Multidisciplinary Investigation of the Evolution of Persistently Active Basaltic Volcanoes

by

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Abstract

The size, shape, location and/or chemical evolution of basaltic magma plumbing systems at most volcanoes is not well constrained. Having this information beneath active systems allows scientists to target areas which will likely be the first to display volcanic unrest. With these constraints and datasets that cover long periods of time or include anomalous topographic features, we can start to investigate how a volcanic system has changed over time. To accomplish this, geochemical and geophysical studies at Masaya volcano (Nicaragua) and Mauna Loa volcano (Hawaii, USA) were conducted.

Melt Inclusions were collected from Masaya volcano to investigate the processes within the magma chamber. The almost unchanging chemistry of the whole rock, crystals and melt inclusions regardless of which eruptive cone sampled suggests that the system is buffered in both temperature and chemistry. A large deep reservoir with rapid transit times to the surface could explain the data.

Bouguer gravity mapping data at Masaya and Mauna Loa volcanoes were collected, processed and inverted to constrain the location and volumes of density anomalies at depth. Beneath Masaya volcano, the gravity data provides evidence of a very large intrusive complex (< 900 km³) at 4-9 km depth as well as several small shallow anomalies perhaps due to ring dykes around a buried caldera rim. This study strengthens arguments that Masaya does not have a large shallow magmatic system and that shallow endogenous growth is minimal. Gravity mapping and inversions from Mauna Loa provide evidence for relatively rapid rift zone migration most likely caused by a large edifice destabilizing event. The massive Ālika debris flows are contemporaneous with the age of rift zone migration suggesting that mass wasting is the cause.

Keywords: Volcanology; Geochemistry, Geophysics; Magnetics; Gravity; Melt inclusions
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Whenever I needed to bounce ideas around or needed some kind of code or software help, my lab mates and friends were there to save the day. Nathalie you were always available to answer my chemistry questions and point me towards possible solutions. Patricia, besides partaking in the banter, crafting of field songs and plain silliness, having a different perspective around helped my own research looking at survey design in a more quantitative way. Craig without you I swear it would have taken me an extra two months to finish this document, it was as if you knew what thing I was going to be doing next and forged ahead producing easy to use scripts. To my missing in action co-author, Nicaragua companion and previous OU PhD student, Guillermo, this document sums up a lot of the discussions and brainstorms we had while collecting this data. I hope it does it justice.

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Introductory Image

Pahoehoe flows Kilauea Hawaii
Chapter 1. Introduction

The pathways and reservoirs that magma resides in or travels through beneath a volcano constitute its magma plumbing system. Some volcanoes have vigorous and extensive magmatic plumbing systems such as Kilauea (Hawaii, USA), [e.g., Cervelli and Miklius, 2003] while others have unconnected magma bodies that occasionally rise to the Earth's surface along preferential pathways such as at Three Sisters (Oregon, USA) [Zurek et al., 2012]. There are many different properties, such as size, connectivity of the magma plumbing system, stress regime, structures, and magma chemistry that have a direct effect on the volcanic activity and the size of eruptions that can be generated. This thesis focuses on obtaining constraints on the size, shape, location and the evolution of basaltic magma plumbing systems via studies at two different volcanoes, Masaya (Nicaragua) and Mauna Loa (Hawaii, USA).

1.1. Magma plumbing systems

1.1.1. Basaltic volcanoes

The majority of all volcanic eruptions consist of basalt, and basaltic volcanoes can be found in all tectonic environments. There are different ways to categorize and group basaltic volcanoes but typically it is done based on morphology. Persistently active basaltic volcanoes are categorized as either a shield volcano or a stratovolcano based on the slope of the edifice.

There are also two main chemical subdivisions of basalt, tholeiitic and calc-alkaline, based on their evolution from a parent melt. Calc-alkaline basalts are higher in the alkali and alkaline earth metals, are usually found in arc settings, erupt to form stratovolcanoes and are typically more evolved than tholeiitic basalt. The degree of evolution is in part due to the effect of higher concentrations of water (greater than 2 %)
and its effect on fractional crystallization. Calc-alkaline fractionation, in the range of basalt to andesite, is characterized by decreasing iron and magnesium leading to the enrichment of alkaline and alkali metals [e.g., Irvine and Baragar, 1971; Sheth et al., 2002; Fig. 1-1]. Calc-alkaline basalt also tends to have higher viscosities favouring steep stratovolcanoes and Strombolian eruptions [e.g., Vergniolle and Mangan, 2000].

![Ternary diagram showing the general fractional crystallization trends in both tholeiitic and calc-alkaline melts.](image)

Figure 1-1 Ternary diagram showing the general fractional crystallization trends in both tholeiitic and calc-alkaline melts.

Tholeiitic basalt dominates in both hot spot and ocean rift environments and exists in arc settings usually in immature oceanic volcanic fronts, near fracture zones in the subducting plate and wherever there is rifting associated with arc volcanism. Due to the anhydrous nature of these melts (less than 2 % H\textsubscript{2}O), magnesium-rich phases crystalize first causing an enrichment of iron in the melt [e.g., Zimmer et al., 2010] (Fig. 1-1). Tholeiitic lava flows also typically have low viscosity allowing the flows to spread out over a wide area producing a low sloped edifice or shield volcano [e.g., Vergniolle and Mangan, 2000].

The classic examples of shield volcanoes come from the Hawaii Islands (i.e., Mauna Loa and Kilauea volcanoes). Each Hawaiian volcano has two well defined rift zones that extend up to 250 km away from the summit caldera [Clague and Sherrod, 2014]. Eruptions are concentrated at the rift zones and the summit of each volcano, and are termed “Hawaiian eruptions”, meaning they are continuous in character and generate lava fountains [Vergniolle and Mangan, 2000]. The current understanding is that all
magma flux into the magmatic plumbing systems of Hawaiian volcanoes first enters a reservoir beneath the summit before traveling to an eruption site [Cervelli and Miklius, 2003]. This is based on observations of aseismic deformation events at Kilauea volcano, where the shallow magma reservoir deflates and then inflates over 24 to 72 hours [Cervelli and Miklus, 2003]. These events are first detected at the summit within the shallow magmatic system and then seen to propagate down Kilauea’s East Rift Zone to the eruption site over several hours. Most shield volcanoes however, do not fit the Hawaiian model with rift zones extending from a centralized caldera. Another ocean island chain of volcanoes, the Galapagos volcanic chain, are instead built via radial dike swarms extending from a central caldera [Chadwick and Howard, 1991].

Subduction related shield volcanoes are variable as they erupt a wider variety of chemistries, including small quantities of rhyolite. They are also structurally variable as many do not have the classical central caldera [e.g., the Tolbachik volcanic complex, Russia; Fedotov et al., 2011] and none have well defined rift zones. The increase in diversity is likely due to the increased likelihood of crustal contamination causing a more evolved melt. Examples include Medicine Lake and Newberry volcanoes [e.g., Grove et al., 1982; Higgins, 1973] in the Cascade Volcanic Arc of the USA and western Canada.

1.1.2. Composition

Under normal circumstances magma generated from partial melting of the mantle undergoes significant chemical changes before it erupts. Through fractional crystallization, crystals precipitate from the magma removing elements from the melt. If this is the only process occurring, the chemistry of the melt will gradually evolve along either the tholeiitic or calc-alkaline fractional crystallization trend [e.g., Irvine and Baragar, 1971; Sheth et al., 2002; Fig. 1-1]. In both trends, the most primitive and least modified magma will crystalize magnesium-rich olivine (forsterite) [e.g., Basaltic Volcanism Study Project, 1981]. As basaltic melt evolves, the chemistry of the crystalizing olivine changes along its solid solution series, from magnesium-rich forsterite (Mg$_2$SiO$_4$) to iron-rich fayalite (Fe$_2$SiO$_4$). To quantify this evolution, the ratio of magnesium to iron plus magnesium is used. When values are close to one (greater than 0.9) they are considered primitive meaning the magma is representative of mantle-derived melt [e.g., Basaltic Volcanism Study Project,
To erupt a primitive melt requires short residence times, no contamination, and no significant mixing with other bodies of magma before eruption.

Contamination involves the incorporation of material external to where the melt was generated. An obvious source of potential contamination is the host rock. This is particularly prevalent in continental arc settings as granitic material has a lower melting point [~700 °C; e.g., Sawyer et al., 2011] than hot primitive basaltic melt [> 900 °C; e.g., Herzberg, 2011]. Volcanic plumbing systems can have different bodies of magma evolving along different paths due to differences in storage time, parent melt differences or amount of contamination. These different melts can intersect within the crust and mix, creating a melt with an intermediate composition between the two.

The details of melt evolution are further complicated by physiochemical properties which may control which phase crystalizes such as, but not limited to, oxygen fugacity, temperature, water content and pressure. Some physiochemical properties such as oxidation and water concentration are quite dissimilar in different tectonic environments in part due to fluids from the subducting slab [e.g., Sato, 1978; Mysen, 1991; Holtz and Johannes, 1994].

1.1.3. Current understanding

Little is known about the size, shape and location of most active magmatic plumbing systems. This is due to the fact that researchers have no direct access and therefore must use geophysical and geochemical techniques to infer all physical aspects. To interpret the data these techniques produce, simplifying assumptions such as spherical sources are often used. The location of active magma chambers are primarily constrained using seismic [e.g., Detrick et al., 1987; Gudmundsson et al., 1994] and deformation data [e.g., Sigmundsson et al; 1992; Yun et al., 2006]. Magma dynamics such as storage depth [e.g., Spilliaert et al., 2006], magma mixing [e.g., Moune et al., 2012; Venugopal et al., 2016], and degassing [e.g., Métrich and Wallace, 2006; Martin et al., 2010], can be constrained using a variety of geochemical data.

At Kilauea volcano, the analysis of seismic data has identified regions where long period seismic events originate [e.g., Battaglia et al., 2003], as well as areas with
anomalously low seismic wave velocities [e.g., Okubo et al., 1997] both of which are thought to be caused by fluid movement. Long period events have been identified at other basaltic volcanoes including Masaya [Métaxian et al., 1997], however, at most volcanoes the seismic coverage required to locate long period events or conduct seismic tomography is lacking. At best, long period events suggest a point where magmatic fluids exist, and seismic tomography provides the maximum areal constraint of magma reservoirs. The resolution of seismic data is typically too coarse to provide more than a general area of where there is a significant concentration of magma.

Measured surface deformation related to active magma plumbing systems can be modeled to constrain the shape, volume and depth of the source [e.g., Yun et al., 2006]. This has been performed at a large number of basaltic volcanoes [e.g., Hekla volcano: Sigmundsson et al., 1992; Sierra Negra volcano: Yun et al., 2006] using GPS and InSAR techniques. However, interpretation of the data can be difficult as the results from modeling and inversions are non-unique.

Other geophysical surveys such as magnetotellurics [e.g., Newman et al., 1985] and gravity [e.g., Johnson et al., 2010] can also infer the locations of magma chambers, however; they are used less often and have limitations. Gravity data are used in this thesis (Chapters 5 and 6) to map out density contrasts and structures pertaining to intrusive material at both Mauna Loa and Masaya volcanoes. This technique does not directly detect the presence of melt; however, magma reservoirs can be denser than the host rock and form part of a dense intrusive complex [e.g., Rousset et al., 1989]. Dense intrusive complexes, if cool, also provide a glimpse of the past magmatic history [e.g., Rousset et al., 1989; Zurek and Williams-Jones, 2013]. Another geophysical survey used in conjunction with gravity in this thesis is total field magnetics. While total field magnetic data cannot detect melt, they can image structures below their Currie temperature (the temperature at which material loses its permanent magnetism) which can aid in identifying whether a density anomaly could be a part of an active magmatic system [e.g., Vine and Matthews, 1963].

Magma storage depth is primarily inferred through melt inclusion geochemistry studies [e.g., Spilliaert et al., 2006]. Melt inclusions (MIs) are created when a crystal traps
a small amount of melt as it grows and crystalizes. That entrapped melt is thus a snap shot of the magma chemistry at that time (see Chapter 2 and Chapter 3). Volatile data (H₂O, SO₂ and CO₂) and major element data from MIs can be entered into solubility models to determine the approximate pressure or depth of melt entrapment. A cluster of similar pressures would suggest that there is a storage reservoir at that pressure.

It is also possible to infer whether magma mixing has occurred through fractional crystallization models and trace element data. Using the most primitive melt composition obtained through MI geochemistry, phenocryst chemistry, and estimates of temperature, oxygen fugacity and pressure, one can model how a melt is expected to evolved via fractional crystallization [e.g., Danyushevsky and Plechov, 2011]. If the chemical evolution trends recorded in a sample’s MIs cannot be recreated using fractional crystallization models, then magma mixing likely occurred [e.g., Venugopal et al., 2016]. However, fractional crystallization trends of the major elements are insensitive to mixing between two magmas of similar composition. To rule out magma mixing as a major component of the evolution requires trace element data, which can vary significantly even though the major element concentrations are similar [e.g., Moune et al., 2012; Venugopal et al., 2016]. A MI study at Cerro Negro and El Hoyo volcanoes in Nicaragua was able to show that Cerro Negro lavas generally have a simple evolution described by fractional crystallization [Venugopal et al., 2016]. This is in contrast to the eruptive products of its neighbour El Hoyo, which are due to mixing between Cerro Negro magmas and a more evolved melt.

With measurements of both erupted volatile flux and melt inclusion volatile concentrations, a degassing budget can be inferred. In the case of persistent activity and long term measurements of eruptive volatile flux, it is possible to use the undegassed concentration, obtained from MIs, to calculate a magmatic flux to the system. This has been done at a number of volcanoes such as Masaya [0.05 to 0.2 km³ yr⁻¹; Stoiber et al., 1986; Martin et al., 2010] and Etna [0.28 km³ yr⁻¹ between 1975 and 1995; Allard, 1997].
1.1.4. Persistent activity

Persistent volcanic activity can refer to fumeroles, near continuous eruptions, acid crater lakes, and lava lakes with prolonged thermal and gas flux [e.g., Stevenson and Blake, 1998]. Here, however, I focus on basaltic volcanoes which continuously degas with the presence of surface to near-surface lava in the form of lava lakes, glowing vents and eruptions. Persistently active volcanoes have robust magmatic systems with enough flux and vigorous convection to keep the system hot and prevent solidification. With a high magmatic flux, there is a potential to create a complex system of pathways and chambers. Kilauea is perhaps the best example as the shape and nature of its plumbing system has been the subject of numerous studies; however, we still do not have robust constraints on its shape and size. It has been suggested that Kilauea has at least 3 separate areas of melt storage beneath the summit [located at ~20 km depth, 2 - 3 km depth and 400 m - 1 km depth; Clague and Sherrod, 2014], as well as regions of storage in the rift zones [e.g., Cervelli et al., 2002].

In addition to high magmatic flux, persistently active basaltic volcanoes also degas more magma than they erupt. Based on gas flux measurements and volatile MI analyses, the difference in erupted versus degassed magma can be orders of magnitude different [Etna; e.g., Allard, 1997; Harris et al., 2000; Kilauea; e.g., Elías et al., 1998]. To explain this discrepancy, endogenous growth has been proposed. At Kilauea, dyke intrusions and south flank movement can account for most of the non-erupted material; however, at other volcanoes such as Stromboli or Masaya, there are no deformation indicators to suggest where the bulk of the non-erupted material is stored. In these cases, it is suggested that degassed material is emplaced back at a depth, below detection limits.

1.1.5. The importance of constraining magmatic plumbing systems

This thesis primarily focuses on expanding our understanding of the structures, processes and evolution of basaltic volcanic systems through case studies at Masaya and Mauna Loa volcanoes. At each volcano geophysics is used to infer the location of intrusive complexes to constrain the locations of present and past magma plumbing structures. Melt inclusion geochemical data from Masaya provides further constraints on the depth of possible magmatic storage reservoirs, magmatic flux, primary volatile concentrations and
magmatic transport. Constraining the location(s), size and shape of magmatic storage systems beneath persistently active volcanoes provides the information required to implement a comprehensive monitoring network. It also provides the basic information to investigate many different volcanic processes such as cooling, recharge, convection, endogenous growth and equilibrium volatile concentrations [e.g., Giberti et al., 1992]. This information may provide the critical insight required to forecast future volcanic behaviour.

1.2. Masaya volcano, Nicaragua

1.2.1. Location and tectonic setting

Masaya volcano is a basaltic shield within a large NE-SW oblong (~ 6 km by 10 km) caldera which forms part of a larger complex of nested calderas called the Las Sierras-Masaya complex; although the entire complex is often referred to as Masaya volcano. Masaya is located in Nicaragua and part of the Central American Volcanic Arc created by the subduction of the Cocos Plate beneath the Caribbean Plate (Fig. 1-2 A). The volcano is approximately 20 km south east of the capital, Managua, of just over 2 million people and the city that shares its name, Masaya, is located on the edge of the caldera with a population of over 100,000 (Fig. 1-2). The possibility of loss of property or life due to the proximity of urban areas, places Masaya as a top priority for eruption monitoring and study.

There are 18 potentially active volcanoes in Nicaragua and of these, Masaya volcano stands out for several reasons. The most striking is Masaya’s persistent activity, as it has been in a nearly constant state of unrest since being first described by Spanish Conquistadors in 1524 [Maciejewski, 1998]. It is also a shield volcano unlike the majority of the volcanoes in Nicaragua and for over 30,000 years, the chemistry of the eruptive products has remained essentially the same [Walker et al., 1993]. The most surprising thing that sets Masaya apart is that it has produced 3 large basaltic Plinian eruptions in the last 7000 years [e.g., Williams, 1983]. By themselves, the processes that can cause
the different possible eruptive activities are difficult to conceptualize and it is further compounded by a complex local tectonic setting (Fig. 1-2 B). Masaya is located within the Nicaraguan depression, defined as a half graben [McBirney and Williams, 1965], and on
the southern end of the Cofradia fault which bounds the south-eastern edge of the Managua graben. The interplay between tectonic stresses, the Nicaraguan depression, the Managua graben and volcanic activity is unclear [Burkart and Self, 1985; van Wyk de Vries, 1993; van Wyk de Vries and Merle, 1998]. The most promising avenue for linking all tectonic features together is suggested by Girard and van Wyk de Vries [2005] whereby a large intrusion beneath the Las Sierras-Masaya complex creates a ductile region creating pull apart features in a transtensive regional stress regime.

1.2.2. Historical activity

Only two relatively small effusive eruptions (in 1670 and 1772) have occurred since Masaya was first discovered by Spanish Conquistadors (Fig. 1-2 C). The first, in 1670, was due to overflows from an active lava lake in Nindiri crater which flowed down the northern slope of Nindiri cone (Figs. 1-2 C, 1-3). The 1772 eruption took place from a fissure on the north side of Masaya cone. While there has not been a significant eruption involving juvenile material in over 300 years, Masaya volcano has been persistently degassing during this time. In addition to persistent degassing, small explosive events occasionally occur at Masaya, the largest of which occurred in 2001 [Martin et al., 2010].

Through qualitative observations, 4 different periods (1852-1859, 1902-1906, 1919-1927 and 1947-1959) of increased degassing rates or degassing crises have been identified before modern gas measurements began in 1972 [Stoiber et al., 1986]. Since 1972, degassing levels have not been constant with two more degassing crises occurring in 1979-1985 and 1993-2006; part of the last event has been documented using COSPEC, miniDOAS, and FLYSPEC ultraviolet spectrometers [e.g., Horrocks et. al., 1999; Burton et al., 2000; Duffell, 2003; Williams-Jones et al., 2003; Nadeau and Williams-Jones, 2006]. Unlike the total SO2 flux, gas ratios have not varied significantly over time [e.g., Martin et al., 2010] suggesting that changes in magmatic flux are the sole cause of the variable SO2 flux. Estimates of the magmatic flux based on melt inclusions and SO2 flux suggest that ~ 10 km³ of magma has degassed between 1875 and 1985 [Stoiber et al., 1986]. This is unusual as there has been no significant deformation recorded and no effusive eruptions over this time frame.
Figure 1-3 Top: Aerial photo of Masaya caldera with each major cinder cone labeled except Arenoso located just off the image north of Cerro Montoso along the dashed circle which represents an annular structure [Mauri et al., 2012; Caravantes 2013]. Bottom: Aerial photo of Masaya and Nindiri cones with each pit crater labeled.
1.2.3. Plinian eruptions and Masaya’s caldera

Originally, Masaya’s caldera was thought to be formed solely through collapse similar to that of Hawaii or Piton de la Fournaise, Réunion island [McBirney and Williams, 1965]; however, the discovery of basaltic Plinian deposits sourced from Masaya volcano changed our understanding of Masaya’s formation [Williams, 1983]. Three different large basaltic Plinian eruptions have been identified to have occurred within the last 7 ka years, with the most recent at ~1.8 ka [e.g., Williams, 1983; Pérez and Freundt, 2006; Pérez et al., 2009; Kutterolf et al., 2008]. The largest of the three was 6 ka and produced the San Antonio Tephra which accounts for more than half (~ 14 km$^3$) of the total volume of all three basaltic Plinian eruptions (~ 24 km$^3$) [Kutterolf et al., 2008].

While the caldera is the product of three different Plinian eruptions, there are no obvious discernible surface structures to differentiate between each Plinian event. Any direct evidence is likely buried under younger lava flows. A geophysical self-potential study inferred the presence of hydrothermal upwelling and diffuse degassing structures under Comalito, Masaya, Nindiri, Cerro Montoso and Arenoso, suggesting the semicircular set of cones have either individual shallow intrusions or are connected via faults and dykes to a deeper source [Fig. 1-3; Mauri et al., 2012]. Total magnetic and VLF (Very Low Frequency) profiles also show anomalies crossing the circular structure [Caravantes, 2013]. With the addition of InSAR data, implying a slow but persistent deflation within the centre of the circular structure, Caravantes [2013] suggested that this feature is likely a small caldera associated with the most recent Plinian eruption ~1.8 ka that has been subsequently filled.

1.2.4. Chemistry

Previous analysis of whole rock data shows that Masaya’s eruptive chemistry has been essentially the same for 30,000 years [Walker et al., 1993]. This includes the last three Plinian eruptions. Stable eruptive chemistry is an important result; however, it raises many questions about the processes occurring within Masaya’s magmatic plumbing system. What magmatic processes are present to maintain a stable evolution? Are there changes occurring at depth, such as mixing, that are not seen in whole rock chemistry? Can magmatic processes explain the discrepancy between the historical volume of
magma degassing versus the amount erupted? Are there large fluctuations in volatile levels over time that can at least partially explain Masaya's Plinian eruptions?

Chapter 3 addresses these questions through a melt inclusion geochemistry study. Melt inclusions are inclusions of glass within a host crystal. During crystallization, crystal growth can be irregular and cause the entrapment of some of the surrounding melt. If the crystal is then erupted before the melt inclusion has time to crystalize, the chemistry of the magma where the crystal formed can be preserved. By analyzing many different melt inclusions, it is possible to infer processes that occurred at depth.

1.2.5. Density structure

Knowledge of the locations of structures and magma reservoirs allow volcanologists to focus monitoring equipment and resources to improve eruption forecasting. However, there is little known about the internal structure of Masaya volcano other than what can be inferred from surface geology, and that it has a small and vigorous shallow magma reservoir system beneath Nindiri cone [e.g., Rymer et al., 1998; Williams-Jones et al., 2003]. Previous Bouguer gravity studies, conducted before GPS technology was available, inferred the presence of a large intrusive body overlapping the northern caldera edge [Connor et al., 1989; Metaxian 1994]. These studies, however, have large vertical errors and low measurement density which reduces their resolution and accuracy.

Chapter 4 uses both total field magnetics and Bouguer gravity to infer different structural features beneath Masaya volcano. Total field magnetics infer differences in the magnetic properties of the subsurface to a maximum depth coinciding with the Currie point isotherm. Bouguer gravity is able to identify areas of anomalous density which may pertain to hydrothermal alteration, faults or intrusive material.
1.3. Mauna Loa volcano, Hawaii, USA

1.3.1. Location and Hawaiian Volcanism

Mauna Loa is one of five volcanoes that make up the Big Island of Hawaii in the Pacific Ocean along the Emperor–Hawaiian hot spot trend. It is also the largest volcano on Earth in both mass and volume, at approximately 75,000 km$^3$ (Fig. 1-4, 1-5) [Lockwood and Lipman, 1987]. Like all Hawaiian volcanoes that make up the main islands, Mauna Loa is a shield volcano generally erupting low viscosity basalt with two rift zones, the Southwest Rift Zone (SWRZ) and the Northeast Rift Zone (NERZ).

Hawaiian volcanoes are described to have life stages to their eruptive histories [e.g., Chen and Frey, 1985; Peterson and Moore, 1987]. Pre-shield stage is the first, representing the birth of the volcano and is characterized by infrequent eruptions of alkali basalts. Loihi is the youngest of the Hawaiian volcanoes, under water off the east coast of Hawaii, and is in its pre-shield stage or transitioning into its shield building phase [e.g., Moore et al., 1982; Garcia et al., 1993]. Next, the shield building stage occurs when the volcano is receiving the most heat/material from the mantle plume, accounting for 80 – 95% of the total volume of the volcano [e.g., Clague and Sherrod, 2014]. Mauna Loa has been erupting for over 700,000 years and is likely nearing the end of it shield building phase [e.g., Kurz and Kammer, 1991; Lipman, 1995]. The waning two stages are post-shield [e.g., Mauna Kea and Hualalai volcanoes on the Big Island; Peterson and Moore, 1987] and rejuvenation [e.g., Haleakala on Maui; Clague et al., 1987] stages characterized by alkali eruptive products and lengthy repose times.
Two subparallel trends of Hawaiian volcanoes have been identified stretching from the Big Island back at least 2 million years to Moloka‘i [Fig. 1-4; e.g., Dana, 1849; Jackson et al., 1972; Clague and Dalrymple, 1987]. Named after the largest volcanoes in each trend, they are the Loa and Kea trends. The Kea trend volcanoes on the Big Island of Hawaii include Kilauea, Mauna Kea, and Kohala whereas Loa trend volcanoes include Hualalai, Mauna Loa and Loihi. Trend-specific eruptive products can be typically identified by lead isotope \((\text{Pb}^{208/204} \text{ and } \text{Pb}^{206/204})\) ratios and suggest that the mantle plume beneath Hawaii is chemically zoned [e.g., Abouchami et al., 2005]. Furthermore, the isotopic trends can be used to identify which volcano an eruptive unit came from which could be important in areas where edifices overlap.
1.3.2. Recent activity and topography

Mauna Loa, like its neighbour Kilauea, has two rift zones which have created a topographic ridge that connects to its central summit caldera (Fig. 1-5). In general, rift zones at Hawaiian volcanoes are linked to gravitational spreading and rift orientations are related to both gravitational instability of the edifice and the effects of overlapping volcanoes [e.g., Delaney et al., 1998]. All historical eruptions, 33 eruptions since 1843 [Lockwood and Lipman, 1987], have taken place from either rift zone and/or the summit area. The volcano’s voluminous lava flows have extensively repaved its surface, with 90% of the surface younger than 4000 years [Lockwood and Lipman, 1987]. The most recent eruption in 1984, occurred from a series of fissures on the Northeast Rift Zone and sent lava flows 29 km downhill. One flow stopped 6.5 km from the town of Hilo, the island’s largest city. The relatively slow moving basaltic lava flows and the low explosivity of eruptions minimize direct risk to humans; however, recent residential developments have been built directly on the Southwest Rift Zone. It is conceivable that the proximity of the new subdivisions to potential eruption sites may cause the next eruption to have Mauna Loa’s first historical human casualty.

Following the 1984 eruption, Mauna Loa was quiet with few deformation or seismic events until 2002. In May 2002, monitoring networks began to record a radial deformation pattern indicating inflation as the magma reservoir beneath the volcano began to refill [e.g., La Marra et al., 2015]. A seismic swarm occurred in 2004 with a peak of over 100 events per day. Inflation continued but significantly slowed in 2009 and had stopped by mid-2013. Mauna Loa’s activity again started to change in 2015 with seismic levels above background and in late summer, the deformation network began to record inflation. This prompted an increase of the aviation warning level from green to yellow in mid-September 2015. The increase in activity, at the time of this writing, does not indicate that an eruption is imminent [http://hvo.wr.usgs.gov/].

Mauna Loa’s high eruption rate leads to difficulties when attempting to understand its evolution, as features that once existed at the Earth’s surface have been buried by younger lavas. To explore Mauna Loa’s past, it is necessary to locate anomalous structural features, or use geophysical techniques to gain insight into what exists at depth. The summit caldera and rift zones dominate the topography; however, there are several
anomalous topographic features. The smallest of these are radial vents on the southwest and northern flanks, likely due to a tensional stress regime causing volcanic spreading [Clague and Sherrod, 2014]. An impressive fault scarp, also associated with gravitational slumping [e.g., Lipman, 1980], the Kahuku pali, is situated just east of Mauna Loa’s SW rift zone beginning at a right lateral offset in the SWR and extending along the submarine portion of the rift. Subaerially, the Kahuku pali has an average height of 120 m and a maximum offset of 170 m; however, the submarine portion is more than 1600 m high. The underwater portion of the scarp has also been the focus of a recent study that showed
Figure 1-6 Map showing where the Ninole hills outcrop (in green) on the southwestern flank of Mauna Loa and local water drainage in blue. Both roads used for Bouguer gravity surveying are also shown, which include State Highway 11 and Cane Haul Road [Modified from Lipman et al., 1990].
that the Southwest Rift Zone has been active for over 400 ka [Jicha et al., 2012]. The area known as the Nīnole Hills (Fig. 1-5, 1-6) rounds out the identifiable anomalous topographic regions and is the focus of Chapter 5.

1.3.3. Previous studies - The Nīnole Hills

The Nīnole Hills rise out of the southwest flank of Mauna Loa and dominate the local topography (Figs. 1-5, 1-6). The hills also represent the oldest exposed subaerial rocks on the volcano at 100 - 200 ka [Lipman at al., 1990]. Due to the age of the rocks, the Nīnole Hills have attracted interest and generated numerous hypotheses on their formation as they are a potential window into Mauna Loa’s past [Hitchcock, 1906; Stearns and Clark, 1930; Lipman et al., 1990; Morgan et al., 2010]. The earliest hypothesis to explain the Nīnole Hills suggested that they were the remnants of a different volcano or an older summit of Mauna Loa [Hitchcock, 1906]. Geochemical analysis meant to test this showed no appreciable difference beyond alteration products between the Nīnole basalts and that of known Mauna Loa samples [Lipman et al., 1990]. In the absence of a magmatic chemical difference, Lipman et al. [1990] proposed that faulting and landslides may have created the Nīnole Hills; however, no off shore slides have been identified to be sourced from the Nīnole Hills. Other studies have suggested that the SWRZ has migrated westward due to large mass wasting events from the western flank and faulting towards the west-southwest [Fig. 1-5; Fisk and Jackson, 1972; Lipman, 1980; Lipman et al., 1990]. An island wide Bouguer gravity study proposed a previously unrecognized orientation for the SWRZ or a proto-rift zone south of Mauna Loa’s summit to explain an elongated gravitational high [Kauahikaua et al., 2000]. The most recent suggestion for the formation of the Nīnole Hills is that they are the result of a failed rift zone that began at Mauna Loa’s summit through the Nīnole Hills and continuing offshore [Morgan et al., 2010]. Chapter 5 sets out to determine how Nīnole Hills fits within Mauna Loa’s structural history and evolution by providing concrete evidence as to the formational mechanism of the Nīnole Hills.
1.4. Thesis Structure

Geophysical and geochemical approaches were undertaken at two different volcanoes to look at the evolution of basaltic volcanism over time scales that span 1000s of years. In Chapter 3, melt inclusions from 8 different eruptive deposits from Masaya volcano are geochemically probed to investigate magmatic processes through time. Chapter 4 utilizes both total magnetics and Bouguer gravity to identify subsurface structures which may be controlling the volcanic activity of Masaya volcano. Chapter 5 presents a study on the south flank of Mauna Loa volcano, where Bouguer gravity was used to image density structures beneath a topographic anomaly called the Ni'hole Hills.

The main chapters of this thesis (Chapters 3 – 5) are written in manuscript format as they have either been published or will be submitted for publication and therefore, can be read as standalone papers. This approach leads to some repetition and overlap between chapters. As with most publications, each chapter has a number of co-authors who have significantly contributed in some way, however, the majority of the work presented here is my own. Breaking this down by chapter:

- **Chapter 3**
  - 9 samples were collected by Moune [2007]. All other samples were collected and processed by me.
  - Interpretation, modeling and analysis are my own.

- **Chapter 4**
  - Approximately 30 % of total magnetics was collected by Caravantes [2013] without input from myself. I collected or aided in the collection of the remaining 70 %.
  - I collected or aided in the collection of ~30% of the Bouguer gravity data, the rest was collected by Caravantes [2013].
  - All processing, modeling, and interpretation are my own.
• Chapter 5
  
  o Geological structural information was collected by USGS scientists and supplied by Frank Trusdell.
  
  o All other data collection, processing and interpretation are my own.

  In addition to the main article-style chapters, there is one methodology chapter (Chapter 2) and Appendix A. Chapter 2 provides background and a more detailed description of survey procedure and methods used in the core chapters. Appendix A discusses a solution to accomplish geophysics across harsh and hazardous terrain from the air using unmanned aerial vehicles (UAV) or drones.

  Regardless of the field location or the technique used, the aim of this thesis is to obtain physical constraints on the magma plumbing systems at depth and investigate volcanic evolution. Hopefully this information can one day help lead to better forecasting of eruptive behaviour.

1.5. References


Dana, J. D. (1849), Geology, United States Exploring Expedition: During the years 1838, 1839, 1840, 1841, 1842, Under the command of Charles Wilkes, U.S.N: Hauptbd, 10, 756.


La Marra, D., M. Poland, A. Miklius, and V. Acocella (2015), SBAS-InSAR analysis of a decade of surface deformation at Mauna Loa (Hawai‘i): Preliminary results, *EGU General Assembly Conference Abstracts*.


Zurek, J. M. (2010), Characterizing the plumbing systems of active volcanoes through potential field studies, Masters dissertation, Simon Fraser University
Chapter 2.  General Methodology

This chapter discusses each technique used in this thesis by first providing some basic background information and then details the survey methods used. Any deviations from best practices (e.g., due to terrain, logistics, etc.) or assumptions are discussed within the second section for each technique. Furthermore, all location data used, presented or collected in this thesis is in the datum WSG 84 for consistency. All figures with gridded data used the minimum curvature gridding algorithm.

2.1. Magnetics

2.1.1. Background

A magnetic field measurement at the Earth’s surface is made up of three different components: 1) The core magnetic field, a large and slowly varying magnetic field generated by the Earth; 2) Diurnal variations, small and rapidly varying external components; 3) The magnetic response of the ground and surrounding material. The large and slowly varying magnetic field created within the Earth is described by a magnetic dynamo in the outer core (e.g., Alldredge and Hurwitz, 1964). While the details and perturbations to this field are complex, it can be well approximated by a simple dipole. The magnetic field lines are usually described in spherical coordinates; inclination, declination and radius (radius is reported as elevation). Near the equator the field lines are approximately horizontal (inclination of 0) and at the poles they are vertical (inclination of -90° at the North Pole and +90° at the South Pole). Field lines are most concentrated at the poles and therefore the poles have a higher background magnetic field (~65,000 nT), while at the equator the magnetic field lines are sparse and therefore a smaller background field (~25,000 nT) [Finlay et al., 2010]. In addition to positional variation in the Earth’s magnetic field, it is also variable over time. While negligible over time frames of less than a year, it should be removed if merging data spanning many years.

The second component, diurnal variations, is due to external sources most of which are caused by the Sun. The intensity of electromagnetic energy from the Sun that
affects the survey area will change due to solar processes (solar storms, sunspot activity), and position of the Earth (time of the day, or tilt/seasons). The shortest time scale and largest amplitude changes that can occur due to the Sun is solar storm activity. Solar flares coupled with a coronal mass ejection can cause variations of over 700 nT over seconds [Gonzalez et al., 1994]. Surveying to determine the third component, ground response, of Earth’s field is impossible during these events as the target anomalies are usually smaller than the storms variability.

The third and final component is the magnetic response of the ground and the purpose of the total magnetic survey presented in Chapter 4. The response of the rock beneath a survey is made up of two different parts, remanent and induced magnetism. A material with remanent magnetism is able to retain a magnetic moment like a permanent magnet (e.g., magnetite). The second, induced magnetism, is the response of a material to an external field which can be positive (paramagnetic) aligning with the external field or negative attempting to repel the external field (diamagnetic). The ground component can be significant as near magnetic ore bodies the field can almost double in size. At basaltic volcanoes like Masaya, the primary minerals responsible for the ground response are iron oxides, particularly magnetite. A complicating factor at volcanoes is the possibility that material beneath the survey area may be above its Currie point temperature; the temperature at which a material loses all of its remnant magnetism.

2.1.2. Survey procedures

Total magnetic data is presented in Chapter 4 and was collected using two different types of instruments, proton precession magnetometers and a flux gate magnetometer (in 2013). The two different instruments measure external fields differently. Proton precession magnetometers measure the magnitude of the Earth’s magnetic field by subjecting a hydrogen rich fluid (i.e., rich in protons) to a burst of radio frequency electromagnetic energy causing the protons to become polarized in the direction of the net magnetic field. When the burst of energy ceases, the proton’s magnetic moment precesses or wobbles about the external field at a rate proportional to its strength [Tyagi et al., 1983]. Using a package of 3 fluxgate sensors, one can measure the vector of the Earth’s magnetic field
instead of just the magnitude of the field. For a detailed discussion on flux gate magnetometers see Appendix A.

Under perfect circumstances, a magnetic survey is conducted with at least 2 different magnetometers, one as a base station and the other as the surveying instrument. The base station should be located within the survey area to record any diurnal variations in the Earth’s magnetic field over a survey day. While the magnetic field that is generated by the Earth can be considered static on the time scale of a few days, there are external factors, such as the Sun, which cause measurable variations in the magnetic field. The base station records these variations which are then ideally removed from the data set. Due to logistical difficulties of importing scientific equipment into Nicaragua, there was only one magnetometer available each year. To characterize the amount of error in the dataset due to lack of a base station, a proton precession magnetometer was set up within the survey area to measure the diurnal variations when not surveying. The average diurnal variation over 7 days spread across 2 years was 50 nT which is small compared to the measured anomalies of magnitudes over 1000 nT.

Surveys that cover significant elevation changes and areal extent should remove the changing magnitude of the Earth’s core field due to position. This is done by removing a theoretically calculated magnetic field, the International Geomagnetic Reference Field [IGRF; Finlay et al., 2010], using position (latitude, longitude and elevation) and time (decimal year). The residual data set is called the Total Magnetic Intensity (TMI). If sufficient data is available to characterize and remove the local regional field, the resulting data set is referred to as the residual Total Magnetic Intensity. The magnetic data collected and processed in Chapter 4 has had the IGRF removed from it using hand held GPS locations for each measurement; however, there was insufficient data to properly characterize the local regional field so presented data is the TMI and not the residual. Location data collected in 2009 and 2014 used a GPS programmed to record a position every 5 seconds instead of every point to allow for rapid surveying. The location of each measurement was then obtained by extrapolation of the GPS and magnetometer time stamps. Data in 2010 and 2012 was collected slightly differently with each point separately recorded. The hand held GPS has a horizontal accuracy of 3 to 15 m and a vertical accuracy of 6 to 30 m. This leads to a possible error of ~ 5 nT.
To obtain the best results in a ground magnetic survey, stations or survey lines are planned in a grid across the area of interest. Given the size of Masaya’s caldera, extensive `a`a lava flows, dense forest, and topography, it was not feasible to survey the whole caldera as a grid. To cover the entire area, data was collected in profiles that utilized the park trail system. Where there were no trails, short traverses across `a`a lava flows were done to gain better coverage. Where it was feasible, small magnetic grids were surveyed at the summit and on the slopes of Masaya and Nindiri cones.

The data was collected over several years (2009, 2010, 2011, and 2013) with different operators and the last set of data collected with a flux gate magnetometer in 2013. Magnetometers may have a different response to different operators and instruments. To remove this effect, it is necessary to have some overlap between surveys (Chapter 4, Fig. 4-3). This can be done by surveying the same points in the survey area or creating a set of leveling stations. Creating permanent leveling stations within Masaya caldera is not feasible due to vandalism; therefore, overlap with previous campaign years was used.

2.2. Gravity

2.2.1. Background

Measuring the gravitational field of the Earth is principally done in two ways, as an absolute measurement or a relative measurement. An absolute measurement, measures the magnitude of a gravitational field by determining the acceleration of an object falling in a vacuum [e.g., Niebauer et al., 1995]. The apparatus is complex and requires up to 24 hours of data collection to obtain one measurement. Relative gravity measurements, however, are quick (2 to 10 minutes) allowing for many measurements in a single day. The drawback is that measurements are relative to that specific meter; specifically, to the calibration of its spring. Relative or spring gravimeters operate using Hooks law:

\[ F = -Kx \]
where K is a constant, x is the length of the spring and F is the force applied to the spring. In the natural world, a spring’s length cannot be zero even if the force is zero and the relationship between force and length is not linear for the entire range of the spring. To overcome these limitations, the spring is designed such that it obeys Hooke’s law over the expected operating range of force the spring is subjected to. It is also important to note that every meter’s spring will have a slightly different spring constant and therefore the measurements between meters will not be the same. To merge the data from two different spring gravimeters requires measuring a range of values over the same location with both meters to characterize the calibration factor between the two meters.

In this thesis, the Earth’s gravitational field is mapped to investigate anomalous changes in density beneath the Earth’s surface referred to as Bouguer gravity. For this application, only the magnitudes of the gravitational anomalies are required as the goal is to characterize the excess or deficit in mass within a volume (i.e., density). Two different types of relative meters were used, LaCoste and Romberg meters (Fig 2-1) were used at Masaya volcano and a Sintrex CG5 meter (Fig 2-2) was used at Mauna Loa. Both are similar as they use springs, however the spring and sensor package is different. In a LaCoste and Romberg meter, the length of a zero length spring is measured to determine the relative field. In a Sintrex meter, the spring is made from quartz and instead of measuring the length, the capacitance of the sensor package is measured which changes due to the amount of force on the quartz spring.

An additional issue that affects all spring gravimeters is instrumental drift and data tares. Instrumental drift occurs over time due to changes in the instrument’s spring caused by inelastic stretching over its life time. Although all gravimeters drift over time, newer meters drift significantly more. It is important to note that over short time scales, instrumental drift can be approximated as linear, however over months and years drift is not linear. Repeat measurements at single location, spaced out over a survey day, are used to characterize the drift for LaCoste and Romberg meters. While this procedure is still done using Sintrex’s CG5 to look for data tares, the drift is characterized before surveying (at least every 3 months) by allowing the meter to record continuously in a quiet place for at least 24 hours. The Sintrex CG5’s onboard software then removes the calculated drift for each subsequent measurement. Data tares are offsets in the data.
A tare can manifest itself in two different ways, a recoverable tare or an irrecoverable tare. A recoverable tare has an initial offset from the original spring position followed by decay back to the starting point before the tare. An irrecoverable tare occurs when the spring length is irreversibly changed creating a permanent offset in the data. The vast majority of all tares are smaller than 0.1 mGal and to obtain accuracies better than 0.05 mGal requires many repeat measurements throughout a survey day to identify and characterize any tares [Rymer, 1989].

Figure 2-1 Schematic of a LaCoste and Romberg gravimeter
There are a number of factors other than instrumental drift that should be removed before any interpretation of Bouguer gravity is done.

1. Tidal correction: The removal of Earth Tides which are the tidal force due to the Moon and Sun [as much as 0.3 mGal; Telford et al., 1990].

2. Free air correction: The effect of elevation or distance from the centre of the Earth. With each metre of elevation gain, the gravitational field decreases on average by the theoretical free air gradient (0.3086 mGal m⁻¹), however there are areas where the actual free air gradient is significantly different from the theoretical value [e.g., Johnson, 1992].

3. Bouguer slab correction: Compensates for a rock layer, with an assumed density, between the measurement elevation level and the reference level which was sea level for both surveys [Telford et al., 1990].

4. Latitude correction: Removes the effect of the shape of the Earth calculated by a reference ellipsoid (Geoid – WSG84).

5. Terrain correction: Removes the effect of local topography an excess mass above the station (nearby hills) or a mass deficiency below the station (nearby valleys) causing a reduction in the gravitational field measured. Regardless of method used, a density value for the surrounding terrain must be assumed or obtained to calculate this correction. A clinometer is used to determine the slope of the ground.
near the station (0 to 20 m from measurement location) with is used in Hammer tests to determine the near terrain correction [Hammer, 1939]. To determine the effect of terrain farther from the station, a digital elevation model and the slope prism method is used [e.g., Olivier, and Simard, 1981].

6. Regional field: Where there is enough data to determine the gravitational field across a significantly larger area then the area of interest, it is important to remove the regional field to accentuate the anomalies of interest [Telford et al., 1990].

The terrain densities used for both the Bouguer slab correction and the terrain correction should be the same and represent the average density of the terrain from the surface to a base level (sea level is the base level used in both Chapters 4 and 5). At Masaya volcano, 2400 kgm\(^{-3}\) was assumed as it was the medium value between that of low density cinder cones and dense solid lava flows in the northern caldera. At Mauna Loa, 2330 kgm\(^{-3}\) was used as it is the accepted average density for subaerially erupted basalt at Hawaii.

2.2.2. Satellite Gravity Data

Satellite data for Nicaragua was downloaded [Förste et al., 2015; from http://icgem.gfz-potsdam.de/ICGEM/modelstab.html] to investigate if the ground data matches with larger arc wide structures. However, integrating satellite and ground data requires significantly more ground data covering the wide area. Since a regional ground based survey does not exist, the satellite gravity data is treated separately.

2.2.3. Surveying

Both Bouguer gravity studies presented in this thesis (Chapters 4 and 5) follow the same survey procedures. At the beginning and end of each day, a base station is measured within the survey area to characterize the closure (difference between base station measurements after removing Earth tides) which proxies for instrumental drift. During most survey days, only two base station measurements are made in order to cover as much area as possible. Distance between stations was determined based on what was reasonable due to time constraints and target resolution. A hand held GPS was used to
choose the location for every station and a differential GPS system was used to accurately record the position for processing. Wherever possible, the rover GPS was left recording in a dynamic mode to improve the position during occupations.

There is a slight difference in the processing of location data for the two different studies. The Mauna Loa data used a rover and permanent GPS station, which gave absolute vertical accuracies of better than 10 cm. At Masaya volcano, there was no permanent GPS station available, so a base station was set up and taken down every survey day over a point near the Park visitor center (Fig. 4-8, Chapter 4). The base station, given all the occupations, has an absolute location accuracy of approximately 15 cm. The rover GPS data is processed using the most accurate base station location providing a relative accuracy of better than 15 cm. The absolute accuracy of the gravity station locations is difficult to directly calculate. The gravity data is however, relative and hence, not having a calculated absolute accuracy on the GPS data has no effect on the results.

At each station, a clinometer was used to determine the slope of the ground close to the station. This was done by estimating an area out to 40 m and measuring the slope from the operator’s eye to that point on the ground. Before the slope data is processed to provide a terrain correction, all slopes are recalculated to remove the height of the operator (ground to eye level).

2.3. Melt inclusion geochemistry

2.3.1. Background

Crystal growth within magma is often not a perfect homogenous process which occasionally leads to the entrapment of a small amount of melt (melt inclusion) within the crystal. It is usually assumed that melt inclusions are closed systems and therefore, protected from degassing and evolution during magma ascent. Once erupted, if lava cools quickly enough, the melt inclusion will quench to a glass preserving its composition (majors and volatiles) at the time of entrapment. Melt inclusions can provide constraints on temperature, pressure, pre-eruptive chemistry, as well as whether magma mixing or fractional crystallization is dominant.
The simple assumption that melt inclusions are closed systems and do not change after entrapment is usually false. There are different mechanisms which can change melt inclusion post-entrapment, the most common being diffusion. With long magma residence times, a melt inclusion can at least partially re-equilibrate with the melt via diffusion through the host crystal overprinting the entrapment chemistry with that of the storage reservoir. Crystallization of the melt inclusion will also alter the chemistry and it is not always apparent to what extent this has affected the melt inclusion. A visible case would be the crystallization of daughter minerals inside the melt inclusion. Crystallization of the host mineral around the rim of the inclusion, however, can go unnoticed. Through equilibrium crystallization experiments and models, olivine hosted melt inclusions can be corrected for continued crystallization around the rim, however, robust equilibrium models required for other host crystals are lacking [e.g., Roeder and Emslie, 1970; Spear, 1980]. During cooling, thermal shrinkage of the inclusion can create a void space, or shrinkage bubble, into which volatiles may partition [e.g., Lowenstern et al., 2015]. Due to external stresses, a host crystal can also develop fractures that intersect a melt inclusion and allow volatiles to escape [Roedder and Ribbe, 1984].

2.3.2. Data collection

Samples from 2011 were collected from just below the surface or from exposures created through quarrying activities in order to reduce the effects of weathering. Each sample was documented at the time of collection by field sketches, abundance of minerals in the hand sample, handheld GPS coordinate and marking the location on a field map to ensure all major eruptive features were sampled at least once. Some of the features sampled in 2011 had already been collected by Moune [2006] and were not processed further. One sample (Qv5) was collected separately and processed as part of this work.

All samples were crushed to separate the crystals from the tephra (mostly lapilli). Individual crystals were hand separated with tweezers and microscope, however, the phenocryl-poor nature of the samples required a method to concentrate the phenocrysts and remove the unwanted material. A number of techniques were tried but the most efficient proved to be the use of a gold pan and picking the crystals from a water-filled petri dish under transmitted light with a microscope. Individual crystals were set aside until dried.
and then mounted in epoxy. Once the epoxy was set (24 to 48 hours), each crystal was polished until it was possible to determine if any glassy melt inclusions were present. Crystals that appeared to have glassy inclusions were carefully polished until the inclusion was intersected and exposed with a clean unblemished surface.

2.3.3. Analysis Methods

The concentration of major elements and volatiles (S, Cl and F) were measured at the Laboratoire Magmas et Volcans (LMV, Clermont-Ferrand, France) using a SX-100 CAMECA electron microprobe (EMP). It ran a 15 kV accelerating voltage for all analyses and a sampling current of 15 nA for major elements (minerals and silicate glass) and 80 nA for volatiles. The beam was defocused to 5 or 10 µm to reduce the excitation and loss of sodium. To further reduce the sodium and volatile loss, the beam was blanked or covered by a Faraday cup whenever measurements were not being collected. Each host mineral was probed at least twice in different locations, one of which was near the inclusion, to detect any possible zoning and compositional homogeneity. Volatile acquisition time was 50 s using a large pentaerythritol diffraction (PET) crystal for Cl and S and a 300 s acquisition time with a thallium acid phthalate (TAP) crystals for F. The precision of the electron microprobe is better than 5 % for major elements, excluding MnO, Na₂O, K₂O and P₂O₅ which have precisions < 10 %. The 2σ precision for Cl, S and F is 8, 10 and 66 %, respectively. Due to the relatively low concentration of F together with the large 2σ error, F concentration is not used although it is reported in Chapter 3.

In addition to using the electron microprobe, the largest inclusions also underwent Raman spectroscopy and trace element analysis. The trace measurements were performed at LMV with a Resonetics M50 EXCIMER Laser with a wavelength of 193 nm coupled to an Agilent 7500cs ICP-MS (LA-ICP-MS). Raw trace element data were processed with Glitter Software [Griffin et al., 2008] using CaO content as an internal standard. Raman spectroscopy was performed at the Laboratoire de Géologie de Lyon (France) using a Labram HR800 vis Jobin Yvon Horiba semi-confocal microscope.

Whole rock and ground mass data were also obtained to characterize the bulk chemistry and degassed volatile levels after eruption. Thin sections of the whole rock were
probed using the same method used for melt inclusions and the whole rock geochemical analysis for major elements and water was done by ICP AES at LMV with errors of ~ 10%.

2.4. References


Chapter 3. Insights On Volcanic Evolution Of Las Sierras-Masaya Volcano, Nicaragua, By Melt Inclusions

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3.1. Abstract

The Las Sierras-Masaya volcanic complex has had at least 4 basaltic Plinian eruptions over the last 30,000 years, while maintaining a stable, nearly unchanging whole rock chemistry. Melt inclusions from 8 different eruptive features spanning ~7000 years were probed for volatile, major and trace element concentrations to investigate magma processes not characterized by whole rock geochemical data. Direct sulphur and chlorine measurements together with volatile balancing using gas ratios from the erupted gas plume suggests that prior to degassing, Masaya’s magma consists of 440 ppm S, 600 ppm Cl, 1.7 wt% H₂O, and 0.145 wt% CO₂. These concentrations together with historical SO₂ flux measurements suggest a magmatic flux of ~0.17 km³ yr⁻¹ which is similar to other persistently active basaltic volcanoes. As there has been no significant effusive eruption for over 300 years and essentially no surface deformation, the degassed magma must be stored at depth to hide the un-erupted flux (~50 km³). Furthermore, fractional crystallization modeling can reproduce the variations measured in olivine hosted melt inclusions; however, this requires a thermally buffered system with a low degree of cooling in transit to the surface. The fractionation trends and the stability of the starting chemistry for 30,000 years require a very large and interconnected magma reservoir(s) at depth (greater than 5 km).
3.2. Introduction

Chronicling volcanic evolution is important to understanding and recognizing changes in volcanic activity and processes which may be a precursor to cataclysmic eruptions. It is thus crucial to use geochemical and geophysical methods together with field relationships to detail how volcanic products, magma, and the structure of a volcano have evolved through time. This is of paramount importance at volcanoes which are capable of large Plinian eruptions and have repose times that are much longer than existing data sets.

The integration of techniques to identify the processes in order to better understand and identify the precursors to explosive volcanic activity are being used at systems around the world such as Yellowstone, USA [Morgan, 2007], Campi Flegrei, Italy [Piochi et al., 2013] and Kilauea, Hawaii [e.g., Poland et al., 2009]. For many volcanoes, this is not possible as the data required has not been collected. The Las Sierras-Masaya volcanic complex (more commonly known as Masaya volcano) is an intermediary example as historical activity is well studied but studies spanning prehistoric activity are clustered around several large explosive eruptions [e.g., Williams, 1983; Kutterolf et al., 2008; Pérez et al., 2009]. In order to study the processes controlling activity at Masaya volcano and how or if they have changed over time, data from prehistoric eruptions of Masaya are necessary. Melt inclusions, which are small droplets of magma which were trapped at depth within growing phenocrysts, are an excellent tool as they preserve the pre-eruptive melt chemistry. Here we present melt inclusion geochemical analyses from 8 eruptive units within Masaya’s caldera spanning approximately 6000 years and discuss how they relate to gas and geophysical studies.

3.3. Geologic setting

Masaya volcano is one of 18 located in Nicaragua and is part of the Central American Volcanic Arc (Fig. 3-1). It is an active basaltic volcanic centre which has been in a state of near-continuous activity since its discovery in 1524 by the Spanish Conquistadors [Maciejewski, 1998 and reference therein]. Masaya is also situated within two large active local tensional tectonic features, the Nicaraguan depression and the
Managua graben. All active volcanic centres in Nicaragua are situated within the Nicaraguan depression, a half graben [McBirney and Williams, 1965], parallel to the volcanic arc. The Managua graben is approximately perpendicular to the Nicaraguan depression with its southern extent (Fig. 3-1), the Cofradias fault, intersecting or ending at Masaya caldera. Analog modeling suggests that large dense ductile volcanic intrusions beneath the northern edge of the caldera and the regional stress regime are the cause of the Managua graben and therefore the graben likely has a direct effect on the volcanic activity [Girard and van Wyk de Vries, 2005].

Masaya volcano is part of a larger complex of nested calderas. It is also among a select group of volcanic centres known to have produced basaltic Plinian eruptions [e.g., Tarawera, New Zealand; Walker et al., 1984, Etna, Italy; Coltelli et al., 1998]. The older and larger, Las Nubes caldera, formed approximately 30,000 years ago in a large Plinian
Figure 3-1  Inset map of Central America with a map of Nicaragua, Masaya represented by a red star and the Central American Volcanic Arc indicated by dashed red line and the Nicaraguan Depression highlighted in green. B) Location of large tectonic features near Masaya volcano. Modified from Girard and van Wyk de Vries, [2005]. C) Topographic map of Masaya volcano with each major cone marked by an orange triangle and dashed circle represents an identified annular structure [Mauri et al., 2012; Caravantes 2013]. Contour interval 25 m. Modified from Zurek [2010].
eruption [Williams, 1983]. Masaya caldera, the most recent caldera of the complex, formed via three Plinian eruptions over the last 7000 years [e.g., Williams, 1983; Perez and Freundt, 2006; Pérez et al., 2009]. The most recent eruption was ~1.8 ka and ejected ~2.8 km$^3$ of tephra [Perez and Freundt, 2006]. The processes that led to these eruptions are unknown.

Masaya caldera hosts many different cinder cones with the majority occurring in a semi-circle within the northwestern part of the caldera (Fig. 3-1C). The two largest cones, Masaya and Nindiri, are the only locations where historical activity is recorded. Masaya cone is the site of the last effusive eruption in 1772 when a fissure opened on its northern flank. After 1772, all activity has been centred on Nindiri cone consisting of lava lakes, pit crater formation, passive degassing and small explosive vent clearing eruptions [e.g., Stoiber et al., 1986; Williams-Jones et al., 2003; Mauri et al., 2012]. Present activity has been focused within the Santiago pit crater of Nindiri cone since its formation in 1858-1859 [McBirney 1956; Rymer et al., 1998] and is currently characterized by persistent degassing.

3.4. Previous work

Being the site of Nicaragua’s first national park, the volcano is easily accessed via a paved road into the caldera, which continues up to the summit of Nindiri and Masaya cones. The ease of access has led to this volcano being relatively well studied although long time-series datasets are still limited. For example, there are many different degassing studies that use a variety of remote sensing techniques [e.g., Stoiber et al., 1986; Nadeau and Williams-Jones, 2009; Martin et al., 2010]. Combining several studies of SO$_2$ flux [Stoiber et al. 1986; Horrocks et. al., 1999; Burton et al., 2000; Duffell, 2003; Williams-Jones et al., 2003, Nadeau and Williams-Jones, 2009], provides the longest temporal data set at Masaya with the earliest published measurements in 1972. These measurements record periods of higher degassing in 1979-1984, 1993 and 1999 and historical observations suggest periods of increased gas flux in 1852-1859, 1902-1906, 1919-1927 and 1947-1959 [Stoiber et al., 1986]. The majority of the SO$_2$ measurements were collected along roads downwind from the volcano using UV spectrometers [COSPEC, Mini-DOAS and FLYSPEC; Horrocks et. al., 1999; Burton et al., 2000; Duffell, 2003;
Williams-Jones et al., 2003; Nadeau and Williams-Jones, 2009]. The error on this type of data is typically large due to uncertainties of the plume speed overhead [Nadeau and Williams-Jones, 2009]. With this in mind, taking all published sources, the average SO$_2$ flux since 1972 is approximately 1030 metric tonnes per day (td$^{-1}$) with a standard deviation of ~30% [Stoiber et al., 1986; Horrocks et al., 1999; Burton et al., 2000; Duffell, 2003; Williams-Jones et al., 2003; Nadeau and Williams-Jones, 2006], which when even considering the error, is quite stable with time. Previous estimates of magma supply rates based on melt inclusions and degassing estimates for 1875 to 1985 suggest a supply rate of 0.091 km$^3$ yr$^{-1}$ or a total of ~10 km$^3$ during that time period [Stoiber et al., 1986]. This is a perplexing result as no significant effusive eruption has occurred since 1772 [Maciejewski, 1998], requiring a large magmatic storage system.

Gas species ratios within Masaya’s plume have been characterized by both open-path Fourier transform infrared spectroscopy (averaged over a campaign lasting approximately a week) [Horrocks et al., 1999; Burton et al., 2000; Duffell et al., 2001; Duffell, 2003; Horrocks et al., 2003; Martin et al., 2010] and gas filter packs [Stoiber et al., 1986, Allen et al., 2002; Mather et al., 2004a, 2006; Witt et al., 2008; Martin et al., 2010]. Except for data in 2001, when there was a minor phreatic eruption and reenergizing of the hydrothermal system, gas ratios from 1998 to 2009 (H$_2$O/SO$_2$, CO$_2$/SO$_2$, SO$_2$/HCl, HCl/HF and SO$_2$/HF; Table 3-1) have remained stable within error. The long term stability is remarkable as it suggests that volatile concentrations have not noticeably changed in character in over a decade of persistent degassing.

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<td>CO$_2$/SO$_2$</td>
<td>2.5 ± 0.2</td>
<td>2.3 ± 0.2</td>
<td>1.5 ± 0.4</td>
<td>2.9 ± 0.2</td>
<td>3.5 ± 0.4</td>
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<td>H$_2$O/SO$_2$</td>
<td>69 ± 9</td>
<td>66 ± 10</td>
<td>62 ± 3</td>
<td>30 ± 4</td>
<td>63 ± 7</td>
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<td>SO$_2$/SO$_4$</td>
<td></td>
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<td>190 ± 10</td>
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<td>SO$_2$/HCl</td>
<td>1.57 ± 0.05</td>
<td>1.68 ± 0.05</td>
<td>1.784 ± 0.003</td>
<td>4.589 ± 0.009</td>
<td>2.0 ± 0.03</td>
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<td>HCl/HF</td>
<td>4.5 ± 0.2</td>
<td>4.5 ± 0.2</td>
<td>6.09 ± 0.04</td>
<td>6.17 ± 0.04</td>
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The only long-term time dependant geophysical dataset is microgravity that has measured changes in mass flux from 1993 to 2016 [Rymer et al., 1998; Williams-Jones et
al., 2003; Williams-Jones, pers. com.]. During times of increased gas flux, microgravity was seen to decrease and interpreted as the development of a thick gas-rich foam layer at the top of shallow reservoirs centred under and restricted to Nindiri cone [Williams-Jones et al., 2003]. Periods of reduced gas flux show gravity increases thought to be indicative of densification of the shallow magmatic system through reduction in the frothy foam layer [Williams-Jones et al., 2003].

Any model that is applied to Masaya’s current activity must take into account the volcano’s persistent activity, a vigorous shallow magmatic system and large volumes of magma needed to represent the amount of gas erupted. One such model put forward by Stix [2007] suggested that the best way to explain Masaya volcano’s steady-state persistent degassing activity is through the combination of two different degassing models (Fig. 3-2). The first consists of convection within a conduit where low density, volatile-rich magma rises in the centre of a conduit and denser, gas poor magma descends around the edges [Fig. 3-2A; e.g., Jaupart and Vergniolle, 1989]. The second involves the development of a foam layer at the top of a magma reservoir via the accumulation of gas bubbles which degasses through a conduit at the roof of the reservoir [Fig. 3-2B; Kazahaya et al., 1994; Stevenson and Blake, 1998]. In Stix’s [2007] model, a convective conduit from depth empties into a reservoir which develops a foam layer at the top and degasses through a conduit at the top of the reservoir. The combination of the two models can explain both the microgravity results, suggesting periodic thickening of a foam layer, and the large volumes of degassed magma not being erupted. It does not, however, address how or where this material is stored.
Figure 3-2 (A) Conduit degassing [Jaupart and Vergniolle 1989] and (B) foam layer accumulation [e.g., Kazahaya et al., 1994; Stevenson and Blake, 1998] are two different degassing models that combined are proposed to explain volcanic behaviour at Masaya Volcano. Figure from Stix [2007].

Figure 3-3 Surface geology of Masaya volcano overlain with 25 m topographic contours; for simplicity, not all originally identified stratigraphic units from Williams [1983] are labeled. Symbols represent sample locations; stars are sample locations from this study where melt inclusions were found and analyzed. Simplified from Walker et al. [1993].
Williams [1983] mapped the surface geology of Masaya caldera and the surrounding area and was able to develop relative stratigraphic relationships (Fig. 3-3; units defined by relative age). Based on the relative stratigraphy, Walker et al. [1993] sampled the eruptive products from the Las Sierras-Masaya volcanic complex spanning 30,000 years and found whole rock chemistries had only minor variations (Fig. 3-4) and plot near low pressure cotectics suggesting shallow processes dominate. These studies suggest that the small observed compositional variation is the result of fractional crystallization in a shallow convecting magma chamber with periodic injection of more primitive basaltic magmas [Williams, 1983; Walker et al., 1993].

Several melt inclusion studies from olivine hosts with a forsterite range of 76 to 71 have been conducted on samples from Nindiri cone [Horrocks, 2001; Sadofsky et al., 2008; Wehrmann et al., 2011; de Moor et al., 2013]. Melt inclusion major element concentrations from these studies have a narrow range of K2O (1.34-1.49 wt%) and broadly represent an evolved basaltic composition. Volatile concentrations display a wide range with sulphur: 92 to 448 ppm, chlorine: 440 to 1531 ppm, fluorine: 10 to 672 ppm, water: 1.39 to 1.91 wt% and a single measurement of CO2 of 369 ppm. The large variations in volatile concentrations are likely due to varying degrees of pre-eruption degassing. However, these studies do not infer the initial undegassed volatile content of the melt. There may also be open system processes occurring that mask true concentrations where material may degas at depth although never erupt. More specifically, Wehrmann et al. [2011] suggests CO2 fluxing from depth is occurring at Nicaraguan volcanoes (including Masaya), agreeing with other studies that suggest CO2 fluxing is a common process in basaltic systems [e.g., Métrich and Wallace, 2008; Blundy et al., 2010; Johnson et al., 2010]. A detailed study investigating sulphur degassing constrained melt temperatures to 1097 to 1127 °C via the olivine-liquid geothermometer of Putirka [2008] and melt inclusion oxygen fugacity to be ∆QFM +1.7 (± 0.4) (as determined through micro XANES spectra of sulphur speciation), which is typical of basalt in arc settings [de Moor et al., 2013]. Prior to our study, there has not been a concerted effort to look at melt inclusions away from the active vent in Santiago crater.
Figure 3-4 Major elements from whole rock geochemical analysis [Carr et al., 2014] where yellow stars represent chemical analyses of Plinian deposits [Williams, 1983; Walker et al., 1993]. Na$_2$O plus K$_2$O versus SiO$_2$ has ranges from Nicaraguan volcanoes. Telica in green, Cerro Negro in Blue, and composite volcanoes with large ranges in composition (Cosiguina, San Cristobol, Mombotombo, Mombacho, and Conception) in orange tholeiitic-calc alkaline rock series and categories plotted over top.
3.5. Sample description

In total, 12 tephra samples from different eruptive units and one glass sample (Pele’s hair from Nindiri cone) were collected for analysis (Fig. 3-3). Each sample was named (Table 3-2) after either the eruptive feature it was collected from or the relative stratigraphic unit it belongs to as described by Walker et al. [1993]. Samples where glassy melt inclusions were not found are not listed in the results section presented below.

<table>
<thead>
<tr>
<th>Sample name * used for analysis</th>
<th>Relative stratigraphic unit [Fig. 3-2, Walker et al., 1993]</th>
<th>Eruptive feature</th>
</tr>
</thead>
<tbody>
<tr>
<td>QaW1*</td>
<td>QaW1</td>
<td>Deposit in caldera wall</td>
</tr>
<tr>
<td>Qv5</td>
<td>Qv5</td>
<td>From cinder cone just outside of the caldera</td>
</tr>
<tr>
<td>Errant</td>
<td>?</td>
<td>Errant cone</td>
</tr>
<tr>
<td>Arenal</td>
<td>Qv7</td>
<td>Arenal cone</td>
</tr>
<tr>
<td>Sentepe</td>
<td>Qv7</td>
<td>Sentepe cone</td>
</tr>
<tr>
<td>Cerro Montoso*</td>
<td>Qv 7</td>
<td>Cerro Montoso Cone</td>
</tr>
<tr>
<td>Qv9*</td>
<td>Qv9</td>
<td>Masaya cone</td>
</tr>
<tr>
<td>Visitor center*</td>
<td>Qi11</td>
<td>Visitor center cone</td>
</tr>
<tr>
<td>Comalito*</td>
<td>Qi11</td>
<td>Comalito cone</td>
</tr>
<tr>
<td>Qa16*</td>
<td>Qa16</td>
<td>Masaya and Nindiri saddle</td>
</tr>
<tr>
<td>Qa24*</td>
<td>Qa24</td>
<td>Near the base of Nindiri cone</td>
</tr>
<tr>
<td>MS1997*</td>
<td>Modern</td>
<td>Bomb erupted 1997 from Nindiri</td>
</tr>
<tr>
<td>Pele’s hair*</td>
<td>Modern</td>
<td>Pele’s Hair collected 2002</td>
</tr>
</tbody>
</table>

Table 3-2 List of all samples collected including those that contained no glassy melt inclusions, in relative chronologic order [Williams 1983; Walker et al., 1993]. Errant cone is not described as belonging to a specific relative stratigraphic unit.

Qv5, Sentepe, Arenal, and Errant either have no glassy melt inclusions or they are too small for analysis. While all samples are crystal poor (less than 10 % by volume), the lapilli in Sentepe, Arenal, and Errant cones were especially aphanitic and glassy,
containing oxides and plagioclase (<3 % by volume combined), and pyroxene (trace). The Errant sample was also the most oxidized and red in colour, perhaps due to Errant cone being older than the other eruptive features. The scoria collected from Qv5 has a slightly higher amount of phenocrysts with ~3 % by volume plagioclase and oxides with trace amounts of olivine and pyroxene. Peculiarly, in Qv5, Comolito and Cerro Montoso, inclusions of anorthite-rich plagioclase were found in olivine and pyroxene. The Pele’s hair sample was collected from the floor of Nindiri crater between San Pédro and Santiago pit craters to provide information on degassed volatile levels. Below are sample descriptions for each sample where melt inclusions were analyzed with the oldest material (based on relative stratigraphy of Williams [1983]) described first.

QaW1 is the oldest material sampled [Williams, 1983] from the San Judas formation, sampled near the village of Barillo Panama, just outside of Masaya caldera. The layer sampled contains ash and scoria air fall deposits erupted before the collapse of the caldera [Williams, 1983]. Bice [1985] also suggests that the deposit is from a basaltic Plinian eruption. Phenocryst-poor, the sample contains olivine, plagioclase and trace amounts of pyroxene and oxides.

The Cerro Montoso sample was collected on the northwest side of the cone. A hole approximately 60 cm deep was dug to get below the forest soil and access the unaltered lapilli. Extremely phenocryst-poor with a trace amount of plagioclase (~1 %), olivine and pyroxene (< 1 %).

Qv9 is scoria 2-5 cm in diameter taken from the south-eastern slope of Masaya cone in an old road/path cut. The sample was likely erupted after the collapse of Fernando crater and before the collapse of Masaya crater. Phenocrysts include a trace amount of olivine and pyroxene, with ~ 3 % plagioclase.

The national park’s visitor centre sits upon either an eruptive ridge connected to Comalito cone or a completely separate cone. The “Visitor Center” sample was taken from an area near the visitor centre beside a periodically used weather station on the cones northern side where the substrate has not been modified by park activities. Although the cone belongs to the same relative age unit as Comalito cone and is not well described in
the literature, we treat this sample as belonging to a separate eruptive feature. This unit contains 1-3 % by volume plagioclase with trace amount of olivine and pyroxene.

The area around Comalito is the site of a broad low temperature hydrothermal area which has significantly altered the erupted products of the cone. The sample (QI11) was collected from an area with no visible alteration. This sample contains 1-3 % plagioclase and trace amounts of olivine.

Qa16 was collected above the crater of Santiago near a lookout on Masaya cone. The deposit is 1-2 m thick of ash and scoria deposit which erupted from Masaya cone before the main collapse created Masaya crater [Williams, 1983]. Phenocryst-poor, the sample contains olivine, plagioclase, pyroxene and oxides.

Qa24 caps Nindiri cone and likely was deposited before the cone’s collapse which created Nindiri crater [Williams, 1983]. The Qa24 sample is scoria from the walls of Santiago crater containing olivine, plagioclase pyroxene and oxides.

The MS1997 sample is from a volcanic bomb erupted between November 17, 1997 and September 14, 1998 via small explosions from the Santiago crater. Bombs over 50 cm in diameter were ejected from Nindiri crater during this minor explosive eruptive episode. Phenocrysts include olivine, plagioclase and pyroxene.

All analyzed melt inclusions appear to be primary inclusions (trapped during the initial growth of the crystal) due to their random locations within the crystals [Roedder, 1984]. The absence of daughter minerals in the melt inclusions and the scarcity of thermal shrinkage bubbles suggests rapid cooling [e.g., Lowenstern, 2003]. Examples of the host crystals and their melt inclusions are found in Figure 3-5 and a detailed breakdown of the host crystals and the melt inclusion populations is presented in Table 3-3.
Figure 3-5 One crystal under reflective light from each sample collected in 2011. The intersected melt inclusions are circled in red although not all identified melt inclusions were probed. The two images on the left show melt inclusions hosted in olivine crystals and the right in plagioclase.
<table>
<thead>
<tr>
<th>Properties</th>
<th>QaW1</th>
<th>Qa16</th>
<th>Qa24</th>
<th>Ms1997</th>
<th>Comalito</th>
<th>Qv9</th>
<th>Cerro Montoso</th>
<th>Visitor Center</th>
</tr>
</thead>
<tbody>
<tr>
<td>Olivine</td>
<td>100-250 µm</td>
<td>250-600 µm</td>
<td>250-600 µm</td>
<td>250-600 µm</td>
<td>-</td>
<td>300-500 µm</td>
<td>350-1000 µm</td>
<td>200-600 µm</td>
</tr>
<tr>
<td>Mls in Olivine</td>
<td>Size</td>
<td>20-50 µm</td>
<td>20-125 µm</td>
<td>40-140 µm</td>
<td>10-160 µm</td>
<td>-</td>
<td>15-60 µm</td>
<td>16-150 µm</td>
</tr>
<tr>
<td></td>
<td>Bubble</td>
<td>None</td>
<td>Rare ~5 % vol</td>
<td>None</td>
<td>None</td>
<td>-</td>
<td>None</td>
<td>None</td>
</tr>
<tr>
<td></td>
<td>Post entrapment Minerals</td>
<td>None</td>
<td>None</td>
<td>None</td>
<td>None</td>
<td>-</td>
<td>None</td>
<td>None</td>
</tr>
<tr>
<td></td>
<td>Fractures</td>
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<td>None</td>
<td>None</td>
<td>None</td>
<td>-</td>
<td>None</td>
<td>None</td>
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<td></td>
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<tr>
<td>Plagioclase</td>
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<td>250-600 µm</td>
<td>250-600 µm</td>
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<td>No zoning</td>
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<td>No twinning</td>
</tr>
<tr>
<td></td>
<td>Size</td>
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<td>15-30 µm</td>
<td>10-35 µm</td>
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<td>10-80 µm</td>
<td>10-160 µm</td>
<td>-</td>
</tr>
<tr>
<td></td>
<td>Bubble</td>
<td>Rare ~2 % vol</td>
<td>Rare ~2 % vol</td>
<td>Frequent ~5 % vol</td>
<td>Frequent ~5 % vol</td>
<td>Frequent ~5 % vol</td>
<td>Frequent ~5 % vol</td>
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</tr>
<tr>
<td></td>
<td>Post entrapment Minerals</td>
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<td>None</td>
<td>None</td>
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</tr>
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<td>Fractures</td>
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</tbody>
</table>

Table 3-3 Host crystal and melt inclusion properties in each analyzed sample. Zoning is a reflection of several probes within the crystal and is thus only a qualitative assessment.
3.6. Methods

While every effort was made to acquire fresh and unaltered samples, the material from Comalito cone, as previously described, was collected from an area of broad low-temperature fumaroles (< 80 °C). To further reduce the effect of post entrapment crystallization, diffusion and re-equilibration, all samples, except MS1997 (Table 3-8), were taken from glassy lapilli which would have cooled rapidly. The samples were crushed and crystals then separated by hand. Individual crystals were mounted in epoxy and if melt inclusions were present, the crystals were ground and polished until the melt inclusions were intersected. Thin sections were prepared from each whole rock sample to measure major element and volatile concentrations of any glassy groundmass, however; only 3 samples (Comalito, Table 3-9; Qv9, Table 3-10; and VIS, Table 3-11) contained sufficient glassy groundmass for accurate measurements.

Of the 8 samples which contained glassy melt inclusions, 40 melt inclusions in olivine crystals and 57 in plagioclase crystals were chosen for analysis. The major and volatile element compositions (S, Cl, and F) of melt inclusions, host crystals and matrix glasses were analyzed at the Laboratoire Magmas et Volcans (LMV), Clermont-Ferrand, France using a SX-100 CAMECA electron microprobe (EMP) with an accelerating voltage of 15 kV. Mineral analyses used a 1 µm focused beam; for glass analyses the beam was widened to 5-10 µm to reduce Na loss. However, Na loss is observed in the smallest melt inclusions where a beam size of 2 µm was used. Volatile analyses were made using an 80 nA sampling current and a 50 s acquisition time using a large pentaerythritol diffraction (PET) crystal for Cl and S and a 300 s acquisition time with a thallium acid phthalate (TAP) crystals for F. To reduce volatile loss during analyses, measurements were taken at 20 s intervals. During sulphur analysis, variations in the wavelength of sulphur Kα X-ray, a function of its oxidation state in silicate glasses, was considered. The precision of the EMP is better than 5 % for major elements, excluding MnO, Na₂O, K₂O and P₂O₅ which have precisions less than 10 %. The 2σ precision for Cl, S and F is 8, 10 and 66 %, respectively.
Of the 97 melt inclusions only the largest melt inclusions (12 in olivine and 6 in plagioclase) and the ground mass glass in thin sections from the Visitor, Comalito and Qv9 samples were analyzed for water content using Raman spectroscopy. Using the average of two different measurements, the error is considered to be ~10 % [e.g., Losq et al., 2012]. To investigate the possibility that magma mixing is a controlling factor, trace element data were obtained through in-situ trace element analysis. The measurements were performed at LMV with a Resonetics M50 EXCIMER Laser with a wavelength of 193 nm coupled to an Agilent 7500cs ICP-MS (LA-ICP-MS). Raw data were processed with Glitter Software [Griffin et al., 2008] using CaO content as an internal standard. Analytical precision and accuracy of measurements was highly variable but most have a σ1 error of less than 10 %; elements that have a σ1 greater than 10 % are in bold in Table 3-13.

3.7. Post-Entrapment Modifications

Melt inclusions (MIs) can be affected by post-entrapment processes, which are mainly controlled by temperature. An example of a post-entrapment modification is crystallization of the host around the melt inclusion’s rim, which can significantly modify the chemistry of the enclosed glass [e.g., Gaetani and Watson, 2000]. Olivine hosted melt inclusions were corrected for post-entrapment crystallization around the rim of the glassy inclusions using the well-established equilibrium constant KD defined as 

\[(\text{FeO/MgO})_{\text{ol}}/(\text{FeO/MgO})_{\text{melt}}\]  

[Roeder and Emslie, 1970]. The relative amounts of FeO and Fe₂O₃ are unknown in our inclusions but can be estimated using ratios from previous studies, 0.232–0.262 [de Moor et. al., 2013]. The KD used was 0.30 as calculated from the equilibrium model of Toplis [2005]. The result shows post entrapment crystallization (PEC) ranges from 0 to 8.5 %; corrected results are found in Tables 3-5 through 3-12. Since the KD between melt and plagioclase is not well-known [e.g., Spear, 1980], no corrections of plagioclase crystallization at the rim of melt inclusions were undertaken so there is no propagation of errors. Therefore, the reported values are those directly measured via the microprobe.

Other specific post-entrapment modifications that can occur are the crystallization of daughter minerals within the glassy melt inclusion as it cools, fracturing of the host crystal and MIs (volatile leakage), bubble formation and diffusion. Daughter minerals can
be the same as the host (parent) or different, significantly changing the chemistry of the remaining melt inclusion. There are no daughter minerals present in the melt inclusions chosen for analysis. Fractures can develop due to overpressures in the melt inclusion creating pathways for volatiles to escape and create opportunities for secondary melt inclusions to form if the newly created fracture anneals [e.g., Lowenstern, 2003]. Melt inclusions with visible fractures were therefore avoided. Bubbles form when the melt inclusion cools and thermal contraction of the silicate glass creates a void space. Volatiles like $\text{H}_2\text{O}$ and $\text{CO}_2$ preferentially partition into the void space reducing volatile concentrations in the remaining glass [e.g., Moor et al., 2015]. However, we preformed volatile analysis in MIs with and without bubbles and there was no sign of volatile leakage. Diffusion from the melt inclusion to the host crystal and/or the surrounding magma is also a concern. Studies have shown that elements, especially H and Fe, can be lost over the residence time associated with typical arc magmatic systems [e.g., Danyushevsky et al., 2000; Spandler et al., 2007; Gaetani et al., 2012]. However, there is no obvious Fe loss in our data as our equilibrium KDs are not abnormally high (0.3) and the Fe content within the MIs are not abnormally low compared to the whole rock geochemistry. Even though likely minimal, we cannot rule out diffusion between plagioclase and the melt inclusions.

3.8. Results

3.8.1. Host Crystal

All olivine crystals used for analysis have forsterite (Fo) contents between 71 and 78 with samples from Cerro Montoso, Masaya, Nindiri (Qa24) and Visitor Center cones having a narrow Fo range of 76 to 78. Both MS1997 (a volcanic bomb erupted in 1997) and QWa1 have more evolved Fo contents of 72-73 and 71-72, respectively (Fig 3-5). Since the highest Fo content is 78, the erupted melt and crystals are evolved and have been significantly modified from the mantle-derived/primary parental melt. The range in anorthite (An) content in plagioclase crystals is between 65 and 87, with samples of higher anorthite content plagioclases correlating to olivines with higher forsterite content. Cerro Montoso, Masaya, Nindiri (Qa24) and Visitor Center cones have a narrow range of An numbers from 83 to 87 (Fig. 3-5). Consistent with the olivine, plagioclase in samples QWa1
and MS1997 show a more evolved chemistry with An numbers 78 to 81 and 65 to 78, respectively. Olivine crystals showed no zoning and the plagioclase showed only slight zonation with the largest measured difference within an individual crystal of An +/−5.

![Figure 3-6 Anorthite and Forsterite contents of plagioclase and olivine crystals which hosted melt inclusions.](image)

**3.8.2. Melt inclusions: Major elements**

All melt inclusions have a tholeiitic basaltic composition with SiO₂ between 48 and 54 wt %. The vast majority of melt inclusions fit in a narrow range of total alkali between 3.5 and 6 wt% and MgO content between 4 and 7 wt%. This is consistent with what has been reported by other studies for both whole rock [Walker et al., 1993] and melt inclusions [Horrocks, 2001; Sadofsky et al., 2008; Wehrmann et al., 2011; de Moor et al., 2013].

Potassium in mafic magmas behaves as an incompatible element [e.g., Neumann et al., 1999]; therefore, as the melt evolves and crystalizes, the concentration of potassium will increase as other elements are removed from the melt through crystallization. Plotting K₂O vs other major elements in the melt provides the ability to analyze melt evolution assuming the lowest K₂O values are the most primitive or least evolved. Figure 3-7 shows how major elements differ with different K₂O concentrations. Melt inclusions within plagioclase do not show well defined linear trends with the exception of CaO/Al₂O₃ vs K₂O.
(Fig. 3-7) which shows a linear trend in both olivine and plagioclase-hosted Mls. Even though their trends overlap, it is possible to interpret the Mls hosted in plagioclase and olivine separately as those hosted in plagioclase show a crystallization trend indicative of plagioclase and pyroxene fractionation, while the Mls in olivine are in a tighter cluster suggesting plagioclase fractionation dominates. Unlike the Mls hosted in plagioclase, those hosted in olivine do show identifiable general trends (Fig. 3-7) as both SiO$_2$ and TiO$_2$ increase with increasing K$_2$O while MgO and Al$_2$O$_3$ are inversely proportional, decreasing with increasing potassium. In the cases of FeO$^*$ vs K$_2$O, the variance is too large to clearly identify a trend. There is essentially no change in Na$_2$O over the entire range of K$_2$O within olivine hosted melt inclusions except in sample QAW1; this most likely represents sodium loss due to the high intensity of the microprobe beam.

3.8.3. Melt inclusions: Volatiles

The maximum and minimum concentrations of chlorine within the dataset come from olivine hosted melt inclusions with 590 and 380 ppm. The maximum S concentration measured was 540 ppm and 14 % of the samples had concentrations of 400 ppm or higher. The lowest concentration measured in Mls, 70 ppm, was from MS1997 a volcanic bomb that was erupted in 1997. Likewise, water has a maximum concentration of 1.4 wt% and similar to S, the lowest measured H$_2$O concentration occurs in the MS1997 sample with 0.12 wt%. While the concentration of fluorine within the melt inclusions were measured, the values obtained were at or below the detection limits for the EMP. Therefore, we are only confident in the maximum value obtained, 550 ppm, and cannot describe the variability beyond this. To measure a degassed melt, S and Cl were measured in the groundmass glass of Comalito, Qv9, and Visitor Center samples. The range of S was 15 to 18 ppm and Cl was 422 to 484 ppm (Tables 3-5 through 3-12). A sample of Pele’s hair from Nindiri crater was also probed showing slightly higher S in the glass at 40 ppm and lower Cl at 340 ppm. When the individual volatile species are plotted as a ratio over K$_2$O versus K$_2$O, the olivine hosted melt inclusions form a nearly vertical line indicative of degassing and little no to crystallization (Fig. 3-8). Except for H$_2$O, the plagioclase hosted Mls show significant variability suggesting crystallization and evolution while degassing.
Figure 3-3-7 Melt inclusion major elements versus $K_2O$ with the exception of CaO and Al$_2$O$_3$, which is plotted as a ratio versus $K_2O$. Data from previous studies comes from Sadofsky et al. [2008]; Wehrmann et al. [2011], and de Moor et al. [2013]. Units are wt%. Arrows show possible crystallization trends.
Figure 3-8 The ratio of volatile concentrations and K$_2$O as a function of K$_2$O (wt%). Gray regions represent where both crystallization and degassing is taking place, however, there seems to be 2 different pure degassing trends with respect to Cl. Data from previous studies comes from Sadofsky et al. [2008]; Wehrmann et al. [2011], and de Moor et al. [2013].

3.8.4. Trace elements

Trace element data was obtained through laser ablation on 17 different melt inclusions, 4 of which were hosted in olivine (Fig. 3-9; Table 3-13). The results broadly agree with what is reported in the whole rock chemistry [Walker et al., 1993; Carr et al., 2014]. The majority
of the trace element measurements are within $3\sigma$ regardless of which crystal the melt inclusion was hosted in.

Figure 3-9 Trace elements normalized to depleted mid-ocean ridge basalt plotted on a logarithmic axis. Green lines are from olivine hosted melt inclusions, black from plagioclase.

3.8.5. Chronological trends

Correlating to the geologic mapping work of Williams [1983], there is almost no chemical/compositional variation greater than $2\sigma$ from one relative stratigraphic unit to another using the average of each major element species in the melt inclusions. The only differences that can be seen have already been described as MS1997 and Qaw1 are the most evolved with the lowest An and Fo numbers.
3.9. Discussion

3.9.1. Magmatic processes

Each sample shows nearly the same melt inclusion chemical trends and has similar whole rock chemistries. The melt inclusions also lie along the same crystal fractionation trend (Fig. 3-7, CaO/Al₂O₃ vs K₂O). This suggests that the fractional crystallization of plagioclase and pyroxene control melt evolution. The validity of fractional crystallization as the main control on melt evolution has been tested through the use of geochemical modeling software such as Petrolog [Danyushevsky and Plechov, 2011].

There are a number of parameters that can be controlled in the software beyond a starting chemistry including pressure, temperature/pressure gradient, water concentration, oxidation state, amount of crystallization and the controlling geochemical models for each mineral. Based on the melt inclusion trends, a 4 mineral phase model is required. Testing with starting conditions of the least evolved melt inclusion, only a few of the models were able to produce olivine or plagioclase with the appropriate chemistry (Fo or An number) and two of these models that fit the olivine starting conditions were not chosen as they were not calibrated over a range of chemistries that included our data. Note, we do not have pyroxene data and therefore chose a pyroxene model that can best reproduce the data. Therefore, the models of Beattie [1993], Pletchov and Gerya [1998], Ariskin and Barmina [1999] and Nielsen [1998] were chosen for olivine, plagioclase, magnetite and clinopyroxene, respectively. Using the oxidation state from de Moor et al. [2013], the buffer QFM +2 was selected and water was set to 1.4 wt% as found in this study. The program allows the user to keep the pressure fixed which results in the modeling of fractional crystallization in a magma chamber. This did not reproduce the trend we see in our data nor is this model currently suggested for Masaya as no shallow magma chamber has been identified. Instead we used a pressure/temperature gradient (dP/dT) and a starting pressure (0.5 kbar). This opens some complexity into the modeling
as eruption temperatures for Masaya lavas are already close to the liquidus of the system at eruption [de Moor et al., 2013]. This requires a relatively high pressure/temperature gradient of ~300 bar/°C (versus the default of 0.1 bar/°C) to keep the temperature high while crystalizing.
The Petrolog [Danyushevsky and Plechov, 2011] result fits the data very well (Fig. 3-10), however to obtain the fit in TiO, 0.05% CrO was added to the model. Without CrO, the titanium in the model becomes compatible to substitute into magnetite as it crystallizes. However, the whole rock data previously collected [Walker et al., 1993; Carr et al., 2014] shows essentially zero CrO suggesting that other species such as BaO may be substituting instead of TiO. This exercise reaffirms that fractional crystallization can reproduce the general chemical variability in olivine hosted melt inclusions. The trace element data also reinforces the Petrolog [Danyushevsky and Plechov, 2011] result of fractional crystallization as all melt inclusions show the same trace element trends necessitating one source.

Irrespective of any differences in plagioclase and olivine hosted melt inclusions, the narrow range of Fo (76 to 78) and An (83 to 87) numbers for the majority of the host crystals, and stable chemistry of melt inclusions and whole rock requires a set of stable processes buffering and evolving the melt over the last 30,000 years. The simplest mechanism to explain this would be a single large interconnected magma storage system within the crust that buffers the chemistry and temperature of the system. Based on the Petrolog modeling, temperature is perhaps the most important variable within Masaya’s magmatic system as significant deviations in temperature produce very different fractional crystallization trends. Furthermore, the pressure/temperature gradient reproduced the data without any stalls or steps, which suggests that the crystallization records a constant ascent that begins after the magma has left a deeper reservoir.

3.9.2. Volatile concentrations and gas flux

Since Masaya’s melt has been essentially stable for 30,000 years and historical gas chemistry measurements have been stable [FTIR: Horrocks et al., 1999; Burton et al., 2000; Duffell et al., 2001; Duffell, 2003; Horrocks et al., 2003; Martin et al., 2010; SO$_2$: Horrocks et. al., 1999; Burton et al., 2000; Duffell, 2003; Williams-Jones et al., 2003, Nadeau and Williams-Jones, 2009], we assume that all collected samples are derived from the same chemically stable magmatic system and therefore, have the same approximate initial or un-degassed volatile concentrations. This assumption allows the use of MI volatile concentration from prehistoric eruptive sequences as a proxy for melt
concentrations spanning the entire time span. Unlike previous studies that focused on Nindiri cone, the dataset presented here enables the recognition of degassed from an undegassed sample and thus allows for better constraints on the original volatile concentrations.

It is important to first determine if the volatile content of our melt inclusions represent an un-degassed magma. There are a number of ways to estimate the original un-degassed concentrations of volatiles. The simplest method is to take the highest repeated concentration in a particular sample suite as the initial concentration in the melt. For S, 440 ppm is the highest repeated value within the study, for Cl it is 590 ppm and for H$_2$O it is 1.4 wt% (fluorine was near or below detection limit and not used). As expected, these concentrations also correspond to the highest volatile K$_2$O ratios recorded and displayed in Figure 3-7. However, due to the possibility that degassing may have occurred prior to crystallization and entrapment, these values should be considered as minimums. It is important to note that these values broadly agree with the ranges reported by other studies with the exception being the chlorine data which is significantly higher in Wehrmann et al. [2011]. The higher Cl is anomalous as it is not repeated in other studies [Sadofsky et al., 2008; de Moor et al., 2013], therefore the Cl data from Wehrmann et al. [2011] is treated as an outlier.

The pressure at which each glassy inclusion was entrapped can be estimated via geochemical models such as Papale et al. [2006], which can then determine whether S would have begun to degas. This requires knowledge of both H$_2$O and CO$_2$ concentrations; however, CO$_2$ data was not obtained in this study. The minimum entrapment pressure, determined from H$_2$O (with 1.4 wt%) alone is 16.4 MPa or at ~600 m depth. At this depth, sulphur would have already begun to exsolve and degas [e.g., Scaillet and Pichavant, 2005]. Wehrmann et al. [2011], however, reports that a single melt inclusion, measured using secondary ion mass spectrometry (SIMS) showed 369 ppm CO$_2$, 1.39 wt% H$_2$O, 448 ppm S, 1278 Cl, and 475 ppm F. While the chlorine is much higher than measured in this study, the other volatile species are consistent with the least degassed melt inclusion. Using the single major element and volatile measurement from Wehrmann et al. [2011], with a melt temperature of 1127 oC estimated by de Moor et al.
[2013], we can calculate both pressure of entrapment and the mole fraction (gas/fluid) of H2O and CO2 using the methodology of Papale et al. [2006] (Table 3-4).

<table>
<thead>
<tr>
<th></th>
<th>Inputs and results using Papale et al. [2006]</th>
</tr>
</thead>
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<tr>
<td>Temperature</td>
<td>1127 C (^b)</td>
</tr>
<tr>
<td>CO2</td>
<td>0.037 wt% (^a)</td>
</tr>
<tr>
<td>H2O</td>
<td>1.39 wt% (^a)</td>
</tr>
<tr>
<td>Mole fraction H2O Gas/fluid</td>
<td>0.18</td>
</tr>
<tr>
<td>Mole fraction CO2 Gas/fluid</td>
<td>0.82</td>
</tr>
<tr>
<td>Pressure</td>
<td>122 MPa</td>
</tr>
</tbody>
</table>

Table 3-4 \(^a\) Wehrmann et al. [2011] \(^b\) de Moor et al. [2013]

Using the methodology of Papale et al. [2006], this one measurement suggests a pressure of 122 MPa or ~ 5 km depth. At this depth, sulphur is still stable as a dissolved volatile in the melt and likely represents, barring post-entrapment processes, the concentration at depth. While this is only one data point and may be an outlier, it supports the suggestion that those melt inclusions that have near the maximum measured H2O concentration (~1.4 wt%) have accurately captured the concentration of sulphur at depth.

While sulphur concentrations determined from melt inclusions may faithfully represent the levels from a deeper body, other volatile species may have already begun to degas. The measured H2O and an estimate of CO2 concentrations in the deep melt can be verified by comparisons with gas ratio measurements if we assume that gas ratios measured in the plume are the same as the melt at depth. Many volcanic systems have large scale, vigorous hydrothermal systems with fumarole activity that can scrub water soluble gasses or provide pathways for degassing away from centralized vents. In these cases, the assumption that plume gas ratios are the same in the deeper reservoirs are likely false. Masaya does not have a vigorous hydrothermal system with only limited examples of low temperature fumaroles (Comolito, Fig. 3-1) and cold CO2 seeps [Mauri et al., 2012] suggesting that plume gas ratios likely mimic deeper melt. Open path FTIR measurements of the gas plume from Santiago crater show the molar ratios of a number of different gasses including SO2/HCl, H2O/SO2, and CO2/SO2 [Table 3-1; Horrocks et al., 1999, Burton et al., 2000; Duffell et al., 2003 and Martin et al., 2010]. Assuming that FTIR
gas ratios hold at depth, it is possible to constrain how much H₂O and CO₂ exist in the deeper melt by volatile balancing. Using the highest repeated value for S (440 ppm) in MI as the concentration at depth and assuming that S, H₂O and CO₂ were completely degassed before eruption, we can use the balance equations below to derive an expected original melt concentration of H₂O and CO₂.

\[
\frac{(H_2O/SO_2)_G}{(CO_2/SO_2)_G} = \frac{[(H_2O - H_2O^v)/(S^i - S^v)] \epsilon}{[(CO_2 - CO_2^v)/(S^i - S^v)] \epsilon} = 19.4 \quad \text{where} \quad \epsilon = 0.5 \quad (\text{eq. 2})
\]

\[
\frac{(H_2O/CO_2)_G}{(SO_2/CO_2)_G} = \frac{[(H_2O - H_2O^v)/(S^i - S^v)] \epsilon}{[(SO_2 - SO_2^v)/(S^i - S^v)] \epsilon} = 1.65 \quad \text{where} \quad \epsilon = 0.5 \quad (\text{eq. 3})
\]

Where indices i is the original concentration in the melt and v is the concentration in the degassed glass and \( \epsilon \) is a mass ratio to balance the equation (note all ratios in equations 2 and 3 are weight ratios). The result suggests that 1.7 wt% H₂O and 0.145 wt% CO₂ exist in the un-degassed parent melt. While the majority of CO₂ is likely degassing from the central vent, 0.145 wt% should be treated as a minimum due to CO₂ seeps located in several locations [Mauri et al., 2012]. The calculated water concentration is close to the measured one (1.4 wt%). The same approach for the HCl/SO₂ ratio yields volatile concentrations close to the maximum measured within the melt inclusions; therefore, maximum measure concentrations represent an un-degassed magma. Using these volatile numbers with melt inclusion chemistry, the model of Papale et al. [2006] calculates a stable depth of ~11 km for a melt with this concentration of volatiles.

### 3.9.3. Magma flux

With both SO₂ flux and the concentration of S in the deep melt, we can estimate a magmatic flux. Previous estimates have large ranges (1.3 to 5.4 x10¹¹ kg yr⁻¹ or 0.05 to 0.2 km³ yr⁻¹ assuming melt density of 2600 kg m⁻³) due to the lack of constraints on S concentrations as well as using only a subsection of the available SO₂ flux data [Stoiber et al., 1986; Martin et al., 2010]. Combining all available SO₂ data from 1972 to 2012 gives a long term average of 1030 td⁻¹ with an average standard deviation of 30%. To account for all S that Masaya releases would require correcting for other S compounds, namely H₂S and SO₄⁻² aerosols. Martin el al. [2010] measured SO₂/SO₄⁻² at 190 +/-10; however, there is only limited data on SO₂/H₂S with one reported value of ~1590 [Stoiber et al., 1986]. Here we only correct for SO₄⁻² (~5400 kg d⁻¹) due to the lack of data on H₂S and its
low abundance in comparison to SO$_2$. Using the measured 440 ppm S as the concentration at depth and the above constraints on Masaya’s sulphur budget yields $1.2 \times 10^9$ kg d$^{-1}$ of magma required to degas in order to reproduce the measured SO$_2$ flux. Assuming a melt density of 2600 kg m$^{-3}$, the magmatic flux for Masaya volcano is 0.17 km$^3$yr$^{-1}$. Although higher, this value is similar to what has been calculated for Kilauea [$0.1$ km$^3$yr$^{-1}$; e.g., Dvorak and Dzurisin, 1993] and Stromboli [0.03 - 0.15 km$^3$yr$^{-1}$; e.g., Harris and Stevenson, 2012]. Importantly, Masaya has not had a significant effusive eruption since 1772.

Based on the stability of the nature of Masaya volcano’s activity, gas emissions, whole rock geochemistry and pre-eruptive magma compositions through time, we use the magmatic flux to calculate a total volume spanning 40 years of historical measurements (1972-2012) to be 6.8 km$^3$. If the same flux is used to characterize the entire period from the last effusive eruption in 1772 (240 years), a total of 41 km$^3$ of magma would have degassed. While errors with this type of analysis are large, it is clear that there is a significant mismatch between volume of degassed and erupted magma. Intrusion and endogenous growth is a well-documented process at many volcanoes [e.g., Francis et al., 1993]; however, at Masaya no significant deformation has been reported. Microgravity measurements also show that changes in mass and density are only occurring in and adjacent to Nindiri cone suggesting, a small shallow source [Rymer et al., 1998; Williams-Jones et al., 2003]; therefore, degassed material must be stored at a depth sufficient to not cause surface deformation.

The transport of 0.17 km$^3$ yr$^{-1}$ of magma, to and from the shallow magmatic system is a problem that must fit with the melt inclusion data. For the shallow system, Stix’s [2007] (Fig. 3-2) suggestion of a two stage model works well for the historical observations of persistent degassing. The major element data from the melt inclusions does not contradict the two stage model and the degassed nature of MS1997 (a volcanic bomb) directly supports it. Transport via the second part of the model, a conduit carrying the degassed material around the edges and hot gas rich magma in the centre, is slightly harder to reconcile due to the differences in plagioclase and olivine hosted melt inclusions.
As previously discussed, each sample shows nearly the same melt inclusion chemical trends, and lie along the same crystal fractionation trend (Fig. 3-7, CaO/Al₂O₃ vs K₂O). This suggests that fractional crystallization of plagioclase and pyroxene control melt evolution. However, the relatively simple fractionation trend cannot explain the apparent duality in the data with respect to TiO and volatiles. TiO has two different trends where it stays within the melt as an incompatible species or is removed from the system. The diffusion of Ti is higher in plagioclase than olivine, so it could diffuse across the crystal effectively reducing the Ti within the melt inclusion or once more compatible with magnetite, it could fractionally crystalize within the system as titanomagnetite and/or ulvöspinel after olivine has finished crystalizing. The majority of olivine hosted MIs sit on the incompatible trend while the majority of plagioclase hosted MIs are located along the crystallization trend. This is similar to plots of the ratio of volatile species and K₂O versus K₂O where olivine hosted MIs show a degassing trend without melt evolution and plagioclase hosted MIs show evolution which could be due to post entrapment processes such as diffusion. It is also possible that the phenomenon of bringing material to the near surface and storing it at depth is not restrictive to the historical period. It is unlikely that the transport of material many kilometres to a reservoir at depth is 100% efficient and material sinking will be occasionally entrained with the upwelling material. Due to the lower density of plagioclase versus olivine it should be preferentially entrained. If this is occurring, the upwelling and downwelling material would be approximately the same chemistry so the variations would be slight based on either a longer period of time for post-entrapment processes to modify MI chemistry or provide a slightly different evolution. This process also potentially explains the presence of inclusions of plagioclase within olivine and pyroxene in Qv5, Comolito and Cerro Montoso samples.

The possibility of a large interconnected magmatic system present within the crust beneath Masaya volcano is supported by Bouguer gravity studies which infer the presence of a 6 km thick and 10-15 km diameter intrusive body, 3-9 km beneath the northern rim of the caldera [Fig. 3-11; Connor and Williams, 1989; Métaxian, 1994]. Analogue modeling of this intrusion as a ductile body, also suggests there is a feedback system between the opening of the Managua graben and intrusion at depth [Girard and van Wyk de Vries, 2005]. While there is no direct evidence of a magma storage system as part of the large intrusion suggested to be 190 to 280 km³, the active pull-apart tectonics could create the
necessary space at depth to store the > 6 km$^3$ of magma degassed since the start of historical measurements. The constant injection of heat from fresh material may also be the reason this intrusion is still ductile and able to sustain the continual opening of the Managua graben.

Figure 3-11 Cartoon summarizing approximating the locations of magma reservoirs and the processes that are likely occurring at different locations in Masaya’s magmatic system.

It is surprising that there is not more evidence from melt inclusions showing a deeper reservoir or storage system in the form of a cluster of inclusions with a water concentration near or greater than 1.4 wt%. This may be due to deeper reservoirs acting like sieves with the velocity of the melt leaving the deeper chamber slower than the settling of the denser olivine crystals [crystal settling; e.g., Huppert and Sparks, 1980].
3.10. Conclusions

Melt inclusion analysis from 8 different eruptive features of Masaya volcano directly agrees with whole rock geochemical data that shows stable whole rock chemistry for at least the last 30,000 years. Major and trace element trends in olivine hosted melt inclusions do not show evidence for magma mixing and suggest that the fractionation of plagioclase and pyroxene is a primary process within the shallow magmatic system. Fractional crystallization modeling also suggests that a low degree of cooling occurs during magma ascent to the surface and without significant pauses in transit, which is a major factor in maintaining this stable chemistry.

The data set of direct electron microprobe measurements has allowed for the identification of un-degassed versus degassed samples for S (440 ppm), Cl (600 ppm), and H2O (1.4 wt%). Volatile balancing equations provide a more accurate estimation of water concentrations at depth of 1.7 wt% and CO2 of 0.145 wt% from melt below Masaya’s shallow plumbing system. Using previous gas studies, this leads to a magmatic flux of 0.17 km³ yr⁻¹. To transport this amount of magma and not erupt it, the magmatic plumbing system must be made up of at least two parts, a large well connected reservoir deep enough not to cause deformation and an open pathway for material to move up and degas shallowly before sinking back down. A large reservoir will buffer the chemistry of the upwelling magma and the temperature which is required to reproduce the fractional crystallization trends.

It is rare that we can describe the nature of a volcano’s magmatic plumbing system over a prolonged period of time; however, for Masaya this is possible due to it being in a steady state. The new understanding of Masaya’s larger magmatic system provides the basis for future studies to investigate the volcanic activity in relation to heat budgets, reservoir dynamics and basaltic Plinian eruptions.
Table 3-5 Major element and volatile data from sample Qwa1. Fo or An refers to the chemistry of the host crystal where Fo is forsterite content in olivine and An is anorthite content in plagioclase.
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<th>FeO*</th>
<th>MnO</th>
<th>MgO</th>
<th>CaO</th>
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<th>P₂O₅</th>
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Table 3-6 Major element and volatile data from sample Qa24. Fo or An refers to the chemistry of the host crystal where Fo is forsterite content in olivine and An is anorthite content in plagioclase.
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Table 3-7 Major element and volatile data from sample Qa16. Fo or An refers to the chemistry of the host crystal where Fo is forsterite content in olivine and An is anorthite content in plagioclase.
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<th>MgO</th>
<th>CaO</th>
<th>N₂O</th>
<th>P₂O₅</th>
<th>K₂O</th>
<th>Cl</th>
<th>S</th>
<th>Fl</th>
<th>H₂O</th>
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Table 3-8 Major element and volatile data from sample MS1997. Fo or An refers to the chemistry of the host crystal where Fo is forsterite content in olivine and An is anorthite content in plagioclase.
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<th>MgO</th>
<th>CaO</th>
<th>N₂O</th>
<th>P₂O₅</th>
<th>K₂O</th>
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<th>S</th>
<th>H₂O</th>
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Table 3-9 Major element and volatile data from sample MS1997. Fo or An refers to the chemistry of the host crystal where Fo is forsterite content in olivine and An is anorthite content in plagioclase.
Table 3-10 Major element and volatile data from sample MS1997. Fo or An refers to the chemistry of the host crystal where Fo is forsterite content in olivine and An is anorthite content in plagioclase.
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<th>MgO</th>
<th>CaO</th>
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<th>P$_2$O$_5$</th>
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Table 3-11 Major element and volatile data from sample MS1997. Fo or An refers to the chemistry of the host crystal where Fo is forsterite content in olivine and An is anorthite content in plagioclase.
Table 3-12 Major element and volatile data from sample Cerro Montoso. Fo is the forsterite content in the host olivine.

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Table 3-13 Trace element data, those in bold have $\sigma_1$ greater than 10%.
3.11. References


Chapter 4. The structure of Masaya volcano as revealed by potential field methods

Jeffrey Zurek1, Guillermo Caravantes2, Glyn Williams-Jones1, Hazel Rymer2

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To be submitted to Journal of Geophysical Research

4.1. Abstract

Total magnetic and gravity data were collected at Masaya volcano, Nicaragua, to provide constraints on subsurface structures such as the location of magma reservoirs and intrusive complexes. The magnetic data shows three different types of anomalies. The first pertains to surface features such as different outcropping lava flows. The second group of anomalies correlates to areas previously identified with rising hydrothermal fluids. The third is a deeper seated anomaly overlapping the northeastern section of the caldera. The gravity data shows a broad positive ~25 mGal anomaly overlapping the northern caldera edge including the same region as the deeper magnetic anomaly. Inversions of the gravity anomaly suggest that it begins shallowly (~500 m depth) and continues to almost the bottom of the model at 3 km depth. Satellite gravity data shows that the Masaya – Mombacho section of the volcanic arc is underlain by a dense root and preliminary inversions suggest the density structure imaged via the ground data is likely connected to this deeper root. Masaya’s magma reservoir(s) is most likely at depth beneath the northern caldera rim as part of an intrusive complex inferred by gravity inversions.
4.2. Introduction

Identifying magma pathways and the structural controls that define them is important to understanding volcanic systems. The location of intrusions, faults and magma reservoirs is important information to properly design and implement a monitoring strategy at volcanoes. For volcanoes where significant resources are being spent in monitoring and the above information exists, accurate eruption forecasting is feasible. Kilauea volcano is an example where forecasting small changes in behaviour of the eruption is possible on a day to day basis as tilt meters, seismometers, GPS and gas sensors are continuously recording data from Kilauea’s summit and rift zones [Poland et al., 2009]. This allows scientists to detect small changes in the magmatic system which when put together with a long record of eruption observations, enables accurate forecasting.

In contrast, Masaya volcano is an example where little is known about its internal structure and magmatic pathways. This information is of paramount importance as approximately 6 million people live within 23 km of Masaya volcano which has experienced 3 large Plinian eruptions in the last 7000 years [e.g., Williams 1983, Kutterolf et al., 2008]. Current monitoring efforts do not allow for the identification of magma reservoirs, conduits or active faults away from the persistently degassing Santiago crater. This study addresses this lack of information about subsurface structural controls through modeling magnetic anomalies in the top 500 m of the crust and density anomalies between depths of 500 m to 10 km. To accomplish this, total magnetic and Bouguer gravity data were acquired during a ground survey campaign in and around Masaya caldera and augmented with satellite gravity data [Förste et al, 2015].

4.3. Geologic setting

Masaya volcano is part of the bigger volcanic centre known as the Las Sierras-Masaya volcanic complex, a series of nested calderas located in western Nicaragua and part of the Central American Volcanic Arc (Fig 4-1). The entire caldera complex was formed through Plinian eruptions, the most recent of which was ~1800 years ago [Pérez and Freundt, 2006]. Masaya’s eruptive products range from tholeiitic basalt to basaltic
andesite and have built a shield in the northern part of the caldera centred on its two main cinder cones, Masaya and Nindiri.

Masaya’s tectonic setting is complex due to the interaction of two large structures. The largest of which is the Nicaraguan depression, a half graben that begins in El Salvador and stretches south nearly the entire length of the country and where all 18 of Nicaragua’s historically active volcanoes lie (Fig. 4-1). The second large tectonic feature, the Managua graben, is a pull-a-part structure with its eastern boundary, the Cofradías fault, intersecting or ending at the Las Sierras-Masaya volcanic complex. There has also been a suggestion that the graben structure was created due to the combination of tectonic stresses and a large intrusive complex that likely includes Masaya’s magmatic system [Girard and van Wyk de Vries, 2005; Chapter 3].

The historical activity of Masaya volcano is dominated by persistent degassing with occasional lava lakes hosted in a pit crater within Nindiri cone [e.g., Maciejewski, 1998; Rymer et al., 1998]. There has been a negligible amount of effusive activity with only two lava flows outside of Nindiri cone in the historical record. The first in 1670...
Figure 4-1 A) Inset map of Central America with a map of Nicaragua, Masaya represented by a red star and the Central American Volcanic Arc indicated by dashed red line and the Nicaraguan Depression highlighted in green. B) Location of large tectonic features near Masaya volcano (Modified from Girard and van Wyk de Vries, [2005]). C) Topographic map of Masaya volcano with each major cone marked by an orange triangle and labeled dashed circle represents an identified annular structure [Mauri et al., 2012; Caravantes 2013]. Contour interval 25 m. Modified from Zurek [2010].
[Maciejewski, 1998] was an overflow from a lava lake hosted in Nindiri crater, while the second and more voluminous flow took place in 1772 when an eruptive fissure opened on the north side of Masaya cone [Maciejewski, 1998]. Other historical activity includes one fissure eruption, explosive vent clearing eruptions when degassing vents become blocked, and the formation of Santiago and San Pedro pit craters (Fig. 4-2). Both craters were formed through collapse in 1858-59 and since its creation, Santiago crater has been the site of all historical activity [Rymer et al., 1998].

As previously stated, Masaya is also among a small group of volcanoes known to have basaltic Plinian eruptions [e.g., Tarawera, New Zealand; Walker et al., 1984, Etna, Italy; Coltelli et al., 1998]. The older and larger, Las Sierras caldera, formed approximately 30,000 years ago in a large Plinian eruption. Three large Plinian eruptions during the last 7000 years have formed Masaya caldera [e.g., Williams, 1983; Kutterolf et al., 2008] with the most recent eruption taking place 1800 years ago [Pérez and Freundt, 2006]. Unfortunately, the processes the led to these eruptions are not understood.

4.4. Previous Studies

There have been two previous Bouguer gravity studies conducted in the region [Connor and Williams, 1989; Métaxian, 1994]. The first by Connor and Williams [1989] was exploratory with sparse station coverage, however, they concluded that there is a shallow intrusion overlapping the northern caldera margin. Métaxian [1994] covered significantly more ground with higher station density and suggested that a 6 km thick intrusive complex exists, 3-9 km beneath the northeastern rim of the caldera. Both of these studies were conducted before GPS was available for high accuracy elevation control and therefore relied on topographic maps and barometers (accuracy ~ 6 m). While these initial studies identified areas for further investigation, resolution issues and the lack of powerful inversion software at the time means that more can be done to investigate the density structures beneath Masaya volcano.
Figure 4-2 Top: Aerial photo of Masaya caldera with each major cinder cone labeled except Arenoso located just off the image along the dashed circle (representing an annular structure; Mauri et al., 2012; Caravantes, 2013) north of Cerro Montoso. Bottom: Aerial photo of Masaya and Nindiri cones with each pit crater labelled.
Other studies have investigated the mismatch of persistent degassing with lack of effusive eruptions and have found that as much as 41 km$^3$ of magma has degassed and not erupted since the last effusive eruption in 1772 [Stoiber et al., 1986; Chapter 3]. Since there is no significant measurable uplift in or around Masaya volcano and the shallow magma reservoir system is small [Williams-Jones et al., 2003], it is assumed the degassed magma is stored at depth as part of the intrusive complex identified by Métaxian [1994]. Stix [2007] suggested a mechanism for transporting material to and from the shallow system by combining two different degassing models [conduit; Jaupart and Vergniolle, 1989; and foam layer development, e.g., Kazahaya et al., 1994; Stevenson and Blake, 1998]. The first part is a conduit, which allows material to flow both towards the surface and downwards to a crustal magma reservoir. Hot material ascends within the centre of the conduit and denser degassed material sinks along the edges. This conduit is connected to the second part, a shallow magma chamber with a vent at its top from which it degasses.

4.5. Methods and data corrections

4.5.1. Total Field Magnetic data

Magnetic data were collected during a series of campaigns beginning in 2009 with an Overhauser Magnetometer primarily in the summit region of Nindiri cone. In 2010 and 2011, a Proton Precession Magnetometer was used for profiles throughout the caldera. In 2014, a three axis fluxgate magnetometer was used to expand previous coverage. Given the rough and dangerous terrain with cactus, `a`a lava flows, thick forest and topographic variations, a regular grid over the whole area was not realistic. However, smaller local grids were surveyed where feasible, namely in the summit region of Nindiri cone and within the un-vegetated zone at the southern base of Nindiri cone.

Operating an additional instrument as a base station in each survey day was not possible. To minimize the potential effect of magnetic changes due to diurnal variations, surveying was only conducted on days with no solar storm activity and compared to the record kept at the Managua airport when data was available. A stationary magnetometer was maintained on non-surveying days away from human magnetic sources.
(12°00'02.64706"N, 86°09'03.62968"W; Fig. 4-4) to determine the average diurnal difference (~50 nT). As the data were collected in different campaign years (Fig. 4-4), it is necessary to correct and remove global changes to Earth’s dynamic magnetic field in both time and space. This is achieved by subtracting the theoretical total field as calculated by the International Geomagnetic Reference Field for each station location [IGRF; Finlay et al., 2010]. Overlap in different sections, between magnetic campaigns, is then used to determine if any offset exists between different instruments and/or operators. A section of unused dirt road within the caldera was surveyed in both 2010 and twice in 2014. 2D plots of each profile (Easting vs IGRF corrected mag, Fig. 4-3) shows an offset of 300 nT between the 2010 and 2014. There is an additional 200 m overlap near the summit of Masaya cone which was collected in 2009, 2010 and 2014. This overlap confirms that the data from 2010 and 2014 has an offset of ~300 nT and that 2009 has an offset with 2014 of ~275 nT, or an ~25 nT offset between 2009 and 2010. The largest sections of overlap are with 2014 and therefore individual campaigns are corrected to 2014. There is only one complete profile in 2011 and while it does not overlap with previous campaigns, the operator and the instrument were the same as in 2010 and thus an offset of -300 nT is applied to the data.
Figure 4-3 Top: Profiles along an unused dirt road within Masaya caldera (location see Fig 4-4) that were surveyed twice in 2014 (red and black) and once in 2010 (blue). The green profile is after an offset of -300 nT has been added to the 2010 data. Bottom: Overlapping profiles on Masaya cone (location see Fig 4-4) that were surveyed in 2009 (brown), 2010 (blue) and 2014 (red). The green and orange profiles are after offsets of -300 and -275 nT have been added to the 2010 and 2009 data, respectively.

To improve the ability to model and interpret magnetic data collected at low latitudes, the data is processed using Reduction to Pole (Fig. 4-5C). Reducing the magnetic data to pole recalculates the total magnetic field as if the survey was conducted at the magnetic north pole with an inclination of 90° [e.g., Hansen and Pawlowski, 1989]. The operation transforms dipolar anomalies into monopole anomalies positioned directly above their source. At magnetic latitudes near the equator, Reduction to Pole can cause selective amplification to north-south trends in the calculated field [Blakely, 1996]. The magnetic inclination at Masaya is significantly north of the magnetic equator at 39.67°, declination of -0.26 and field strength of 36163 nT [calculated via IGRF; Finlay et al., 2010]; greatly reducing any north-south amplification due to Reduction to Pole. The one underlying assumption is that all rocks in the survey are magnetized in the same direction as the Earth’s magnetic field. The assumption is valid here as the vast majority of the rocks
in and beneath the survey area are young (<7000 years) with remanent magnetizations in the same direction as the current field. The TMI around Nindiri cone, where there is higher resolution, was reduced to pole separately at a higher resolution (20 m node spacing) to investigate short wavelength anomalies (Fig. 4-5D).

The last step is to determine the effect of topography. In terrain where the surface geology does not have high remanent magnetization or high magnetic susceptibility, topography is not a significant concern [Telford et al., 1990]. At higher latitudes, the magnetic response to topography with high magnetic susceptibility will mimic the change in elevation. However, in lower latitudes where magnetic field lines are close to horizontal, they interact with topography significantly and with greater complexity than at higher latitudes. For example, measurements that are taken at the base of a cliff will be affected by the material directly above and beside the sensor. Likewise, at the top of a cliff, the lack of magnetised material beside and below the sensor will disrupt local magnetic field lines changing the total field. In theory, it should be possible to create a model to correct for the effect of topography; however, this requires detailed knowledge of the magnetic susceptibility and the topography of the ground surface. Volcanic terrain, like that found at Masaya volcano, is extremely heterogeneous and it is not possible to constrain magnetic properties sufficiently to produce a model which can be used to accurately correct the dataset. However, forward models using geologically reasonable parameters can be qualitatively applied to identify areas where topographically-induced anomalies are located and their approximate amplitudes. Figure 4-6 is a forward model created via a software suite MAG3D [MAG3D, 2002] using a 35 m DEM and a magnetic susceptibility of 0.7 SI units for all rock included in the model. Remanent magnetization is not included in the forward model due to software limitations. The magnetic susceptibility is larger than what is typically reported for basalt [0.2 – 175 x 10\(^{-3}\) SI units; Clark and Emerson, 1991], however, Masaya lavas may contain more magnetite than average as it is a major crystalizing phase [Chapter 3]. Furthermore, the use of higher magnetic susceptibility will provide the maximum possible response due to topography allowing for easier identification of areas susceptible to topographic induced errors.

It is important to note that while each instrument is capable of obtaining measurements down to, or better than 1 nT accuracy, the error levels in this combined
dataset are significantly larger. First, due to the difficulties of importing a second magnetometer to Nicaragua, measurements of diurnal variations were only taken during non-survey days recording an average error of ~30 nT. Different days also had different baseline intensities ranging ~ 110 nT with an average variation of ~40 nT. Second, three different instruments were used in four different years with different operators. As shown in Figure 4-3, the overlap in the repeated profiles is not perfect. The differences are likely due to differences in physical position, GPS error (horizontal accuracy of 3-6 m), and slightly different responses from the different instruments and operators. Given all sources of error after corrections, we suggest that the error for this survey is less than 250 nT away from sharp topographic features. In basaltic volcanic terrains, magnetic anomalies are typically 1000s of nT in size [e.g., Hildenbrand, et al., 1993], therefore, the error/noise will be less than 25 % of any significant anomaly.

Figure 4-4 Magnetic profiles colour coded to the different years collected
Figure 4-5 A) Topographic shaded relief overlaid with 25 m contours and the location of each magnetic measurement. Areas used to determine offset between campaign years in yellow. B) Total Magnetic Intensity (TMI) overlain with 25 m contours. Star shows location of magnetic base station C) The result of applying the reduced to pole to the TMI. Dashed box approximately represents the area in D. Star shows location of magnetic base station D) Zoomed in view of the area around Nindiri cone where terrain allowed for detailed mapping. Outlines of lava flows outcropping on the surface are approximated by long dashed and arrows show flow direction.
4.5.2. Gravity

Bouguer gravity data were collected in 2009, 2010 and 2011 using different LaCoste and Romberg spring gravimeters (G-513, G-403, G127 and D-61) at 475 different gravity stations within and surrounding Masaya caldera. The gravity survey methods employed in this study have been extensively described in the literature [e.g., Rymer and Brown, 1986; Berrino et al., 1992; Battaglia et al., 2008] and therefore will only be briefly described here.

As with the magnetic data, it was not possible to collect gravity measurements in a grid due to the terrain. Instead the survey was conducted by using the network of trails throughout the caldera to obtain the best coverage possible within a reasonable amount of time. Therefore, the resolution across the survey area varies depending on station spacing and profile density. Station spacing along profiles ranged from 50 m near the summit of Nindiri and Masaya cones to 400 m away from the caldera. Thus the minimum resolvable wavelength along profile is 100 m to 800 m based on the Nyquist frequency. Determining the actual resolution of the survey is non-trivial due to changing station
spacing and error within the data; therefore, unless within an area of high station density, we assume any anomaly with spatial coverage less than 500 m to be unresolvable.

An issue that can affect spring gravimeters is a recoverable or irreversible offset in the spring length due to a shock to the meter, called a tare. Changes in spring length directly affect the gravity value and if the change is too great with respect to the target accuracy, the data will be un-useable. Base station measurements were made at least twice a day, before and after daily surveying, to identify if large data tares occurred. The base station, to which all measurements are normalized, is the same bench mark used as a base station for dynamic gravity studies [survey bench mark A1; Rymer et al., 1988; Williams-Jones et al., 2003]. The difference between the two base measurements, after being corrected for Earth tides (changing gravitational force of the Moon and Sun), is referred to as the closure. No daily closure was greater than 0.095 mGal. When surveying in remote areas, where repeating A1 at the beginning and end of the day was not reasonable, a local base station was used. All local base stations were tied to A1 so measurements could be normalized to the base station after correcting for Earth tides. Normalizing to a single station, which is assumed to be stable and unchanged, allows for the data from multiple meters and different years to be merged.

Each meter is a zero length spring gravimeter and therefore measures relative rather than absolute gravity. Furthermore, each spring is slightly different so the measurement made by each meter will be different. The variability between all meters used for this study is less than 0.1%, determined using data collected at a range of elevations from an ongoing dynamic gravity study [Rymer et al., 1998; Williams-Jones et al., 2003]. The range measured across the survey area after corrections is ~20 mGal, therefore, the maximum difference between the gravimeters is less than 0.02 mGal, well within the noise of the survey.

Before data can be interpreted and inverted to investigate subsurface density contrasts, a series of corrections are required to remove the effects of elevation, the shape of the Earth, the gravitational force of the Moon and Sun (Earth tides) and topography. The mathematics behind each correction is described in detail in the literature [e.g.,
Telford et al., 1990], therefore the discussion will focus on why each is necessary and the inherent assumptions required to use them.

The gravitational force felt at the Earth’s surface theoretically decreases by 0.3086 mGal per meter gained in elevation (also call the free air gradient or FAG); therefore, an accurate elevation for each station is essential. A Leica differential GPS system consisting of a base station GPS within the caldera and a rover GPS was used to obtain elevation control. The GPS base station, although also used for the dynamic gravity station, is not a precisely known point; therefore, daily positions over the survey were averaged to obtain one position. The absolute vertical accuracy for the base station is better than 1 m. Since the Bouguer gravity data is relative, only the positional accuracy relative to the GPS base station is a concern. Rover occupation times were greater than 5 minutes providing a vertical accuracy better than 10 cm relative to the average base station location. The gravity measurement at each station was corrected to sea level.

The gravitational effect of the rock must then be removed from the station elevation to sea level, also called the Bouguer correction. This is done by calculating the gravitational effect of an infinite slab of rock with a density equal to the average density of the terrain within the survey area. A previous gravity study estimated the average density within Masaya caldera to range from 2200 kg m$^{-3}$ (in the summit area) to 2600 kg m$^{-3}$ (NE area of the caldera) [Métxian, 1994] using the Parasnis [1952] method. The two extremes represent the difference between the cinder cones, primarily consisting of scoria, and the massive lava flows in the northeastern part of the caldera. Without further constraints, we use the median density of 2400 kg m$^{-3}$ to represent the terrain.

The Earth is an ellipsoid where the distance to the centre of the Earth is greater at the equator then at the poles. Therefore, the gravitational force is higher at poles and varies with latitude. This is particularly important for datasets that span large areas. The Latitude correction used here removes this effect using the Geodetic Reference System (GRS-1967) formula [Kovalevsky, 1971].

To remove the effect of topography, we use two different corrections, one which removes terrain effects within 60 m of the station (near terrain correction) and the other removes the effects of terrain further than 60 m (far terrain correction). To remove the near
terrain correction, a clinometer was used to estimate the height of the ground at 2 m, 20 m, 40 m and 60 m in 6 directions starting with North (and rotating 60° with every measurement). This data is then applied to Hammer charts to determine the gravitational effect of the near terrain on the gravity measured at the station. The far terrain (60 m to 5 km) was removed using the flat top prism method [e.g., John and Green, 1967] with a 35 m DEM.

Satellite Bouguer anomaly data was obtained from Förste et al., [2014] to investigate how the larger tectonic structures like the Managua graben and the Central American Volcanic Arc interact with the local density structure. The resolution of the data is significantly less than the ground survey, with a resolution of approximately 10 km [Barthelmes, 2013]; however, it covers a much wider area. It can also be used to remove the regional field and better image the shallow structure.

4.6. Results

4.6.1. Magnetic Results and Interpretation

The corrected Total Magnetic Intensity (TMI) data show a number of magnetic anomalies within the caldera (Fig. 4-5C). Many of the positive magnetic anomalies correspond directly to lava flow or unit boundaries. One of the clearest examples of this is near Nindiri’s summit on the northern side where a lava flow overflowed from a lava lake hosted with Nindiri crater in 1670. The flow creates a clear positive magnetic anomaly in contrast to the tephra surrounding it (Fig. 4-5D). Other locations where lava flows have created positive magnetic anomalies are south of the saddle between Masaya and Nindiri cones, and the western edge of the un-vegetated zone southwest of Nindiri cone. The only negative anomaly that can be directly traced to a surface feature begins on the northern flank of Masaya cone and extends to the area around Comalito cone. This area is a low temperature hydrothermal area with steam fumaroles, warm ground, CO₂ seeps and obvious surface alteration [e.g., Mauri et al., 2012]. While the majority of anomalies cannot be traced directly back to surface features, their wavelengths suggest the causative bodies are shallow. The anomalies with longer wavelengths and therefore likely a deeper
causative body are in the northwest, overlapping the caldera edge, and overlapping the southern caldera edge (Fig. 4-5).

To verify the qualitative interpretation of the shallow anomalies being due to short wavelengths, a 1-D power spectrum of the IGRF corrected data is calculated using the method of Spector and Grant [1970] (Fig. 4-7). This method correlated the average depths of a source to rate of decay of the power spectra from a minimum curvature grid with cell size of 80 m. The power spectrum shows that the vast majority of all magnetic sources are within 400 m of the ground surface with prominent clusters at 152, 71 and 35 m below the ground surface. Before making further interpretation, it is important to note that Masaya volcano and its surroundings are not likely to conform to simple layered geometries. Overlapping cones, varying flow thicknesses, faulting and previous Plinian eruptions will create horizontally and vertically heterogeneous geology.

Figure 4-7 1D power spectrum calculated from a gridded TMI map (gridding algorithm minimum curvature) before the reduction to pole operation is applied. Red lines are the 4 different slopes used to calculate depth information. Very little confidence is put into the slope D due to noise and is treated as a maximum depth.

The first slope calculated from the 1D power spectrum section may be just noise as there is some instability at the beginning of the plot. Therefore, instead of attempting to find a reasonable geologic constraint that could account for a 400 m depth, we assume that the vast majority of all magnetic sources detected by the data are within 400 m of the Earth’s surface. As to the other calculated depths, there are other datasets that give some
insight as to what might cause magnetic sources to cluster at different depths. Mauri et al. [2012] collected self-potential (SP) data throughout the caldera and identified hydrothermal systems beneath each major cone. Multi wavelet tomography was used to identify the depth to the causative bodies creating the SP anomalies attributed to hydrothermal systems. 50 % of all derived depths for the hydrothermal waters were 50 – 100 m and 40 % for 100 – 200 m. These depths overlap with the 71 and 152 m depths calculated from the 1-D power spectrum (Fig. 4-7). It is plausible that the clustering as calculated by the 1D power spectrum is due to hydrothermal alteration.

![Figure 4-8 Upward continuation of the TMI over Masaya caldera to remove/filter shorter wavelength anomalies. Overlain with 25 m elevation contours.](image)

The data was also upward continued 1000 m to filter all short wavelength anomalies and focus on two identified longer wavelength anomalies. The resulting potential field map shows the northern anomaly as a bullseye with an ~ 4 km wavelength and the southern anomaly as a ridge overlapping the caldera edge and edge of the survey area (Fig. 4-8). Simple 2D inversions suggest a depth between 600 and 300 m for the
northern anomaly which agrees with the lowest depth calculated from the one dimensional power spectrum.

### 4.6.2. Gravity results and Interpretation

The corrected Bouguer anomaly data (Fig. 4-9A) shows a 20 mGal high centred over the Northern caldera rim. The overall shape and position is approximately the same as described by previous studies [Conor and Williams, 1989; Métaxian, 1994], though with more station density within the caldera. The satellite data (Fig 4-9B) [Förste et al., 2014] shows a positive anomaly beginning north of Masaya and trending SE along the volcanic arc. The resolution of the satellite data is ~10 km [Barthelmes, 2013] and the ground survey covers approximately 12 by 12 km; therefore, there is essentially no overlap between the datasets. To evaluate the ground data with respect to the satellite data, an interpolated grid (using minimum curvature) of the satellite data was sampled at the same locations as the ground measurement locations and subtracting it from the corrected data to obtain Figure 4-10. It still has a positive Bouguer anomaly overlapping the northern caldera; however, its extent and magnitude is reduced and an artifact is introduced in the northeastern section of the field area. While this process does not remove the regional field, it does show that the majority of the terrain corrected gravity anomalies are shallower then 10 km, based on the satellite resolution. Attempts to remove the regional field by fitting a linear function or derivatives, created edge artifacts; therefore, without further constraints and the lack indicators of a deeper regional field directly impacting the data, no regional field is removed in the data.
Figure 4-9 A) Terrain corrected Bouguer gravity map, overlain with 50 m topographic contours and outline of the caldera, shows a positive 20 mGal anomaly over lapping the northern caldera margin. Black dots are the gravity station locations and the star is the base station A1. Solid white shape outlines long wavelength magnetic anomaly (Fig. 4-5). B) Satellite Bouguer gravity data overlain with 250 m topographic contours showing a positive long wavelength anomaly beginning just north of Masaya stretching southwest along the volcanic arc. The white box represents the field area.
Figure 4-10 Result from removing the satellite gravity data from the corrected ground data to evaluate the differences between the two. Shows a slightly smaller positive anomaly than seen in the terrain corrected data.

Gravity techniques are non-unique, which means there are an infinite number of possible subsurface structures or solutions to the inversion problem that can reproduce the gravitational field. Limiting inversion models to geologically feasible solutions is only as good as the geologic data available to constrain the model. We have few geological constraints, as drill hole data is lacking; and therefore, must rely on unconstrained inversion algorithms. Applying filters to analyze the depth or wavelength of the gravity anomalies provides another tool, however both inversion models and filters cannot resolve multiple anomalies that have the same horizontal position but different depths. In an attempt to provide some constraints due to the non-uniqueness of the problem, both the ground and satellite gravity data are first inverted and treated separately to investigate any
differences due to scale. The residual data is then inverted to specifically investigate shallow structures.

4.6.3. Gravity Inversions

Inversions were performed using Growth 2.0 [Camacho et al., 2000] as this software handles unconstrained models in a volcanic setting better than most inversion software suites. This is due to how Growth 2.0 solves the inversion problem by growing anomalies with relatively sharp boundaries which is geologically realistic as boundaries between intrusive and host rock are expected to be abrupt. The inversion algorithm includes a number of parameters that a user can change that will affect the inversion, such as homogeneity, balance and density contrast limits. The most important parameter is the density contrast limit. Growth 2.0 solves the inversion problem by growing anomalies until the field is reproduced, therefore, choosing density bounds effectively sets the density of the anomalies that will be grown during the inversion [Camacho et al., 2000]. Consequently, to effectively use Growth 2.0 requires knowledge of the expected density contrasts at depth or to run many different inversions to provide a range of possible solutions. Homogeneity (between 0 and 1) allows the user to determine how homogenous the anomalies are at depth. Low values will create homogenous anomalies with sharp boundaries whereas a homogeneity value near 1 will produce anomalies with gradual boundaries approaching the pre-set density limit in the centre of the anomaly. The balance parameter prioritizes model smoothness at high numbers and fit at low. Care must be taken not to over fit the data using an overly low balance factor.
Figure 4-11 Depth slices from inverting the terrain corrected Bouguer gravity data using Growth 2.0 with density constraints of +/- 400 kg m$^{-3}$. The top left depth slice is overlain by 50 m contours and black dots showing the measurement locations.
Volcanic areas are known for their heterogeneous geology both laterally and vertically due to the difference in the types of eruptive products; therefore, obtaining accurate geologic constraints can be difficult. Density maxima and minima can be approximated by looking at expected densities beneath Masaya. Surface density at Masaya varies from 2200 to 2600 kg m$^{-3}$ with an average of approximately 2400 kg m$^{-3}$ directly related to differences in eruptive products [Métaxian, 1994]. At depth, density anomalies could be related to differences in the basement rock, faulting, eruptive products and intrusive material. While perhaps unexpected, the highest density that may occur beneath Masaya volcano would be from olivine cumulates at 3300 kg m$^{-3}$ providing an upper limit. Taking the lowest density suggested for the terrain corresponding to tephra and that of olivine provides the largest possible density contrast of 1100 kg m$^{-3}$. In Growth 2.0 that would correspond to +/- 550 kg m$^{-3}$. Intrusive basalt is expected to have densities between 2800 and 3000 kg m$^{-3}$, therefore the lowest density contrast possible would be between solid lava flows (2600 kg m$^{-3}$) and intrusive basalt (2800 kg m$^{-3}$) yielding +/- 100 kg m$^{-3}$.

Inversions of the ground collected Bouguer gravity data

Both the terrain corrected Bouguer and the residual of the ground satellite data are inverted and presented here in order to investigate if/what artifacts are induced through the inversion process. On both data sets, three separate inversions are attempted using the limits +/- 550, 100 and 400 kg m$^{-3}$, the last of which represents the most geologically reasonable range based on the average value used for terrain and expected densities. The lowest density contrast inversion, +/- 100 kg m$^{-3}$, did not converge to a solution for either data set; the lowest that did is +/- 200 kg m$^{-3}$. Only inversions using the most geologically reasonable density contrast of +/- 400 kg m$^{-3}$ are displayed below (Fig. 4-11, Fig. 4-12, Fig. 4-16).
Figure 4-12 Depth slices from inverting the residual ground data using Growth 2.0. with density contrast of +/- 400 kg m\(^{-3}\). Top left depth slice is overlain with 150 m contours and each slice has an outline of Masaya caldera, Masaya cone and Nindiri cone.

The inversion of the residual gravity with +/- 200 kg m\(^{-3}\) is chaotic with the vast majority of the model at the extremes of the density contrast. Both the +/- 400 kg m\(^{-3}\) (Fig.
4-12) and +/- 550 show similar density structures with dense material north of Masaya cone overlapping the northern caldera rim. As the inversions reach 1500-1700 m a.s.l., the inversions image a positive density contrast striking northwest-southeast in the northwest quadrant of the inversion model.

**Inversion of regional satellite data**

Terrain densities are expected to be muted in the satellite gravity data set as the resolution is ~10 km, there is not a significant amount of topographic relief way from the volcanic edifices and the depth of the basement beneath the Nicaraguan depression is suggested to be 2 to 3 km [e.g., Elming and Rasmussen, 1995]. The density of the basement rock of the Nicaraguan depression has been proposed to be 2600 to 2705 kg m$^{-3}$ [Elming and Rasmussen, 1995]. In this region of the crust, significant anomalies are likely due to the volcanic arc or changing crustal thickness. Positive anomalies would likely be due to mafic intrusions with the densities between 2800 and 3300 kg m$^{-3}$. The three starting inversions that were attempted using Growth 2.0 for the satellite gravity use density contrasts of +/- 100, 250, and 400 kg m$^{-3}$.

The +/- 100 kg m$^{-3}$ inversion did not converge to a solution and the lowest density contrast Growth 2.0 was able to invert for was +/- 150 kg m$^{-3}$ (Fig. 4-13). Inversion residuals for each inversion are similar in shape suggesting each is a reasonable solution (Fig. 4-16). Each inversion shows a positive density contrast beneath nearly the entire volcanic arc (Fig. 4-13, 4-14 and 4-15). Regardless of the chosen density contrast, the density structure beneath Masaya caldera is larger than that to the NW or SE along arc such that at density contrasts of +/- 400 kg m$^{-3}$ (Fig. 4-15), the positive density contrast beneath the Masaya-Apoyo section of the arc becomes separated at depths shallower than -9000 m a.s.l. The dense root of the arc also appears to be slightly dipping to the northeast.
Figure 4-13 Depth slices from inverting the Bouguer satellite data using Growth 2.0, with density contrast of +/- 150 kg m$^{-3}$. Top left depth slice is overlain with 150 m contours and each slice has an outline of Masaya and Apoyo calderas.
Figure 4-14 Depth slices from inverting the Bouguer satellite data using Growth 2.0 with density contrast of +/- 250 kg m$^3$. Top left depth slice is overlain with 150 m contours and each slice has an outline of Masaya and Apoyo calderas.
Figure 4-15 Depth slices from inverting the Bouguer satellite data using Growth 2.0. with density contrast of +/- 400 kg m⁻³. Top left depth slice is overlain with 150 m contours and each slice has an outline of Masaya and Apoyo calderas.
Figure 4-16 Inversion residuals for the terrain corrected ground data (density contrast +/- 400 kg m$^{-3}$) and satellite gravity data (density contrast +/- 250 kg m$^{-3}$)
4.7. Discussion

4.7.1. Masaya’s shallow structure

The geophysical data presented here spans 3 different ranges of depth within the crust, with some overlap between each. The total magnetic dataset primarily gives information of the top 400-600 m of the crust, relating to the surface geology and the shallow hydrothermal system. The models and interpretation here reinforce the results from a self-potential study [Mauri et al., 2012] showing that the hydrothermal system in the region of Comalito is connected to a larger system that includes the northern flank of Masaya cone. Furthermore, magnetic lows at Comalito, Cerro Montoso and Santepe cones in the same area where self-potential multi-wavelet tomography identified hydrothermal waters. The 1-D power spectrum also provides depths that broadly match those found to represent the stable hydrothermal system at Masaya. While the magnetic results provide an independent verification that Masaya’s hydrothermal system(s) extends beyond the visible low temperature fumaroles, the magnetic data resolution is insufficient to model these detailed structures. The results add further evidence that the cones form part of a dominant annular structure as previously proposed by Mauri et al. [2012]. Preliminary InSAR results suggest that the centre of the “ring” is subsiding at ~ 1 cm yr⁻¹ [Caravantes, 2013], however this is at the edge of the detection limit of the technique. The circular ring of cones may delineate a ring fault or perhaps represent a relic of a caldera formed in the most recent Plinian eruption ~1800 years ago [Pérez and Freundt, 2006].

Only two magnetic anomalies are not associated with surface geology, the hydrothermal system or topography. Both are long wavelength anomalies located next to the caldera rim although on the opposite sides (north and south; Fig. 4-8). The southern anomaly could be due to the buried caldera rim or lava flows ponding up against the caldera margin. It is also possible that there are dikes that have intruded along the caldera rim in this region although there is no corresponding gravity signal. The northern magnetic anomaly is more interesting as the anomaly has better coverage, and thus while the resolution is poor, its areal extent is better known. The gravity inversions (Figs. 4-11, 4-12) also suggests the presence of dense material beginning with the top slice -130 m a.s.l. and continuing to at least -1000 m a.s.l. The positive density contrast most likely indicates
intrusive material and is supported by the magnetic data as solidified basaltic intrusions typically have a higher magnetic response than eruptive products [e.g., Hildenbrand et al., 1993]. Magnetite is the main magnetic mineral in basalt and when heated beyond its Currie point of \( \sim 550 \, ^\circ C \) [Zablocki and Tilling, 1976], the magnetite loses its remanent and induced magnetism. This means that at minimum, the top of the imaged intrusive body must be cooler than 550 \(^\circ C\).

### 4.7.2. Masaya’s regional structure

Each ground-satellite residual inversion model (Fig. 4-12) is dominated by a body that stretches out of the northwest from \(-1000 \, \text{m a.s.l.}\) to the bottom of the model. This is an artifact likely due to long wavelength anomalies being introduced through the satellite data. The best inversion is the terrain corrected Bouguer gravity inversion using the most geologically reasonable density bounds \(+/- 400 \, \text{kg} \, \text{m}^{-3}\); Fig. 4-11). In the inversion a dense body begins shallowly and nearly pinches completely out by the bottom of the model at \(-3000 \, \text{m a.s.l.}\) (Fig. 4-11).

Overall, the models contain a significant density anomaly corresponding to \(\sim 2700 \, \text{kg} \, \text{m}^{-3}\) (assumes zero density contrast is \(\sim 2400 \, \text{kg} \, \text{m}^{-3}\)) in the top 3 km of the crust beneath the northern region of the caldera. It is likely a safe assumption that the suggested intrusion does not contain a magma chamber of significant size in the upper 1 km of the body, as at minimum, its upper section must be below its Currie point. One way to test this would be to place tilt meters and seismometers near the anomaly to record potential episodic signals similar to those seen at Kilauea [Poland et al., 2009]. As mentioned previously and thoroughly discussed in Chapter 3, Masaya volcano must have a large storage system as the amount of erupted gas is equal to \(\sim 0.17 \, \text{km}^3 \, \text{y}^{-1}\) of magma [Chapter 3]. Based on the shallow structure of the gravity and magnetic data, the best place to explore for a shallow chamber would be in this region below \(-1000 \, \text{m a.s.l.}\). However, it is just as probable, due to the Currie point constraint, that the intrusive material is caused by old (cooled) intrusions near caldera bounding faults, a pre-caldera solidified magma chamber and/or an older plumbing system associated with several cinder cones just north of the caldera.
There is a gap in the resolvable depth between the satellite and ground Bouguer gravity data. Therefore, it is impossible to determine whether the shallow to intermediate density structure (Fig. 4-11) connects with the larger arc (Fig. 4-13, 14, 15). It is likely they are directly connected as the top resolvable slice in the satellite data shows dense material in the same region. This suggests that a dense structure beneath the northeastern rim of the caldera exists shallowly in the crust and connected to a dense root underneath the volcanic arc. This includes the previous inferred 3-6 km thick intrusive complex by Métaxian [1994].

![Satellite Bouguer gravity map](image)

**Figure 4-17** Broad view of the satellite data with the locations of the major volcanic centers, lakes (blue lines), suggested break in the volcanic arc as a white dashed line (Stoiber and Carr, 1973) and the Nicaraguan depression overlain on top.

The most notable feature in the satellite Bouguer inversions is the linear positive density contrast along this section of the volcanic arc from northwest of Masaya volcano to Mombacho volcano in the southeast (Fig. 4-17). Previous researchers have subdivided
the Central American Volcanic Arc into separate segments based on changes in the trend of the volcanic front, chemistry and eruption style, with one of those breaks occurring just northwest of Masaya volcano [e.g., Stoiber and Carr, 1973]. The satellite gravity data corroborates that there is a change in the arc in this region as the linear anomaly does not propagate further northwest. The tectonic features and faults north of Masaya are complex with different interpretations [e.g., Dewey and Algermissen, 1974; Genrich et al., 2000; Frischbutter, 2001; La Femina et al., 2002] that cannot explain the presence of all of the active structures. One promising suggestion is that the Managua graben (Fig. 4-1B) formed due to a large ductile intrusion associated with Masaya volcano and perhaps both the graben and the volcano should be monitored together [Girard and van Wyk de Vries, 2005]. The satellite gravity supports this idea as there are likely large hot ductile intrusions as part of the density structure although the size of the ductile region could be significantly larger than previously thought. This section of arc has also been the site of some of the largest recent prehistoric explosive eruptions in Nicaragua [e.g., Williams, 1983; Freundt et al., 2006; Kutterolf et al., 2008]. The caldera preceding Masaya, Las Nubes caldera, formed in a large Plinian eruption ~ 30,000 years ago and Masaya has had 3 eruptions in the last 7000 years erupting a total of ~24 km³ [Kutterolf et al., 2008]. Apoyo caldera has had two Plinian eruptions approximately 25,000 years ago with the largest erupting ~43 km³ [Kutterolf et al., 2008]. While not a direct link, the density structure could indicate a concentration of crustal magma chambers capable of producing these large eruptions and perhaps provide the instability necessary to cause large basaltic Plinian eruptions.

Field geology together with tectonic reconstructions suggests that the volcanic front moved seaward due to a steepening of the subducting slab during the Pliocene and the Pleistocene [e.g., Weinberg, 1992]. It is curious, but perhaps not surprising, that there are not more positive density anomalies east of the volcanic front that would record this migration. Looking at a wider view of the satellite gravity (Fig. 4-17), the arc within Nicaragua does not generally have high positive gravity anomalies. The wider view also suggests that the density structure beneath Masaya to Mombacho has undergone an offset near the beginning of Lake Nicaragua [deepest point 26 m; Freundt et al., 2007] and approximately follows the shore of Lake Nicaragua rather than the volcanic front. It also has a linear positive anomaly trending NE at the offset point. This large crustal feature does not correlate to visible structures on the Earth’s surface. One possible mechanism
to achieve large modifications in crustal structures without affecting the surface is a decoupling of the upper and lower crust due to rheological contrast. One active example is the southeastern Tibetan plateau, which is currently rotating clockwise without the crustal shortening that is expected due the collision of the Indo-Australian Plate with Eurasia [e.g., Royden et al., 1997]. The suggested reason for the lack of crustal shortening is the complete decoupling of the upper and lower crust allowing for different deformation fields in each [e.g., Royden, 1997]. The crust beneath Nicaragua is thin (~20 km under the volcanic arc; Harmon et al., 2013] and hot so it is possible that there is a rheological contrast at depth allowing for large crustal structures to be modified without affecting the subsurface. To fully utilize the regional satellite data further requires more ground data to merge the two datasets and crustal modeling which is currently beyond the scope of this study.

4.8. Conclusions

Masaya’s magnetic structure is dominated by its surficial geology and the hydrothermal system. This reinforces that the hydrothermal system extends beyond the low temperature fumaroles on Comalito cone and exists beneath each major cone within the caldera. This provides further evidence that the semicircular ring of cones are all part of the same structure, and may be a ring fault associated with a previous Plinian eruption.

The deepest anomaly (~400 m) the magnetic survey detected corresponds to the top of a body which is also imaged by the Bouguer gravity. Gravity inversions suggest that this structure connects to a deeper volcanic arc scale feature. The shallow portion is most likely the top of an intrusive complex, that must be below it Currie point (~580 °C) as it was detected via magnetic surveying. More broadly however, the density structure at depth represents the best place to look for Masaya’s yet unidentified magma reservoir system.

The satellite Bouguer gravity enables the investigation of gravity anomalies on a tectonic scale. A large positive gravity anomaly exists under the arc from Masaya to Mombacho and reinforces other observations that there is in fact a change in the Central American Volcanic arc just northwest of Masaya where this anomaly ends. South of Mombacho, the anomaly offsets towards the trench and continues to run parallel to the
coast southward into Costa Rica (Fig. 4-17). The nature of the positive anomaly on the satellite Bouguer gravity raises the possibility that the lower crust may be decoupled from the upper as there are no surface features that correspond to the offset.

Future work via seismic, magnetotellurics and more detailed gravity work is needed to begin to image the large broad satellite gravity anomaly to place it into a tectonic framework. Only then may it be possible to understand the changes in eruption style along arc.

4.9. References

Barthelmes, F. (2009), Definition of functionals of the geopotential and their calculation from spherical harmonic models: theory and formulas used by the calculation service of the International Centre for Global Earth Models (ICGEM), Scientific Technical Report STR, German Research Centre for geosciences


John, V. P. S. and R. Green (1967), Topographic and isostatic corrections to gravity
surveys in mountainous areas, *Geophys. Prospect.*, 15(1), 151-162, doi:

Kazahaya, K., H. Shinohara, and G. Saito (1994), Excessive degassing of Izu-Oshima
volcano: magma convection in a conduit, *Bull. Volcanol.*, 56(3), 207-216, doi:
10.1007/BF00279605.


Kutterolf, S., A. Freundt, and W. Peréz (2008), Pacific offshore record of Plinian arc

La Femina, P. C., T. H. Dixon, and W. Strauch (2002), Bookshelf faulting in Nicaragua,

of Open-path Fourier transform infra-red spectroscopy (OPFTIR) and Correlation
spectroscopy (COSPEC), Doctoral dissertation, Open University.

MAG3D; A Program Library for Forward Modelling and Inversion of Magnetic Data
over 3D Structures, version 3.1 Developed under the consortium research project
Joint/Cooperative Inversion of Geophysical and Geological Data, UBC-
Geophysical Inversion Facility, Department of Earth and Ocean Sciences,
University of British Columbia, Vancouver, British Columbia

Mauri, G., G. Williams-Jones, G. Saracco, and J. M. Zurek (2012), A geochemical and
geophysical investigation of the hydrothermal complex of Masaya volcano,
Nicaragua, *J. Volcanol. Geotherm. Res.*, 227-228(0), 15-31, doi:
10.1016/j.jvolgeores.2012.02.003.

Geophys. Union.*, 37, 83-96.


Dynamisme interne et structure de la Caldeira Masaya Nicaragua., Doctoral
dissertation, University of Savoie.

Parasnis, D. S. (1952), A Study of Rock Densities in the English Midlands,
*Geophysical Supplements to the Monthly Notices of the Royal Astronomical

Pérez, W. and A. Freundt (2006), The youngest highly explosive basaltic eruptions

Poland, M. P., J. A. Sutton, and T. M. Gerlach (2009), Magma degassing triggered by
static decompression at Kilauea Volcano, Hawaii, *Geophys. Res. Lett.*, 36,
L16306.


Chapter 5. The origin of Mauna Loa’s Nīnole Hills – Evidence of rift zone reorganization

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5.1. Abstract

In order to identify the origin of Mauna Loa volcano’s Nīnole Hills, Bouguer gravity was used to delineate density contrasts within the edifice. Our survey identified two residual anomalies beneath the Southwest Rift Zone (SWRZ) and the Nīnole Hills. The Nīnole Hills anomaly is elongated, trending northeast, and in inversions corresponding density anomalies merge at approximately -7 km a.s.l. The positive anomaly, modeled as a rock volume of ~1200 km³ beneath the Nīnole Hills, is associated with old eruptive vents. Based on the geologic and geophysical data, we propose that the gravity anomaly under the Nīnole Hills records an early SWRZ orientation, now abandoned due to geologically rapid rift-zone reorganization. Catastrophic submarine landslides from Mauna Loa’s western flank are the most likely cause for the concurrent abandonment of the Nīnole Hills section of the SWRZ. Rift zone reorganization induced by mass wasting is likely more common than currently recognized.
5.2. Introduction

Mauna Loa volcano is the largest active volcano on Earth, and has erupted 15 times since 1900. Its voluminous lava flows have repaved 90% of its surface in the past 4000 years [Lockwood and Lipman, 1987], obscuring its internal structures and earlier evolution. To open a window into Mauna Loa’s history and internal structure, geophysical techniques such as Bouguer gravity are needed to image concealed structures and intrusions. Our study focuses on an anomalous surface feature, the Nīnole Hills, which forms topographic highs on the southeast flank of Mauna Loa and represents the oldest exposed subaerial rocks on the edifice dated at 100-200 ka [Lipman et al., 1990; Jicha et. al., 2012].

Several theories have been proposed to explain the formation of the Nīnole Hills [Hitchcock, 1906; Stearns and Clark, 1930; Lipman, 1980; Lipman et al., 1990; Kauahikaua et al., 2000; Morgan et al., 2010]. Hitchcock [1906] proposed that the hills were the remnants of an older summit of Mauna Loa or a different Hawaiian volcano (Mohokea proto-volcano). Geochemical characterization of the lava flows comprising the hills showed no appreciable difference, beyond supergene alteration, between the Nīnole Hills and Holocene Mauna Loa lava flows; this led Lipman et al. [1990] to propose faulting and landslides as the formation mechanism. Several studies have suggested similar hypotheses revolving around the idea that the Nīnole Hills were once the site of an active rift zone. Lipman [1980] and Kauahikaua et al. [2000] proposed the existence of a previously unrecognized orientation of the SWRZ or a proto-rift zone south of Mauna Loa’s summit to explain an elongated gravitational high. More recently, seismic velocity modeling suggested that the Nīnole Hills may be a failed rift zone stretching from Mauna Loa’s summit through the Nīnole Hills and out past the coastline [Fig. 5-1; Park et al., 2007; Morgan et al., 2010]. Each of these mechanisms will leave a distinct geophysical signature due to differences in subsurface modification. This study presents the results of the modeling of Bouguer gravity data to infer subsurface density structures and distinguish between formational mechanisms for the origin of the Nīnole Hills.
5.3. Data reduction and residual gravity field

Bouguer gravity data were collected in April and May 2013 at 367 locations (using a Sintrex CG-5 gravimeter, Fig. 5-2) covering an area of ~600 km² following standard survey procedures [e.g., Rymer and Brown, 1986; Berrino et al., 1992; Battaglia et al., 2008]. Daily closures were on average 80 µGal and elevation control was obtained via...
kinematic GPS at each location. The GPS data was corrected using a nearby continuous GPS base station (PIIK, Fig. 5-1) providing a vertical positional error <10 cm. The largest daily closures occurred when surveying over jungle paths, dirt roads, and large changes in elevation where the gravimeter was subject to vibration and opportunities to repeat base station measurements were limited. The travel induced data tares are the most probable cause for the large closure errors observed [e.g., Crider et al., 2008; Zurek et al., 2012]. Data reduction included Bouguer corrections, terrain corrections, bathymetry, free air correction, removal of solid Earth tides, Earth curvature, and removal of the regional field. Terrain corrections were made using a clinometer to estimate the average slope of the terrain within 20 m of the measurement location and a prism method using a digital elevation model for distances greater than 20 m [Hammer, 1939]. Both sets of terrain corrections use an assumed density of 2330 kg m\(^{-3}\) for topography, based on previous gravity studies on the Island of Hawai‘i [Kauahikaua et al., 2000]. Bathymetry was corrected using a density of 1028 kg m\(^{-3}\) and a 50 m resolution bathymetric elevation model (www.soest.hawaii.edu/HMRG). To correct for elevation, a free air gradient of -322 +/- 6 µGal m\(^{-1}\) measured at the survey base station was used; this is similar to those measured at the summit of Kīlauea [-327.3 µGal m\(^{-1}\), Johnson, 1992; -330.25 µGal m\(^{-1}\), Kauahikaua and Miklius, 2003].

Removal of the regional field was accomplished with a two-step inversion process using GRAV3D [GRAV3D, 2007]. This first involved the creation of a density contrast model through the inversion of a regional dataset [Kauahikaua et al., 2000]. The density contrast model was then forward modeled, with each block in the survey area (down to 15 km depth) set to a 0 kg m\(^{-3}\) density contrast, effectively removing the influence of any density variations within the survey area. The resulting forward model represents the survey area’s regional field. The application of the corrections described above and removal of the regional field from the measured gravity data gives the residual Bouguer anomaly field (Fig. 5-2). This study primarily uses a different inversion suite, GROWTH2.0 [Camacho et al., 2011], to invert the corrected dataset. However, GRAV3D was chosen for removal of the regional field due to its flexibility and ease of use manipulating large density models [see Auxiliary Martials after references].
Figure 5-2 Residual Bouguer anomaly map contoured using kriging. Black dots represent gravity survey locations and the gray diamond represents the location of the gravity base station. Map area is approximately the same as the dashed box in Figure 5-1. Elevation contour interval of 100 m

The residual Bouguer map (Fig. 5-2) shows positive anomalies in two locations, one in the region of the Nīnole Hills (center of the figure) and the other under Mauna Loa’s SWRZ (lower left potion of the figure). No significant (>12 mGal) positive anomalies are present off the strike of the SWRZ or the Nīnole Hills. The lowest relative gravitational field (0 mGal; Fig. 5-2) is south of the Nīnole Hills, on the edge of the survey area and located away from obvious topographic features. Although on the edge of the survey area, the gravity low is unlikely to be comprised of only edge effects or error in data reduction as it is present in the data prior to terrain corrections and regional field removal. Furthermore, Moore and Chadwick [1995] identified slump and landslide deposits just off shore (Fig. 5-1) and the Ka‘ōiki-Honu‘apo fault system is also in this region, both of which would be consistent with low gravitational field.
Due to difficult terrain and limited access, gravity measurements were concentrated near trails and 4WD-accessible roads. This led to an unequal spatial distribution of measurements and thus the resolution varies depending on survey point density. Individual roads and paths have station spacing of 250 m, 500 m or 1000 m, and therefore have minimum resolvable wavelengths of 500 m, 1000 m, and 2000 m, respectively. Due to the lack of station coverage, it is not possible to determine the extent of the anomaly beneath the Nīnole Hills to the northeast. However, given that the anomaly is over 5 km in length and there is no mapped geologic structure that would truncate it, it is likely that it does not abruptly end and continues beyond the edge of the survey area. Likewise, it is not possible to determine conclusively if the SWRZ and the anomaly beneath the Nīnole Hills connect due the lack of data in this region, although they likely do.

To place the positive anomaly beneath the Nīnole Hills in context, it is important to compare it to other positive anomalies that have been described on the Island of Hawai‘i [Kauahikaua et al., 2000]. Due to differences in data reduction, the size of different volcanic edifices and their cumulate cores, comparing the horizontal gradient of anomalies rather than their magnitude is a meaningful approach. The positive anomaly associated with the Nīnole Hills has a maximum seaward horizontal gradient of -3 mGal km\(^{-1}\) (averaged over 10 km). The perpendicular seaward horizontal gradient of positive anomalies associated with all identified rift zones on the Island of Hawai‘i are -3 to -4.5 mGal km\(^{-1}\) (averaged over 10 km). Positive anomalies associated with the summits of volcanoes, with steeper slopes, Kohala, Hualālai, and Mauna Kea on the Island of Hawai‘i have on average between -4.5 and -5 mGal km\(^{-1}\) (averaged over 10 km). Kīlauea and Mauna Loa summit anomalies are outliers with -3.5 and -6.5 mGal km\(^{-1}\) (averaged over 10 km), respectively. This comparison suggests that the positive Bouguer gravity anomaly beneath the Nīnole Hills is more similar to those at rift zones then anomalies found at the summits.

### 5.4. 3D Inversions

Two gravity inversion suites, GRAV3D [GRAV3D, 2007] and GROWTH2.0 [Camacho et al., 2011], were used to invert the Bouguer gravity data in order to obtain density contrast models beneath the survey area. Only results from GROWTH2.0 are
presented due to its greater ability to handle data sets with irregular station spacing; a detailed discussion of the two inversion suites and inversion sensitivity can be found in the supplemental material [Auxiliary Martials after references]. The size of each cell (height 520 m to 1500 m and areal extent of 0.1 to 6 km²) in the inversion models were determined by GROWTH2.0 based on distance from the edge of the data set and data resolution. GROWTH2.0 was able to invert the data with density contrast end members from +/- 185 kg m⁻³ to +/- 900 kg m⁻³, each inversion resulted in positive linear anomalies beneath the SWRZ and the Niñole Hills. Density bounds outside of these two end members resulted in a failure to converge on an inversion solution. The resulting inversion from density bounds of +/- 185 kg m⁻³ contained many short wavelength perturbations forcing a fit to much coarser data and also produced the largest volume, ~3200 km³ [Auxiliary Martials after references], of positive density contrast from beneath the Niñole Hills to the bottom of the model. In contrast, inversions using the maximum density bounds, +/- 900 kg m⁻³, showed only sparse bodies and produced the lowest anomalous volume of ~500 km³ [Auxiliary Martials after references]. In the optimized inversion model (based on the stability of inversion structures and approximately half way between end member bounds, Fig. 5-3; +/- 400 kg m⁻³), the two distinct density anomalies beneath the SWRZ and the Niñole Hills begin at 0 m a.s.l., and merge by -8.2 km a.s.l. The Niñole Hills density structure in the optimized inversion is also 1220 km³ in size. The merging of these two anomalies suggest that both the SWRZ and the Niñole Hills anomalies share the same root. The linearity of the model density contrasts beneath the Niñole Hills is persistent in every inversion regardless of the starting density bounds suggesting that a linear structure runs under the Niñole Hills at depth.
Figure 5-3 Smoothed density contrast depth slices (starting at -1000 m a.s.l.) from the optimized GROWTH2.0 inversion with density bound of +/- 400 kg m\(^{-3}\) and residuals for each station. Positive density variations shown in red correspond with interpreted intrusive material based on a zero density contrast obtained from drill holes [e.g., Keller et al., 1979; Moore, 2001]. Elevation contour interval of 100 m.
5.5. Anomalous mass, density contrast and intrusion size

The calculated anomalous mass in each inversion model is $\sim 1 \times 10^{15}$ kg with each differing by less than 2% from each other. To calculate the size of any intrusions suggested by the inversion models requires assumptions of the average density with depth. Due to the lack of density information within the Nīnole Hills, the only constraints available are from drill holes [e.g., Keller et al., 1979; Moore, 2001]. From these studies we assume subaerially erupted basalts have an average density of 2300 kg m$^{-3}$ (for dry basalt) to 2500 kg m$^{-3}$ (basalt below the water table) and submarine basalt from 2700 kg m$^{-3}$ to 2900 kg m$^{-3}$. Intrusive complexes are then assumed to have densities in excess of 2900 kg m$^{-3}$. Without making assumptions regarding subsidence due to volcanic loading near the Nīnole Hills, we apply both ranges to the optimized inversion model. Furthermore, due to data resolution, intrusive volume calculations begin at -1000 m a.s.l. Based on these assumptions, the density anomalies map out the 2900 kg m$^{-3}$ or 3100 kg m$^{-3}$ isosurfaces and contain $\sim 1200$ km$^3$ which represent a significant additional mass, requiring a mechanism to concentrate it beneath the Nīnole Hills.

5.6. Formational mechanisms

It has been suggested that the Nīnole Hills were created by a number of mechanisms including: the eruptions of a proto-volcano [Hitchcock, 1906], landslide/fault blocks [Lipman et al., 1990] and a separate Nīnole Hills rift zone [Park et al., 2007; Morgan et al., 2010]. Each mechanism has a different expected gravitational field; therefore, the processed gravity data should enable us to discriminate between these hypotheses. If the Nīnole Hills were the location of a centralized proto-volcano summit, the expected gravity signal would be circular and centered above a dense intrusive complex [e.g., Zurek and Williams-Jones, 2013]. The measured gravity field cannot exclude a proto-volcano summit as it appears circular; however, the Bouguer gravity anomaly centered on the Nīnole Hills is at least 5 km in length (Fig. 5-2), may extend further to the northeast and 3-D inversions allude to a connection with the SWRZ at depth at its SW extent, suggesting a more elongate structure. Geochemical work by Lipman et al. [1990] shows that the chemistry of the hills are not appreciably different from other Mauna Loa lavas and therefore, rule out the possibility that the Nīnole Hills was a separate and different volcano. Furthermore,
seismic data shows a P-wave velocity anomaly stretching from Mauna Loa’s summit to the Nīnole Hills and a gravity anomaly [Kauahikaua et al., 2000] striking N-S from the summit (Fig. 5-4). Taking all the data together suggests that the Nīnole Hills was not the site of a proto-volcano.

Figure 5-4 Top left: Island wide Bouguer gravity data from Kauahikaua et al. [2000] near the summit of Mauna Loa also used for regional field removal. Top right: P wave isosurfaces of seismic tomography from Figure 6 in Park et al. [2007]. Bottom: Residual Bouguer gravity from this study. Solid black lines represent the approximate locations of Mauna Loa’s SWRZ and NERZ. Dashed lines outline the range of the location of the suggested proto Southern rift zone. Elevation contour interval of 250 m.

The Nīnole Hills rift zone was first proposed by Park et al. [2007] to explain seismic tomography data. However, if the proposed rift zone existed, residual gravity measurements should increase towards the center of the structure due to the likelihood of denser intrusive material [e.g., Broyles et al., 1979; Kauahikaua et al., 2000]. Two parallel
gravity profiles along roads (Fig. 5-2) that would have approximately traversed the proposed rift zone perpendicularly, did not detect an anomalous density; therefore, the gravity data does not support the existence of a separate failed rift zone beneath the Nīnole Hills (Fig. 5-3, 5-4). A recent review on the growth of Hawaiian volcanoes suggests that the original interpretation by Park et al. [2007] and Morgan et al. [2010] tied the high velocity zone beneath the Nīnole Hills to the continuation of Kilauea’s SWRZ [Lipman and Calvert, 2013]. Furthermore, Jicha et al. [2012] provides age constraints for the submarine extension of the SWRZ that suggests it has been active for more than 470 ka; including the time when the Nīnole Hills formed. In order for both the proposed Nīnole Rift zone and the SWRZ to be active simultaneously, a junction angle between them approaching 120° would be expected [Oertel, 1965; Reches and Dieterich, 1983]. However, the approximate orientation of the two rift zones would have been between 45 and 90° which has not been documented in any geologic setting.

Lastly, if the Nīnole Hills represent slump blocks and landslide scarps, characteristically basaltic flows and faults are low density and should generate a gravitational low. Although there is a negative anomaly in the southern portion of the survey area (Fig. 5-2), it is not situated beneath the Nīnole Hills. A large dense body beneath the Nīnole Hills is not consistent with the topographic feature being created by faulting and landslides but is consistent with intrusive material. While the Bouguer gravity data suggests that faulting and landslide blocks are not the primary cause of the Nīnole Hills, the hills would almost certainly have been subsequently modified by the widespread mass wasting documented across the southeast flank of Mauna Loa [e.g., Lipman et al., 1990].

5.7. Rift zone migration or reorganization

The substantial Bouguer gravity anomaly beneath the Nīnole Hills, elongated SW-NE, is consistent with a large intrusive complex that likely connects with the SWRZ. Furthermore, previously documented north-trending dikes exposed on the south flank of one of the Nīnole Hills [Lipman et al., 1990] together with recent geologic mapping by
Trusdell and Lockwood [2015] are consistent with the former presence of a rift zone since the number of dikes decreases rapidly away from rift zones [Walker, 1987].

The data presented here does not support the previous suggestions for the creation of the Nīnole Hills such as a previous volcanic summit, Nīnole Hills rift zone, or faulting and landslides but rather can be explained more comprehensively by rift zone migration or reorganization of Mauna Loa’s SWRZ. The migration of Hawaiian rift zones was first suggested by Swanson et al. [1976]. Lipman [1980] followed with detailed work on the possible migration of Mauna Loa’s SWRZ supported by geologic field evidence. Rift zone migration was also invoked to explain the large bends in Hawaiian rift zones [e.g., Lipman, 1980; Swanson, 2014].

Bouguer gravity surveys along Kīlauea’s East Rift Zone clearly show an asymmetrical gravitational field centered on the present axis of the rift zone due to rift zone migration [Broyles et al., 1979]. If the migration of Mauna Loa’s SWRZ was slow and steady we would expect to see a constant westward increase in the gravitational field due to the shallowing of intrusives as one reached the axis of the modern rift zone. However, a constant increase in the gravitational field towards the modern (or current) rift zone axis was not observed. Instead, we see two gravitational highs beneath the modern SWRZ and the Nīnole Hills. Furthermore, the imaged anomaly directly beneath the Nīnole Hills could be up to 1200 km³ in volume.

If continuous rift zone migration is the cause of the high density structure beneath the Nīnole Hills, it would require long-lived activity. Sustained activity in the vicinity of the Nīnole Hills would build a large intrusive complex and potentially produce the observed gravity signal. However, there is no supporting evidence to suggest why the Nīnole Hills would have been a preferential pathway for magma ascent or why it would have ceased to be one as the rift zone migrated. If rift zone migration does not occur at a constant velocity, then relatively rapid migration would leave behind areas with larger intrusive volumes creating a positive Bouguer anomaly in the gravitational field towards the axis of the rift zone.

Gravitational instability of volcanic piles has been shown to cause edifices to spread through slow deep seated gravitational slope deformation, faulting, and landslides.
[e.g., Lipman et al., 1988; Clague and Denlinger, 1994; Delaney et al., 1998; Benz et al., 2002; Park et al., 2009]. Variations in rift zone migration velocities necessitate rapid changes to the gravitational stability of the edifice. While not yet recognized for Hawai‘i volcanoes, a geologically instantaneous reorganization of rift zones of Anaga volcano (Tenerife, Canary Islands) has been suggested due to a large mass wasting event [Walter et al., 2005]. Based on \(^{40}\text{Ar}/^{39}\text{Ar}\) dates, geologic mapping, and aerial photograph lineament mapping, Walter et al. [2005] argues that Anaga volcano originally had straight east west rift zones. A large landslide occurred on the northern flank of the volcano which caused the rift zone to bend as it conformed to the scarp. A third rift likely formed later due to the stresses associated with a curved rift zone. Large landslides have been discovered off the coast of many Hawaiian Islands [Moore et al., 1989] and have been recognized as being associated with volcanic spreading [e.g., Oehler et al., 2005]. Applying this new information to Mauna Loa’s Southwest Rift Zone, suggests that it was positioned directly under the Niihau Hills 100-200 ka ago. An event capable of changing the edifice stress field, such as a large mass wasting event like the Ālika debris flows, then caused a geologically instantaneous reorganization of Mauna Loa’s SWRZ. Given the number of large mass wasting deposits identified around the ocean Island volcanoes, we suggest that reorganization and rapid movement of rift zones are not uncommon; however generally unrecognized.

Events capable of causing drastic changes to Mauna Loa’s internal stress field in the last 200 ka include large mass wasting events and possibly the buttressing of Mauna Loa by Kīlauea Volcano [e.g., Lipman et al., 1988; Coombs et al., 2006]. Mass wasting, such as the Ālika-1 and Ālika-2 debris flows, on the west flank of Mauna Loa, have together transported 200 to 600 km\(^3\) of material [e.g., Lipman et al., 1988] and been dated at 200-240 ka [Felton et al., 2000; Rubin et al., 2000] and 127 ka, respectively [McMurtry et al., 1999]. These are thus approximately contemporaneous with the Niihau Basalt [100-200 ka; Lipman et al., 1990; Jicha et al., 2012]. The removal of more than 100 km\(^3\) of material due to either of the Ālika debris flows [Lipman et al., 1988; Felton et al., 2000; Morgan et al, 2010] would have changed the stress/strain on the edifice potentially causing Mauna Loa’s SWRZ to jump northward or migrate faster than before the event. The effect of Kīlauea buttressing Mauna Loa’s south flank would have also significantly affected the stress/strain fields within Mauna Loa, potentially changing deformation patterns causing
relatively rapid changes to any SWRZ migration. The most recent model for the age of Kīlauea suggests it is ~275 ka [e.g., Calvert and Lanphere, 2006] and the timing of buttressing is unconstrained. It is not currently possible to determine which mechanism, buttressing, landslides, or both, could have caused the change in gravitational stability of Mauna Loa to lead to changing its rift zone orientation. However, of the above processes, instability due to mass wasting would be the most catastrophic and consequential.

Geologically fast rift zone migration or orientation change can also explain the preservation of the Ninole Hills, as a stable rift zone will create a topographic high along its axis. When the rift zone moves westward, the old rift axis will shelter land seaward from lava inundation. Although it is not possible to conclusively determine the mechanism responsible for the imaged density anomaly beneath the Ninole Hills, we interpret the gravity data as evidence of an early Mauna Loa Southwest Rift Zone [Lipman 1980, Kauahikaua et al., 2000]. This study suggests that migration or offsetting of rift zones by large landslides is likely more important to the development and modification of basaltic island volcanism then previously recognized.

5.8. References


GRAV3D, (2007), A program library for forward modeling and inversion of gravity data over 3D structures, developed under the consortium research project Joint/Cooperative Inversion of Geophysical and Geological Data: UBC-Geophysical Inversion Facility, Department of Earth and Ocean Sciences, The University of British Columbia, Vancouver, version 20070309.


Park, J., J.K. Morgan, C.A. Zelt, and P.G. Okubo (2009), Volcano-tectonic implications of 3-D velocity structures derived from joint active and passive source tomography of the island of Hawaii, *J. Geophys Res.*, 114, B09301, 19


5.9. Auxiliary Material

Residual gravity data from this study was inverted with both GROWTH2.0 [Camacho et al., 2011] and GRAV3D [GRAV3D, 2007] to investigate the inherent non-uniqueness of gravity inversions. Each software suite has strengths and weaknesses when dealing with different datasets. For instance, GRAV3D inversions attempt to create density contrasts that are smooth in three directions and match observations consistent with data error levels and degree of misfit [GRAV3D, 2007]. For datasets with irregular station spacing, 3D anomalies will appear smeared to maintain smoothness across the model; however, for datasets with a regular grid, GRAV3D balances smoothness, data misfit and resolution well. GROWTH2.0 treats inversion problems differently, using a least squares inversion approach [Tarantola, 1988] that “grows” positive and negative density contrasts with predetermined density values. This method will produce relatively sharp contrast boundaries as model smoothness is less of a priority then in GRAV3D. This process works well with irregular station spacing as imaged density structures are not excessively smoothed.
To remove the effect of density contrasts outside of the survey area, the regional gravitational field must be removed. There are several techniques using filters and derivatives to accomplish this, however, they often leave edge effects unless the dataset is resampled to a smaller area. If a much larger and wider dataset exists, a two-step inversion process can be performed to remove the regional field reducing edge effects and allowing the use of the full dataset. This is done by taking the density model created by inverting the large dataset and setting the area corresponding to the survey to a zero density contrast. Forward modeling the modified density will result in the regional field in the area of the survey which can then be removed; the Kauahikaua et al. [2000] island wide dataset was used for this purpose. GRAV3D was used to complete this step due to its flexibility and ease of use when manipulating large density models. GRAV3D inversion models can be directly imported into MeshTools, an associated software which can display, edit and forward model gravity and magnetic models. At this time, GROWTH2.0 inversion models are not easily modified and forward modeled.

Both inversion suites can reproduce the residual gravitational field with approximately the same amount of model misfit and the same amount of required additional mass (within 1 to 2 %). However, GRAV3D inversions show an intrusive volume of 1.5 km$^3$ to 6.5 km$^3$ beneath the Ninole Hills; much less than the 500 km$^3$ to 3200 km$^3$ obtained from GROWTH2.0. The large discrepancy in intrusive volumes is indicative of how GRAV3D smears density contrasts due to poor station coverage. While the intrusive volumes calculated by GROWTH2.0 are likely too high and those calculated by GRAV3D are likely too low, the value range obtained by GROWTH2.0 is likely more accurate based on its effectiveness in solving inversions with irregular station spacing. Even though GRAV3D and GROWTH2.0 solve inversion problems differently, leading to different concentrations of dense material, the resulting shapes of the density contrasts are similar. The anomaly imaged beneath the Ninole Hills by GROWTH2.0 is linear and elongated in a NE-SW direction and the GRAV3D (Fig. 5-5) inversions show a more oval body elongated in the same direction. Even though the inversion method is inherently non-unique, two different inversion suites, using different approaches to inversion problems, show a concentration of dense material beneath the Ninole Hills elongated in a SW-NE direction.
Figure 5-5 Smoothed density contrast depth slices from the optimized GRAV3D inversion with density bound of +/- 400 kg m\(^{-3}\) and the residuals for each station. Black dots in the -2190 m a.s.l. depth slice represent gravity measurement locations. Contours at 100 m.
Figure 5-6 Smoothed density contrast depth slices from the optimized GROWTH2.0 inversion with density bound of +/- 185 kg m\(^{-3}\) and the residuals for each station. Positive density variations shown in red approximately correspond with interpreted intrusive material based on a zero density contrast obtained from drill holes [e.g., Keller et al., 1979; Moore, 2001]. Contours at 100 m.
Figure 5-7 Smoothed density contrast depth slices from the optimized GROWTH2.0 inversion with density bound of +/- 900 kg m\(^{-3}\) and the residuals for each station. Positive density variations shown in red approximately correspond with interpreted intrusive material based on a zero density contrast obtained from drill holes [e.g., Keller et al., 1979; Moore, 2001]. Contours at 100 m.
Chapter 6. Conclusions

6.1. Masaya

6.1.1. Geochemistry

Melt inclusions were analyzed from 8 different eruptive events from Masaya caldera. Although the samples spanned a significant time period of ~6000 years, there was essentially no difference in their trace elements or major chemistry. Together with geochemical modeling showing that the minor differences between samples can be explained by fractional crystallization, this suggests that all samples are derived from the same source. Furthermore, magma mixing of different chemical batches is minimal or non-existent.

The conditions required to explain the data via fractional crystallization have their own implications. First, Masaya crystalizes as a four component system with high anorthite plagioclase, 71-78 forsterite olivine, clinopyroxene and magnetite. And secondly, the crystallization occurs on route to the surface without major pauses at a nearly constant temperature. The geochemical modeling also shows that a significant decrease in temperature (cooling during ascent) will result in drastically different fractional crystallization curves due to lowering of the forsterite and anorthite contents in the olivine and plagioclase. This suggests that the temperature of the system has not varied appreciably over the time period covered by the samples. This requires a large magmatic system with a continual flux of material to maintain the stable temperature of the system.

Due to the chemical stability of the melt inclusions, whole rock and historical gas measurements, the differences between samples with respect to volatiles are interpreted to be due to different amounts of degassing and not a difference in volatile concentrations coming from depth. With this assumption, minimum volatile concentrations are determined for the un-degassed magma coming from depth (440 ppm S, 590 ppm Cl, 1.6 wt% H₂O, 0.145 wt % CO₂). Using the historical SO₂ data with the melt inclusion data, an average magmatic flux over the period of historical measurements is estimated at 0.17 km³ yr⁻¹.
This is comparable with other persistently active basaltic systems around the world, however Masaya has not had an effusive eruption since 1772.

This leads to the conclusion that Masaya must have a large interconnected crustal magma chamber that has been essential unchanged over the last 6000 years. Furthermore whatever the cause of the basaltic Plinian eruptions at Masaya, the magmatic system is in this steady state before and after the events with no obvious perturbations.

6.1.2. Masaya Geophysics

Magnetic data confirms previous research that there are hydrothermal waters away from the only surface exposures on the northern flank of Masaya cone and surrounding Comalito cone under Cerro Montoso. The magnetic anomaly relating to the deepest detected magnetic body corresponds to an area that gravity inversions identify as dense material that connects to larger arc scale density structures. This material is most likely intrusive material and must be below it Currie point (~580 °C), however the density structure at depth represents the best place to look for Masaya’s yet unidentified magma reservoir system.

At the volcanic arc scale, the satellite gravity data reinforces other observations that there is in fact a change in the Central American Volcanic arc just northwest of Masaya. A large positive gravity anomaly exists under Masaya to Mombacho and then offsets towards the trench and continues to run parallel to the coast southward into Costa Rica. This density structure stops just northwest of Masaya where the suggested break is thought to occur based on eruption style, chemistry and strike of the volcanic front. The nature of the positive anomaly on the satellite Bouguer raises the possibility that the lower crust may be decoupled from the upper as there are no surface features that correspond to the offset.

6.2. Mauna Loa and the Ninole Hills

The anomalous topographic feature that dominates the local landscape on the southwest flank of Mauna Loa, the Ninole Hills, hosts a large dense intrusive structure at
depth. The most reasonable explanation for the linear intrusion beneath the Hills is that it was once the site of Mauna Loa’s Southwest Rift Zone. The nature of the overall Bouguer gravity anomaly shows that the current SWRZ and the Nīnole Hills are separate distinctive anomalies, and therefore, does not suggest a slow continual migration process like what is seen at Kilauea for its East Rift Zone. Instead, the structural evolution of Mauna Loa was likely much more dramatic than previously recognized as the Bouguer data suggests that the SWRZ must have jumped or moved northward. To shift a stable rift zone kilometers requires an event capable of drastically changing the internal stresses and stability of the edifice. The most likely candidate is the Alika debris flows that removed over 200 km³ of material from the northwest flank of Mauna Loa. The unloading of material from the flank may have caused the SWRZ to undergo a geologically instantaneous shift towards the scarp of the mass wasting events. Given the number of large mass wasting events identified along the Hawaiian Archipelago this dramatic reorganization of volcanic structure may be more common than recognized.

6.3. Future Work

The chapters contained in this thesis have expanded what is known about both Masaya and Mauna Loa volcanoes and proposed a method of geophysical surveying which will be safer, cover large areas and have a low cost. There is however, much more to investigate as this work has also posed new questions.

While Chapters 3 and 4 were not able to answer why or how Masaya volcano has gone through several large basaltic Plinian eruptions, they do show the state (steady state) of the magma reservoir(s) before and after these events. Future studies will need to take this into account when developing models and plausible mechanisms to explain the Plinian eruptions.

The satellite Bouguer gravity shows a tantalizing picture of an arc scale density structure. To expand on this study, more regional ground data is needed to merge the ground and the satellite datasets, which will then allow for rigorous inversions with seismic constraints. This type of work may provide an explanation as to why the Central American Volcanic Arc has segment breaks.
As stated above, in Chapter 5 we propose that the Nīnole Hills was the site of Mauna Loa’s SWRZ ~120 ka. Further gravity and geologic mapping in conjunction with the locations of other large mass wasting events along the Hawaiian Archipelago, should allow for the identification of similar rift zone reorganizations. With the identification of more events, numerical modeling can be used to investigate edifice/island stability and how volcanic activity changes due to these events.
Appendix A.

New method of total magnetic surveying - the development of a micro Unmanned Arial Vehicle

Introduction

The ability to access dangerous areas safely has always been a goal particularly when referring to volcanology. The hazards in volcanic terrains are numerous from those associated with activity (e.g., heat, lava flows, pyroclastic flows, explosions, caustic gas) to the rugged unstable terrain itself (e.g., cliffs, unstable slopes, sharp and blocky lava flows, voids, fissures, shelly pahoehoe). Being able to obtain useful data while ensuring personal safety is extremely important. One possibility to reduce the exposure to risk in rough hazardous terrain is to use drone technology. Here we present the trials and errors in developing a micro unmanned aerial vehicle (UAV) to conduct magnetic surveys.

Other UAV applications

There has been a group of dedicated individuals for decades that have built and flown their-own remote control (RC) aero planes. The first RC planes were built in the late 1930s, however it was only recently, ~2004, that this technology became mainstream with consumers; as RC aerial vehicles were on toy shelves everywhere. Today (summer 2015), it is even possible to buy a small RC helicopter at local convenience/gas store for about $30 CAD.

The industry which has been most affected by this technology thus far is film. With the addition of GPS, barometers, advanced stability control and gyros, it is possible to fly and keep modern high definition cameras steady. Therefore, cinematography and sequences that used to be filmed with a crew inside a helicopter are now being taken from drones. It is also possible to get vantage points that were previously unattainable due to obstacles such as buildings. They are cheaper and reduce the possible hazards and risks associated with operating and crewing a large helicopter. They have become so pervasive, it is now difficult to find a non-animated film that does not use drones in their filming.
Slowly, small UAV drones are being utilized for a variety of geologic and surveying uses; thus far LIDAR and photogrammetry are the most common. LIDAR is a remote sensing technique that uses lasers and the reflections from objects to determine distance. By mounting one to a UAV with dual channel GPS capabilities, it is possible to map elevations of the ground in wide swaths in a short amount of time. Photogrammetry uses precise GPS location with a calibrated SLR camera to take photos at many different locations/angles around a subject (rock face, gravel pile, etc.) which are used to build a 3D model. Both of these techniques can be used to investigate a large number of problems including, but not limited to, determining volumes of rock/gravel piles, slope stability, engineering, forestry to estimate canopy height, and agriculture to look at maximum sun exposure due to small changes in topography.

Within the volcanology community, there have already been some inroads to using this technology. Japanese volcanologists have been using large gas powered drone helicopters to deploy and retrieve equipment in dangerous areas for several years. Other researchers have equipped gas sensors and flown them into plumes from volcanoes to measure SO₂ [Corrales et al., 2012]. FLIR have also been mounted below multirotor UAVs to investigate changing temperature with respect to volcanic activity [Amici et al., 2013].

Currently, it is the dedicated hobbyists that are constantly pushing the bounds of what these machines can do and improving upon them. Open source software and CPU modules have been developed by the hobbyist community allowing for more sophisticated flight control and autonomous flight (Arduino). While there are commercial builds and software that have more support and perhaps ease of use, the open source option has just as many abilities with the option of programing or modifying the flight control programs. This project utilizes the work done by this community to create a powerful yet inexpensive drone that can in theory fly geophysical surveys.

**The Fluxgate Magnetometer**

Unfortunately, geophysical equipment for the most part has not kept up with the miniaturization of technology and therefore most geophysical surveys cannot be conducted via small drones. One piece of equipment that has gone through significant technological advancement is the fluxgate magnetometer. The original flux gate was
developed in the 1930s and during the Second World War, and was used to detect submarines [Fig. A-1; Primdahl, 1979; Ripka, 1992]. Now virtually every new technological device or new car utilizes a small fluxgate to determine orientation and heading by measuring the Earth’s magnetic field.

Figure A-1 Fluxgate magnetometer that was used during the Second World War to detect submarines [http://geomag.nrcan.gc.ca/lab/vm/fluxgate-en.php?pedisable=true].

While fluxgates can be built directly into a circuit board (Fig. A-2), to be able to measure the magnitude of Earth’s magnetic field at an accuracy suitable for geologic investigation requires 3 sensors, one for each component of the field, which are slightly larger than those built into circuit boards [Primdahl, 1979; Ripka, 1992].

Figure A-2 Example of a small commercial single axis fluxgate magnetometer used in electronic devices [S. Zurek, Encyclopedia Magnetica, CC-BY-3.0]

While the size of an individual sensor can be drastically different, they all use the same principles. There are two different styles (rod, Fig. A-2; toroidal, Fig. A-3) but both have a magnetically permeable, ferrous solid core wrapped using two different wires, drive
winding and sense winding (Fig A-3). The drive winding runs a known frequency/wave which cyclically saturates and unsaturates the magnetically permeable core. The constantly changing magnetic field generates an induced current sense wire which is measured by a detector. If there is no external field the induced current will have the same phase as the input signal. When in presence of an external magnetic field the phase will be different as the core will saturate easier in the same direction of the external field and resist saturating in the opposite direction. With careful tuning and calibrating the amplitude and phase, direct measurements of the strength of an external magnetic field can be done [Primdahl, 1979; Ripka, 1992].

Fluxgate magnetometers meet all the requirements for small drone surveys as they are small, lightweight and do not have unreasonable power consumption. They also have the added benefit of being able to measure each component of the magnetic field instead of just the total field like proton precession magnetometers. However, sensor orientation is important if each component is to be measured and analyzed. There are a handful of companies that currently

![Figure A-3 Schematic of a toroidal fluxgate magnetometer with the core as green and blue. Arrows show direction of the magnetic fields external and within the core.](image)

produce fluxgate magnetometers which are used for geologic investigation. Our UAV prototype uses a SENSYS FGM3D three-axis sensor capable of measuring +/- 100,000 nT, weighting just 115 g, and requiring +/- 15 v supply voltage (Fig. A-4).
Figure A-4 SENSYS FGM3D three-axis sensor Fluxgate magnetometer ( ~10 cm long by 2 cm tall by 2 cm wide and weight of 115 g) used to build UAV prototype.

The Build and components

There are many of different build possibilities for a complete system (RC plane or multirotor helicopter) directly from hobby sites. For this project, we chose not to use a RC plane, even though they have longer flight times, as they need a smooth flat area for landing and takeoff which is not a luxury we typically have at volcanic field sites. Instead we choose to build a multirotor helicopter due to its maneuverability, stability and ease of landing and takeoff. In general, multirotor helicopters come in a variety of configurations
Figure A-5 Common multirotor frame setups involving 4 (quadcopter), 6 (hexacopter) and 8 (Octocopter) motors.

using 3 to 8 motors. Figure A-5 schematically depicts the most common set-ups with the most common being the Quadcopter X using 4 motors.
There are pluses and minuses to each rotor configuration in Figure A-5. With more rotors one can use smaller blades and still achieve the same amount of lift. Smaller blades spinning faster also increase flight stability; however, with more rotors and smaller blades, efficiency is decreased and flight times are reduced. Whether the frame is laid out in the X or + format depends on personal preference or the payload the UAV is carrying. For example, for filming the “x” configuration is preferred as it provides the widest uninterrupted field of view. This projects main goal is to develop a multirotor UAV for magnetic surveying, however its secondary one is to provide a UAV platform which can be easily changed and adapted to other scientific purposes. With this in mind we choose to use a 3DR Hexacopter kit bought directly online (Fig. A-6).

Figure A-6 How the original 3DR ArduCopter hexacopter looked before modifications were made. Blue arms indicate the “front” of the helicopter.

Electronics, batteries and CPU

Figure A-6 shows the completed hexacopter unit, however, it does not show the electronics and sensors as they are mostly covered by plastic at the centre the UAV. The
most important part of the electronics is the Arduino board or ArduPilotMega (APM; Fig. A-7). This controls all flight actions and includes 3 axis gyros, barometer and accelerometer sensors. It also has ports for external GPS, telemetry, and other Arduino compatible sensors. The pins on one end of the APM are used to direct the motors and the other end to receive or send signals using the radio. The APM is also the reason that the system (with a GPS) can be autonomous. There are other flight control boards available such as like the newer Pixhawk which have slightly different functionality and allow for modification in different ways; however fundamentally, both perform the same for the average user.

![Image of hexacopter components]

**Figure A-7** Top of the hexacopter and the different electronic components. The Power distribution board in the lowest enclosed area with the APM directly atop it and beneath the GPS and Radio.

The power for the whole unit comes from one (although you could use two) Lithium-ion polymer (Lipo) battery. The power is distributed to each motor via the power distribution board and to the other components via the APM. Motors and electronic speed controllers (ESCs discussed further in the next section) are rated to a specific amperage and/or voltage, so it is important to pick the right battery. However, even with rating as a guide, there are a wide range of batteries to choose from as well from voltage (number of cells),
amperage (capacity), max burst discharge and max continuous discharge (designated with just a C). Our off the shelf, starter kit can handle 2 to 4 lipo batteries (11.4 V - 14.7 V) and the ESCs are rated to 30A.

**Motors, ESCs and props**

The motors that came with our starter kit are D2836 Out Runner Brushless motors with 880Kv, where Kv stands for the motor velocity constant measured in revolutions per minute (RPM) per volt. The Kv rating of a brushless motor is the ratio of the RPM to the maximum voltage of the back EMF (the voltage on the wires connected to the motor coil) when not under load. The higher the Kv, the faster the motor is able to spin for the same input voltage and shorter the propellers that are used. Motors that are designed to work with and use larger propellers have lower Kv values, use bigger batteries, generate more lift and spin slower. For instance, multirotor motors with ~100 Kv are typically designed to work with propellers ~ 20 inches in length (~51 cm) and batteries with between 8 and 12 cells (29.6 to 44.4 V). For multirotor, it is also a necessity that the motors can spin in either direction (clockwise or counter clockwise) to be able to do more than just move up and down. The direction of spin is controlled by the ESCs.

ESCs or electronic speed controllers regulate the power from the battery to the motor controlling the speed based on the messages from the APM. They have an amperage rating in our case of 30 A. Even though the motors for our build work with either 3 or 4 cell Lipos, it is not possible to use the majority of 4 cell batteries and even some 3 cell due to the amperage limitation. As described above, batteries have maximum burst amperage that can be drawn. In the case of a 4 cell 5000 mAh 45C battery the maximum burst amperage is 450A and a continuous rate of 225 A. At high throttle speeds (typical of takeoff) each ESC will be flooded with greater than 37.5 A and up to, but less than, 75 A. The ESCs are only rated to 30 A and therefore when we first attempted to fly the helicopter with larger batteries the ESCs failed and caused the unit to crash.

As mentioned above, motors are built with specific propeller (also referred to as props or blades) sizes that are in line with the Kv of the motor. In general the bigger the propeller, the greater the lift; however, using the wrong propellers with the wrong motor can cause loss in efficiency and create more shaking of the craft in flight. There is also an
increase to personal risk when using larger blades as it is easier to come into contact with them causing serious injury. The propellers that came with the starter kit are 10x4.5. The first number is the length, 10 inches, and the second, 4.5 inches, is the pitch of the blade. A higher blade pitch reduces lift or efficiency, however, a lower pitch will reduce top speed and maneuverability. A pitch of 4.5 is the standard for multirotor builds of this size. In our build the motor the manufacturer states that the best prop when using a 3 cell Lipo is 10 inches long and 9 inches when using a 4 cell Lipo.

**Radio, controller and telemetry**

The ability to send commands and control the UAV is done through a radio transmitter and receiver. Our original kit came with a 6 channel Spektrum DX6i which can handle manual control of the helicopter. However, 6 channels, due to the lack of 3 position switches on the controller, were not sufficient to use the full functionality of the craft. The APM allows for a number of flight modes, the most often used are stabilize (normal manual operating mode), auto (for autonomous flight), loiter (hover in one spot), land, alt hold (hold elevation but allow horizontal movement), and RTL (return-to-launch). Since the APM is compatible with a wide variety of radio/control systems, we installed a popular radio system called FrSky 2.4GHz ACCST Taranis X9D Digital Telemetry Radio System. With up to 16 channels and capable of telemetering data, including video, back to the launch site this is a very powerful controller and gave us the full range of inflight modes and flexibility.

To reduce power consumption and weight, we did not use telemetry but it can provide information including video of the flight in real time. This can be useful when pushing to the maximum flight time, allowing the operator to monitor battery levels or ensure the field of view is correct for the best photo. Given time, a working prototype and more funding, we may have needed to use simple telemetry to monitor battery levels and survey speed.

**Magnetometer power supply and Data Logger**

In order to power and record the data from the magnetometer, we needed to either buy an expensive and heavy data-logger/power system from the manufacturer or design our own system. To start with, we ordered a GigaLog S which has a ADC resolution of 24
bit, 0 to 1.2 V input and can record 16 channels simultaneously with the ability to write text files to a SD memory card. First, all non-essential parts were stripped off the data logger leaving a circuit board, SD card slot, and USB mount for sending instructions to the board via computer. Power was wired in using a single 9 V battery that can keep it operational for at least 6 hours. To power the fluxgate magnetometer, four 9 V batteries were used such that 2 in parallel would deliver +18 V and the other two -18 V. The voltage from the batteries was stepped down to provide the magnetometer the +/- 15 V input it requires. The three outputs (X, Y, and Z) from the magnetometer are +/- 10 V (where every volt represents 10,000 nT) so before reaching the data-logger, the voltage is stepped down and converted to 0 to 1.2 volts. The calibration of this process as well as the finished product is displayed in Figure A-8.

The motors and electronics of the multirotor UAV represent a large magnetic noise source. To reduce the effect the fluxgate magnetometer must be mounted a minimum distance of 2 m from the motors. A closer mount would require filters that can remove the noise which also requires characterizing the nature of the noise (frequency and amplitude) which will likely change during flight due to changes in throttle and therefore, motor rotation. To start, we tried “long lining” the magnetometer using lightweight and strong twine at 2.5 m below the UAV. With this setup any non-smooth movements of UAV would cause the long line to sway. The onboard software would attempt to correct for this sway, causing the fluxgate to swing in the other direction. The swaying then intensified until the unit crashed. In an attempt to eliminate this problem, we mounted a lightweight extendable camera tripod leg underneath the UAV. This provided a ridged place to attach the fluxgate to reduce inflight sway. While this reduced the amount of sway, it did not eliminate it and the unit continued to crash.
Figure A-8 Three graphs show the calibrations for each component of the field as outputs from the fluxgate magnetometer. The image on the bottom right is the data-logger and power box for the magnetometer (built by the SFU electronics shop).

Issues and limited data

As discussed above, we had problems keeping the craft in the air when under load in autonomous mode. There seems to be an issue with where the unit thinks its center of mass is. I believe that if the mass below begins to wobble or sway the onboard software over compensates and begins an ever increasing pendulum effect until the craft crashes. Numerous attempts have been made to error trap but at this time all attempts have found nothing leaving us to point to the autonomous flight algorithms.

On one very still, clear winter day, we were able to get one profile of ~150 m before the pendulum effect brought down the UAV. The profile was flown in pasture in Langely
British Columbia away from large sources of magnetic noise. What is special about this pasture is that the Kinder Morgan pipeline runs directly across the field near its northern end. The goal was to survey the field with the UAV system then ground survey using the proton precession magnetometer and compare the results. The short profile shows a relatively stable magnetic field until just before the UAV came down. It is possible that the large magnetic swing at the end of the profile is due to the buried pipeline (Fig. A-9); however, the UAV was moving side to side and the sensor may be recording motor noise due to swinging too close to the motor. Unfortunately, the amount of data is just too little to interpret.

![Magnetic profile flown over a pasture in Langely, British Columbia, Canada, above the Kinder Morgan pipe line to test how well the UAV system surveys. The sinusoidal nature of the profile could be due to the UAV’s pendulum motion.](image)

**Figure A-9** Magnetic profile flown over a pasture in Langely, British Columbia, Canada, above the Kinder Morgan pipe line to test how well the UAV system surveys. The sinusoidal nature of the profile could be due to the UAV’s pendulum motion.
Next steps

The next steps have already started, however by other people. An engineering group at California Polytechnic State University have adopted the project to work out the issue with having a moving center of mass. This process has shown that this type of technology has significant promise for the use in surveying.

References


