unrest

Caldera unrest can be hazardous even if no eruption occurs. It is uncommon but possible that caldera unrest results in fatalities. A review of the literature indicates that gas poisoning has resulted in the most deaths at calderas during unrest (see section 1.3.3). Injuries can also occur from caldera unrest, such as from building failure if large earthquakes occur. However most episodes of caldera unrest do not result in any casualties, and many episodes show only one or two of the following unrest phenomenon, at varying levels of severity.

Volcanic hazards relating to eruptions are described in, for example, Johnston et al. (2011), Blong (1984) and the Yellow Book series (for example Nairn, 1991). Johnston and Nairn (1993) apply two eruption scenarios from Okataina Volcanic Centre and describe the potential impact on infrastructure and society. Becker et al., (2010) describe mitigation measures for volcanic hazards in New Zealand utilising land use planning. For social issues resulting from caldera unrest, refer to section 2.2.

1.3.1 Ground shaking

Earthquakes precede and accompany most, if not all volcanic eruptions. Volcanic processes generate a wide variety of seismic activity. These may be reflecting sub-surface processes like the movement of magma, signatures generated by eruptive activity or post-eruption readjustment (McNutt, 2000). Volcanogenic earthquakes are thought to rarely exceed magnitude 5 (Richter scale), but buildings within the volcano area may be subject to shaking damage (Johnston, 1997).

Earthquakes are the most common expression of volcanic unrest and eruptive activity (Newhall & Dzurisin, 1988). In some cases they are only detected if monitoring is adequate, while in other cases they will be felt locally and may cause some alarm. They can occur in swarms (many earthquakes occurring close together in time and space, usually of a similar size), or can be isolated events, affecting localised areas. Volcanic earthquakes largely have the same impacts as tectonic earthquakes. Ground shaking can cause building, structure and infrastructure damage or collapse, endangering lives. Earthquakes during eruptions have caused deaths at overseas volcanoes due to building collapse (or partial collapse) (Blong, 1984), however it is unusual for earthquakes during unrest to be large enough to cause fatalities. Collapsing brick chimneys can fall through building roofs; the rupturing of gas lines and electrical circuits may lead to a fire; and broken water pipes can cause flooding (Blong, 1984). Liquefaction (the upwelling of water and silt from ground shaking, as occurred in Christchurch during the 2010-12 earthquakes) can occur in areas with sand and gravel substrates, especially near low gradient waterways, if the earthquakes are of sufficient magnitude. Fault lines and cracks can be formed on the ground surface, potentially causing damage or destruction of buildings and underground services.

The GeoNet website (www.geonet.org.nz) contains a catalogue of historical earthquakes and a description of the seismic monitoring network in New Zealand. The main method of mitigation of earthquake damage is the enforcement of seismic building codes to protect structures against earthquake damage. Reinforcing chimneys, securing furniture, bracing structures and other seismic protection methods are recommended for areas surrounding volcanoes. Incorporating known fault lines into land use planning is also recommended (Kerr et al., 2003).

1.3.2 Ground deformation

Deformation (ground movement) at volcanic centres can occur as a result of magma moving beneath the ground surface, before, during and after eruptions. Movement of magma doesn’t necessarily result in an eruption, as it can often stall at depth and not reach the surface. As volcanoes lie in active tectonic environments they may also be influenced by regional deformation, such as rifting. Uplift or subsidence can cause damage to structures and infrastructure but is not directly life threatening. The deformation can range from millimetres to metres of uplift or subsidence, can affect a wide area, and may cause fissures (large cracks in the ground).

The most adverse effects of ground deformation are subsidence causing flooding (as seen in Taupo in 1922 and in Campi Flegrei over a number of centuries). Potential disruption to underground lifelines infrastructure in affected areas, such as gas, water, electricity and communication networks due to pipe or cable breakages can occur. Buildings, bridges, hydropower dams and geothermal power stations can be structurally damaged and roads cracked. Uplift and subsidence can cause flooding through altered water courses, subsidence below the water table or sea level or from hydropower dam and control gate failures.

There is very little which can be done to mitigate the effects of ground deformation. Depending on the form and location of ground deformation existing plans for flooding and landslides could be consulted, and areas deemed as dangerous could be evacuated and cordoned off as necessary. Gas, water and electricity pipelines in affected areas could be disconnected to minimise leakages and fires. Structures and infrastructure on areas of ground deformation should be regularly checked for safety.

1.3.3 Gas poisoning

Emissions of volcanic gases occur during eruptions, but are also common events between eruption episodes at many volcanoes and geothermal areas where they may be vented from the main crater, from fumaroles or diffusely through soil (Hansell et al., 2006). Volcanic gases include carbon monoxide (CO), carbon dioxide (CO_2), sulphur dioxide (SO_2), hydrochloric acid (HCl), hydrofluoric acid (HF), hydrogen sulphide (H_2S) and radon (Rn) (Parfitt & Wilson, 2008).

Documented health effects include discomfort and/or asphyxiation due to the accumulation of carbon dioxide (which is denser than air) in topographic lows (probably the most common life-safety hazard); deaths from hydrogen sulphide poisoning, primarily in geothermal areas; and respiratory effects (and occasionally deaths) from exposure to acidic sulphate aerosols formed from sulphur dioxide. High levels of volcanic gas in soil during unrest commonly causes areas of dying vegetation, as seen at Long Valley, Campi Flegrei and Rabaul during unrest episodes. During an eruption, the volcanic gas hazard is generally higher than during unrest. Wind tends to disperse gases to a point where they are at low concentrations and therefore are no longer hazardous, however they can still cause general discomfort. For a comprehensive review of health hazards from volcanic gases, refer to Hansell and Oppenheimer (2004) and the International Volcanic Health Hazard Network (www.ivhhn.org) (guidelines/gas pdf).

In Rotorua Caldera, about 14 people have been killed by gas poisoning in the past century. This is due to the geothermal nature of this region, and having buildings and hot pools sited over hot water bores and steaming areas. These deaths have mostly been attributed to asphyxiation by H_2S and CO_2 gases in low-lying, confined spaces such as geothermal spa pools, telecommunication trenches and workshop pits (Durand & Scott, 2005). This is in contrast to Rotorua's much less recent magmatic eruption history compared to the active (in terms of magmatic eruptions) Okataina and Taupo Calderas. In Cameroon in 1986, approximately 1700 people were killed by volcanic gas when a build-up of carbon dioxide was released suddenly from the waters of Lake Nyos (at the summit of a volcano) and flowed down the slopes to a nearby town (Baxter et al., 1989). At Rabaul in 1990 CO_2 gas killed six people who were collecting eggs in a small crater last active 50 years previously (as described by Rabaul Volcano Observatory bulletins, Global Volcanism Program website (www.volcano.si.edu)).

Mitigation of the volcanic gas hazard includes wearing face masks to provide protection for toxic gases, and ideally covering the eyes. During a volcanic gas hazard event, basements and other low lying, sealed areas may have to have restricted access or be entered with caution. Gas flux can be monitored from the active volcanic vent to provide an indication of the level of risk, and equipment can be installed in at-risk areas to monitor local gas concentrations if necessary.

1.3.4 Hydrothermal system changes

During caldera unrest, geothermal areas are susceptible to changes, including to the flow, temperature and chemistry of fumaroles and springs. A growing magma body can act as the catalyst providing additional heat to the overlying hydrothermal system, as well as releasing gas through it. As the temperature and/or pressure of the hydrothermal system changes, surface emissions can increase, and if large enough, can form small steam eruptions, called hydrothermal explosions (Browne & Lawless, 2001). Groundshaking (from tectonic or volcanic processes) can also change underground cracks and pressure systems, resulting in hydrothermal changes (Vandemeulebrouck et al., 2008). Most commonly no magma is erupted (Nairn, 2002; Nairn et al., 2005), however large hydrothermal eruptions at Rotomahana in an extension of the 1886 Tarawera eruption (Okataina Caldera) did include new magma (Nairn, 1979; Simmons et al., 1993).

As a complication, hydrothermal explosions can also occur without the influence of magma due to normal hydrothermal system processes, or to exploitation (drilling). For example, rainfall can influence the status of hydrothermal systems causing a hydrothermal explosion to occur. Another example of a cause of an increase in hydrothermal hazards not likely to be due to magma systems is inadequate borehole maintenance causing failure of the casing and hot fluids leaking at shallow depths, as occasionally occurs in Rotorua (Figure 2).

Hydrothermal systems are also located in the TVZ away from known calderas, e.g. at Kawerau, but unrest at these systems is likely to generate similar uncertainty with respect to possible future magmatic eruptions. This is because new rhyolite, and occasionally caldera eruptions occur outside of known past calderas from time to time.

Hydrothermal explosions occur fairly regularly in the Rotorua area (Scott et al., 2005). Recent significant events were at Kuirau Park in January 2001 and December 2006. Other significant hydrothermal eruptions have also occurred at Waimangu between the Okataina and Kapenga Calderas, with eruptions from Waimangu Geyser (1900-1904), Mud Rift (1906), Echo Crater (1915), Frying Pan Flat (1917), Frying Pan Lake (1924, 1973) and Raupo Pond Crater (1981) (Scott, 1994). Several of these resulted in fatalities. Geothermal fields near Taupo Caldera have also experienced several hydrothermal eruptions, such as in 2000 and 2001 (Bromley & Clotworthy, 2001). The eruption at Raoul Island in 2006, which caused one fatality, was a hydrothermal explosion. Hydrothermal eruption deposits have also been observed at Tahunaatara and Horohoro (near Kapenga), and at the Ongaroto, Ngatamariki, Rotokawa and Kawerau geothermal fields (Leonard et al., 2010).

Figure 2 St. Faith's church (Rotorua) hydrothermal explosion deposits in 2011, caused by the leakage of a bore, seen below the window on the right-hand side. Photo by A. Somerville.

Mitigating against hydrothermal explosions is difficult due to the unpredictability of the hazard. They tend to only occur in established geothermal fields, which generally already have restricted public access. Isolating dangerous fumaroles, springs and hydrothermal craters will minimise the risk. Restricting the development of land in geothermal fields is likely to minimise casualties and damage.

2.0 CALDERA UNREST MANAGEMENT

Caldera unrest management is a challenging and relatively underdeveloped field combining volcanology, local and regional government, lifelines, media, emergency management and the public. Bridging the gap between scientists and decision making officials for effective unrest management becomes a vital issue potentially affecting lives, property, infrastructure and economies. Future unrest episodes in New Zealand will occur, and scientists and emergency decision makers need to be as prepared for this as possible. Being aware of the past behaviour of the volcanoes will benefit these hazard management processes as an indication of possible future activity.

Eruptive and unrest histories from New Zealand and worldwide calderas are presented in sections 3 and 4 of this report and some scenarios are developed in Appendix 1. Data from these sections are used in the following discussions of caldera unrest and the management issues. It is also important to keep in mind that the central TVZ can be considered a single caldera complex, so semantics of whether unrest is within one known existing caldera or another, or inside or outside of a known caldera at all, should not necessarily or unduly colour the interpretation of what might happen in the future.

The level of risk encountered during caldera unrest is an interaction between the hazard and the exposure of people and structures. During some cases of caldera unrest, both of these factors have high levels of uncertainty and variability. To accurately estimate the level of hazard, any assessment has to examine the probability that unrest may result in a volcanic eruption. This needs to be defined in terms of magnitude, location, timing, style of potential eruption, and the probability of occurrence. To predict all of these factors with any accuracy is very difficult. The number of people exposed to the hazard can also vary according to the season in tourist locations. At the time of the Mammoth Lakes unrest episode in 1982-84, the population of the town was 5,500, while the surrounding area had almost 20,000 permanent residents. This number varied greatly during the winter season, as the number of skiers at Mammoth Mountain was estimated to be 1.2 million people per year, or approximately 15-20,000 people per day on weekends and holidays around the time of the unrest – a very significant increase in transient population to effectively manage (Mader & Blair, 1987). Taupo town's population of over 20,000 people can also multiply during large sporting events and the summer months. This variability in population as well as the future population size affects the level of risk and needs to be considered when planning for caldera unrest.

A high level of coordination between the scientists, response agencies and the public is needed to effectively manage an unrest crisis. Calderas with long periods of quiescence are particularly difficult to manage due to the public, officials and media not recognising the hazards of the nearby volcano, and the potential for eruptions.

GNS Science currently has the legal and contractual responsibility to monitor New Zealand's volcanoes through the Earthquake Commission (EQC) funded GeoNet project, and to communicate the levels of activity to the CDEM sector, media and public.

Natural and technological hazards in New Zealand are managed using the Resource Management Act 1991 (RMA), Building Act 2004 and Civil Defence and Emergency Management Act 2002 (CDEM Act). Local and regional government identify and rank the hazards and develop response plans around these. An indication of how volcanic hazards and caldera unrest are ranked at the various councils is summarised in Table 1. In parallel with this system, four volcano advisory groups have been established to improve management of volcano hazards in New Zealand. They are the Central Plateau Volcanic Advisory Group (CPVAG) for the Tongariro National Park volcanoes, Taranaki Seismic and Volcanic Advisory Group (TSVAG), Auckland Volcano Science Advisory Group (AVSAG) and the Caldera Advisory Group (CAG) with a focus on the central North Island calderas.

Table 1 A summary of how volcanic hazard is currently recognised by the six regional councils that list volcano hazards in their regional plans. Note there is a mix of 1st and 2nd generation plans represented here.

Council	Volcano threat	Likelihood	Consequence	Rating	Ranking
Northland	Local Volcano	Rare	3.4	Moderate/High	8/19
	Distal Volcano	Possible	2.2	Moderate	13/19
Auckland	Distal Volcano	Likely	Major	Very High	3/37
	Local Eruption	Rare	Catastrophic	High	8/37
Waikato	Caldera Unrest			Very High	2/20
	Ashfall source within region			High	5/20
	Ashfall source outside of region			Moderate	7/20
	Geothermal eruption			Low	18/20
Bay Of Plenty	Eruption Local Source			Very High	2/19
	Distal source			High	10/19
	Geothermal eruption			High	11/19
Taranaki	Local eruption	Certain	Major	Extreme	4/10
Horizons	Volcanic activity at Ruapehu				7/15
Hawke's Bay	Ashfall			High	15/38

2.1 Volcanic Alert Levels, Aviation Colour Code and Volcanic Alert Bulletins

The activity at volcanoes is communicated to the emergency management decision makers, media and public using Volcanic Alert Level (VAL) systems. In New Zealand, the VAL system describes the current state of activity and ranges from 0 (normal background activity) to 5 (major eruption in progress) (Table 2). All of the calderas in New Zealand are currently classified as *reawakening volcanoes,* and use the right-hand side of the VAL table, except for the recently active Raoul Island Caldera in the Kermadecs. The VAL system is defined in the Guide to the National Civil Defence Emergency Management Plan by the Ministry of Civil Defence and Emergency Management (2006) (found on the civildefence.govt.nz publications webpage, in section 19.4.2).

Table 2 Volcanic Alert Levels in New Zealand. All calderas in New Zealand, except for Raoul Island in the Kermadecs, currently use the 'reawakening volcanoes' side of the table. From the Guide to the National Civil Defence Emergency Management Plan (Ministry of Civil Defence and Emergency Management, 2006).

Frequently active cone volcanoes White Island, Tongariro-Ngauruhoe, Ruapehu, Kermadecs		**VOLCANIC ALERT LEVEL**	**Reawakening volcanoes** Northland, Auckland, Mayor Island, Rotorua, Okataina, Taupo, Egmont/Taranaki	
Volcano status	**Indicative phenomena**		**Indicative phenomena**	**Volcano status**
Usual dormant, or quiescent state	Typical background surface activity, seismicity, deformation and heat flow at low levels.	**0**	Typical background surface activity; deformation, seismicity, and heat flow at low levels.	Usual dormant, or quiescent state.
Signs of volcano unrest	Departure from typical background surface activity.	**1**	Apparent seismic, geodetic, thermal or other unrest indicators.	Initial signs of possible volcano unrest. No eruption threat.
Minor eruptive activity	Onset of eruptive activity, accompanied by changes to monitored indicators.	**2**	Increase in number or intensity of unrest indicators (seismicity, deformation, heat flow and so on).	Confirmation of volcano unrest. Eruption threat.
Significant local eruption in progress	Increased vigour of ongoing activity and monitored indicators. Significant effects on volcano, possible effects beyond.	**3**	Minor steam eruptions. High increasing trends of unrest indicators, significant effects on volcano, possible beyond.	Minor eruptions commenced. Real possibility of hazardous eruptions.
Hazardous local eruption in progress	Significant change to ongoing activity and monitoring indicators. Effects beyond volcano.	**4**	Eruption of new magma. Sustained high levels of unrest indicators, significant effects beyond volcano.	Hazardous local eruption in progress. Large-scale eruption now possible.
Large hazardous eruption in progress	Destruction with major damage beyond volcano. Significant risk over wider areas.	**5**	Destruction with major damage beyond active volcano. Significant risk over wider areas.	Large hazardous volcanic eruption in progress.

The Aviation Colour Code (Table 3) is defined in International Civil Aviation Organization (ICAO) documents, and is used by the Civil Aviation Authority in New Zealand to alert the aviation industry to changes in the status of volcanoes within the coverage of Wellington Volcanic Ash Advisory Centre (VAAC), which includes a large area of the southwest Pacific. Restrictions on the use of airspace during a volcanic eruption using the New Zealand Volcanic Ash Advisory System (VAAS) is outlined in Lechner (2009). The VAAS is the local enhancement of the International Airways Volcano Watch System. GNS Science, MetService and the Airways Corporation of New Zealand provide input into the VAAS (Scott & Travers, 2009).

Table 3 The ICAO Aviation Colour Code for volcanic activity.

ICAO Colour code		Status of activity of volcano
GREEN		Volcano is in normal, non-eruptive state. *or, after a change from a higher alert level*: Volcanic activity considered to have ceased, and volcano reverted to its normal, non-eruptive state.
YELLOW		Volcano is experiencing signs of elevated unrest above known background levels. *or, after a change from higher alert level*: Volcanic activity has decreased significantly but continues to be closely monitored for possible renewed increase.
ORANGE		Volcano is exhibiting heightened unrest with increased likelihood of eruption. *or,* Volcanic eruption is underway with no or minor ash emission *[specify ash-plume height if possible]*.
RED		Eruption is forecasted to be imminent with significant emission of ash into the atmosphere likely. *or,* Eruption is underway with significant emission of ash into the atmosphere *[specify ash-plume height if possible]*.

Volcanologists at GNS Science have the responsibility of setting the VAL and Aviation Colour Codes for New Zealand's active volcanoes. Responding agencies in New Zealand are notified of changes in volcanic activity, including changes to the VAL and Aviation Colour Code, by the dissemination of Volcanic Alert Bulletins which are issued by GNS Science. Volcanic Alert Bulletins are also issued without a change in VAL to provide additional information such as ashfall forecasts. This information can be used by the responding agencies to help determine decisions and responses. For up to date information on the current status and alert levels for the calderas, visit the GeoNet website (http://www.geonet.org.nz/volcano/).

2.2 Social consequences of caldera unrest

Social effects of caldera unrest include impacts on the national and local economies through the decline of tourism, investments and the real estate industry; media speculation and misreporting; temporary psychological distress, particularly from constant earthquakes; and self-evacuations (Johnston et al., 2002). Mistrust of the scientists and emergency management decision makers from the lack of timely information and high levels of uncertainty can arise during unrest episodes, such as occurred at the town of Mammoth Lakes at Long Valley caldera (California) during unrest in 1982-84 (Mader & Blair, 1987). The outcome of the local elections at Mono County (an area of which includes Long Valley caldera and Mammoth Lakes) in 1983 may have been affected by these factors.

2.2.1 Psychosocial

Initial reactions to the volcanic unrest episode are likely to include fear, confusion and denial, as seen at Mammoth Lakes, Long Valley Caldera in 1982 (Mader & Blair, 1987) and in Pozzuoli during Campi Flegrei (Italy) unrest in 1970 when an evacuation order for 3,000 people was issued (Barberi et al., 1984). Repeated earthquakes can have a detrimental effect on the community, leaving the people on edge and waiting for the seismic swarm to cease so they can respond to damage. Unrest causes a heightened feeling of uncertainty in the community as it is unknown whether the unrest will escalate and culminate in an eruption or die away. This stops life from being lived as it normally would for potentially long periods of time. Education systems may be closed, and some members of the community may leave to gain a sense of normalcy elsewhere. This decreases the workforce, potentially having a flow-on effect on business closure and the local economy.

Perceived effects of the unrest on the community and economy can tempt the public officials and politicians to put pressure on scientists to lower alert levels, or remove the label of volcanic unrest from the situation. Tensions between the two groups can heighten until mistrust occurs. This occurred at Mammoth Lakes, Long Valley Caldera in an attempt by a few of the officials and local business owners to lessen the impact on the tourism and local investment industries (Mader & Blair, 1987). Mistrust of the scientists by the officials and public can cause action delays in situations indicating an evacuation should take place. A high level of interagency communication and public information management is required during caldera unrest to minimise the risk of these issues from occurring.

There is a large demand for information from public officials and scientists by the public and media during unrest, as seen during the 1983-85 unrest episode in Rabaul. This resulted in special arrangements to be made, including establishing a regular newsletter and a Public Information Unit to fulfil this need (Lowenstein, 1988). Daily information meetings were well attended by the public during the 1983 seismic swarm at Long Valley (Mader & Blair, 1987).

2.2.2 Economic

The economic effects of a long period of caldera unrest are varied, and rely on factors such as the duration; magnitude of activity; types, strength and flexibility of businesses; and degree of uncertainty (Johnston et al., 2002). The increase in business uncertainty disrupts the local economy, which can last for weeks to decades. Preparing for a volcanic crisis event can reduce the overall impact on the local and national economies (Shearer Consulting Ltd. & Market Economics Ltd., 2008). The economic consequences of unrest at New Zealand's calderas are not yet well quantified.

The local and national tourism industry can by adversely affected, as experienced by Taupo during and immediately after the 1963-64 episode of unrest (Johnston et al., 2002), and in the ski-season of 1982-83 at Mammoth Lakes, Long Valley Caldera (Mader & Blair, 1987). In the latter example, the effect of unrest on the tourism industry, while easily blamed on the volcanic unrest, is hard to prove or measure due to contributing circumstances including the national recession, coincidental poor weather, and perceived overbuilding at Mammoth Lakes during the early 80's episode (Mader & Blair, 1987). Premature business closure and self-evacuations are likely to affect the image of the town and the confidence of tourists in visiting. Encouraging business owners to remain open during unrest can mitigate this, providing buildings have been deemed safe after earthquakes. The effect on tourism may be short lived if the unrest declines, as shown by the almost record ski season of 1983-84 at Mammoth Lakes, despite the unrest earlier in the year (Mader & Blair, 1987). A marketing campaign by the businesses of Mammoth Lakes in 1984 appeared to be a success in further boosting tourist numbers (Mader & Blair, 1987).

The investment market at Mammoth Lakes was seen to be hit harder by the caldera unrest than the tourism industry (Mader & Blair, 1987). This appeared to be due to the perceived risk on short-term visitors being less than the "constant threat" on long-term property investments. The decline in the real estate market was blamed on the volcanic hazard (Mader & Blair, 1987).

The insurance industry is likely to be affected during caldera unrest, largely due to the repeated and potentially damaging earthquakes. Changes by insurance agencies can include not reinsuring the previously insured once the standing annual contract expires.

Insurance companies may also cancel their cover giving 7-days notice, or they may change what the insurance includes, for example, not cover volcanic hazards. New Zealand's Earthquake Commission (EQC) (http://canterbury.eqc.govt.nz/faq) covers earthquake damage for a portion of the house and contents, provided the owner also has insurance with a private insurance company. After the 1983-85 Rabaul unrest episode, building insurance was restricted and had a high cost, resulting in a lack of finance from lending institutions (Lowenstein, 1988).

2.3 Mitigation measures for caldera unrest

In addition to the considerations mentioned in the above psychosocial and economic sections, a number of actions are recommended to take place during caldera unrest (especially if they haven't already occurred during quiescence). As previously mentioned, the difficulty with caldera unrest is the high uncertainty in the outcome, therefore the range of outcomes must be prepared for. This includes preparing for long periods of damaging unrest, anticipating the needs resulting from potential unrest phenomena (for example preparing to restrict access to hazardous locations, source engineers and equipment for building and infrastructure damage inspections and have cleanup crews ready to clear landslips), preparing for volcanic eruptions of various scales and associated recovery plans. In essence, the CDEM sector needs to be prepared, educated and activated at an appropriate time.

2.3.1 Planning

Emergency plans may need to be activated to prepare the surrounding areas. For example, hazard maps should be drawn and the most vulnerable areas from various unrest hazards (including liquefaction, faulting and flooding) identified, as well as for eruption hazards (for example, Scott & Nairn (1998), Figure 3 below). Evacuation plans should be created and streets and bridges assessed for potential obstacles. Alternative escape routes from isolated communities may need to be created, as occurred at Mammoth Lakes, California in 1983 (Mader & Blair, 1987). Exercises in the form of evacuation drills were carried out during Rabaul's 1983-84 unrest episode, and may have contributed to the successful evacuation of the town 10 years later during the eruption (Finnimore et al., 1995). Preparing for eruption hazards (including pyroclastic flows, tephra falls and lava flows) also needs to take place for a range of scenarios. For further information on preparing for a volcanic crisis, particularly in the Bay of Plenty, refer to Johnston et al., (1996).

2.3.2 Education and communication

Public education and communication is vital during unrest, particularly for events with a wide range of potential outcomes. Information sheets/flyers can be pre-prepared during quiescence and ready to be issued when required. A media plan should be created as the media can be a powerful tool or enemy in these situations. If problems are encountered involving inaccurate media releases, prepaid advertising can be utilised, or a private news-sheet published and distributed, as done in Rabaul (Lowenstein, 1988). Incorrect international reporting during the 1922 unrest episode at Taupo Caldera caused a perceived impact on the Rotorua tourism industry (Evening Post, 7 July 1922). In addition, during the same episode, a San Francisco source reported there had been 60 deaths due to the earthquakes in Taupo, when in fact there had been none, and it was greatly feared by the New Zealand Government and Tourism Department that this would have a detrimental effect on the number of visitors from this area (Evening Post, 10 August, 1922). A publicity officer was appointed for the Government, and in the future “such misrepresentations should be immediately corrected” (Evening Post, 10 August, 1922).

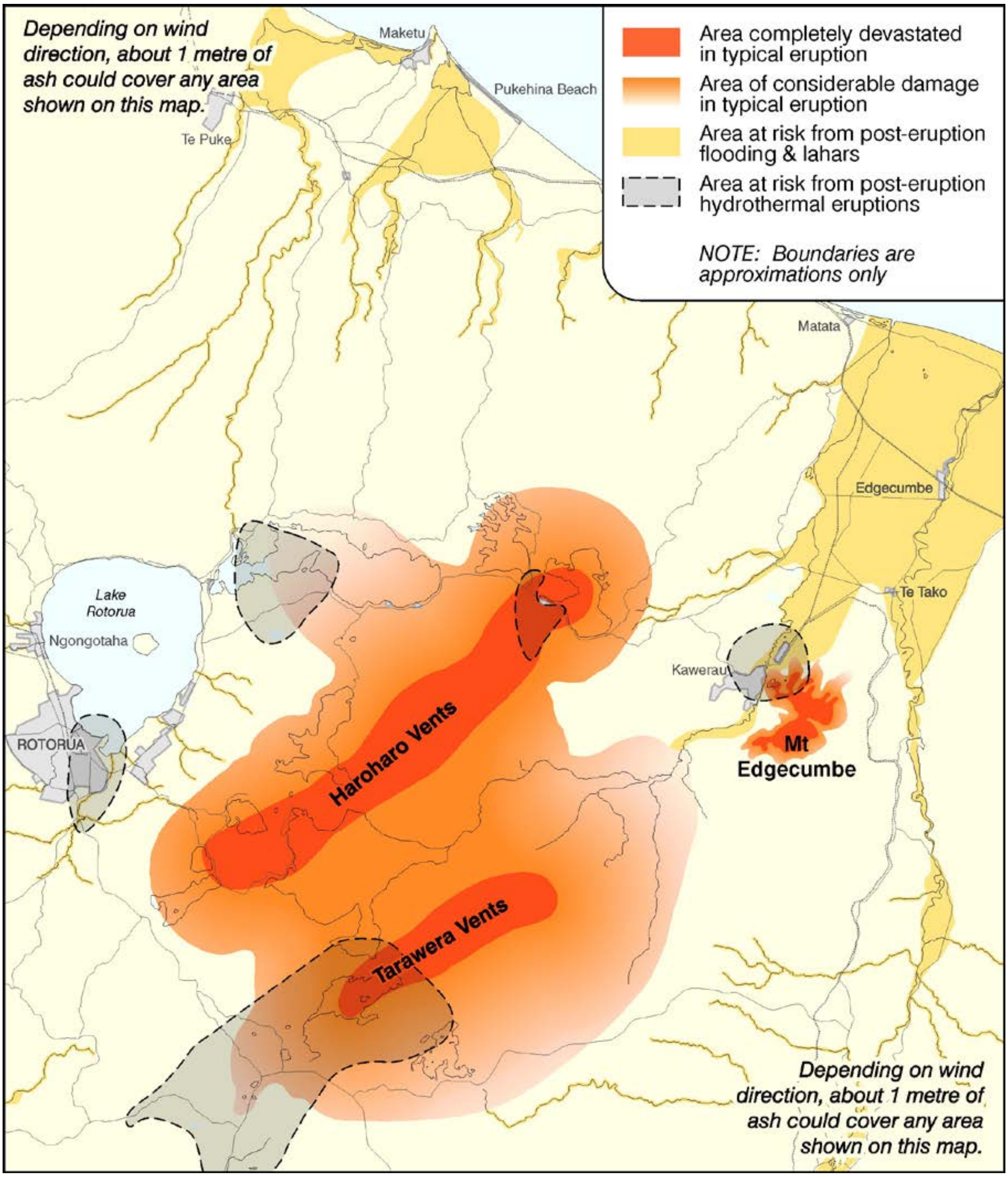

Figure 3 An example of a volcanic eruption hazard map for the Okataina Volcanic Centre (Scott & Nairn, 1998). Areas with orange and red shading have a higher probability of damage, depending on vent location.

2.3.3 Economic

The impact on the local economy may also be lessened with robust but flexible business continuity plans. This will enable the local economy to continue to function throughout even extended periods of unrest, as well as in the period of recovery afterwards. This is particularly the case for larger companies who employ many in the community. Rural, agricultural and industrial sector requirements need to be considered, including hydropower and geothermal power stations. Developing recovery plans for the range of eruptive and non-eruptive scenarios is likely to be beneficial. Lessons learnt elsewhere could be incorporated, including from the repeated Canterbury earthquakes of 2010-12.

2.3.4 Future research

The unrest histories of New Zealand's calderas are largely unknown – only Taupo's history has been (recently) completed. Further research into these histories, the hazards of volcanic unrest and caldera eruptions and increasing awareness of international unrest and eruption events will improve our state of knowledge and capacity to react to future events in New Zealand. Research into the effects of unrest on the community and local and national economies, as well as methods of mitigation will contribute towards the resilience of New Zealand's population who live with restless calderas.

2.3.5 General preparation

As can be seen from the suggested examples of actions which need to occur during unrest, it is going to be very helpful to have everything which can possibly take place during quiescence completed, so that during the event adequate time is left for addressing the media and public, and for focussing on arising issues.

The various sectors, including those related to lifelines, health, agriculture and the environment, as well as individual households can prepare for a caldera unrest event in a similar way to preparing for any natural hazard. In particular, refer to advice given by MCDEM for earthquake and volcanic eruption hazards.

3.0 NEW ZEALAND'S CALDERAS – ERUPTIONS AND HISTORICAL UNREST

The North Island of New Zealand is situated next to a plate tectonic boundary with the Pacific Plate in the east subducting beneath the Australian Plate in the west, with subduction starting from offshore east of the island. The Taupo Volcanic Zone (TVZ) stretches approximately 300 km in length and up to 60 km in width from Mt Ruapehu in the southwest to White Island in the northeast (Houghton et al., 1995a) (Figure 1). It is an area of thinner crust and high heat flow which is more susceptible to large-scale volcanism (Bibby et al., 1995). The middle section of the TVZ contains the Taupo, Whakamaru, Mangakino, Reporoa, Ohakuri, Kapenga, Okataina and Rotorua calderas. Of these calderas, only Taupo and Okataina have erupted in the past 2,000 years, and only Taupo has produced a large caldera-forming eruption in that timeframe. The remainder have not erupted for a very long time, so while the possibility of a future eruption remains, it is less likely that these will erupt than the more frequently and recently active calderas. The TVZ also accommodates a large number of fractures and faults (including the active Taupo Fault Belt) and numerous geothermal fields.

Mayor Island volcano lies just outside of the TVZ boundary (Figure 1). The underlying tectonic plate structure is different to that of the TVZ caldera volcanoes, which influences the magma chemistry, and therefore its eruption styles.

Raoul and Macauley Islands are part of the Kermadec Islands, which lie 750 – 1000 km northeast of the coast of New Zealand (inset of Figure 1). The Kermadec Islands are predominately volcanic, and were formed by the continuation northward of the subduction processes occurring beneath the North Island of New Zealand. Several active submarine volcanoes are also know in the area between the Bay of Plenty coast and the Kermadec islands. A large sea-raft of pumice (Loisels Pumice) was deposited on New Zealand's eastern coastline from two eruptions at approximately 1000-1500 years Before Present (yrs BP) and 650 yrs BP (Shane et al., 1998). The source of these eruptions is thought to be from volcanoes (potentially calderas) in the Kermadec ridge area (Shane et al., 1998). This deposit indicates that future large-scale pumice rafts could affect the coast and ports of New Zealand, particularly on the east coast.

New Zealand's calderas have displayed an enormous range in eruption sizes. For Taupo caldera alone, the eruption size has varied by four or five orders of magnitude in the past 27,000 years (Wilson, 1993). This causes uncertainties in the prediction of future eruptive activity. Further uncertainty exists due to the reliance on the geological record for knowledge of past eruptions. Many small eruptions are likely to be missing or poorly recorded in the geological record as these leave only thin ash deposits (if any at all), which can later be eroded away, destroying any record that the eruption occurred. This is particularly the case for the calderas in the middle of the TVZ which have not been active for hundreds of thousands of years. Therefore many more small eruptions have occurred at New Zealand's calderas than are known.

During an eruption or potential eruption at any of these central TVZ volcanoes, the airspace closure may impact domestic flight paths crossing the area, although due to the dominant wind direction being westerly, it is likely international flight paths will be largely unaffected and the ash will be deposited in the eastern North Island and the Pacific Ocean.

As many of the TVZ calderas are bordered by active networks of faults and geothermal fields, and situated on a back-arc rift formed by a subduction zone, it is difficult to determine the cause of seismic and deformation activity within and surrounding each caldera. Earthquakes and deformation could occur from the tectonic plate collision, regional extension, the local fault belts, geothermal activity or a magmatic intrusion, with the magmatic intrusion the most potentially hazardous. In addition, geothermal systems have their own processes causing fluctuations in activity without further input by volcanic or tectonic processes. Therefore increases in activity at the geothermal fields do not necessarily indicate caldera unrest.

The Earthquake Commission (EQC) funded GeoNet project run by GNS Science manages a GPS, seismic and geochemical (at hydrothermal areas) monitoring network to observe activity at the calderas (Scott & Travers, 2009). In the following section the monitoring, eruptive and unrest history of each caldera is summarised and discussed in more detail.

3.1 Raoul Island

Raoul Island is part of New Zealand's territory, therefore the government is obligated to assess and reduce risks for this area. The island has an area of 30 km^2 and is the surface portion of a volcano approximately 35 km by 20 km in size, rising 1.5 km from its base on the Kermadec Ridge. Raoul Island contains two calderas (shown in oblique view in Figure 4) – Raoul Caldera in the middle of the island is approximately 2 km x 3 km, and Denham Bay Caldera is approximately 3 km x 3 km in size. Past eruptions from Raoul Island, even those involving caldera collapses, have been of similar size to the smaller, non-collapse-related caldera eruptions in the TVZ. This is because Raoul magmas are not rhyolitic. The magma chemistry at Raoul changed from basaltic and andesitic to become dacitic, a more viscous (i.e. flows less easily) and explosive type of molten rock in approximately 2200 B.C. Of the eruptions after this, only one, 1700 years ago was entirely basaltic. No rhyolitic deposits have been found on or originating from Raoul Island (Latter et al., 1992). The island contains two volcanic crater lakes and minor areas of hot springs and fumaroles.

Caldera unrest and eruptions at Raoul Island are likely to be hazardous for visitors to the island, including the Department of Conservation (DOC) workers who run the nature reserve year-round. An eruption at Raoul Island could impact the airspace around it, potentially disrupting flights from New Zealand to the Pacific Islands, and generate rafts of pumice on the sea, disrupting shipping. The most recent eruption at Raoul Island was in 2006.

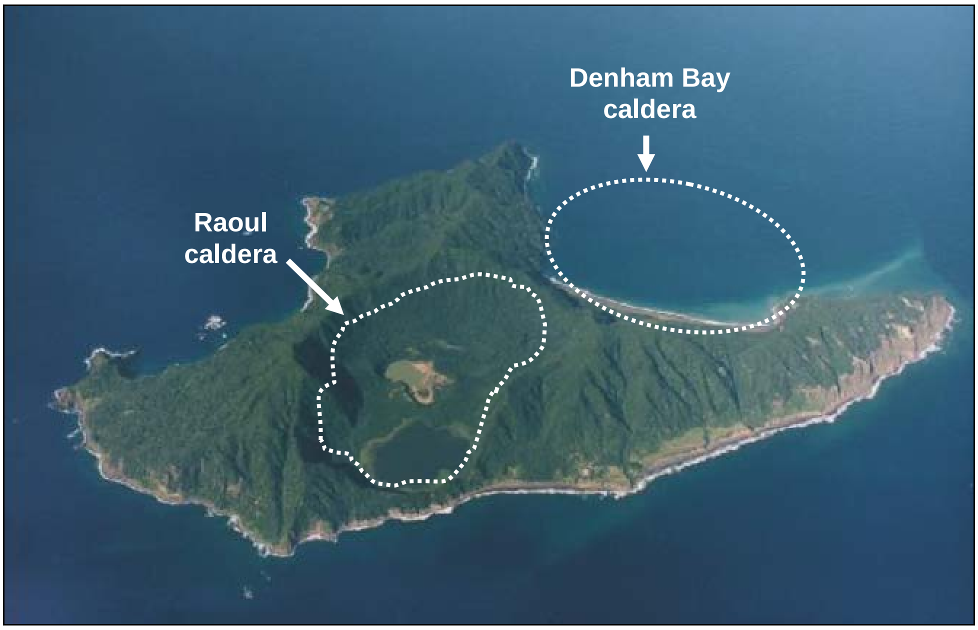

Figure 4 Oblique aerial photo of Raoul Island, looking towards the southwest. The caldera boundaries are indicated. Green Lake is near the centre of Raoul caldera. RNZAF photo.

3.1.1 Eruptions

The oldest volcanic rocks on Raoul Island are up to one and a half million years old (Lloyd & Nathan, 1981; Latter et al., 1992). However Raoul Island didn't emerge from the sea until approximately half a million years ago as a basaltic and andesitic stratovolcano, created by alternating layers of tephra deposits and lava flows. Raoul Caldera began to form in approximately 2000 B.C. during a large eruption, and Denham Bay Caldera formed in approximately 200 B.C. during the largest eruption that has taken place at Raoul in the past 4000 years (Lloyd & Nathan, 1981; Latter et al., 1992). About 16 eruptions have taken place at Raoul in the past 4000 years including the previously mentioned caldera forming eruptions, and historical eruptions in 1814, 1870, 1886 (submarine), 1964 and 2006.

An eruption at Denham Bay on 9 March 1814 was witnessed by a boat which, at the time, was 30 km offshore from Raoul Island. The eruption included the emission of "a strong, fetid, and almost suffocating vapour", and a tall ash column was observed. When the newly formed tephra island was visited two months later, it was "3 miles in circuit, kidney-shaped", and "still smoking" (Lloyd & Nathan, 1981). The island had disappeared by 1854 when it was revisited.

Raoul Island may have erupted many times between 1869 and 1872, but the only confirmed eruptions occurred between June and October 1870 at both Raoul and Denham Bay Calderas (Lloyd & Nathan, 1981). In Raoul Caldera, the eruption began phreatically (steam-driven) with the emission of fine ash, and proceeded to erupt small rock fragments, pumice and mud, killing nearby vegetation. It is thought that no fresh magma was involved in this eruption. A 600 m wide crater was formed, now the site of Green Lake (Figure 4). The eruption lasted for approximately 4 months, endangering the family living on the island at the time. In Denham Bay, the eruption began in June as a submarine eruption, killing fish and causing the water to be discoloured. By July the eruption was forming steam columns estimated to be 600 – 900 m in height and by October of the same year two islands had been formed in Denham Bay. These islands had disappeared by September 1872 (Lloyd & Nathan, 1981).

The submarine eruption in 1886 occurred approximately 8 km from the coast of Raoul Island, probably forming the 240 m high seamount (up to 560 m below the sea surface) currently situated there (Lloyd & Nathan, 1981). Smith (1887, 1888 cited in Lloyd & Nathan, 1981) states that the seismicity and hydrothermal activity at Raoul Island had declined three months after this eruption.

Lloyd and Nathan (1981) describe the 21 November 1964 eruption. It was preceded by constant seismic tremor, lake level changes and ground deformation. The main eruption was centred in Raoul Caldera beginning as a phreatic eruption near Green Lake and lasting for 30 minutes. The eruption included pyroclastic base surges (hot, gassy ash flows), ballistics thrown up to 700 m from the crater and an eruption column up to 1.2 km high. An area of 0.8 km^2 was devastated by the eruption. A pumice slick and discoloured area of water were intermittently seen in Denham Bay from 12 November 1964 until April 1965 and was thought to be caused by gas and possibly fresh lava emerging on the sea floor.

The most recent eruption at any of New Zealand's calderas occurred at Raoul Island in March 2006. It was a very brief (~3 minutes), small phreatic eruption of rocks, pumice debris and lake sediments, centred in Raoul caldera. The eruption was likely to be indirectly triggered by magma (Christenson et al., 2007) however no fresh magma material was found at the surface from this eruption (Rosenberg et al., 2007). This eruption caused the death of a Department of Conservation worker who was sampling in the area at the time of the eruption.

3.1.2 Historical unrest

Due to the lack of population on or near Raoul Island for most of the historic record, the observations of volcanic unrest are very limited and represent an absolute minimum level.

The 1870 eruption was preceded by "almost incessant earthquakes", sulphur fumes and submarine explosions in Denham Bay caldera (Lloyd & Nathan, 1981).

Unrest preceding the 21 November 1964 eruption consisted of 11 days of increased seismic activity up to a magnitude of 5.9 (MM7), including volcanic earthquakes and tremor. The water level in Green Lake (within Raoul Caldera) rose by 6m and the lake temperature increased. Increased activity at springs and high-temperature fumaroles was also reported, and the ground displayed signs of cracking.

An area of vegetation was killed in 1980 near Green Lake due to high ground temperatures (Cole et al., 2006). No eruption occurred.

Between 1989 and 1995, six episodes of unrest consisting of earthquake swarms were recorded which included low-frequency volcanic earthquakes. The individual swarms included more than 300 recorded earthquakes. Monitoring was increased following the 1993 earthquake swarm. There were also changes to the water level of Green Lake during some of these unrest episodes (Scott, 1995).

Precursors to the 2006 eruption are described in Cole et al., (2006). Earthquakes were recorded for five days before the 2006 eruption, opening with a particularly intense swarm lasting for 14 hours, which then died away. Their epicentres were judged to be reasonably far away from the island, and no low-frequency volcanic earthquakes or volcanic tremor were recorded. No other precursory unrest phenomena were observed before the 17 March 2006 eruption.

No significant unrest episodes have occurred since the 2006 eruption.

3.1.3 Potential future activity

Given the frequency of caldera unrest in the past, it seems likely that there will be unrest at Raoul Island in the future. Many unrest episodes have been recorded in the limited period of time with which monitoring networks have been established, with no resulting eruption. This will make predictions of the outcome of future unrest episodes difficult. Eruptions have occurred during historical times with noticeable unrest periods ranging from virtually non-existent to weeks. The unrest behaviour of Raoul Island is likely to be different to the other calderas in New Zealand due to its magma chemistry (i.e. non-rhyolitic). Generally, less silicic volcanoes have shorter and more predictable unrest episodes than rhyolitic volcanoes.

Future eruptions from Raoul Island are unlikely to affect the mainland of New Zealand unless they are exceptionally large (Latter et al., 1992). This would be in the form of tephra deposition over a wide area, and tsunami caused by explosions affecting the water surrounding Raoul Island. Smaller eruptions will endanger all life on the island, and boats and aircraft nearby. Latter et al., (1992) and Blong (1984) further describe possible volcanic hazards.

Today Raoul Island is monitored by GNS Science in conjunction with DOC. Techniques include monitoring seismicity (two sites), deformation using GPS, Green and Blue Lake temperature and water levels and selected hot spring temperatures. The crater lakes and selected hot springs are regularly sampled and sent to the mainland for chemical analysis.

3.2 Macauley Caldera

Macauley Island is the highest point on the mostly submerged Macauley Caldera, located 110 km south-south-west of Raoul Island in the Kermadecs (Figure 1). Macauley Caldera is 13 km x 11 km in size and 1000 m deep and is thought to have been formed during an eruption of magma volume <100 km^3 at 6310 ± 90 yrs BP (Latter et al., 1992; Lloyd et al., 1996). This large eruption was dacitic in composition and produced an ignimbrite deposit as well as widespread tephra (Latter et al., 1992). Basaltic lava eruptions have occurred both before and after the caldera-forming eruption (Latter et al., 1992). Unconfirmed eruptions have been reported in the Macauley area during historical times, including in 1825 and 1887 (Lloyd et al., 1996). Given that Macauley is not rhyolitic it is expected to produce eruptions substantially smaller than TVZ rhyolite calderas, even in caldera-forming eruptions.

Based on the frequency of past eruptions, future eruptions at Macauley Caldera are likely to occur, and will most probably be submarine eruptions from the small cone in the south eastern portion of the caldera. They will probably produce eruption columns which may affect the aviation industry, nearby islands and even mainland New Zealand if there is a northeasterly wind. However the prevailing westerly wind will most likely deposit the tephra into the sparsely populated Pacific Ocean. Pyroclastic flows may enter the sea endangering nearby shipping, and the potential collapse of parts of Macauley Island may cause destructive tsunami.

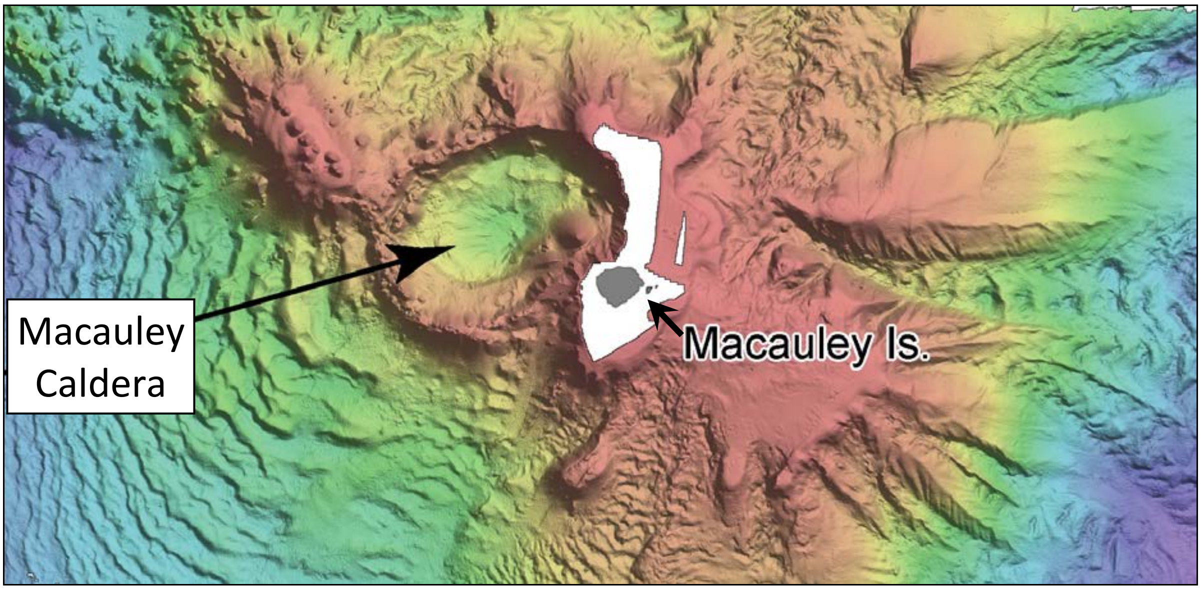

Figure 5 The mostly submerged Macauley Caldera (indicated by the arrow) in the Kermadec Islands (see Figure 1 for location) as seen in a multibeam high resolution image. Macauley Island, shown in grey, is the highest point on Macauley Caldera and the only above-sea portion. Image from NIWA.

3.3 Mayor Island / Tuhua

Mayor Island is located 26 km off the coast of the Bay of Plenty (Figure 1), outside the boundaries of the Taupo Volcanic Zone. The island is the above-surface portion of a 700 m high, 15 km wide shield volcano, which contains a 3 km wide caldera on top (Figure 6). It has erupted on average once every 3000 years for the past 130,000 years (Houghton et al., 1995b). Most major periods of quiescence, such as the one we are in now, have been at least 1000 years in duration, however others have been much longer.

The volcano’s rhyolitic magma chemistry has been very constant for most of its life, but unusually it has displayed a range of eruptive styles and sizes (Houghton et al., 1995b). Almost every style of volcanic eruption has occurred at Mayor Island at some stage in its history, ranging from lava fountaining usually only seen at basaltic volcanoes, through to large explosive, ignimbrite-forming plinian eruptions. The size of eruptions has varied by more than three orders of magnitude (Houghton et al., 1995b).

Figure 6 Oblique aerial view of Mayor Island, looking towards the west at the caldera floor which is covered by younger lava domes. Photo Lloyd Homer, GNS Science.

3.3.1 Eruptions

Three phases of volcanic activity have taken place at Mayor Island (Houghton et al., 1995b). The first, between 130,000 and 36,000 yrs BP, consisted of numerous lava flows and explosive eruptions (including plinian or subplinian, strombolian and phreatomagmatic events, see glossary for details) building up the shield volcano. This phase culminated in the first caldera collapse, possibly caused by several small eruptions (Houghton et al., 1995b).

The second phase occurred between 33,000 and 8,000 yrs BP as the volcanic shield continued to grow within the caldera boundary. Lava domes, lava ponds and pumice cones were created during this time, and at least one explosive (subplinian) eruption occurred, causing minor caldera collapse (Houghton et al., 1995b). Approximately 6,350 yrs BP a second major caldera collapse occurred during a large plinian eruption, which deposited up to 70 cm of pumice on the mainland. Pyroclastic flows entered the sea, probably causing a large tsunami on the mainland.

The third phase of eruptive activity from 6,350 yrs BP until today has included numerous lava flows which have formed domes, and minor explosive activity. The erupted material from this phase has different chemistry to previous deposits, indicating a possible change in the magma source. The most recent eruption occurred less than 1,000 years ago (Houghton et al., 1995b; Buck, 1985).

3.3.2 Historical unrest

No historical unrest is known to have occurred at Mayor Island Caldera. Virtually no earthquakes at all have been located beneath the island in the past ten years.

3.3.3 Potential future activity

Mayor Island is currently in a period of quiescence which has been shorter than previous periods of quiescence (Buck, 1985). This indicates that it is still an active volcano and future activity is possible. Predicting the style of the next eruption at Mayor Island is nearly impossible due to its wide range of styles in the past. Previous periods of quiescence abruptly ended in small but explosive (Vulcanian-style) eruptions, effectively clearing the vent, which then progressed to larger plinian eruptions (Buck, 1985). The most recent eruption probably also started with a small explosive eruption and then became more effusive (containing less gas and therefore less explosive), building a 250 m high lava dome on the caldera floor. It is likely that these styles of eruptive activity will occur again in the future. The entire island is at risk from future eruptive products, especially inside the low-lying caldera, as well as down-wind areas on the mainland during a larger eruption. Tsunami in the Bay of Plenty are a significant hazard during future eruptions (Buck, 1985).

For further details of volcanic hazards resulting from an eruption at Mayor Island, see Houghton et al., (1995b) and a hazard map and recommendations for the emergency management sector are given by Buck (1985).

Mayor Island is monitored for unrest activity by GNS Science using a seismometer. There are two slightly heated crater lakes on the island, and there have been reports of warm springs on western beaches, however no regular sampling currently takes place.

3.4 Okataina Volcanic Centre

The Okataina Volcanic Centre (OVC) has been the source of multiple, mostly rhyolitic eruptions for over 550,000 years. A number of caldera collapses have occurred producing very large eruptions, collectively forming a 16 km x 27 km topographical rim as shown in red in Figure 7 (Nairn, 2002). The complex series of collapses in the north and south of the OVC is called the Haroharo Caldera. More recent intra-caldera eruptions deposited lava domes (such as the voluminous Tarawera and Haroharo dome complexes) and pyroclastics on the caldera floor, covering many of the older individual structure outlines. Lakes lie on the caldera floor between the caldera rim and younger lava domes, in places concealing the caldera rim. Parts of the rim are also hidden by ignimbrites originating from neighbouring calderas, such as in the south-western area of OVC. Tectonic faulting has continued to alter the area (Cole et al., 2010).

The eruptions at Okataina in the last 26,500 years have generally followed a similar pattern – explosive pyroclastic eruptions which have produced widespread ashfalls, followed by the extrusion of thick rhyolitic lava flows with associated near-source block-and-ash flows. Multiple vents were active at the same time, often separated by several kilometres (Nairn, 1991). This is likely to be repeated in future eruptions. However some eruptions show that low viscosity basalt has been involved as well (such as the most recent eruption in 1886). The styles of eruptions can vary when mixed magma types are involved. Very large, caldera-forming eruptions have also occurred in the past. Magma interacting with shallow groundwater or surface water has also occurred at OVC, causing large steam explosions. This occurred in 1886 at Rotomahana during the Tarawera eruption, causing most of the casualties from this event (Nairn, 1991). Periods of quiescence between volcanic activity have varied between 700 and 3000 years (Nairn, 1991).

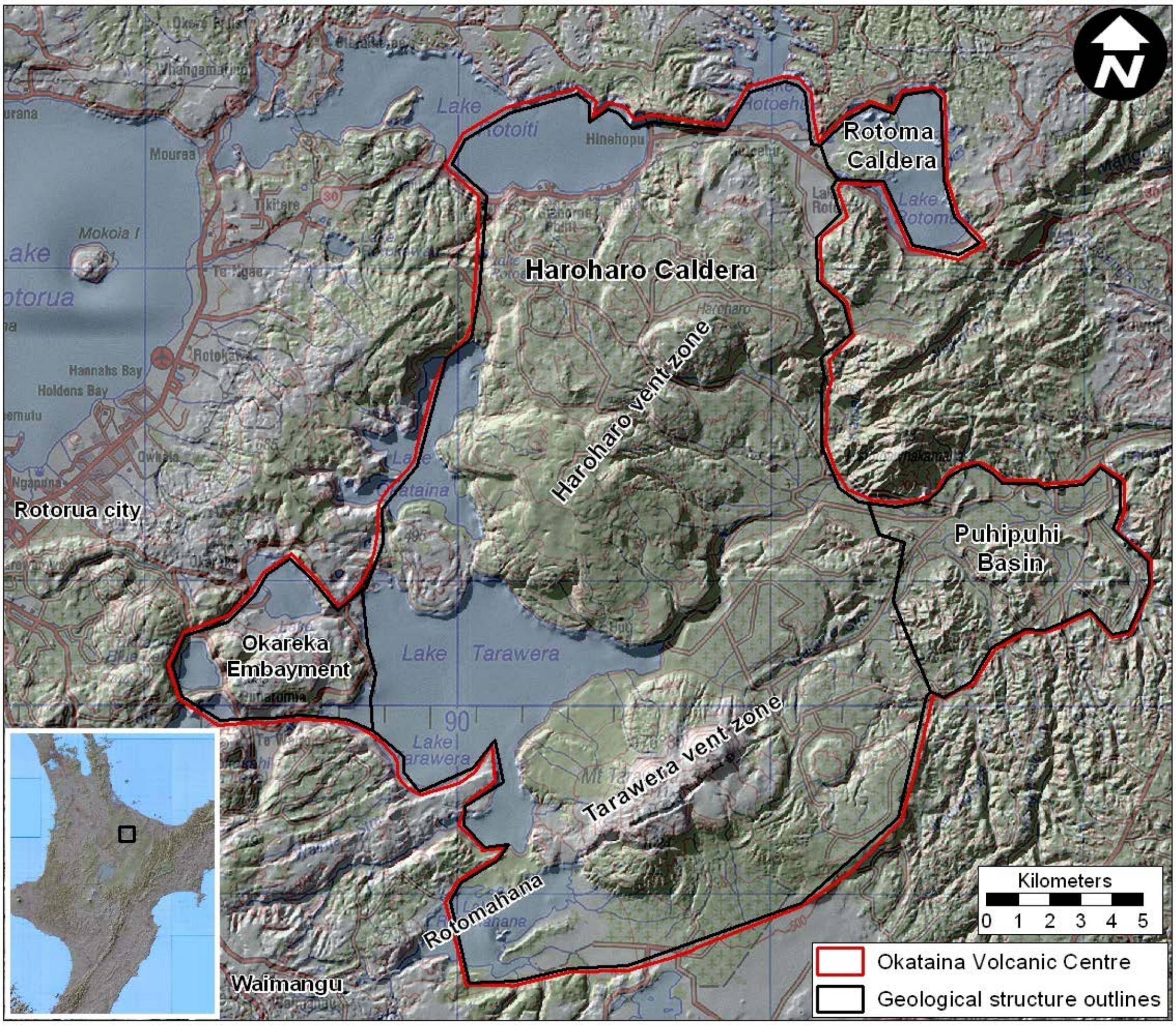

Figure 7 Okataina Volcanic Centre and major geological structures. Inset bottom left shows the map position in the North Island. Based on Nairn (2002).

3.4.1 Eruptions

A large ignimbrite forming eruption (with an erupted magma volume of possibly 90 km^3), which was probably accompanied by caldera collapse, occurred at 550,000 yrs BP with a vent source in the southern part of the OVC and covers earlier dissected lava domes in the OVC area (Cole et al., 2010). Further eruptions occurred 320,000 yrs BP (Leonard et al., 2010), culminating in an enormous 160 km^3 rhyolitic eruption, producing a pyroclastic flow that reached the Bay of Plenty coast, forming the Matahina Ignimbrite deposit (Nairn, 2002). This contributed to caldera collapse of the southern portion of the OVC and possibly the Puhipuhi Basin (Figure 7) (Nairn, 2002). The Puhipuhi Basin was afterwards filled by a lake. This lake deposited sediments which were subsequently uplifted and a dacitic eruption occurred here (date unknown but older than 61,000 yrs BP; Nairn, 2002; Cole et al., 2010).

There are several lava dome and dome complexes preserved at the surface dated between 550,000 and 61,000 years ago (Leonard et al., 2010), with others likely obliterated by younger caldera eruptions or buried. These probably followed broadly similar eruption styles to those of the younger Haroharo and Tarawera lava domes. Following further rhyolitic lava and pyroclastic eruptions, the northern part of the OVC collapsed approximately 61,000 yrs BP in a large (>100 km^3) rhyolitic eruption called the Rotoiti episode and formed part of the

Haroharo Caldera (Cole et al., 2010). This eruption may have been triggered by a basaltic scoria eruption (Pullar & Nairn, 1972; cited in Nairn, 2002). Between 61,000 yrs BP and 26,500 yrs BP, at least 12 plinian eruptions of the Mangaone Subgroup occurred, as well as two pyroclastic flows (Cole et al., 2010). This includes the ~20 km^3 Kawerau ignimbrite eruption, approximately 33,000k yrs BP which may have caused minor caldera collapse in the southern OVC (Spinks, 2005; cited in Cole et al., 2010). Rotoma and Okareka areas collapsed due to lateral magma migration during this eruption (Cole et al., 2010).

Nine rhyolitic eruptions have taken place in the past 26,500 years building the Tarawera and Haroharo dome complexes within the Haroharo Caldera rim, with a combined erupted magma volume of 85 km^3 (Nairn, 2002). These differ from the preceding Mangaone Subgroup in that they have a significant lava volume, rather than being mostly pyroclastic. The various OVC caldera structural outlines have largely been buried by these younger deposits. The Haroharo complex consists of rhyolitic lava flows and domes, and plinian pyroclastic fall and flow deposits (Cole et al., 2010). There was a small basaltic rift eruption near the western margin of the caldera 3400 years ago (Nairn, 2002). There are other small basalt eruption deposits in the Okataina area, and overlapping with the edge of the Rotorua area.

The Tarawera complex (or vent zone, Figure 7) is slightly younger and includes block-and-ash flow deposits. It was the source of eruptions in 1314 AD (Leonard et al., 2010) and 1886 (Cole et al., 2010), the two most recent eruptions from OVC. The rhyolitic Kaharoa eruption in 1314 (AD) had a duration of approximately 4 years, and erupted 4 km^3 of material. It was the largest eruption in New Zealand in the past 1,000 years, and the most recent rhyolitic eruption in New Zealand (Leonard et al., 2010; Johnston et al., 2004). The source of this eruption was an 8 km line of vents across the Tarawera dome complex (Nairn et al., 2001, 2004). The eruption included plinian events, phreatomagmatic explosions, pyroclastic flows, and the extrusion of lava domes which collapsed and caused block-and-ash flows. It is thought to have been triggered by a basaltic intrusion (Leonard et al., 2002). A large breakout flood occurred when the temporary blockage consisting of eruption debris at the lake outlet was breached, after the Kaharoa eruption (Hodgson & Nairn, 2005). A similar but smaller event occurred in November 1904 after the 1886 Mt. Tarawera eruption.

The 10 June 1886 Tarawera Rift eruption produced basaltic scoria from Mt. Tarawera and a mix of basalt and phreatomagmatic mud and breccia from the neighbouring Rotomahana basin and Waimangu Valley (Figure 7). This marked a significant difference in the magma composition compared to a long history of predominantly rhyolitic eruptions. This eruption is the largest to have occurred in New Zealand's recorded history. The eruption began at about 1:30 am from vents along the top of Mt. Tarawera, producing an ash column 10 km high and deposits of basaltic scoria (Nairn, 1991). Within an hour, Rotomahana and Waimangu, both southwest of the Tarawera vent lineation, were also in eruption, totalling 17 km of active rift eruptions. The rift on Mt. Tarawera as it looks today is shown in Figure 8.

Figure 8 Rift on Mt. Tarawera, Okataina Volcanic Centre, formed in the 1886 basaltic eruption, showing the red scoria deposits. View towards the southwest. Photo GNS Science.

The eruptions at Rotomahana were particularly explosive, caused by the interaction of basaltic magma and a large hot hydrothermal system. This generated very violent pyroclastic surges and thick deposits of mud, which caused the collapse of buildings and contributed towards most of the fatalities from this eruption (Nairn, 1991). Extensive lightning took place in the eruption cloud, setting fire to a house and the forest, and fissures in the ground caused travel difficulties post-eruption. Strong winds (possibly eruption blasts) caused trees to be knocked over, and volcanic gases caused breathing difficulties (Nairn, 1991). Tephra was deposited on the land and sea in a north-eastern direction, and caused darkness during the day over a wide area. The eruption ceased at approximately 6 am, after killing 108 people (Nairn, 1991). Hydrothermal explosions continued to occur in the area for several weeks, and steam was emitted from the volcanic vents for months (Simmons et al., 1993). The Rotomahana-Waimangu area is now host to a large new surface expression of a geothermal system (Scott, 1994).

3.4.2 Historical unrest

Research by Leonard et al., (2002), Nairn et al., (2004) and Sherburn and Nairn (2004) on the 1314 AD Kaharoa rhyolitic eruption from Okataina Caldera suggest precursors to this eruption may have been detected up to years in advance had the current monitoring network been in place (Johnston et al., 2004).

In contrast to the interpretation the rhyolitic Kaharoa event, no significant unrest was recorded in the days or months prior to the 1886 basaltic eruption at Mt. Tarawera. “Peculiar waves” were seen on Lake Tarawera, 0.3 m high, 10 days before the eruption which may have resulted from magma-related ground movements (Nairn, 1991). One hour before the onset of the eruption, earthquakes were felt in nearby areas, increasing in intensity until the eruption (Keam, 1988; Nairn, 1991).

Historical unrest of the OVC has not been studied in detail. Volcano monitoring equipment has recorded height changes across the caldera (including approximately 50 mm of subsidence between 1980 and 1984) (Scott, 1989), activity in the Waimangu hydrothermal system, including relatively frequent hydrothermal eruptions as described by Scott (1994), and seismic activity (GNS Science earthquake catalogue). The best recorded seismic swarm occurred in April 1998, centred on the Haroharo vent zone. Over 400 earthquakes >M1.7 were recorded, however only 4 were reported as felt. The maximum earthquake had a magnitude of 4.7, and a felt intensity of MM4.

Another seismic swarm was centred on Rotoehu in 2004, over the boundary of the northern caldera margin. Over 1,300 earthquakes (>M1.7) were recorded from July 1st – August 5th, with magnitudes of up to 5.1. This event was interpreted by GNS Science staff to be a mainshock-aftershock event (Hurst et al., 2008).

3.4.3 Potential future activity

The 1886 Mt. Tarawera eruption was small compared to most OVC eruptions, with a different magma composition, therefore it is unlikely to be representative of future eruptions. It is expected that the next eruption will be larger and more similar to the majority of the eruptions from the past 60,000 years, following the pattern of a rhyolitic pyroclastic eruption (tall ash columns, ash fall and flows), followed by the extrusion of lava flows and domes, with associated block-and-ash flows (Nairn, 1991, 2002). The likely unrest is reviewed by Sherburn and Nairn (2004). A description of the volcanic hazards is in Nairn (1991) and Scott and Nairn (1998) (Figure 3).

Today there are nine seismic monitoring sites and seven cGPS sites monitoring deformation in the OVC. The geothermal systems are monitored by collecting the temperature and water levels or overflows of the large crater lakes at Waimangu and by chemical sampling of selected hot springs.

3.5 Rotorua

Rotorua Caldera collapsed at the end of the very large (145 km^3) Mamaku plateau formation eruption in 240,000 yrs BP (Gravley et al., 2007). This eruption coincided with the Ohakuri eruption 30 km to the south. The Kapenga area collapsed associated with this eruption, probably due to magma withdrawl, and this is the only confirmed collapse episode in the Kapenga area. The ignimbrites from the two sources were emplaced only a few weeks apart, and in some areas overlap. At some stage after these explosive eruptions rhyolitic lava was extruded forming domes including Ngongotaha and Mokoia Island (Leonard et al., 2010). It is unknown when the most recent eruption at Rotorua Caldera occurred, but it appears to be at least 20,000 years ago.

Rotorua Caldera incorporates the Rotorua and Eastern Rotorua geothermal fields (Leonard et al., 2010). Parts of these geothermal fields have been developed as popular tourism attractions, bringing people to these areas. Hydrothermal explosions have occurred in the city of Rotorua a number of times, possibly due to changes to the geothermal system introduced by exploitation. The most recent significant (but relatively small) hydrothermal explosions occurred at Kuirau Park (an inner-city park containing a number of hydrothermal features) in January 2001 (Figure 9) and December 2006. A number of deaths have occurred from people (mainly children) falling into boiling mud pools and hot springs. The main gases emitted by Rotorua's geothermal system are hydrogen sulphide (H_2S) and carbon dioxide (CO_2), both of which are denser than air and toxic (Durand & Scott, 2005). 14 fatalities have occurred from gas poisoning in this district of approximately 70,000 people. These have occurred in small, low, constricted spaces such as natural hot spa baths, when patrons have been overcome by H_2S gas (Durand & Scott, 2005). A study by Durand and Scott (2005) on several Rotorua city buildings showed potentially dangerous and damaging levels of gas (H_2S and CO_2), emitting from cracks in paving, from waste water drains and in narrow, low down spaces, as well as inside buildings.

Land use management can restrict development of geothermal areas at risk of future hazardous activity, during caldera unrest or regular geothermal system processes. Mitigation measures typically include set backs from surface activity for buildings and infrastructure, and restrictions on covering warm and hot ground. Significant geothermal features are typically fenced for safety.

Figure 9 Kuirau Park hydrothermal crater (bottom right) and deposits, 2001, in Rotorua city. Photo by the Daily Post.

No major seismic swarms have occurred inside the Rotorua Caldera in the past ten years, however there is an area of potentially higher seismicity in the southern portion of the caldera (Bryan et al., 1999). Small swarms have been recorded, including those in 1994, 1998, 1999, 2000 and 2001. The 2001 overnight seismic swarm was the largest, with magnitudes of up to 3.2, and over 50 earthquakes recorded in 2 hours, 14 of which were recorded as felt.

GeoNet has four seismometers and one strong motion seismograph within the Rotorua caldera and three GPS stations. They also conduct regular sampling of selected hot springs and bores. The BOP Regional Council has a monitoring programme in place to record the surface geothermal features and borehole pressures and temperatures.

3.6 Kapenga

Gravley et al., (2007) suggest much of the Kapenga collapse structure formed during the Mamaku plateau formation (from Rotorua Caldera) and Ohakuri eruptions 240,000 yrs BP, rather than from an eruption at Kapenga itself. Kapenga has been suggested as the source of a number of ignimbrite and other eruption deposits, but this is unconfirmed (Leonard et al., 2010), and it should be referred to as a volcano-tectonic depression rather than a caldera due to the lack of caldera-collapse vents. This includes ignimbrite-forming eruptions in 300,000 yrs BP and 275,000 yrs BP, the latter with a volume of 100 km^3 of magma (Gravley et al., 2007; Leonard, 2003). Alternatively these could well have been from vents or calderas now buried below the Mamaku plateau.

Smaller, generally rhyolitic eruptions have occurred in the Kapenga area, forming lava domes and scoria deposits. Hydrothermal explosion deposits have also been identified within Kapenga (Leonard et al., 2010).

Due to the length of time this caldera has been dormant, it is unlikely (but cannot be ruled out) that Kapenga will erupt again in the future. No research has been done on historical unrest at Kapenga, therefore no unrest episodes are known to have occurred. Many shallow earthquakes are recorded in the Kapenga area, however it is difficult to determine the source of these events. The Rotorua-Taupo Fault belt runs through the area so much of the seismicity is probably tectonic.

GeoNet has two seismometers in this area and four more nearby that would help locate events. There are also two GPS stations.

3.7 Ohakuri Caldera and Maroa Volcanic Centre

The >100 km^3 ignimbrite-forming Ohakuri eruption probably occurred at ca. 224,000 yrs BP (Gravley et al., 2007). This eruption caused the collapse of Ohakuri Caldera (Leonard et al., 2010). The Ohakuri eruption coincided with a large eruption at Rotorua Caldera which formed the Mamaku ignimbrite (Mamaku Plateau Formation), and it is likely have also caused subsidence of the Kapenga area (Gravley et al., 2007).

Whilst Maroa Volcanic Centre contains no caldera, a brief summary of its volcanic history is included here as it borders Ohakuri caldera closely. There is no clear link between the magmatic systems of these two centres (Leonard, 2003). Volcanism at Maroa was most intense prior to 200,000 yrs BP (Leonard, 2003). In the past 61,000 years, there have been at least four eruptions from this centre, all relatively small (with magma volumes of <0.2 km^3), in 45,000, 43,000 and 40,000 yrs BP (Wilson et al., 2009). Rhyolitic lava domes and dome complexes dot the surface of the Maroa Volcanic Centre, along with ignimbrite deposits. The most recent eruption from the Maroa Volcanic Centre was 16,500 yrs BP, with a volume of 0.14 km^3 (Lloyd, 1972; Leonard, 2003).

Leonard (2003) stated that the probability of a future eruption is at the Maroa Volcanic Centre, based on the rhyolite and basalt eruptive episode history of the centre over the last 100,000 years, approximately 0.7% in an 80 year lifetime, and the most probable eruption size in the future is (<0.1 km^3) based on the more recent eruptions. No known historical unrest has been recorded at Maroa Volcanic Centre except for regional earthquake activity and hydrothermal eruptions at Orakei Korako.

GeoNet has four seismometers in this area and three more nearby that would help locate events. There is also one strong motion instrument.

3.8 Reporoa

Reporoa Caldera collapsed during a single eruptive episode (Nairn et al., 1994; Leonard et al., 2010). The Kaingaroa Formation was deposited in 230,000 yrs BP (Houghton et al., 1995a) in the form of a widespread ignimbrite deposit (Nairn, 2002; Leonard et al., 2010). Rhyolitic lava domes have been erupted in the Reporoa area, dated as both older and younger than the Kaingaroa Formation (Leonard et al., 2010). Small basalt eruptions have occasionally occurred in this area.

Due to the length of time this caldera has been dormant, it is unlikely (but cannot be ruled out) that Reporoa Caldera will erupt in the near future. No research has been done on historical unrest at Reporoa Caldera, therefore no unrest episodes are known to have occurred. The area does experience regional earthquake activity and there are two areas of hot springs.

GeoNet has three seismometers in this area and three more nearby that would help locate events. There is also one strong motion instrument.

3.9 Mangakino

Very large ignimbrite-forming eruptions have been attributed to the Mangakino Caldera, occurring between 1.6 million and 950,000 yrs BP, the latter with a volume of 50 km^3 (Houghton et al., 1995a). Rhyolitic lava domes were also erupted during this period (Leonard et al., 2010).

No research has been done on historical unrest at Mangakino Caldera, therefore no unrest episodes are known to have occurred. Due to the length of time this caldera has been dormant, including a lack of any known smaller intra-caldera eruptions, it is unlikely that Mangakino Caldera will erupt in the near future, but this older caldera system is included here for completeness.

GeoNet has one seismometer in this area.

3.10 Whakamaru

Whakamaru Caldera was the source of a very large eruption about 350,000 yrs BP, erupting 1,500 km^3 of magma (Leonard et al., 2010; Wilson et al., 2009). An eruption 10,000 years later deposited an additional 500 km^3 of magma (Wilson et al., 2009). This is the most recent caldera-forming eruption at Whakamaru. It seems unlikely that Whakamaru will erupt in the near future; this is due to (a) the very long time period since these eruptions, and (b) the presence of the younger Maroa Volcanic Centre and part of Taupo Caldera overlapping the older Whakamaru Caldera – both have different magma chemistry to the older eruptions, suggesting that a quite different magma system configuration in the area now exists.

No research has been done on historical unrest at Whakamaru Caldera, therefore no unrest episodes are known to have occurred. However the estimated Whakamaru Caldera boundary envelops a large area of the TVZ, and includes the Wairakei-Tauhara, Rotokawa and Mokai geothermal fields, and parts of the Orakei-Korako and Atiamuri geothermal fields (Leonard et al., 2010), numerous active fault lines (including the Taupo Fault Belt), and part of the Taupo and Maroa Volcanic Centres. Therefore the Whakamaru Caldera area has been the source of numerous hydrothermal explosions, deformation and seismicity in the geological and historical past, but these events cannot be attributed to (or in fact excluded from) caldera processes.

GeoNet has three seismometers in this area and three more nearby that would help locate events.

3.11 Taupo

Taupo caldera is located in the central North Island of New Zealand and is the southernmost caldera of the TVZ (Figure 1). It has a complex history of both very large and very small eruptions, most of which were rhyolitic in composition. The most recent eruption in 232 AD (A. Hogg, pers. comm., 2010) was very large and destructive, but not representative of the most common size of eruption over the past 27,000 years (Wilson, 1993).

Figure 10 Taupo caldera viewed towards the southwest, with Taupo township located on the north-eastern shore. The Waikato River (in the foreground) is the outlet from Lake Taupo and a source of water and electricity generation for the upper North Island. Photo by Lloyd Homer, GNS Science.

3.11.1 Eruptions

Taupo volcano has been active for at least 330,000 years (Pringle et al., 1992; Wilson et al., 1986). The eruptive history between 330,000 yrs BP and 65,000 yrs BP is poorly understood as the deposits have been either buried or destroyed by subsequent eruptions (Wilson, 1993). Between 65,000 yrs BP and 27,000 yrs BP there were approximately ten eruptions from the Taupo Volcanic Centre, at least five of which were explosive (Vucetich & Howorth, 1976; Wilson et al., 2009). Taupo has an average magma output rate of 0.2 m^3s^{-1} over the past 65,000 years, and is the most productive individual rhyolitic volcano in the world (Crisp, 1984; Wilson, 1993). Small basalt eruptions have also occasionally occurred in the Taupo Volcanic Centre.

The Taupo caldera was largely formed during the cataclysmic Oruanui eruption (Wilson, 1993) at 27,000 yrs BP (Bard, 1998; Wilson et al., 1988). This event had a total magma equivalent volume of 530 km^3 that was erupted episodically over a period of several months (Wilson, 2001). The eruptive style was a complex interaction of magma and water, producing widespread tephra falls interspersed with pyroclastic density currents (Wilson, 2001). Deposits from this eruption formed a dam containing a large volume of water, producing huge floods down the Waikato River when it collapsed.

Twenty-seven of the twenty-eight eruptions that are known to have occurred after the Oruanui eruption formed pyroclastic (tephra and/or pyroclastic density current) deposits, with only one deposit consisting solely of a lava extrusion. Periods of quiescence varied between approximately 20 to 6000 years, and eruption size varied as the volumes of magma erupted ranged from 0.01 (a similar size to Ruapehu's 1995-96 eruption) to 35 km^3 (Wilson, 1993; Wilson et al., 2009). Eruption styles were widely diverse, which will cause difficulties in predictions for future eruptions.

The most recent eruption from the Taupo Volcanic Centre was in 232 ±5 AD (A. Hogg, pers. comm. 2010), which altered the shape of the caldera (Wilson, 1993). This eruption devastated 20,000 km^2 of surrounding land due to widespread tephra falls and ignimbrite-forming pyroclastic flows travelling up to 70 km away from the lake (Wilson & Walker, 1985). It started with a small, wet eruption and increased in size and violence, with the occasional pause of up to three weeks. The majority of the deposits were emplaced in the final stage of the eruption when the magma chamber roof collapsed and a particularly energetic pyroclastic flow travelled at a velocity of 200-300 ms^{-1} radially outwards from the vent in the north-east part of Lake Taupo, lasting about 6.5 minutes (Froggatt, 1981; Walker, 1984; Wilson & Walker 1985). Following this eruption, the lake refilled over several years reaching a level approximately 30 m higher than the present day level for 15 – 40 years (Manville et al., 1999). Once the water cut through the ignimbrite layers damming the lake, it overflowed with a volume of up to 35,000 m^3/s, flooding large areas downstream and lowering the lake level to approximately 10 m higher than the present day level. At this stage, Wilson and Walker (1985) state a further small lava extrusion occurred, possibly forming Horomatangi Reef. Large pumice blocks floated to the surface and came to rest at the lake edge nearby.

3.11.2 Historical unrest

European settlement and the act of writing down events in the Taupo region began in the mid-nineteenth century. There is no unrest record before this time. Four previously recognised episodes of caldera unrest have occurred at Taupo in 1895, 1922, 1964-5 and 1983 (Johnston et al., 2002) as described below. Recent research by one of the authors (Potter) has indicated many more episodes of unrest have occurred at Taupo caldera than had previously been recognised. These episodes range in magnitude from minor unrest (such as earthquake swarms with 15 – 20 earthquakes felt in one day), to months of seismic swarms causing building damage. This research will be published in the near future. Taupo's unrest episodes have included deformation, hydrothermal activity and earthquake swarms, resulting in public alarm, self-evacuations and decreased levels of tourism. They indicate that, had the current VAL system existed during these episodes, they may have been assigned a level of VAL 1 or even up to VAL 2.

Unrest in 1895 began on 17th August with an earthquake of shaking intensity MM8 (Eiby, 1968) striking Taupo and causing widespread damage. Most of the town's chimneys collapsed, bottles and crockery were smashed, and "chaos reigned supreme" (Poverty Bay Herald, 19 and 20 August 1895). Landslides blocked roads around the lake, and residents and visitors camped outside overnight. A 0.6m wave was seen on Lake Taupo and springs in the Hatepe region emitted quantities of fine pumice (Hawke's Bay Herald, 20 August 1895).

Springs changed temperature and tremors continued until at least September 1895 (Poverty Bay Herald, 2 September 1895). It is uncertain whether an event such as this should be classed as unrest, or if it was just a large regional tectonic earthquake.

The largest episode of caldera unrest known to have occurred in New Zealand during historical times (without an eruption) was in Taupo, lasting for 10 months from April 1922 until January 1923. Earthquakes were felt in the Taupo region throughout this period, with the most severe shake on 10th June, and 57 earthquakes felt in 8 hours on 25th June (The Evening Post, 28 June 1922). Fissuring and faulting, landslides and minor changes to activity at hot springs and geysers were reported. Subsidence of 3.7 m caused a sunken shoreline at Whakaipo Bay on the northern side of Lake Taupo, along with hundreds of water fountains emerging from ground cracks, causing flooding (Evening Post, 14 July 1922). Several chimneys collapsed in Taupo, Oruanui and Wairakei, bottles, crockery, books and other items were thrown on the floor, and the Taupo town clock stopped. Tourism was affected in not only Taupo, but also Rotorua due to incorrect international reporting (Evening Post, 7 July 1922).

Earthquakes increased in number and intensity from September 1964, and peaked in December 1964 with magnitudes of up to 4.8 (Eiby, 1966). 140 events per day were reported and over 1100 earthquakes over magnitude 2.7 were felt in two months (Gibowicz, 1973). Seismicity decreased again until February 1965 and a further small swarm occurred in December 1965. The epicentres migrated from Western Bay, Lake Taupo in early December 1964, to northern Lake Taupo by 21st December and then to Horomatangi Reef and Waihaha area by January 1965 (Gibowicz, 1973). Possible uplift of 90 mm near Horomatangi Reef was observed (Grindley & Hull, 1986) otherwise no faulting or deformation was reported.

Seismicity clustered in February and June 1983 with up to 30 tremors recorded a day. Uplift of 55 mm was followed by equivalent subsidence at a block west of Kaiapo fault, which ruptured on 23rd June (Otway et al., 2002). Minor damage from the earthquakes was reported, including cracked chimneys and fallen ornaments (Otway et al., 1984).

Hydrothermal eruptions have occurred in geothermal fields near Taupo, such as at the Wairakei Geothermal Field in July 1960, likely due to geothermal field developments, as well as in 2000 and March 2001. The Tauhara geothermal field had a hydrothermal eruption in June 1981 (Scott & Cody, 1982). It is unknown whether these were indicators of caldera unrest, were part of the normal hydrothermal system processes or induced by the production from the geothermal systems.

3.11.3 Potential future activity

Based on the frequency of unrest at Taupo caldera since European settlement, it seems likely that unrest activity will occur in the future. The lengthy swarms of earthquakes, metres of ground deformation and hydrothermal explosions seen in the past episodes will almost certainly be repeated in the future at varying intensities. Future episodes of unrest may include those larger than previously witnessed during Taupo's short settlement history, reflecting the scale of caldera unrest seen internationally. The town and surrounding areas should be prepared for large, damaging earthquakes and other hazardous unrest phenomena.

Future eruptive activity has been speculated on by Froggatt (1997) in the Civil Defence 'Volcanic Hazards at Taupo Volcanic Centre' publication (in the 'Yellow Book' series). He states that the next Taupo eruption is [more] likely to be on or near the Horomatangi Reef, where 32% of the eruptions in the past 27,000 years have taken place (Froggatt, 1997; Wilson, 1993). The chance of large pyroclastic flows in future eruptions is small, however the likelihood of magma/water interaction is high, which causes more explosive, rather than effusive, eruptions. The eruption size is most likely going to be small to medium and be preceded by significant deformation and geothermal changes (Froggatt, 1997). It is important to consider that nearly 70% of recent eruptions have, however been elsewhere in the northeastern part of the Taupo Volcanic Centre.

GeoNet monitors Taupo caldera with 6 permanent seismographs, 3 additional strong motion seismic sites, and 8 telemetered cGPS sites. There is a network of lake levelling sites around Lake Taupo, which is used as a giant spirit level to monitor tilt of the ground surface.

4.0 INTERNATIONAL CALDERAS – ERUPTIONS AND HISTORICAL UNREST

4.1 Introduction

A number of volcanoes worldwide are similar in magma chemistry, tectonic setting and past eruption styles to New Zealand's calderas. The examples used in this section are rhyolitic calderas of various sizes, some of which have erupted during historical times. New Zealand can learn from these occurrences, particularly from countries with long written records, to supplement the short history of this country. Campi Flegrei Caldera in Italy, and Long Valley Caldera in California (US) have shown moderate levels of unrest in the last few decades, raising concern over the management of these unpredictable and complex volcanoes. Building damage resulting from high levels of seismicity at Campi Flegrei prompted an evacuation of over 40,000 people, while in Long Valley, a nearby tourist resort town suffered economically from the unrest event. Unrest at Yellowstone National Park (US) is carefully monitored due to the very large eruptions in the geological past, and the high number of visitors to the park. Smaller historical eruptions have occurred at Rabaul, Papua New Guinea, following decades of unrest; Chaitén, Chile; Sakurajima in Aira Caldera, Japan; Taal in the Philippines; and the largest rhyolitic eruption in recorded history at Novarupta in Alaska (US). The locations of these calderas are shown in Figure 11, and a comparison of selected caldera activity summarised in Table 4.

4.2 Campi Flegrei (Italy)

Campi Flegrei Caldera is located on the edge of the city of Naples, Italy (Figure 11), which has a population of approximately 3.8 million people in the metropolitan area. It has a reasonably similar eruptive history to Taupo and Okataina, with large plinian, caldera-forming eruptions having occurred in the past. The most recent eruption within the caldera was a small cone-building eruption in 1538 AD. It is one of the only calderas in the world to have had a witnessed eruption (with prior unrest) from a rhyolitic caldera. Campi Flegrei has undergone intense volcanic unrest in the past few decades, with metres of uplift and damaging seismicity, with no resulting eruption. This unrest has caused serious social consequences including mass evacuations, as described below.

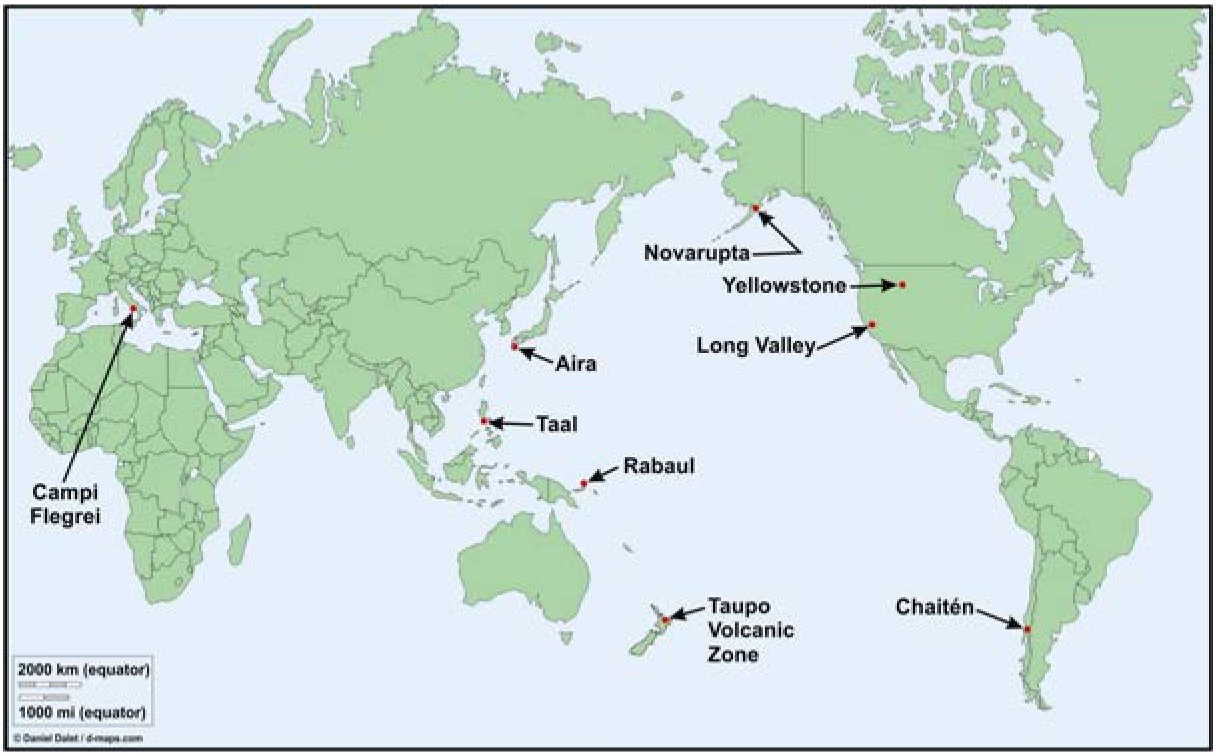

Figure 11 World map showing locations of young case study calderas similar to the Taupo Volcanic Zone calderas.

4.2.1 Eruptions

A large (~150 km^3) caldera-forming eruption occurred at Campi Flegrei approximately 39,000 yrs BP (De Vivo et al., 2001), with a further eruption (~40 km^3) 15,000 yrs BP (Deino et al., 2004). Prior to a period of intense volcanism 4,000 yrs BP, Campi Flegrei exhibited caldera-wide deformation of tens of metres (Isaia et al., 2009). At least 60 smaller eruptions have also occurred, the most recent in 1538 AD (Di Vito et al., 1999).

The 1538 AD eruption created a small cone called Monte Nuovo, centred within the Campi Flegrei Caldera boundary near the harbour town of Pozzuoli (Figure 12). The Campi Flegrei area was home to a few thousand people at the time (Dvorak & Gasparini, 1991). The unrest phenomena observed before this eruption are described in Table 5. The eruption began on 29th September 1538 with an explosive phreatomagmatic phase. It consisted of small pyroclastic flows for two days, and included pumice falling 8 km from the vent (Guidoboni & Ciuccarelli, 2011), followed by four days of only minor explosive activity (Di Vito et al., 1987). On 6th October an explosive eruption of scoria and small pyroclastic flows killed 24 people who were ascending the cone (Di Vito et al., 1987; Guidoboni & Ciuccarelli, 2011). The newly formed Monte Nuovo has an estimated volume of 0.025 km^3 and is 130 m high (Di Vito et al., 1987; Dvorak & Gasparini, 1991). The erupted scoria is phono-trachytic in composition (Piochi et al., 2005) with less silica than rhyolite (more similar to andesite and dacite), but more alkali content resulting in differences in the eruption style.

Table 4 Summary of selected international caldera unrest episodes. Refer to text for more details.

Caldera name	Date of episode	Seismicity	Deformation	Hydrothermal/ Gravity	Social response	Eruption?
Campi Flegrei Refer to section 4.2	1982-84	Hundreds of felt earthquakes, some large (<M4.2). Damaged buildings.	3.5 m uplift at Pozzuoli	Increase in the concentrations of gases such as H_2S and CH_4 at Solfatara, gravity measurements suggested a growing magma chamber.	40,000 evacuated	No
Long Valley Refer to section 4.3	1979-84	Seismicity waxed and waned from1982-84. 3 M6 earthquakes in 1 day, another was large enough to knock out electricity in 1983.	25 cm of uplift in < 6 months in the caldera, a further 7 cm over 4 years.		Anger, uncertainty and frustration felt by the public and some officials over the issue of a Volcanic Hazard Notice and its perceived effect on the economy. Plans updated and an extra access road ploughed.	No
Rabaul Refer to section 4.4	1983-85	Hundreds recorded per hour at times <M5.1. Low frequency earthquakes recorded.	3.5 m uplift from 1971 – 1984. <0.1 m per individual crisis (lasting hours to days).	Gravity changes measured between 1973-83.	Emergency plans developed, evacuation drills undertaken. Previous eruptions memorable so unrest readily accepted.	No (see text for 1994 eruption description)
Chaitén Refer to section 4.5	2008	24 hours of seismicity recorded on distant monitoring equipment before the eruption. Earthquakes large enough to knock objects off shelves in the town of Chaitén 10 km distant. 15-20 recorded per hour.	Not monitored	Not monitored	Town of Chaitén evacuated after start of eruption, subsequently overrun by lahars. Very short lead in time before eruption, but volcano wasn't monitored before the eruption.	Yes

Table 5 A description of the potential caldera unrest reported for 70 years prior to the 1538 AD Monte Nuovo eruption, Campi Flegrei, Italy. Based on Guidoboni and Ciuccarelli (2011).

Date	Seismicity	Deformation	Other
1470 – 1472	Intense seismicity (<MM7), causing damage to buildings in Pozzuoli		Increase in gas emissions, damage to vegetation
1475 – 1499	Occasional moderate sized earthquakes (background levels?)		
1500 – 1511	Large earthquakes (<MM8) in 1505 and 1508	Uplift in the order of meters at Pozzuoli in ~1502 as well as ~1510	
1512 – 1536	Only 1 reported earthquake, in 1520 (<MM7). Unconfirmed reports of a large earthquake and aftershocks in 1534		
1536	MM5 earthquake in Aug., period of seismic activity from Sep – Dec (<MM4)		Increase in gas emissions
1537	Intense earthquakes in Jan and Feb causing building damage (<MM8). Seismic activity continues for the rest of the year (<MM4)		Gas emissions continue to increase
Apr – Aug 1538	Earthquake in Naples on April 20th (MM6). Seismicity progressively intensifies		Earthquakes cause “great fear among the population”
1-27th Sep 1538	Intense seismicity (<MM6). Damage to buildings, residents sleep outdoors.		
28th Sep 1538	25 – 12 hours before the eruption, ~20 earthquakes felt. All buildings badly damaged.	31 hours prior to eruption, uplift (<4.5m[a]) of sea floor begins.	Dry wells fill with water
29th Sep 1538	Increase of seismicity (<MM6)	11 hours prior to eruption, ground subsides by 4m at vent site. 7 hours prior to eruption (until the time of the eruption), subsiding area starts to rise.	A water spring emerges at centre of depression Water emerges from cracks on uplifted area Vent opens in nearby sea floor, the activity advances to the uplifted area over half an hour, increasing in intensity
18:30, 29th Sep 1538	Very strong earthquakes accompany eruption		Eruption begins explosively, and lasts for at least 3 weeks (exact length unspecified)

[a]Uplift prior to this eruption could be as high as 8 m (Parascandola, 1946, cited in Dvorak & Gasparini, 1991).

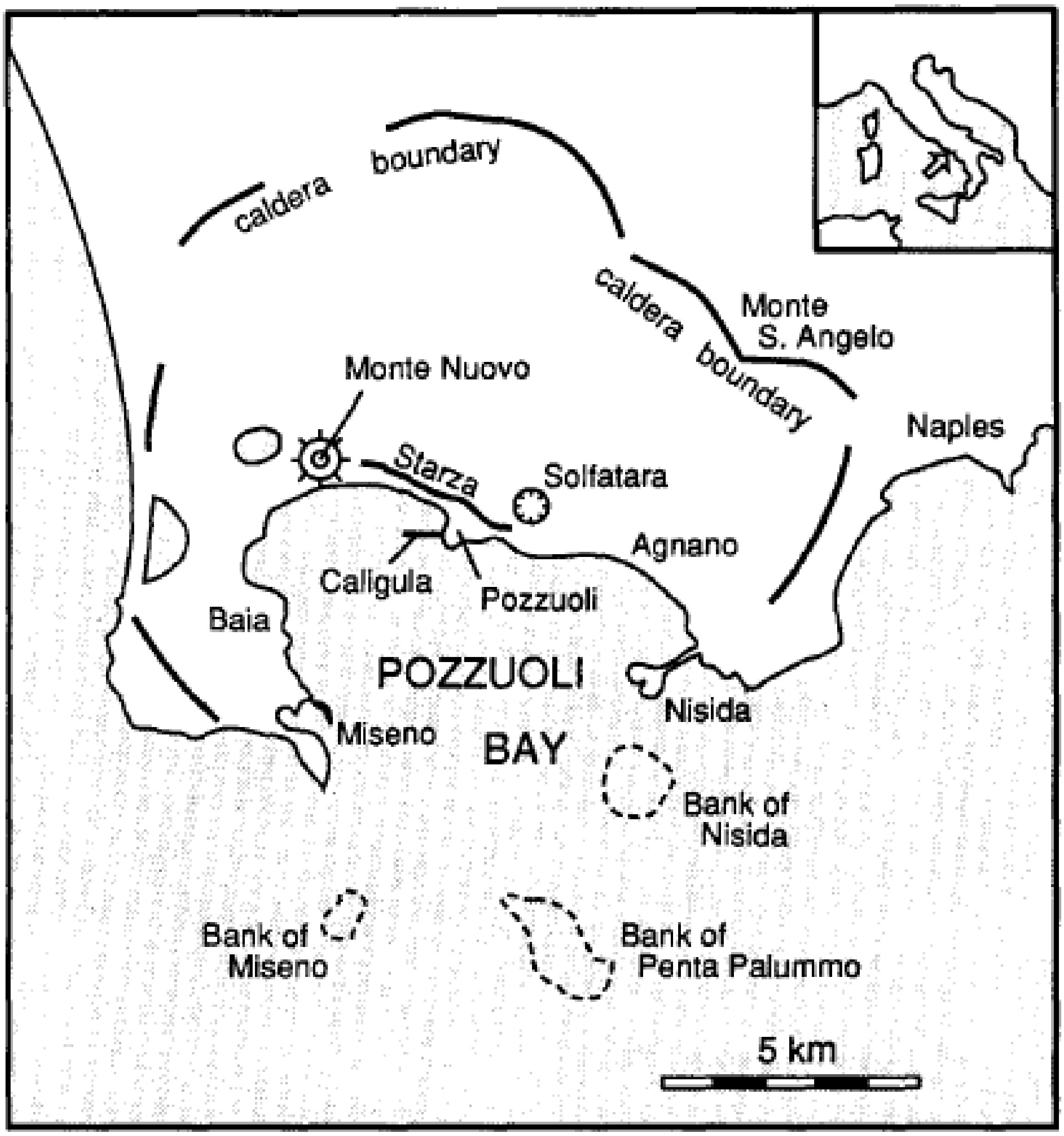

Figure 12 Campi Flegrei Caldera map, showing the location of the cities of Pozzuoli and Naples, the geothermal field Solfatara and Monte Nuovo, formed in the 1538 AD eruption. From Dvorak and Gasparini (1991).

4.2.2 Caldera unrest

A number of unrest episodes have occurred at Campi Flegrei during historical times, involving deformation, hydrothermal system changes, gas emissions and damaging seismicity. After the 1538 AD eruption, further seismic swarms occurred periodically at Campi Flegrei without any eruptions (including in 1564, 1582, 1594, 1970-72, 1983-84) (Dvorak & Gasparini, 1991).

Ground deformation has been recorded unintentionally at the harbour town of Pozzuoli, near the centre of Campi Flegrei caldera by the presence of a Roman marketplace, thought to have been built during the first and second centuries A.D. (Dvorak & Mastrolorenzo, 1991).

The marketplace, called Serapis or Macellum, contains three 13 m high marble columns (Figure 13). Approximately 4 m above the floor of the marketplace, the columns contain a 3 m band of holes created by marine molluscs. This indicates that this area has been submerged into the sea in the past 2,000 years, and then uplifted again. After the 1538 AD eruption, the elevation of the land around Pozzuoli was not substantially altered until the mid-1800's (Dvorak & Gasparini, 1991). Further Roman ruins can be found on the sea floor at Pozzuoli, 14 m below present sea level, demonstrating the overall trend of subsidence over the past few decades (Barberi et al., 1984; Dvorak & Mastrolorenzo, 1991).

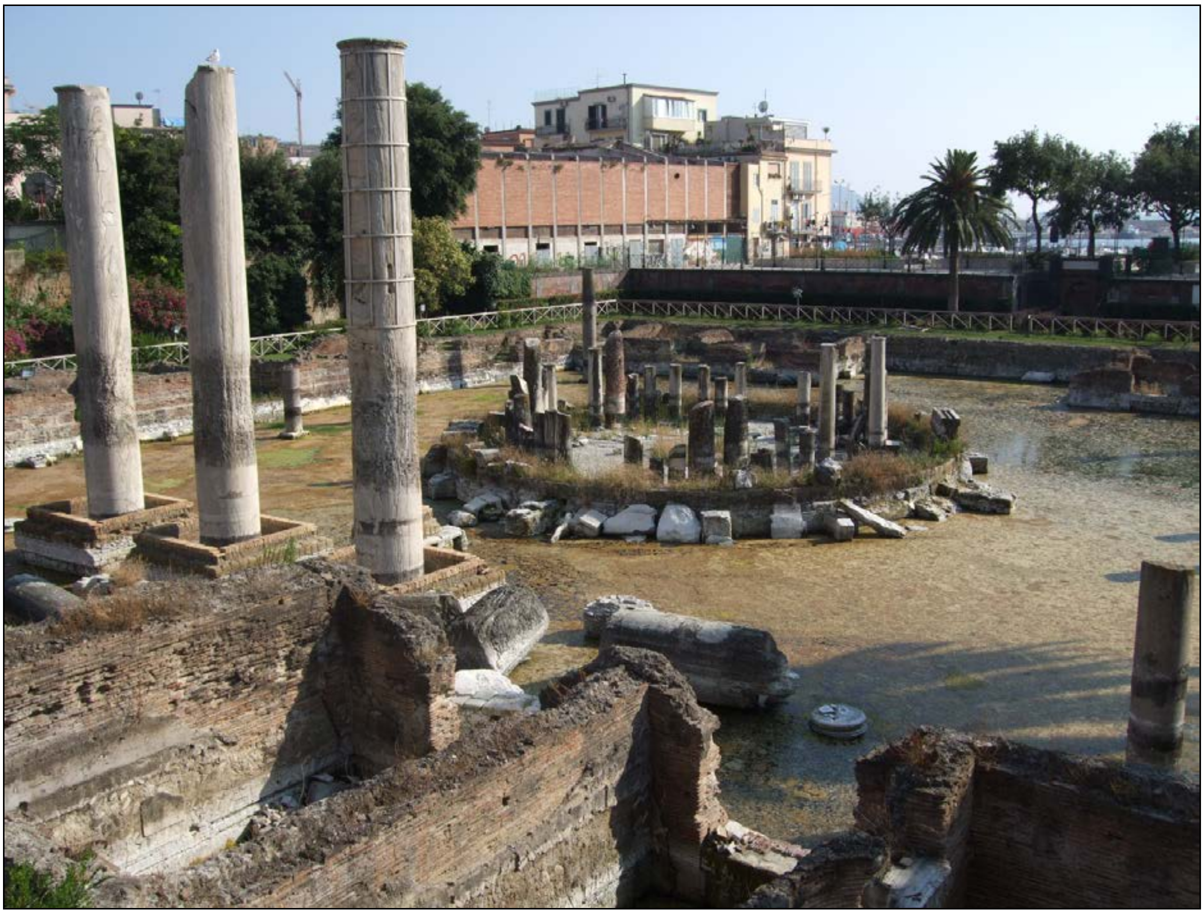

Figure 13 Serapis (or Macellum) Roman marketplace in Pozzuoli, with 13 m tall marble columns (left) showing discoloured bands of holes created by marine molluscs when submerged beneath the sea (which can be seen at the top right of the photo) during the past 2000 years. Photo by S. Potter.

Rapid uplift occurred during unrest episodes in 1969-72 and 1982-84 (Dvorak & Mastrolorenzo, 1991). The unrest episode from 1969-72 consisted of a net uplift of 1.5 m at Pozzuoli, and few felt earthquakes (seismicity was poorly monitored at that time) (Barberi & Carapezza, 1996). In 1970, an evacuation order was issued for 3000 people in Pozzuoli, which caused mass confusion and arguments over the need for evacuation, involving mass-media and lasting for the rest of the decade (Barberi et al., 1984). Different scientific panels were created and the separation of responsibility between the two was unclear, causing conflict. The need for one official Civil Defence Authority, and one official scientific agency with a minimum monitoring standard was identified. Unrest died down for the remainder of the 1970's.

Unrest in 1982-84 began with the local geothermal field (Solfatara) showing geochemical changes in the fumarole emissions, and with increases in the concentrations of gases such as H_2S and CH_4, however the temperatures stayed the same (Carapezza et al., 1984, cited in Barberi & Carapezza, 1996). This occurred before the onset of a net uplift of 3.5 m around Pozzuoli (De Natale et al., 2006). This in turn was followed seven months later by earthquake swarms causing damage to buildings, with magnitudes of up to 4.0 (Barberi & Carapezza, 1996). An evacuation of nearly 40,000 people occurred in 1983, many of whom had lived in the building structures which later collapsed (Barberi et. al., 1984). Gravity measurements were interpreted suggesting the growth of a subsurface magma chamber, however no upward migration of earthquake epicentres occurred (Barberi & Carapezza, 1996). Geochemical changes, seismicity and uplift had begun to decline by autumn 1984 (Barberi & Carapezza, 1996).

Minor unrest again occurred in 2004-06 in the form of small earthquake swarms, including periods of intense long period signals (Saccorotti et al., 2007). Based on the past, future eruptions will most likely be preceded by periods of uplift at Campi Flegrei. Whilst no eruption occurred as a result of these unrest episodes, eruptions at calderas may be a result of cumulative episodes of unrest (Dvorak & Gasparini, 1991). Therefore future episodes of unrest at this caldera will create high levels of uncertainty.

4.3 Long Valley (USA)

The Long Valley Caldera is located in eastern California, USA (Figure 14), and contains the ski resort town of Mammoth Lakes, with the neighbouring popular ski field Mammoth Mountain. Long Valley Caldera is approximately 30 km x 15 km in size, and Mammoth Mountain is a dacitic stratovolcano on the rim of the caldera boundary, which has exhibited unrest in the past few decades (Hughes, 2011). The population of this area is now approximately 8,000 permanent residents, with holiday and weekend tourists drawing an additional 15-20,000 skiers to the area per day (Mader & Blair, 1987). Volcanic eruptions at Long Valley Caldera most recently occurred approximately 250 years ago from the Mono-Inyo craters (Figure 14), and much of the area is covered in deposits from large eruptions in the past (Hildreth, 2004). Caldera unrest during the 1980's caused high levels of concern for the public and business owners due to the noticeable earthquakes as well as the perceived effect of the unrest and the way it was managed.

4.3.1 Eruptions

The Long Valley caldera-forming eruption occurred 760,000 yrs BP (Hill, 2006). Following this, a resurgent dome formed within the caldera. Smaller eruptions have also taken place to the northwest of the caldera, forming the Mono-Inyo domes, including the most recent (andesitic) eruption 250 years ago (Bursik & Sieh, 1989, cited in Hill, 2006). Tephra and ignimbrites have covered the area of the present day location of Mammoth Lakes town, indicating that it may be in danger during future eruption events (Kaye et al., 2009).

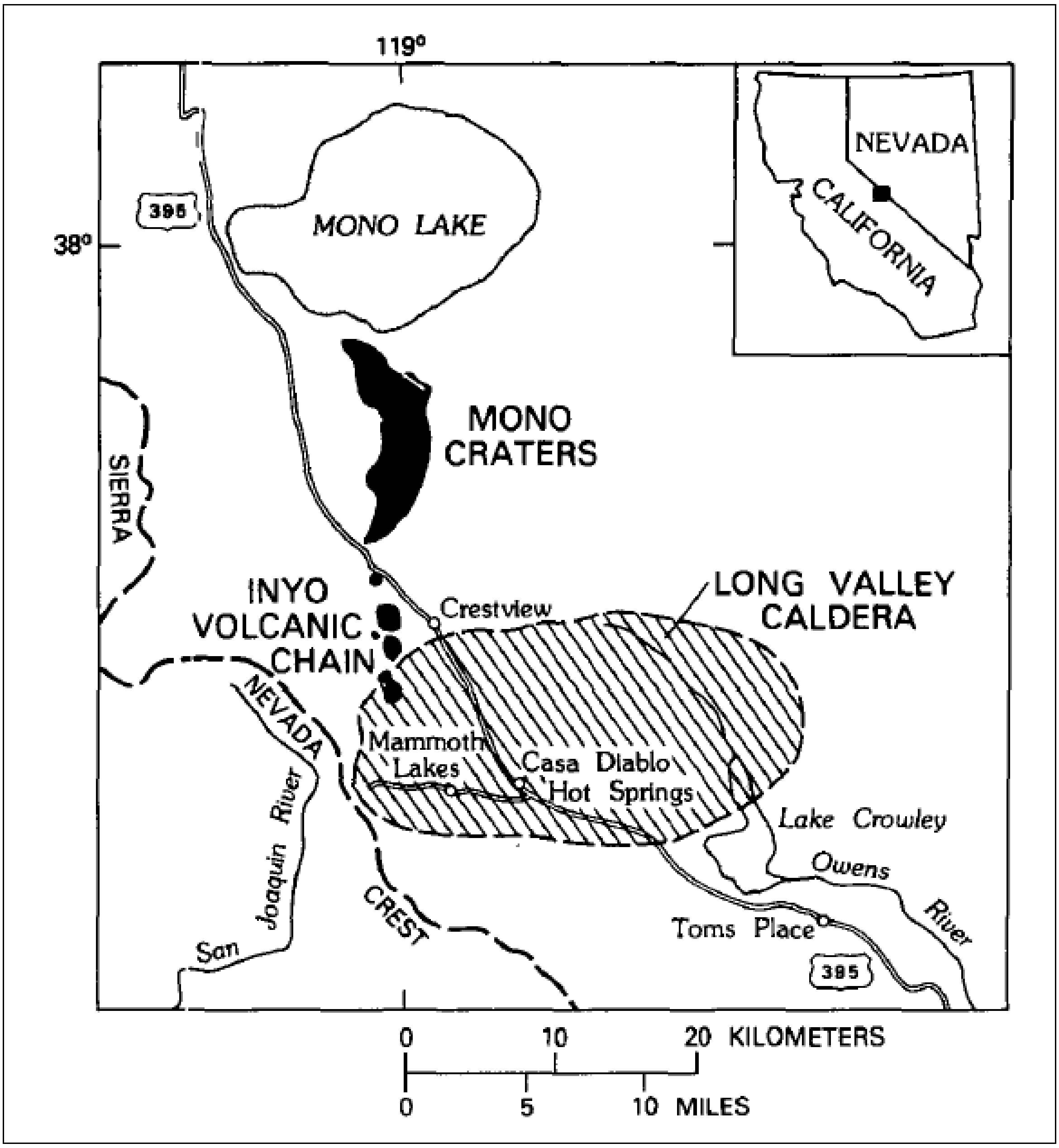

Figure 14 Long Valley Caldera, California map with the more recently active Mono-Inyo craters in black. The town of Mammoth Lakes lies within the caldera boundary on the road leading to the Mammoth Mountain ski area at the caldera margin. From Miller (1985).

4.3.2 Caldera unrest

Intensified unrest at Long Valley began in 1979, with seismicity culminating in three magnitude 6 earthquakes on 25th May 1980 (Mader & Blair, 1987). This prompted an official Earthquake Hazard Watch to be issued for the area two days later by the United States Geological Survey (USGS). Another magnitude 6 earthquake was felt later that day. Between the summer of 1979 and 1980, approximately 25 cm of uplift was recorded within the caldera (Savage & Clark, 1982, cited in Hill, 2006). Uplift continued by small amounts (7 cm) coinciding with seismic swarms until mid-1984 (Savage & Cockerham, 1984, cited in Hill, 2006). This combined with continuing seismicity of up to magnitude 5.9 just 2-3 km south of the caldera (Hill, 2006) resulted in a Notice of Potential Volcanic Hazard to be issued in May 1982.

The Notice caused outcry and alarm from the affected towns (particularly local business owners) due to the impact it had on the economy and tourist industry – the notice was seen as more of a problem than the hazard (Mader & Blair, 1987). This feeling was exacerbated by the Notice being leaked to and published by the media the day before it was officially released to the district officials and local public. It also happened to be Memorial Day Weekend, one of the busiest weekends of the year for the tourist town. The sense of mistrust of the scientists continued throughout much of the episode of unrest, to the point where there were reportedly incidents involving a scientist's car tyres being slashed, and death threats received. There was a perception that the Notice was dissuading tourists from visiting the town, however this is difficult to prove due to contributing factors, including a national recession, the bad weather during the start of the ski season, and the towns perceived negative image (Mader & Blair, 1987). The investment market and real estate industry declines were blamed on the volcanic hazard. Many of the public officials refused to believe their town was in danger from a volcanic event. This feeling continued until frequent earthquakes were felt, including shakes large enough to cause power outages in early 1983, at which time emergency plans including the construction of a second access road to the town were quickly arranged (Mader & Blair, 1987).

A change in the USGS volcanic hazard notification system in 1984 saw the removal of the Volcanic Hazard Notice. Residents saw the removal of this notice as the hazard terminating, rather than the change of system it really was (Mader & Blair, 1987). Since this time, improvements have been made in disaster preparedness and the monitoring network for the Long Valley area. The coordination of agencies and leadership of politicians proved vital during the unrest period. More cohesive systems have subsequently been developed (Mader & Blair, 1987). While unrest has declined since 1999, it continues at lower levels, including regional earthquake sequences and an average of 5 – 10 small earthquakes recorded per day in the area. Relatively high levels of CO_2 gas were emitted from 1989 to 2005 (Hill, 2006; Sorey et al., 2000), killing trees in the region and causing symptoms of asphyxia to be reported. Three members of a ski patrol died from gas poisoning on Mammoth Mountain, Long Valley caldera after falling into a snow cave melted by a fumarole (LVO monthly bulletins, GVP website). The USGS California Volcano Observatory (CalVO) website (http://volcanoes.usgs.gov/observatories/calvo/) describes the activity, of which no significant deformation changes have occurred since 2003.

4.4 Rabaul (Papua New Guinea)

Rabaul Caldera is located on the eastern end of New Britain Island, Papua New Guinea (Figures 11 and 15). It has a history of both very large and relatively small eruptions. Historical eruptions with preceding unrest have been witnessed, and eruptions from the small intra-caldera cone of Tavurvur are still occurring (Global Volcanism Program website). On the northern edge of the caldera is the town of Rabaul. The current population of Rabaul is approximately 8,000 people, however before the 1994-5 eruptions the surrounding area contained 70,000 people (McKee et al., 1985), most of whom were evacuated during the eruptive crisis.

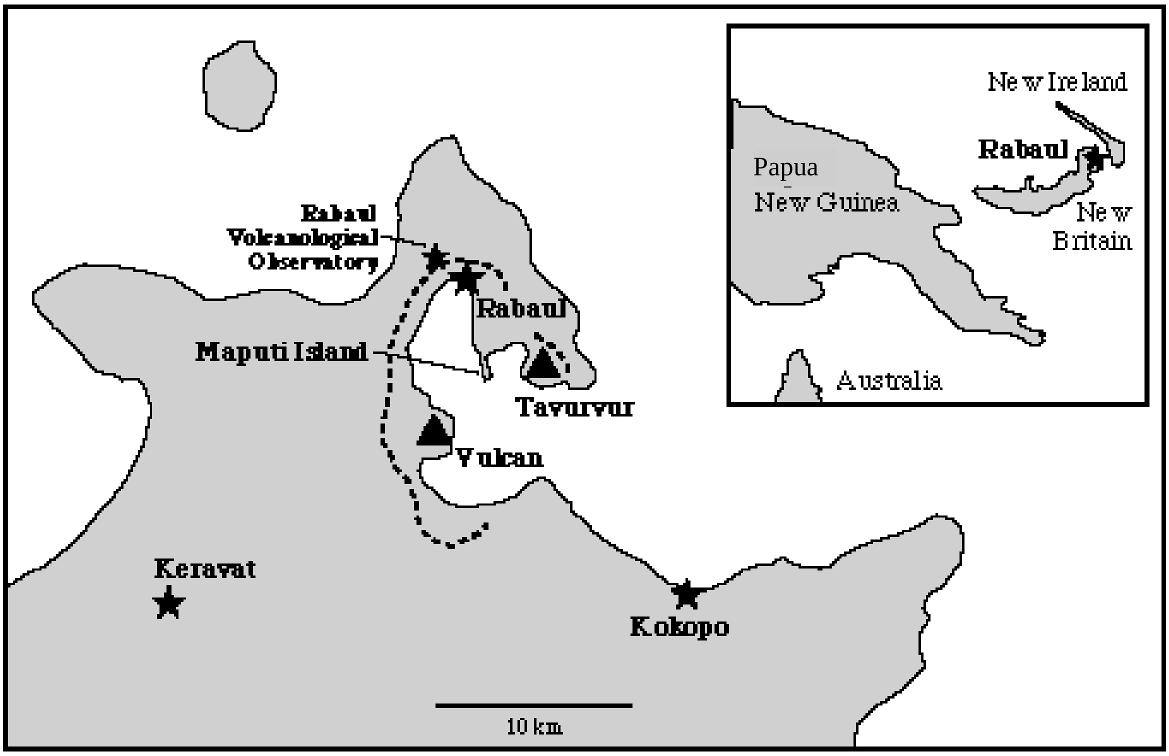

Figure 15 USGS map of Rabaul caldera (dashed line) showing location of Rabaul town and the recently active vents of Tavurvur and Vulcan (http://hvo.wr.usgs.gov/volcanowatch/1994).

4.4.1 Pre-historic eruptions

Rabaul volcano began as a largely basaltic shield, becoming more silicic over time, but it is important to recognise that it is still not rhyolitic and as such is somewhat unlike Taupo Volcanic Zone calderas. At least ten large eruptions have occurred in the 500,000 year life of the volcano (Nairn et al., 1995). The summit caldera formed to its present shape (8 km x 14 km) during the most recent of these large eruptions in 600 AD (Walker et al., 1981; Davies, 1995a). The caldera filled with water and is breached on the eastern side forming an entrance to the harbour of Blanche Bay. All 8 eruptions from within the caldera in the past 200 years have been small (0.3 km^3) (Davies, 1995a; Nairn et al., 1995). These small eruptions have built three basaltic to dacitic cones within the caldera. A number of the past eruptions at Rabaul were rhyolitic, however they are thought to have been triggered by a basaltic magma injection. This is similar to the triggering of eruptions at the Okataina Volcanic Centre in New Zealand (Nairn et al., 1995).

4.4.2 Historical caldera unrest and eruptions

Historical eruptions have occurred at Rabaul in 1767, 1791, 1850, and both Tavurvur and Vulcan erupted in 1878, 1937-43 and 1994 (McKee et al., 1985; Davies, 1995a).

The eruption in 1937 caused more than 500 fatalities in the first part of the four-day eruption (McKee et al., 1985). Precursors to this eruption included several days of intensifying seismicity, including a MM7 earthquake 30 hours before the onset of eruptive activity (Fisher, 1939, cited in McKee et al., 1985). No monitoring equipment was installed so it is not possible to accurately know the frequency and intensity of seismicity, or the amount of deformation. Johnson and Threlfall (1985), however, provide a description of the preliminary earthquakes, which were large enough to cause damage to buildings, and injure a small number of people. Rapid uplift of the harbour floor also occurred in the hours before the eruption (Johnson & Threlfall, 1985). This was in the order of metres and located within the caldera, and the resulting island was an object of curiosity that drew spectators and fish collectors to it. While some of these people escaped the eruption which began very soon after, others were killed. The eruption from Vulcan, at the site of the uplift, deposited a thick layer of ash, mud and pumice mainly towards the west. The eruption proceeded swiftly to catch the residents unaware and engulf them in darkness caused by the ash clouds. An informal evacuation of the affected areas occurred in response to the natural warning signs. Fallen trees and abandoned vehicles became obstacles as residents attempted to flee in the impenetrable darkness caused by the densely falling tephra. Issues arising during and immediately after this eruption included deep ravines in the ash deposits caused by erosion from heavy rain, flooding, electric shocks as the electricity remained on despite fallen powerlines, and suspected gas poisoning due to emissions from volcanic vents. The threat of disease was also attributed to rotting food left behind in the evacuation and standing puddles contributing towards outbreaks of malaria (Johnson & Threlfall, 1985).

The eruption of Tavurvur in June 1941 was preceded by a rise in ground temperature in the 1878 crater on Tavurvur, a large earthquake a few months before the eruption (which may or may not be related to the volcanic activity), increases in hydrothermal activity and changes in gas chemistry as recorded by the newly installed volcano observatory (Johnson & Threlfall, 1985). The eruption lasted until early to mid-1942, by which stage Rabaul town had switched from being the base for Australian WWII soldiers, to become home to the invading Japanese army. This eruption period had fluctuations in eruptive intensity, with weeks of quiescence. Rocks were thrown more than a kilometre from the active vent, setting fire to surrounding dry grass, and ash and gas clouds covered Rabaul town. No eruptive activity took place from mid-1942 until November 1943. A Japanese seismologist monitoring the volcanoes noted increasing earthquakes and ground tilt of the volcano before this last short eruptive period within the 1941 – 1943 episode (Johnson & Threlfall, 1985).

McKee et al., (1985) describe the unrest events of the 1970’s and 1980’s. Two large (M8.0) tectonic earthquakes occurred in the nearby Solomon Sea in 1971, after which changes at Rabaul Caldera began to be noticed. Uplift, tilt and changes in gravity data were accompanied by seismic swarms containing hundreds of shallow earthquakes (with a maximum magnitude of 5.2 in 1980), creating an elevated background level of activity. The level of unrest escalated in 1983, perhaps in relation to a M7.6 earthquake 200 km east of Rabaul in March 1983. The unrest consisted of short periods described as “crises”, which typically contained hundreds of recorded earthquakes in the space of an hour, with a maximum felt intensity of MM3-4 in Rabaul. Uplift within the caldera increased from a background level of 8 mm per month in the 1970's, to an average rate of 50 mm per month from November 1983 until May 1984. The maximum amount of uplift during an individual crisis was 100 mm. There was a total uplift of 3.5 m between 1971 and 1984. Gravity changes and horizontal deformation were recorded. Low frequency earthquakes were recorded, however no shallowing trend of any seismicity was observed and the crisis periods did not continue to intensify.

According to Davies (1995b), this unrest episode caused some preparedness actions to take place. An additional airstrip and wharf were constructed, and the private sector took measures to protect their equipment by storing it in safer areas. Response plans and legislation were updated. For the next decade, occasional episodes of increased seismicity and deformation occurred (Nairn & Scott, 1995). In 1990 six people were killed when overcome with CO_2 gas poisoning while collecting bird eggs in a small crater on the side of Tavurvur cone (as described by Rabaul Volcano Observatory bulletins, Global Volcanism Program website (www.volcano.si.edu)).

In September 1994 two vents within the caldera (Tavurvur and Vulcan) began erupting just 27 hours after the most recent onset of unrest in the form of two M5.1 earthquakes. Uplift of up to 6 m was observed just hours before the eruption commenced, and there was a 2 – 3 m tsunami in the harbour. There were concerns that the small eruptions could lead into a large scale eruption (Davies, 1995b). 45,000 people were evacuated, and the eruption claimed five lives (Davies, 1995a). The eruption from the Tavurvur cone (Figure 16) has remained intermittent since 1994, albeit at a smaller scale than the initial outbreak of activity. The Global Volcanism Program website contains updated information on the Rabaul eruption.

Figure 16 Tavurvur in eruption in November 2008 with the town of Rabaul in the foreground. The topography forming the southern caldera rim is in the background. Photo B.J. Scott.

Volcanic hazards including air-fall tephra and the fall of mud-rain, pyroclastic flows and surges, pumice rafts in the water, volcanic gas discharges, lightning strikes, tsunami, earthquakes, torrential runoff and lahars (volcanic mudflows) have been involved in previous eruptions at Rabaul, and are likely to be hazards in future eruptions. Very large caldera-forming eruptions could also occur in the future, causing large areas of destruction from eruption products, ground shaking and tsunami.

After the 1937 eruption, a recommendation to move the town of Rabaul was made, but the attraction of the port overrode this, and the town suffered again in the 1994 eruption (Davies, 1995b). During historical times, eruptions have occurred at Rabaul at intervals of between 24 and 59 years (Davies, 1995b), however the eruption which began in 1994 has continued (intermittently) until the present day.

Due to the rarity of the rhyolitic eruptions it is unknown how the precursory activity at these basaltic to dacitic volcanic eruptions (from within a large, previously rhyolitic volcano) differ from precursors before a predominately rhyolitic eruption, which is most likely to occur at Taupo Caldera and the Okataina Volcanic Centre.

4.5 Chaitén (Chile)

Chaitén volcano is located in the Chilean Andes (Figure 11), 10 km from the coastal town also named Chaitén. It is a rhyolitic volcano with a small (2.5 km x 4 km) summit caldera, which contains rhyolitic lava domes. Chaitén's most recent, and first historical eruption in 2008 is estimated to be the largest volcanic eruption in the world since the 1991 eruption of Hudson, also in Chile, and the largest explosive rhyolitic eruption in the world since the 1912 eruption of Novarupta, Alaska (Martin et al., 2009). The effects of this 2008 eruption included the evacuation of over 5000 people from surrounding areas as lahars swept through the town of Chaitén, airborne tephra causing airport closures disrupting international and domestic flights and impacts on the eco-tourism and aquaculture industries (Carn et al., 2009). Chaitén volcano was not scientifically monitored prior to the 2008 eruption, therefore low levels of unrest preceding the eruption would not have been recorded. Obvious signs of unrest (felt earthquakes) were only recognised for 24 hours prior to the onset of the eruption, implying unusually fast rates of rhyolitic magma movement beneath the surface. This is also supported by studies on the chemistry of the rocks (Castro & Dingwell, 2009). This has implications for hazard mitigation at rhyolitic volcanoes, particularly those not well-monitored, due to the very limited warning time.

4.5.1 Eruptions

The caldera-forming eruption at Chaitén occurred at approximately 9,370 yrs BP (Naranjo & Stern, 2004). This eruption was small to medium in size and consisted of a pyroclastic surge and tephra fall, followed by the deposition of mafic scoria (Naranjo & Stern, 2004). Following this eruption the caldera was partially filled by a rhyolitic lava dome.

The historical eruption of Chaitén began either in the evening of 1st May (Carn et al., 2009; Castro & Dingwell, 2009) or on the morning of 2nd May 2008 (Martin et al., 2009; Global Volcanism Program (GVP) monthly reports). Plinian eruptions with ash columns up to 21 km in height continued for a week before the growth of a new lava dome in the caldera began, and continued for nearly 3 years (Castro & Dingwell, 2009; GVP website: Chaitén monthly reports). A section of the lava dome collapsed in February 2009, resulting in a lateral blast, pyroclastic flows and ashfall in surrounding areas (Carn et al., 2009). Further block and ash flows have occurred as portions of the lava dome continued to collapse. Within a few days of the onset of eruptive activity the entire town of Chaitén was evacuated. The town was subsequently overrun with lahars. Growth of the lava dome diminished throughout 2011 (GVP website: Chaitén monthly reports).

4.5.2 Caldera unrest

Prior to the 2008 eruption at Chaitén, earthquakes were observed on monitoring equipment up to 300 km distant from the volcano on the evening of 30th April. The seismicity included volcano-tectonic earthquakes, with up to 15-20 per hour seen in retrospective analysis (Carn et al., 2009). These earthquakes were felt in the town of Chaitén, and were strong enough to knock objects off shelves (Castro & Dingwell, 2009). No hydrothermal, gas or deformation measurements were made. Castro and Dingwell (2009) analysed crystals from the deposits of the following eruption and interpreted the results to conclude the magma ascended from a depth of 5 km to the surface in just 4 hours. This very rapid movement of magma has not been documented at a rhyolitic volcano before. It has implications on the amount of warning time which can occur before an eruption takes place. While the Chaitén summit caldera is quite small compared to Taupo and Okataina calderas, it is a rhyolitic volcano with a history of explosive eruptions, one of which occurred in the recent past. Unlike the more recent eruptions at Rabaul, Long Valley, Campi Flegrei and Okataina volcanic centres, the Chaitén historical eruption consisted of rhyolitic magma. Its rapid onset of eruptive activity after only one day of unrest demonstrates the possibilities for future behaviour at New Zealand's calderas.

4.6 Yellowstone (U.S.A.)

Yellowstone Plateau volcanic field is located in Wyoming, United States (Figure 17). It contains three calderas and a number of smaller vents outside of the caldera boundaries. The Yellowstone National Park surrounds this area, drawing millions of tourists every year. Regular caldera unrest continues to occur, requiring multi-agency coordination to manage the potential volcanic hazards. The geographical size of Yellowstone's calderas are comparable to those of the Taupo Volcanic Zone (Figure 17), and both have a similar discharge rate of magma in the past 2.2 million years (Houghton et al., 1995a). However the TVZ has had more frequent and smaller eruptions than Yellowstone (Houghton et al., 1995a).

4.6.1 Eruptions

The volcanic history of the Yellowstone Plateau volcanic field, including its calderas, is described by Christiansen (2001). The first caldera-forming rhyolitic eruption at Yellowstone Plateau volcanic field occurred just over 2 million yrs BP, erupting a volume of approximately 2,500 km^3. Further large eruptions occurred 1.3 and 0.64 million yrs BP, with volumes of 280 km^3 and 1000 km^3 respectively. This most recent caldera-forming eruption affected the Earth's climate by reducing the intensity of solar radiation entering the Earth's atmosphere (Dzurisin et al., 1995).

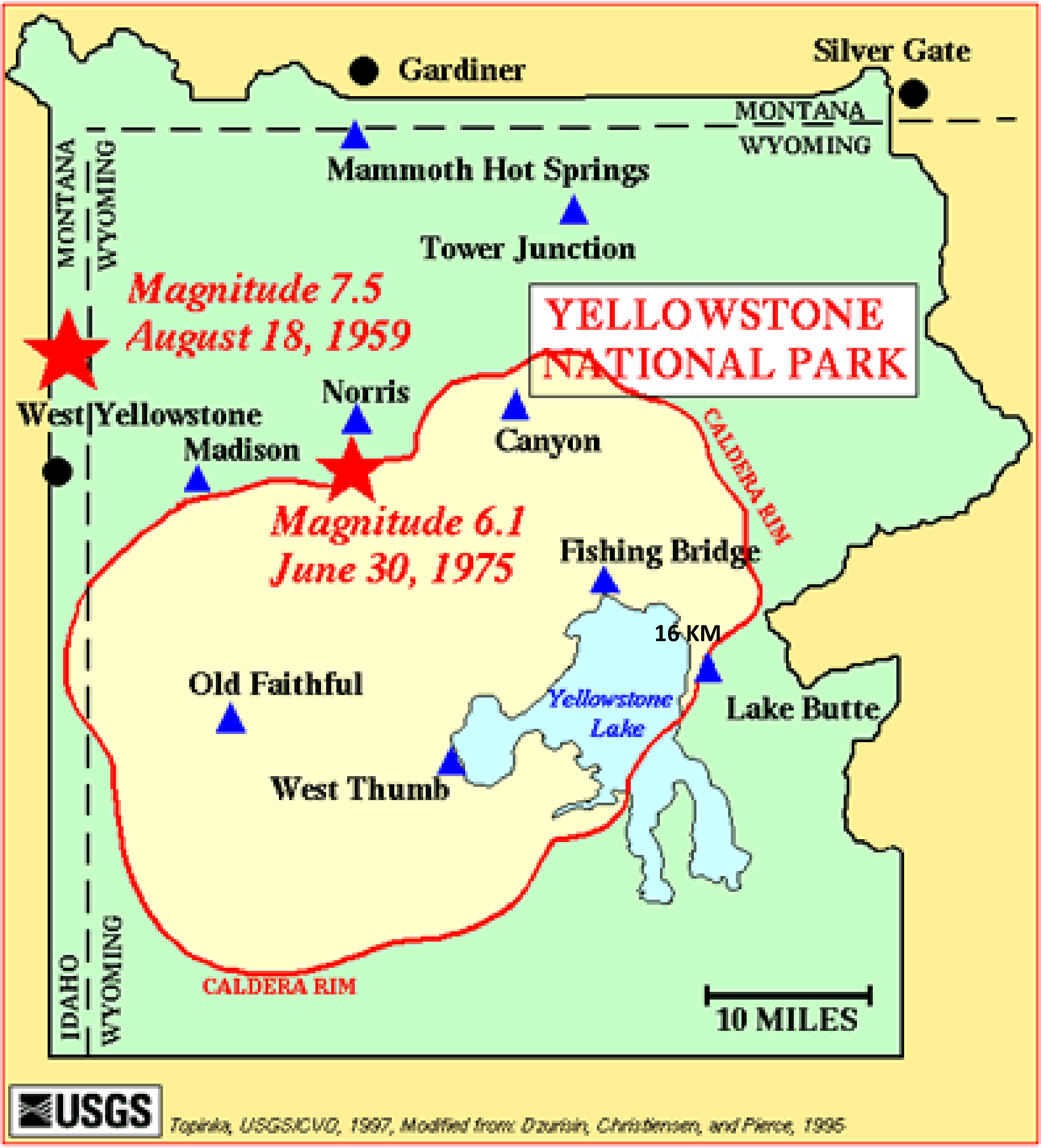

Figure 17 USGS map of Yellowstone National Park and the caldera boundary from the 0.64 million yrs BP eruption. It also shows epicentres of large earthquakes in the past, and major hydrothermal features.

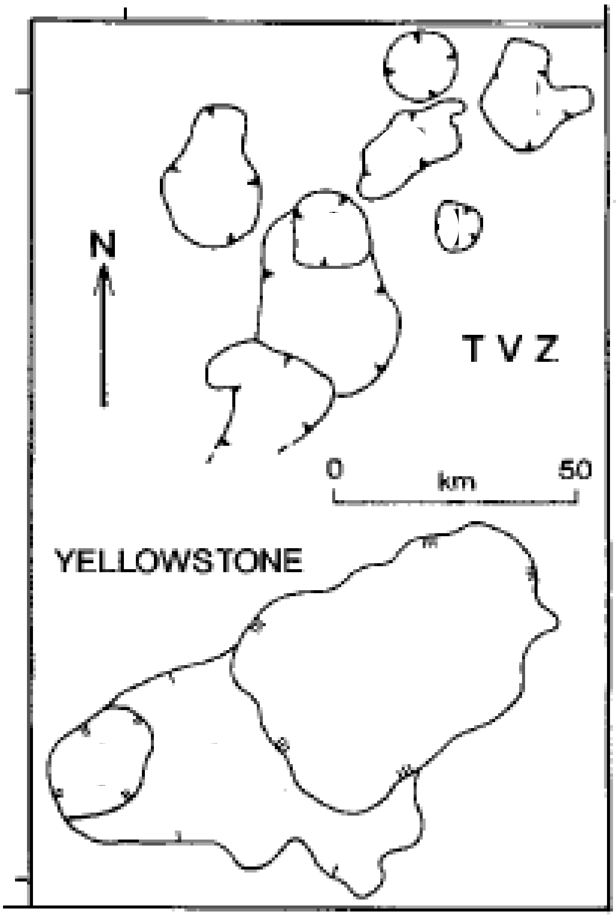

Figure 18 A comparison in size of calderas in the Taupo Volcanic Zone (TVZ), New Zealand (top, see Figure 1 for further details), to Yellowstone (bottom). From Houghton et al., (1995a).

Prior to each of these large eruptions, rhyolitic lavas (with volumes of over 10 km^3) were extruded. Following the youngest of these caldera-forming eruptions, rhyolitic magma was injected beneath the caldera floor to form resurgent domes. Basaltic lava has also been erupted around the edges of the calderas. The most recent eruption from within the Yellowstone Caldera was a rhyolitic lava extrusion, at approximately 72,000 yrs BP (Christiansen et al., 2007).

4.6.2 Caldera unrest

The region is intensely monitored for volcanic unrest by the USGS at the Yellowstone Volcano Observatory (YVO) and the University of Utah. The Yellowstone area undergoes regional seismicity including potentially large earthquakes from local fault lines (such as a M6.1 earthquake in 1975) and neighbouring mountain ranges (such as a nearby M7.5 earthquake in 1959 which costed 28 lives; Dzurisin et al., 1995). Smaller earthquake swarms within the caldera area are also experienced. The largest seismic swarm to be recorded at Yellowstone occurred in 1985, coinciding with subsidence of the caldera (Waite & Smith, 2002; cited in Christiansen et al., 2007)). Further swarms were recorded in 2004, 2009 and January-April 2010. This latter swarm included 16 earthquakes >M3.0, a few of which were felt (YVO 2010 news archive).

Ground deformation has also occurred at Yellowstone and in the surrounding area. Deformation during historical times at Yellowstone has included uplift of 23 mm per year from 1976-83, then subsidence of up to 35 mm per year until 1987 (Dzurisin & Yamashita, 1987; Dzurisin et al., 1990, both cited in Christiansen et al., 2007). Subsidence and uplift at rates of up to 60 mm per year continued to occur; most recently slight subsidence is being experienced (YVO October monthly update; Christiansen et al., 2007). In the past 10,000 years, the centre of Yellowstone caldera has moved up and down by about 20 m at least three times (Dzurisin et al., 1995).

In more than 130 years of historical records for Yellowstone National Park, at least 25 hydrothermal eruptions have occurred. Hydrothermal eruption craters several kilometres across have formed within Yellowstone National Park, however none have been associated with a volcanic event (Christiansen et al., 2007).

Due to its geological past and unrest during historical times, Yellowstone Plateau has similarities with the calderas of the Taupo Volcanic Zone, despite the Yellowstone Plateau's tectonic setting as a hotspot rather than on a subduction zone. Yellowstone has a different situation regarding risk, as it has a very low permanent population, but a high number of seasonal tourists during the day.

4.7 Aira Caldera and Sakurajima (Japan)

Aira Caldera is one of the largest volcanoes in the southern Japanese island of Kyushu. It forms the northern end of Kagoshima Bay, and contains the post-caldera cone of Sakurajima, one of the most active volcanoes in Japan. Kagoshima city, with a population of over 600,000, is located near the south-western caldera boundary (Figure 18). The residents have adapted to living with the effects of frequent tephra fall from this reawakened rhyolitic caldera.

4.7.1 Eruptions

Aira Caldera formed approximately 22,000 yrs BP during a large, ignimbrite-forming eruption (Kigoshi et al., 1972; cited in Aramaki, 1984). This rhyolitic eruption included widespread pumice falls and pyroclastic flows, with a total erupted volume of >140 km^3 (Aramaki, 1984). In the 22,000 years after this eruption, three small post-caldera cones and a small caldera were formed, which lie submerged on the caldera floor in Kagoshima Bay, at least one of which shows vigorous fumarolic activity releasing CO_2 gas (Aramaki, 1984). Sakurajima stratocone is the largest and most active post-caldera vent. This andesitic to dacitic cone formed on the southern caldera rim at least 13,000 yrs BP (Fukuyama, 1978; cited in Aramaki, 1984), and is now one of Japan's most active volcanoes. The larger historical eruptions from Sakurajima cone occurred in 1471-76, 1779 and 1914.

Figure 19 Sakurajima post-caldera volcano, located in Aira caldera, emitting steam. Kagoshima city is in the foreground.

Historical eruptions from Sakurajima have caused frequent deposits of tephra on the city of Kagoshima, located 8 km from the summit of the active cone across Kagoshima Bay. Sakurajima has had ongoing eruptions since 1955 with frequent but relatively small ash columns. The city of Kagoshima has adapted to the frequent eruptions from this nearby volcano by implementing measures to deal with volcanic hazards. Many of the buildings have large, overhanging roofs covering balconies to limit the build-up of tephra on these weak structures. Some roofs don't have gutters, but have a channel on the ground beneath the roof to enable easy cleaning of the tephra off the roof. Hard hats have been issued to children walking to school, and firefighters make regular patrols during eruptions. Further examples of Kagoshima adapting to regular eruptions are on the Taranaki Blowout exercise webpage (http://www.trc.govt.nz/taranaki-blowout-background-info). Lessons for New Zealand taken from Kagoshima on coping with frequent ashfall is described in a GNS Science report by Durand et al., (2001).

4.7.2 Caldera unrest

Aira Caldera has undergone frequent unrest during historical times, as described by Newhall and Dzurisin (1988), particularly before and after each eruption at Sakurajima cone. An example of this is before and after the 1914 eruption of <2 km^3 of dacitic magma from Sakurajima. In the preceding approximately 50 years, uplift occurred on Sakurajima's west coast of at least 1.5 m, and on the northwestern shoreline of Kagoshima Bay of 1 m. The rate of uplift increased until the eruption. In June 1913, earthquake swarms were centred 55 km and 15 km from Kagoshima, and springs changed temperature and flow rate on the edge of Sakurajima Island. Seismic activity continued in the week before the eruption, increasing in intensity in the final 30 hours at Kagoshima city, while ten times as many were felt on Sakurajima. On the morning of the onset of eruptive activity, hot and cold springs emerged around Sakurajima island, some spouting to a height of 1 m. The eruption began small, and one of the largest known earthquakes associated with volcanic activity occurred 8 hours after the start of the eruption, with a magnitude of 7.0 (Abe, 1981; cited in Newhall & Dzurisin, 1988). Dacitic lava flowed down Sakurajima's flanks during this eruption, joining the island to the mainland.

After the 1914 eruption, the caldera floor subsided by up to 6 m, and then started to slowly uplift for the remainder of the century. In the months following the eruption, areas of hot soil killed vegetation 600 m from the vent, and volcanic gases killed an ox and made people ill (Omori, 1916; cited in Newhall & Dzurisin, 1988). After a year the soil temperature returned to normal. Unrest continues to occur at Aira Caldera, accompanied by eruptions from its frequently active Sakurajima cone.

4.8 Taal (Philippines)

Taal caldera is located in southwestern Luzon in the Philippines, and formed between 100,000 and 500,000 yrs BP (Listanco, 1994; cited in Lowry et al., 1991). The 15 km x 25 km caldera contains Lake Taal. Within Lake Taal is the 5 km wide Volcano Island, the source of all of the historical eruptions, and home to several thousand people. On Volcano Island is a small (3 km) caldera lake (called Main Crater Lake) which itself has a small island, a remnant of historical eruptions. At least 33 eruptions have been witnessed at Taal since the 16th Century (Punongbayan & Tilling, 1989; cited in Bartel et al., 2003), including pyroclastic flows and surges which have caused many fatalities, especially from villages on Volcano Island. These eruptions have been basaltic to dacitic in composition (Bartel et al., 2003). An eruption in 1911 from Volcano Island killed about 1335 people from pyroclastic flows (Blong, 1984). The most recent eruption ceased in 1977 (GVP website).

Unrest occurred at Taal Caldera in 1992 and 1994, without resulting in an eruption (Bartel et al., 2003). Rates of uplift in 1992 were up to 21 cm per day (Gabinete, 1999; cited in Bartel et al., 2003). Other unrest phenomena during this and the 1994 episode included heightened seismicity, and changes in the lake water chemistry and temperature. Both of these unrest episodes have been attributed to dike intrusions (Bartel et al., 2003). Geysering has been observed, including in 1998 and 1999. Deformation has fluctuated in the past couple of decades between inflation and deflation, each trend lasting on the order of months (Bartel et al., 2003). Volcanic earthquakes occurred in August 2008, which were heard and felt by the island residents. Between April and June 2010 the number and intensity of earthquakes increased, there was a slight inflation noted, gas emissions changed, fumaroles intermittently increased output and the temperature of the Main Crater Lake increased by a few degrees (according to the Philippine Institute of Volcanology and Seismology (PHIVOLCS) as seen on the GVP website). Residents were advised (but not ordered) to leave, however most did not comply (*Philippine Daily Inquirer*, cited on the GVP website). Further changes to the volcanic parameters continued at lower levels throughout 2010 and into 2011, before abating in mid-2011 (GVP and PHIVOLCS (http://www.phivolcs.dost.gov.ph) websites).

4.9 Novarupta and Katmai (U.S.A)

The largest eruption in the world during the 20th Century took place from 6 – 9 June 1912 in a remote area of Alaska, U.S.A, and is described by Hildreth (1983, 1991). Due to this area's highly remote location, unrest phenomena preceding this rhyolitic eruption unfortunately remain unknown.

The eruption formed a new vent called Novarupta. Approximately 15 km^3 of magma was erupted, covering the valley floor in up to 250 m of pyroclastic material including ignimbrite sheets. The valley is surrounded by five dacitic to andesitic stratovolcanoes, and was named the 'Valley of Ten Thousand Smokes' due to the hot ignimbrite deposits issuing steam through cracks for many years. Mt Katmai lies 10 km to the east of the 1912 vent. During the

Novarupta eruption, the Mt Katmai summit caldera (with a 1.5 km diameter) formed due to magma withdrawal, and this eruption has often mistakenly been attributed to Mt Katmai rather than Novarupta. The first half of the Novarupta eruption was rhyolitic, depositing pumice and destructive pyroclastic flow deposits. The remaining deposits indicate dacitic magma (Hildreth, 1991).

No recorded unrest has been observed at Novarupta since it formed in 1912. Nearby volcanoes, such as the stratovolcano Trident have however erupted during historical times. The permanent monitoring network at Novarupta was installed in the 1990s by the USGS Alaska Volcano Observatory.

5.0 GLOSSARY

Andesite	(Or andesitic) Volcanic rock (or lava) containing 54 to 62% silica and moderate amounts of iron and magnesium.
Ash	Fine particles of pulverized rock (tephra) erupted from the vent of a volcano. Particles smaller than 2 mm in diameter are termed as ash, and may be solid or molten when first erupted.
Ashfall	Volcanic ash that has fallen through the air from an eruption cloud.
Ballistic	Large tephra particles with diameters of over 64mm. Includes blocks and bombs.
Ballistic projectile	A block or bomb explosively ejected from the vent that is not carried upwards by the eruption column.
Basalt	(Or basaltic) Volcanic rock (or lava) containing less silica than andesite, commonly producing more effusive, runny and less explosive lava.
Base surge	Volcanic density current pulse that moves laterally outwards formed of a dilute, turbulent mixture of hot gas (steam), water and solid ejecta.
Block	Angular chunk of solid rock ejected during an eruption, with diameters of over 64 mm.
Block and ash flow	An avalanche of ash, hot gas and potentially large blocks from oversteepening of a lava front or dome. These can travel at tens of kilometres per hour, can be hundreds of degrees in temperature and cover distances of several kilometers.
Bomb	Fragment of molten or semi-molten rock, with a diameter of over 64 mm. Because of their plastic condition, bombs are often modified in shape during their flight or upon impact.

Caldera	A volcanic depression with a diameter many times larger than the size of the individual vents, usually formed during large volcanic eruptions.
CDEM	Civil Defence and Emergency Management
CO	Carbon monoxide.
CO_2	Carbon dioxide.
Conduit	A passage followed by magma within a volcano.
Country Rocks	The existing rock intruded by and surrounding an igneous intrusion (magma).
Crater	A commonly circular depression formed by either explosion or collapse at a volcanic vent, from which volcanic material is ejected.
Dacite	(Or dacitic) Fine-grained rock intermediate in composition between andesite and rhyolite.
Debris Avalanche	A rapid and unusually sudden sliding or flow of unsorted rock and other material (such as fragmented cold and hot volcanic rock, water, snow/ice and trees).
Deformation	Ground movement in a vertical or horizontal direction.
Dome	A steep-sided mass of viscous lava extruded from a volcanic vent. Its surface is often rough and blocky as a result of fragmentation of the cooler, outer crust during growth of the dome.
Ejecta	Material that is thrown out by a volcano, including tephra.
Eruption Column	The cloud of gases, steam and tephra rising from a crater or other vent, driven by thermal convection and gas pressure. If it is of sufficient volume and velocity, this column may reach many kilometers into the stratosphere, where winds may carry it long distances. Eruption columns can collapse and form pyroclastic density currents.
Eruptive Vent	The opening through which volcanic material is emitted.
Extinct Volcano	A volcano that is not presently erupting and is not likely to do so for a very long time in the future, if ever.
Extrusion	The emission of magmatic material at the earth's surface. Also, the structure or form produced by the process (e.g. lava flow, volcanic dome).

Fault	Fracture or zone of fractures along which displacement takes place or has taken place in the past.
Fissures	Elongated fractures or cracks on the slopes of a volcano. Fissures can host eruptions, which typically consist of runny lava flows and fountains, but pyroclastics (tephra) may also be ejected.
Flank Eruption	An eruption from the side of a volcano (in contrast to a summit eruption.)
Fumarole	A vent or hole through which steam and other gases emit.
Gravimetric	The measurement of microgravity, which can indicate the presence of a subsurface magma body.
H_2S	Hydrogen Sulphide, a poisonous gas.
Harmonic Tremor	A continuous release of seismic energy typically associated with the underground movement of magma. It contrasts distinctly with the sudden release and rapid decrease of seismic energy associated with the more common type of earthquake caused by slippage along a fault.
Hydrothermal eruption	Explosion driven by the transformation of hot groundwater to steam.
Igneous	The type of rocks formed during volcanic activity, both above and below the ground surface.
Ignimbrite	The rock formed by the widespread deposition and consolidation of hot pyroclastic flows. The term was originally applied only to densely welded deposits but now includes non-welded deposits.
Intensity	A measure of the effects of an earthquake at a particular place. Intensity depends not only on the magnitude of the earthquake, but also on the distance from the epicenter and the local geology.
Intrusion	The process of emplacement of magma in pre-existing rock. Also, the term refers to the igneous rock mass so formed within the surrounding rock.
Lahar	A flow of water-saturated, typically dense volcanic material, resembling a flow of wet concrete. Lahars usually follow topographical lows, however may overtop banks. They may be unaccompanied by an eruption by remobilisation of volcanic material.
Lapilli	Literally, "little stones." Round to angular erupted rock fragments (tephra) measuring 2 to 64 mm in size in diameter, which may be ejected in either a solid or molten state.

Lava Magma which has reached the surface through a volcanic eruption. The term is most commonly applied to the flowing rock that emits from a crater or fissure, however it also refers to cooled and solidified rock formed this way. Lava varies in viscosity (runniness and therefore speed of movement), chemistry and temperature.

Lava Dome Mass of sticky lava, that has built a dome-shaped pile at a vent.

Liquefaction A saturated soil loses strength and behaves as a liquid due to an applied stress, usually earthquake shaking.

Lithic Particle of previously formed rock.

Mafic Magma with a silica content of less than about 55%.

Magma Molten rock beneath the surface of the earth. Magma that reaches the surface erupts as lava or pyroclasts.

Magma Chamber The underground reservoir containing the molten magma beneath a volcano.

Magnitude Earthquake magnitudes in this report represented by a single 'M' (i.e. M5.0) refer to the Richter Magnitude scale.

Mantle The zone of the earth below the earth's crust and above the core.

MM Modified Mercalli earthquake intensity scale (see the GeoNet website: http://www.geonet.org.nz/earthquake/modified-mercalli-intensity-scale.html).

Phreatic Eruption (Or phreatically) An explosion caused when water and heated volcanic rocks interact to produce a violent expulsion of steam and pulverized rocks. Magma is not involved.

Phreatomagmatic An explosive volcanic eruption that results from the interaction of surface or subsurface water and magma.

Plinian An eruption with a powerful, convecting column reaching up to 45 km high, usually requiring the eruption of high viscosity magma (such as dacite and rhyolite). Plinian eruptions often lead to the formation of pyroclastic density currents.

Ppm parts per million.

Pumice Light-coloured, frothy volcanic rock, formed by the expansion of gas in erupting, sticky lava during an eruption. Pumice commonly floats on water and can travel further than other rocks of a similar size during an eruption due to their low density.

Pyroclastic	Erupted material which starts out hot (pyro) and consists of fragmented rock (clastic) material formed by a volcanic explosion.
Pyroclastic [Base] Surge	A type of pyroclastic density current which has high gas content, is turbulent and the material is well mixed.
Pyroclastic Density Current	A gravity-controlled, laterally moving mixture of pyroclasts and gas.
Pyroclastic Flow	A turbulent mixture of hot gases and rock fragments that can move at high speed (up to 900 km an hour) down the sides of the volcano. A type of pyroclastic density current, which usually follows topographical lows. Generated by the collapse or partial collapse of an eruption column.
Quaternary	The period of Earth's history from about 2 million years ago to the present; also, the rocks and deposits of that age.
Quiescence	The periods of time between eruptions.
Rhyolite	Volcanic rock, light coloured, with a high silica content.
Scoria	A pyroclast that is irregular in form and generally very vesicular. It is usually heavier, darker, and more crystalline than pumice.
Seismograph	An instrument that records seismic waves (earthquakes).
Seismologist	Scientists who study seismicity (earthquakes).
Silica	(Or silicic) A chemical combination of silicon and oxygen (SiO_2)
SO_2	Sulphur dioxide (gas)
Stratocone	See stratovolcano
Stratovolcano	A volcano composed of both lava flows and pyroclastic material.
Strombolian	Basaltic (low viscosity magma) eruptions, including a series of explosions.
Subduction Zone	The zone of convergence of two tectonic plates, one of which usually overrides the other.
Subplinian	Lower magnitude and intensity versions of the plinian eruption, can result in pyroclastic density currents.
Surge	A cloud of gas and suspended pyroclastic material that moves radially outward at high velocity from the base of a vertical eruption column accompanying a volcanic eruption.
Swarm	A group of many earthquakes of similar size occurring closely clustered in space and time with no dominant main shock.

Tephra Solid materials of all types and sizes that are erupted from a crater or volcanic vent and travelling through the air.

Tilt The angle between the slope of a part of a volcano and some reference. The reference may be the slope of the volcano at some previous time.

Tremor Low amplitude, continuous earthquake activity often associated with magma movement.

Tsunami A great sea wave produced by a submarine earthquake, volcanic eruption, or large landslide.

Vent The opening at the earth's surface through which volcanic materials emit, or emitted in the past.

Vesicle A small air pocket or cavity formed in volcanic rock during solidification.

Viscosity A measure of resistance to flow in a liquid (water has low viscosity while honey has a higher viscosity.)

Volcano A vent in the surface of the Earth through which magma and associated gases erupt, and the form or structure that is produced by the ejected material.

Volcanogenic A process attributed to a volcano or volcanic activity.

Vulcanian An eruption style of an explosive event of <1 km^3 in volume, but with an eruption column reaching 10-20 km high.

6.0 REFERENCES

Aramaki, S. (1984). Formation of the Aira Caldera, southern Kyshu, approximately 22,000 years ago. Journal of Geophysical Research, 89(NB10), 8485-8501.

Barberi, F. and Carapezza, M. L. (1996). The problem of volcanic unrest: the Campi Flegrei case history. In R. Scarpa & R. I. Tilling (Eds.), *Monitoring and Mitigation of Volcano Hazards.* (pp. 771-786): Springer, Berlin.

Barberi, F., Corrado, G., Innocenti, F., Luongo, G. (1984). Phlegraean Fields 1982–1984: brief chronicle of a volcano emergency in a densely populated area. Bulletin of Volcanology, 47(2), 175-185.

Bard, E. (1998). Geochemical and geophysical implications of the radiocarbon calibration. Geochimica et Cosmochimica Acta, 62(12), 2025-2038.

Bartel, B. A., Hamburger, M. W., Meertens, C. M., Lowry, A. R., Corpuz, E. (2003). Dynamics of active magmatic and hydrothermal systems at Taal Volcano, Philippines, from continuous GPS measurements. Journal of Geophysical Research-Solid Earth, 108(B10).

Baxter P. J., Kapila M., Mfonfu D. (1989). Lake Nyos disaster, Cameroon, 1986: the medical effects of large scale emission of carbon dioxide? British Medical Journal 298 (6685): 1437.

Becker, J. S., Saunders, W. S. A., Robertson, C. M., Leonard, G. S., Johnston, D. M. (2010). A synthesis of challenges and opportunities for reducing volcanic risk through land use planning in New Zealand. Australasian Journal of disaster and trauma studies, 2010-1: 24 p.

Bibby, H. M., Caldwell, T. G., Davey, F. J., Webb, T. H. (1995). Geophysical evidence on the structure of the Taupo Volcanic Zone and its hydrothermal circulation. Journal of Volcanology and Geothermal Research, 68: 29-58

Blong, R. J. (1984). Volcanic hazards: a sourcebook on the effects of eruptions. Australia: Academic Press, Inc. 424p.

Bromley, C. J. and Clotworthy, A. W. (2001). Mechanisms for water level declines in Alum Lakes, Wairakei. Paper presented at the 23rd New Zealand Geothermal Workshop 2001, The University of Auckland.

Browne, P. R. L. and Lawless, J. V. (2001). Characteristics of hydrothermal eruptions, with examples from New Zealand and elsewhere. Earth-Science Reviews, 52: 299-331.

Bryan, C. J., Sherburn, S., Bibby, H. M., Bannister, S. C., Hurst, A. W. (1999). Shallow seismicity of the central Taupo Volcanic Zone, New Zealand: its distribution and nature. New Zealand Journal of Geology and Geophysics, 42(4), 533-542.

Buck, M. D. (1985). An assessment of volcanic risk on and from Mayor Island, New Zealand. New Zealand Journal of Geology and Geophysics, 28, 283-298.

California Volcano Observatory (CalVO) USGS website (http://volcanoes.usgs.gov/ observatories/calvo/) – accessed 21 February 2012.

Carn, S., Pallister, J., Lara, L., Ewert, J., Watt, S., Prata, A., et al. (2009). The unexpected awakening of Chaitén volcano, Chile. Eos, 90(24), 205-212.

Castro, J. M. and Dingwell, D. B. (2009). Rapid ascent of rhyolitic magma at Chaitén volcano, Chile. Nature, 461(7265), 780-783.

Christiansen, R. L. (2001). The Quaternary and Pliocene Yellowstone plateau volcanic field of Wyoming, Idaho, and Montana. US Geological Survey Vol. 729.

Christiansen, R. L., Lowenstern, J. B., Smith, R. B., Heasler, H., Morgan, L. A., Nathenson, M., et al. (2007). Preliminary assessment of volcanic and hydrothermal hazards in Yellowstone National Park and vicinity (No. 2007-1071). Reston, Va.: U.S. Geological Survey.

Christenson, B. W., Werner, C. A., Reyes, A. G., Sherburn, S., Scott, B. J., Miller, C., Rosenberg, M. J., Hurst, A. W., Britten, K. A. (2007). Hazards from hydrothermal sealed volcanic conduits. EOS 88(5).

Cole, J. W., Cowan, H., Webb, T. H. (2006). The 2006 Raoul Island Eruption – a review of GNS Science's Actions. GNS Science Report 2006/07. 38p.

Cole, J. W., Spinks, K. D., Deering, C. D., Nairn, I. A., Leonard, G. S. (2010). Volcanic and structural evolution of the Okataina Volcanic Centre; dominantly silicic volcanism associated with the Taupo Rift, New Zealand. Journal of Volcanology and Geothermal Research, 190(1-2), 123-135.

Crisp, J. A. (1984). Rates of magma emplacement and volcanic output. Journal of Volcanology and Geothermal Research, 20(3-4), 177-211.

Darby, D. J., Hodgkinson, K. M., Blick, G. H. (2000). Geodetic measurement of deformation in the Taupo Volcanic Zone, New Zealand: The north Taupo network revisited. New Zealand Journal of Geology and Geophysics, 43(2), 157-170.

Davies, H. (1995a). The 1994 Rabaul eruption. University of Papua New Guinea.

Davies, H. (1995b). The 1994 Eruption of Rabaul Volcano - A Case Study in Disaster Management. Port Moresby: University of Papua New Guinea.

De Natale, G., Troise, C., Pingue, F., Mastrolorenzo, G., Pappalardo, L., Battaglia, M., et al. (2006). The Campi Flegrei Caldera: unrest mechanisms and hazards. In G. De Natale, C. Troise and C. R. J. Kilburn (Eds.), *Mechanisms of activity and unrest at large calderas* (Vol. 269, pp. 25-45). London: The Geological Society of London.

De Vivo, B., Rolandi, G., Gans, P. B., Calvert, A., Bohrson, W. A., Spera, F. J., et al. (2001). New constraints on the pyroclastic eruptive history of the Campanian volcanic Plain (Italy). Mineralogy and Petrology, 73(1), 47-65.

Deino, A. L., Orsi, G., de Vita, S., Piochi, M. (2004). The age of the Neapolitan Yellow Tuff caldera-forming eruption (Campi Flegrei caldera-Italy) assessed by 40Ar/39Ar dating method. Journal of Volcanology and Geothermal Research, 133(1-4), 157-170.

Di Vito, M. A., Isaia, R., Orsi, G., Southon, J. R., D'antonio, M., De Vita, S., et al. (1999). Volcanic and deformation history of the Campi Flegrei caldera in the past 12 ka. Journal of Volcanology and Geothermal Research, 91(2-4), 221–246.

Di Vito, M. A., Lirer, L., Mastrolorenzo, G., Rolandi, G. (1987). The Monte Nuovo eruption (Campi Flegrei, Italy). Bulletin of Volcanology, 49(4), 608–615.

D'Oriano, C., Poggianti, E., Bertagnini, A., Cioni, R., Landi, P., Polacci, M., et al. (2005). Changes in eruptive style during the AD 1538 Monte Nuovo eruption (Phlegrean Fields, Italy): the role of syn-eruptive crystallization. Bulletin of Volcanology, 67(7), 601-621.

Durand, M. and Scott, B. J. (2005). Geothermal ground gas emissions and indoor air pollution in Rotorua, New Zealand. Science of The Total Environment, 345(1-3), 69-80.

Durand, M., Gordon, K., Johnston, D. M., Lorden, R., Poirot, T., Scott, J., Shephard, B. (2001). Impacts of, and responses to ashfall in Kagoshima from Sakurajima Volcano: lessons for New Zealand. Lower Hutt: Institute of Geological & Nuclear Sciences science report 2001/30. 53 p.

Dvorak, J. J. and Gasparini, P. (1991). History of earthquakes and vertical ground movement in Campi Flegrei caldera, Southern Italy: comparison of precursory events to the A.D. 1538 eruption of Monte Nuovo and of activity since 1968. Journal of Volcanology and Geothermal Research, 48(1-2), 77-92.

Dvorak, J. J. and Mastrolorenzo, G. (1991). The mechanisms of recent vertical crustal movements in Campi Flegrei Caldera, southern Italy. Geological Society of America - Special Paper 263, 47p.

Dzurisin, D., Christiansen, R. L., Pierce, K. L. (1995). Yellowstone: restless volcanic giant. Open file report 95-59, U.S. Geological Survey.

Earthquake Commission (EQC) website (http://canterbury.eqc.govt.nz/faq) – accessed on 21 February 2012.

Eiby, G. A. (1966). Earthquake swarms and volcanism in New Zealand. Bulletin of Volcanology, 29(1), 61-73.

Eiby, G. A. (1968). An annotated list of New Zealand earthquakes, 1460-1965. New Zealand Journal of Geology and Geophysics, 11(3), 630-647.

Finnimore, E. T., Low, B. S., Martin, R. J., Karam, P., Nairn, I. A., Scott, B. J. (1995). Contingency planning for and emergency management of the 1994 Rabaul volcanic eruption, Papua New Guinea: results of a fact-finding visit. Wellington: Ministry of Civil Defence.

Froggatt, P. C. (1981). Stratigraphy and nature of Taupo pumice formation. New Zealand Journal of Geology and Geophysics, 24, 231-248.

Froggatt, P. C. (1997). Volcanic hazards at Taupo Volcanic Centre (Vol. 7). Wellington: Volcanic Hazards Working Group of the Civil Defence Scientific Advisory Committee.

GeoNet website (http://www.geonet.org.nz) and (http://www.geonet.org.nz/volcano) (GNS Science) – accessed 21 February 2012.

Gibowicz, S. J. (1973). Variation of frequency-magnitude relationship during Taupo earthquake swarm of 1964-65. New Zealand Journal of Geology and Geophysics, 16(1), 18-51.

Global Volcanism Program (GVP) website (www.volcano.si.edu) – accessed 20 February 2012.

Gravley, D. M., Wilson, C. J. N., Leonard, G. S., Cole, J. W. (2007). Double trouble: Paired ignimbrite eruptions and collateral subsidence in the Taupo Volcanic Zone, New Zealand. Geological Society of America Bulletin, 119(1-2), 18-30.

Grindley, G. W. and Hull, A. G. (1986). Historical Taupo earthquakes and earth deformation. Bulletin of the Royal Society of New Zealand, 24, 173.

Guidoboni, E., and Ciuccarelli, C. (2011). The Campi Flegrei caldera: historical revision and new data on seismic crises, bradyseisms, the Monte Nuovo eruption and ensuing earthquakes (twelfth century 1582 AD). Bulletin of Volcanology, 73(6), 655-677.

Hansell, A. L. and Oppenheimer, C. (2004). Health hazards from volcanic gases: a systematic literature review. Archives of Environmental Health: An International Journal, 59(12): 628.

Hansell, A. L., Horwell, C. J., Oppenheimer, C. (2006). The health hazards of volcanoes and geothermal areas. Occupational and Environmental Medicine 63(2): 149-156.

Hildreth, W. (1983). The compositionally zoned eruption of 1912 in the Valley of Ten Thousand Smokes, Katmai National Park, Alaska. Journal of Volcanology and Geothermal Research, 18(1-4), 1-56.

Hildreth, W. (1991). The timing of caldera collapse at Mount Katmai in response to magma withdrawal toward Novarupta. Geophysical Research Letters, 18(8), 1541-1544.

Hildreth, W. (2004). Volcanological perspectives on Long Valley, Mammoth Mountain, and Mono Craters: several contiguous but discrete systems. Journal of Volcanology and Geothermal Research, 136(3-4), 169-198.

Hill, D. P. (2006). Unrest in Long Valley Caldera, California, 1978-2004. Geological Society London Special Publications, 269, 1-24.

Hodgson, K. A. and Nairn, I. A. (2005). The c. AD 1315 syn-eruption and AD 1904 post-eruption breakout floods from Lake Tarawera, Haroharo caldera, North Island, New Zealand. New Zealand Journal of Geology and Geophysics, 48, 491-506.

Houghton, B. F. and Wilson, C. J. N. (1986). A1: Explosive rhyolitic volcanism: the case studies of Mayor Island and Taupo volcanoes. In B. F. Houghton & S. D. Weaver (Eds.), *North Island volcanism: Tour guides A1, A4, and C3* (pp. 33-100): New Zealand Geological Survey.

Houghton, B. F., Wilson, C. J. N., McWilliams, M. O., Lanphere, M. A., Weaver, S. D., Briggs, R. M., et al. (1995a). Chronology and dynamics of a large silicic magmatic system - central Taupo Volcanic Zone, New Zealand. [Article]. Geology, 23(1), 13-16.

Houghton, B. F., Wilson, C. J. N., Weaver, S. D., Lanphere, M. A., Barclay, J. (1995b). Volcanic hazards at Mayor Island (Vol. 6). Wellington: Volcanic Hazards Working Group of the Civil Defence Scientific Advisory Committee.

Hughes, G. R. (2011). Reinvestigation of the 1989 Mammoth Mountain, California seismic swarm and dike intrusion. Journal of Volcanology and Geothermal Research, 207(3-4), 106-112.

Hurst, A. W., Bannister, S. C., Robinson, R., Scott, B. J. (2008). Characteristics of three recent earthquake sequences in the Taupo Volcanic Zone, New Zealand. Tectonophysics, 452(1-4): 17-28; doi:10.1016/j.tecto.2008.01.017

International Volcanic Health Hazard Network (IVHHN) website (www.ivhhn.org) – accessed 11 January 2012.

Isaia, R., Marianelli, P., Sbrana, A. (2009). Caldera unrest prior to intense volcanism in Campi Flegrei (Italy) at 4.0 ka BP: Implications for caldera dynamics and future eruptive scenarios. Geophysical Research Letters, 36, 6.

Johnson, R. W. and Threlfall, N. A. (1985). Volcano town: the 1937-43 eruptions at Rabaul. Robert Brown and Associates.

Johnston, D. M. (1997). Physical and social impacts of past and future volcanic eruptions in New Zealand. *Ph.D. thesis, Massey University, Palmerston North*, *New Zealand,* 288 p.

Johnston, D., Becker, J., Jolly, G., Potter, S., Wilson, T., Stewart, C., Cronin, S. (2011). Volcanic Hazards Management at Taranaki Volcano: Information Source Book, GNS Science Report 2011/37 96+iii p.

Johnston, D. M. and Nairn, I. A. (1993). Volcanic Impacts Report. The impact of two eruption scenarios from the Okataina Volcanic Centre, New Zealand, on the population and infrastructure of the Bay of Plenty Region. Bay of Plenty Regional Council Resource Planning Publication 93/6.

Johnston, D. M., Nairn, I. A., Leonard, G. S., Walton, M., Paton, D., Ronan, K. R. (2004). Recovery issues resulting from a long-duration, Kaharoa-type rhyolite eruption on present day New Zealand. Paper presented at the NZ Recovery Symposium 04.

Johnston, D. M., Scott, B. J., Houghton, B. F. (1996). Guidelines for developing a response to a volcanic crisis in the Bay of Plenty. Institute of Geological & Nuclear Sciences Science Report 96/27.

Johnston, D. M., Scott, B. J., Houghton, B., Paton, D., Dowrick, D. J., Villamor, P., et al. (2002). Social and economic consequences of historic caldera unrest at the Taupo volcano, New Zealand and the management of future episodes of unrest. Bulletin of the New Zealand Society for Earthquake Engineering, 35(4), 215-230.

Kaye, G., Cole, J., King, A., Johnston, D. (2009). Comparison of risk from pyroclastic density current hazards to critical infrastructure in Mammoth Lakes, California, USA, from a new Inyo craters rhyolite dike eruption versus a dacitic dome eruption on Mammoth Mountain. Natural Hazards 49: 541-563.

Keam, R. F. (1988). Tarawera: The volcanic eruption of 10 June 1886. Auckland: University of Auckland.

Kerr, J., Nathan, S., Van Dissen, R., Webb, P., Brunsdon, D., King, A. (2003). Planning for development of land on or close to active faults: a guideline to assist resource managment planners in New Zealand. (No. 2002/124): Ministry for the Environment, Institute of Geological & Nuclear Sciences Limited.

Latter, J. H., Lloyd, E. F., Smith, E. E. M., Nathan, S. (1992). Volcanic hazards in the Kermadec Islands, and at submarine volcanoes between southern Tonga and New Zealand (Vol. 4). Wellington: Volcanic Hazards Working Group of the Civil Defence Scientific Advisory Committee.

Lechner, P. (2009). Living with volcanic ash episodes in civil aviation: the International Airways Volcano Watch (IAVW) and the New Zealand Volcanic Ash Advisory System (VAAS). www.caa.govt.nz

Leonard, G. S. (2003). The evolution of Maroa Volcanic Centre. *Ph.D. thesis, University of Canterbury, Christchurch*, *New Zealand,* 322 p.

Leonard, G. S., Begg, J. G., Wilson, C. J. N. (2010). Geology of the Rotorua area: scale 1:250,000. Institute of Geological & Nuclear Sciences 1:250,000 geological map 5. 102 p. + 1 folded map. Lower Hutt: Institute of Geological & Nuclear Sciences Limited.

Leonard, G. S., Cole, J. W., Nairn, I. A., Self, S. (2002). Basalt triggering of the c. AD 1305 Kaharoa rhyolite eruption, Tarawera Volcanic Complex, New Zealand. Journal of Volcanology and Geothermal Research, 115(3-4), 461-486.

Lipman, P. W. (2000). Calderas. In H. Sigurdsson, B. F. Houghton, S. R. McNutt, H. Rymer & J. Stix (Eds.), *Encyclopedia of Volcanoes* (pp. 643–662): Academic Press.

Lloyd, E. F. (1972). Geology and hot springs of Orakeikorako. Wellington: Department of Scientific and Industrial Research. New Zealand Geological Survey Bulletin 85. 164 p.

Lloyd, E. F. and Nathan, S. (1981). Geology and tephrochronology of Raoul Island, Kermadec Group, New Zealand. New Zealand Geological Survey Bulletin 95.

Lloyd E.F., Nathan S., Smith I. E. M., Stewart R. B. (1996). Volcanic history of Macauley Island, Kermadec Ridge, New Zealand. New Zealand Journal of Geology and Geophysics 39: 295-308.

Lowenstein, P. L. (1988). The Rabaul seismo-deformational crisis of 1983-85; monitoring, emergency planning and interaction with the authorities, the media and the public. Paper presented at the Kagoshima international conference on Volcanoes 1988, Kagoshima, Japan.

Lowry, A. R., Hamburger, M. W., Meertens, C. M., Ramos, E. G. (2001). GPS monitoring of crustal deformation at Taal Volcano, Philippines. Journal of Volcanology and Geothermal Research, 105(1-2), 35-47.

Mader, G. G. and Blair, M. L. (1987). Living with a volcanic threat: Response to volcanic hazards, Long Valley, California. Portola Valley, California: William Spangle and Associates.

Manville, V., White, J. D. L., Houghton, B. F., Wilson, C. J. N. (1999). Paleohydrology and sedimentology of a post-1.8 ka breakout flood from intracaldera Lake Taupo, North Island, New Zealand. Geological Society of America Bulletin, 111(10), 1435-1447.

Martin, R. S., Watt, S. F. L., Pyle, D. M., Mather, T. A., Matthews, N. E., Georg, R. B., et al. (2009). Environmental effects of ashfall in Argentina from the 2008 Chaitén volcanic eruption. Journal of Volcanology and Geothermal Research, 184(3-4), 462-472.

McKee, C. O., Johnson, R. W., Lowenstein, P. L., Riley, S. J., Blong, R. J., De Saint Ours, P., et al. (1985). Rabaul Caldera, Papua New Guinea: Volcanic hazards, surveillance, and eruption contingency planning. Journal of Volcanology and Geothermal Research, 23(3-4), 195-237.

McNutt, S. R. (2000). Volcanic seismicity. *In H. Sigurdsson, B. F. Houghton, S. R. McNutt, H. Rymer, J. Stix (Eds.), Encyclopedia of Volcanoes* (pp. 1015-1033): Academic Press.

Miller, C. D. (1985). Holocene eruptions at the Inyo volcanic chain, California: Implications for possible eruptions in Long Valley caldera. Geology, 13(1), 14-17.

Ministry of Civil Defence and Emergency Management (2006). The guide to the National Civil Defence Emergency Management Plan 2006.

Nairn, I. A. (1979). Rotomahana-Waimangu eruption, 1886: base surge and basalt magma. New Zealand journal of geology and geophysics, 22(3): 363-378

Nairn, I. A. (1991). Volcanic hazards at Okataina Volcanic Centre. Volcanic Hazards Information Series Vol. 2 (3 ed.). Ministry of Civil Defence.

Nairn, I. A. (2002). Geology of the Okataina Volcanic Centre, scale 1:50 000 (Vol. 25). Lower Hutt: Institute of Geological & Nuclear Sciences Limited.

Nairn, I. A. and Scott, B. J. (1995). Scientific management of the 1994 Rabaul eruption: lessons for New Zealand. Lower Hutt: Institute of Geological & Nuclear Sciences.

Nairn, I. A., Hedenquist, J. W., Villamor, P., Berryman, K. R., Shane, P. A. (2005). The ~AD1315 Tarawera and Waiotapu eruptions, New Zealand: contemporaneous rhyolite and hydrothermal eruptions driven by an arrested basalt dike system? Bulletin of Volcanology, 67(2): 186-193.

Nairn, I. A., McKee, C. O., Talai, B., Wood, C. P. (1995). Geology and eruptive history of the Rabaul Caldera area, Papua New Guinea. Journal of Volcanology and Geothermal Research, 69(3-4), 255-284.

Nairn, I. A., Self, S., Cole, J. W., Leonard, G. S., Scutter, C. (2001). Distribution, stratigraphy, and history of proximal deposits from the c. AD 1305 Kaharoa eruptive episode at Tarawera Volcano, New Zealand. New Zealand Journal of Geology and Geophysics, 44(3), 467-484.

Nairn, I. A., Shane, P. R., Cole, J. W., Leonard, G. J., Self, S., Pearson, N. (2004). Rhyolite magma processes of the AD 1315 Kaharoa eruption episode, Tarawera volcano, New Zealand. Journal of Volcanology and Geothermal Research, 131(3-4), 265-294.

Nairn, I. A., Wood, C. P., Bailey, R. A. (1994). The Reporoa Caldera, Taupo Volcanic Zone: source of the Kaingaroa Ignimbrites. Bulletin of volcanology, 56: 529-537

Naranjo, J. A. and Stern, C. R. (2004). Holocene tephrochronology of the southernmost part (42^{o}30' - 45^{o} S) of the Andean Southern Volcanic Zone. Revista geológica de Chile, 31(2), 224-240.

Newhall, C. G. and Dzurisin, D. (1988). Historical unrest at large calderas of the World. Washington, D.C., USA: U.S. Geological Survey, Bulletin 1855.

Otway, P. M., Blick, G. H., Scott, B. J. (2002). Vertical deformation at Lake Taupo, New Zealand, from lake levelling surveys, 1979-99. New Zealand Journal of Geology and Geophysics, 45(1), 121-132.

Otway, P. M., Grindley, G. W., Hull, A. G. (1984). Earthquakes, active fault displacement and associated vertical deformation near Lake Taupo, Taupo Volcanic Zone. Report New Zealand Geological Survey, 110, 73.

Parfitt, E. A. and Wilson, L. (2008). Fundamentals of physical volcanology. Blackwell Publishing, Oxford, U.K.

Philippine Institute of Volcanology and Seismology (PHIVOLCS) website (http://www.phivolcs.dost.gov.ph) – accessed 21 February 2012.

Piochi, M., Mastrolorenzo, G., Pappalardo, L. (2005). Magma ascent and eruptive processes from textural and compositional features of Monte Nuovo pyroclastic products, Campi Flegrei, Italy. Bulletin of Volcanology, 67(7), 663-678.

Pringle, M. S., McWilliams, M. O., Houghton, B. F., Lanphere, M. A., Wilson, C. J. N. (1992). 40Ar/39Ar dating of Quaternary feldspar: examples from the Taupo Volcanic Zone, New Zealand. Geology, 20(6), 531.

Rosenberg, M. D., Wilson, C. J. N., Gravley, D. M., Rotella, M., Borella, M. W. (2007). Insights into the 2006 Raoul Island eruption from deposit characteristics and eruption effects. Geosciences New Zealand conference 2007 abstract.

Saccorotti, G., Petrosino, S., Bianco, F., Castellano, M., Galluzzo, D., La Rocca, M., et al. (2007). Seismicity associated with the 2004-2006 renewed ground uplift at Campi Flegrei Caldera, Italy. Physics of The Earth and Planetary Interiors, 165(1-2), 14-24.

Scott, B. J. (1989). Geodetic and geophysical monitoring of the 1886 Tarawera rift. Paper presented at the International Volcanological Congress, New Zealand.

Scott, B. J. (1994). Cyclic activity in the crater lakes of Waimangu hydrothermal system, New Zealand. Geothermics, 23(5/6), 555-572.

Scott, B.J. (1995). Raoul Island: crater lakes, temperatures, deformation and seismicity, 1993. p. 7-11 *In: Scott, B. J. and Sherburn, S. Volcano and geothermal observations 1993*. Lower Hutt: Institute of Geological & Nuclear Sciences. Institute of Geological & Nuclear Sciences science report 95/11; New Zealand Volcanological Record 22.

Scott, B. J. and Cody, A. D. (1982). The 20 June 1981 hydrothermal explosion at Tauhara Geothermal Field, Taupo. Rotorua: New Zealand Geological Survey, Department of Scientific and Industrial Research.

Scott, B. J., Gordon, D. A., Cody, A. D. (2005). Recovery of Rotorua geothermal field, New Zealand: progress, issues and consequences. Geothermics, 34(2): 161-185.

Scott, B. J. and Nairn, I. A. (1998). Volcanic hazards: Okataina Volcanic Centre. Scale 1:100,000. [Whakatane]: Bay of Plenty Regional Council. Resource planning publication / Bay of Plenty Regional Council 97/4. 1 map.

Scott, B. J. and Travers, J. (2009). Volcano monitoring in NZ and links to SW Pacific via the Wellington VAAC. Natural Hazards 51(2): 263-273. DOI 10.1007/s11069-009-9354-7

Shane, P., Froggatt, P. C., Smith, I. E. M., Gregory, M. (1998). Multiple sources for sea-rafted Loisels Pumice, New Zealand. Quaternary Research, 49(3), 271-279.

Shearer Consulting Ltd. and Market Economics Ltd. (2008). Exercise Ruaumoko - report of the economic workgroup: assessment of the impacts of a volcanic eruption on the Auckland economy.

Sherburn, S. and Nairn, I. (2004). Modelling geophysical precursors to the prehistoric c. AD1305 Kaharoa rhyolite eruption of Tarawera Volcano, New Zealand. Natural Hazards, 32(1), 37-58.

Simmons, S. F., Keywood, M., Scott, B. J., Keam, R. F. (1993). Irreversible change of the Rotomahana-Waimangu hydrothermal system (New Zealand) as a consequence of a volcanic eruption. Geology, 21(7), 643-646.

Sorey, M. L., Farrar, C. D., Gerlach, T. M., McGee, K. A., Evans, W. C., Colvard, E. M., et al. (2000). Invisible CO2 gas killing trees at Mammoth Mountain, California (No. 172-96 version 2.0). Menlo Park, CA: U.S. Geological Survey and U.S. Department of the Interior.

Taranaki Blowout exercise webpage (http://www.trc.govt.nz/taranaki-blowout-background-info) – accessed on 21 February 2012.

Vandemeulebrouck, J., Hurst, A. W., Scott, B. J. (2008). The effects of hydrothermal eruptions and a tectonic earthquake on a cycling crater lake (Inferno Crater Lake, Waimangu, New Zealand). Journal of Volcanology and Geothermal Research, 178(2): 271-275.

Vucetich, C. G. and Howorth, R. (1976). Late Pleistocene tephrostratigraphy in the Taupo district, New Zealand. New Zealand Journal of Geology and Geophysics, 19, 51-69.

Walker, G. P. L. (1984). Downsag calderas, ring faults, caldera sizes, and incremental caldera growth. Journal of Geophysical Research, 89(B10), 8407-8416.\

Walker, G. P. L., Heming, R. F., Sprod, T. J., Walker, H. R. (1981). Latest major eruptions of Rabaul volcano. *In Johnson, R. W. ed., Cooke-Ravian volume of volcanological papers.* Geol. Surv. Papua New Guinea Memoir 10: 181-193.

Wilson, C. J. N. (1993). Stratigraphy, chronology, styles and dynamics of late Quaternary eruptions from Taupo volcano, New Zealand. Philosophical Transactions: Physical Sciences and Engineering, 205-306.

Wilson, C. J. N. (2001). The 26.5 ka Oruanui eruption, New Zealand: an introduction and overview. Journal of Volcanology and Geothermal Research, 112(1-4), 133-174.

Wilson, C. J. N. and Walker, G. P. L. (1985). The Taupo eruption, New Zealand I. General aspects. Philosophical Transactions of the Royal Society of London. Series A, Mathematical and Physical Sciences, 314(1529), 199-228.

Wilson, C. J. N., Gravley, D. M., Leonard, G. S., Rowland, J. V. (2009). Volcanism in the central Taupo Volcanic Zone, New Zealand: tempo, styles and controls. Studies in Volcanology: The Legacy of George Walker. Special publication by IAVCEI., 2, 225-247.

Wilson, C. J. N., Houghton, B. F., Lloyd, E. F. (1986). Volcanic history and evolution of the Maroa-Taupo area, central North Island. Late Cenozoic Volcanism in New Zealand, 23, 194-223.

Wilson, C. J. N., Houghton, B. F., McWilliams, M. O., Lanphere, M. A., Weaver, S. D., Briggs, R. M. (1995). Volcanic and structural evolution of Taupo Volcanic Zone, New Zealand - a review. Journal of Volcanology and Geothermal Research, 68(1-3), 1-28.

Wilson, C. J. N., Rogan, A. M., Smith, I. E. M., Northey, D. J., Nairn, I. A., & Houghton, B. F. (1984). Caldera volcanoes of the Taupo Volcanic Zone. Journal of Geophysical Research, 89(B10), 8463-8484.

Wilson, C. J. N., Scott, B. J., Houghton, B. F. (2004). Volcanoes of New Zealand. Tephra, 21: 2-11.

Wilson, C. J. N., Switsur, V. R., Ward, A. P. (1988). A new 14C age for the Oruanui (Wairakei) eruption, New Zealand. Geological magazine, 125(3), 297-300.

www.gns.cri.nz

Principal Location

1 Fairway Drive
Avalon
PO Box 30368
Lower Hutt
New Zealand
T +64-4-570 1444
F +64-4-570 4600

Other Locations

Dunedin Research Centre
764 Cumberland Street
Private Bag 1930
Dunedin
New Zealand
T +64-3-477 4050
F +64-3-477 5232

Wairakei Research Centre
114 Karetoto Road
Wairakei
Private Bag 2000, Taupo
New Zealand
T +64-7-374 8211
F +64-7-374 8199

National Isotope Centre
30 Gracefield Road
PO Box 31312
Lower Hutt
New Zealand
T +64-4-570 1444
F +64-4-570 4657