Episode 4 of the 1783-84 Laki eruption: conduit processes and eruption dynamics of phreatomagmatic phase 1

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Episode 4 of the 1783-84 Laki eruption: conduit processes and eruption dynamics of phreatomagmatic phase

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Abstract

The Laki 1783-4 fissure eruption was a devastating time for Iceland. It featured 10 separate eruption episodes and two of them were phreatomagmatic. The phreatomagmatic (P1) eruption from fissure four produced tuff cone 1 on the SW section of the Laki cone row. Samples collected from the NW and SE side of the cone display a rich abundance of ash and gold pumice and it produced mean vesiculation values between 84-88% with some values as high as 97%. This very high vesiculation mimic values seen in the magmatic phases of the eruption, meaning that the magma, during the P1 eruption, was able to reside near the surface and fragment fully to magmatic foam before external water quenched the already vesiculated clasts. External water had no effect on the explosivity of this eruption phase and its main role was to quench the magma and granulate the fragile clasts. Fuel-coolant interaction processes cannot occur at such shallow conditions because the fragmented foam has a high yield strength making this kind of environment unsuitable to host such an interaction. Grain size data lead to the conclusion that the P1 eruption began on June 27th 1783 where a westerly wind deposited samples on the SE side of the cone and then an easterly wind on June 30th deposited the rest of the samples on the NW side. The SE side of the tuff cone has higher ash (~60-90%) content and golden pumice concentration. This increased ash content is due to cooling-contraction granulation of the vesiculated clasts and melt leading to an additional fragmentation. The availability of external water dwindled leading to the deposit of coarser-grained samples on the NW side of the tuff cone. High effusive rate ($10^7$ kg/s) and high bubble number densities ($10^7$ cm$^3$) consisting of a rich population of small bubbles and bimodal tendencies indicate increased magma ascent rate due to early degassing of CO$_2$ and burst nucleation of H$_2$O, meaning Laki was most likely of subPlinian-Plinian intensity.
Útdráttur

Goshrinur Skaftárelda 1783-84 voru 10 talsins, þar af tvær freatómagmatískar. Freatómagmatíski fasinn sem gaus á sprungu fjögur á Lakagíum og er kjarninn í þessari rannsókn, myndaði hverfjallið vestan Laka ásamt gjóskufallseiningunni P1. P1 er öskurík eining og innihaldur verulegt magma af basískum vikri (golden pumice). Meðal blöðrumang vikurkornanna er 84-88%, en í stökum kornum nær blöðrumagnið 97%. Þessi gildi endurspeglan blöðrumagnið í magmatíski vikrinum frá Laka. Þessar niðurstöður benda til þess að kvikan freatómagmatísku hrinunni á sprungu 4 var full afgösuð og fullþanin, þ.e. var orðin froða, og farin að sundrast í korn af vikurstærð áður en hún komst í snertingu við utanaðkomandi vatn. Hér má draga þá ályktun að megin þáttur utanaðkomandi vatns í goshrinunní var hraðkæling á froðukendum vikurkornum sem varð til þess að þau molnuðu í finan óskusalla. Há kvikuframleiðni (10⁷ kg/s) og hár blöðrúpetleiki (10⁷ cm³) benda til þess að aflað í gosinu hafi verið mikið og keyrt áfram af hvelllosun á H₂O gasi frá kvikunni. Þessi gögna við þá ályktun að sprengfasarnir í Laka hafi verið lá-plínísfir (subPlinian). Athuganir á innri lagskiptingu gjóskueiningarinnar P1 ásamt kornastærðarmælingum á einstökum lögum gefur til kynna að gosin hafi frá hverfjallinu vestan Laka barst fyrst til austurs og síðan í vestur. Þessar niðurstöður ásamt upplýsingum um vindáttir í samtímaheimildum frá þeim tíma sem fjórða goshrina Skaftárelda var í gangi, benda til þess að sprengfasáinn P1 hafi byrjað 27 júní, 1783, en ekki 25 júní eins og lagt hefur verið til í fyrri athugunum.
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Abbreviations

2D: Two-dimensional
3D: Three-dimensional
A\textsubscript{p}: Total image area in pixels (pixels\(^2\))
AR: Axial rift
A\textsubscript{R}: Reference area (mm\(^2\))
A\textsubscript{RM}: Reference area for the magnification (mm\(^2\))
A\textsubscript{RMT}: Total area from the images in the reference magnification (mm\(^2\))
asl: Above sea level
A\textsubscript{T}: Total image area in mm (mm\(^2\))
atm: atmospheric (pressure)
A\textsubscript{V}: Area of the vesicle (mm\(^2\))
CF: Collapsing foam model
CVSD: Cumulative vesicle size distribution
CVVD: Cumulative vesicle volume distribution
D: Diameter of a particle (mm)
E: East
EqD: Equivalent diameter of a vesicle (mm)
EVZ: East Volcanic Zone
EVZ: Eastern volcanic zone
f\textsubscript{bin}: Frequency of vesicles in the bin (unitless)
FCI: Fuel coolant interactions
F\textsubscript{V}: Volume fraction
g: Gravitational force constant (9.81m/s\(^2\))
G\textsubscript{e}: Edge-vesicles mean greyscale value (unitless)
GIMP: GNU Image Manipulation Program
G\textsubscript{p}: Phenocryst greyscale value (unitless)
G\textsubscript{v}: Vesicles mean greyscale value for all vesicles in image (unitless)
H\textsubscript{bar}: Mean projected height by a spinning particle about a randomly oriented axis (mm)
KR: Kolbeinsey Ridge
m: Mass (g)
MAR: Mid-Atlantic Ridge
MFCI: Molten fuel coolant interactions
MFCI: Molten-fuel-coolant interaction
MIB: Mid-Iceland Belt
MIB: Mid-Icelandic belt
$m$: mass of a solid (g)
Mt: Megatons
N: North
NA: Number density of vesicles (mm\(^{-2}\))
$N_A$: Bulk number of vesicles for a unit area (mm\(^{-2}\))
$N_{A_{bin}}$: Number of vesicles in the bin per unit area (mm\(^{-2}\))
$N_A$: Number of vesicles per unit area adjusted for phenocrysts (mm\(^{-2}\))
$N_A$: Total vesicles per area of the clast (mm\(^{-2}\))
n\(_f\): Number of pieces of wax film used on clast
NV: Number density of vesicles per unit volume (mm\(^{-3}\))
$NV_M$: Number density of vesicles per unit volume adjusted for melt (mm\(^{-3}\))
NVZ: Neo volcanic zone or Northern volcanic zone
NVZ: North Volcanic Zone
NW: Northwest
ÖVB: Öræfi Volcanic Belt
P1: Phreatomagmatic phase 1 from fissure 4
$P_f$: phenocryst fraction (unitless)
RR: Reykjanes Ridge
RSD: Rise speed dependent model
RVB: Reykjanes Volcanic Belt
RVZ: Reykjanes Volcanic Zone
S1: Magmatic eruption 1 from fissure 1
S2: Magmatic eruption 2 from fissure 2
S3: Magmatic eruption 3 from fissure 3
SE: Southeast
SEM: Scanning Electron Microscope
SF: Scale factor (pixels/mm)
SG\textsubscript{c}: Specific gravity of the clast
SISZ: South Iceland Seismic Zone
SVB: Snæfellnes Volcanic Belt
SW: Southwest
TFZ: Tjörnes Fracture Zone
TC1: Tuff cone 1
v: Terminal velocity (mm/s\textsuperscript{2})
V: Vesicularity (%)
V: Volume (ml)
V\textsubscript{f}: Vesicle fraction (unitless)
V\textsubscript{ges}: Volume of pycnometer (ml)
V\textsubscript{I}: Vesicles included in reference area
VND: Vesicle number density
V\textsubscript{s}: Volume of sphere with equivalent diameter of Hbar (mm\textsuperscript{3})
VSD: Vesicle size distribution
V\textsubscript{v}: Percentage of vesicles by volume (%)
VVD: Vesicle volume distribution
W: West
w\textsubscript{a}: Weight of clast in air (g)
w\textsubscript{b}: Weight of ballast (g)
w\textsubscript{f}: Weight of wax film (g)
WVZ: West Volcanic Zone
w\textsubscript{w}: Weight of clast in water on ballast (g)
x: Class size number
\alpha: Alpha values (unitless)
\eta: Viscosity
\rho: Density (g/ml)
\rho\textsubscript{c}: Density of the clast (g/cm\textsuperscript{3})
\rho\textsubscript{DRE}: Density of DRE (g/cm\textsuperscript{3})
\rho\textsubscript{w}: Density of water (g/cm\textsuperscript{3})
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1 Introduction

Volcanoes are as dangerous as they are beautiful, but with beauty comes mystery. Volcanology, as a detailed science, has been around for a very short time and has only become popular in the last few decades. There are still many debates as to how volcanoes behave and why. One of those debates being whether or not phreatomagmatic eruptions are unanimously more powerful (=explosive) than magmatic eruptions. There is no doubt that explosive eruptions are more powerful than effusive eruptions (Carey 2005). However, phreatomagmatic and magmatic eruptions have displayed a wide array of intensities and therefore, are still shrouded in mystery as to what role magmatic gases and external water play in regards to eruption intensity (Houghton & Wilson 1989; Costantini et al. 2010; Moitra et al. 2013; Houghton & Schmincke 1986). It is challenging to make an inter-comparison of eruptions that span a range of magma discharge, volatile concentration and/or melt composition, but this is often what we have to work with. The role that external water plays when interacting with magma at various depths in phreatomagmatic eruptions is still debated. Are these eruptions explosive due to fuel coolant interactions (FCI) where steam expands and builds up pressure at the magma-water interface (e.g. Wohletz 1986; Zimanowski et al. 1997; Ort & Carrasco-Núñez 2009) or if external water plays a minor role in controlling the explosive behaviour of these events (e.g. Kokelaar 1986; Houghton et al. 1999; Houghton & Schmincke 1986). There is no doubt that FCI processes are occurring when external water interacts with magma at depth, however, this does not seem to be the case when magma is near the surface. Fortunately, the 1783-84 Laki fissure eruption consisted of magmatic and phreatomagmatic phases where the only stark difference between these phases was the involvement of external water into the later, making this eruption a viable example to show the effects that magma-water interactions have.

The Laki fissure eruption, which erupted from a 27-km-long vent system within the Eastern Volcanic Zone in South Iceland (Figure 1.1), lasted from 8th June 1783 – 7th February 1784. It featured 10 distinct eruption episodes each beginning with a short lived
(0.5 to 4 days long) violent Magmatic or phreatomagmatic phase followed by a longer lasting (days to weeks) effusive phase (Figure 1.2; Thordarson et al. 2003; Thordarson & Self 1993; Thordarson & Self 2003). Here I present a systematic study of the tephra deposit produced by the phreatomagmatic phase (P1) from episode 4, including its grain size, density and vesicularity, in order to assess the processes occurring within the magma conduit prior and during this phreatomagmatic activity. The objective of this study is to answer the research questions listed below.

![Figure 1.1: Laki eruption site and lava flows. The Laki cone row is indicated by the red dashed line and the lava flows stretching around the Síða highlands is shade in grey. The black lines outline the Grímsvötn-Laki volcanic system. The figure shows the geographic division of the Fire districts; Álftaver, Skaftártunga, Medalland, Landbrot, Síða, Síða highlands and Fljótshverfi. (Source: Thordarson and Self 1993).](image)

As indicated by Thordarson and Self (1993) the phreatomagmatic tephra from episodes 4 and 6 of the Laki eruption have significantly higher proportions of ash than their magmatic counterpart from the violent Magmatic phases of episodes 1-3, 5 and 7-10. This evidence indicates significantly higher degree of magma fragmentation during the phreatomagmatic phases.
Figure 1.2: Laki eruption timeline displaying the 10 separate eruptions labeled I-X. The start and end of the eruption is indicated by the black arrows. Each eruption event began with earthquake swarms and a lava surge flowed from each fissure. (Source: Thordarson & Self 2003)

Figure 1.3: Sulphur content vs vesicularity of Laki samples. Phreatomagmatic tephra is indicated by open diamonds and magmatic or Magmatic tephra is indicated by the black triangles. The cross and 2σ in the upper right corner displays the estimated analytical error. (Source: Thordarson et al. 1996)

Moreover, Thordarson et al. (1996) show that the most vesicular clasts in the phreatomagmatic tephra from Laki are compatible to the clasts in the magmatic tephra, implying that magma vesiculation was well underway and that portion of the magma was
fully vesiculated when it came into contact with external water (Figure 1.3). It has also been shown that there is a linear relationship between clast vesicularity and the sulphur content in the groundmass glass, indicating premature arresting of degassing due to quenching by interaction with external water (Metrich et al. 1991; Thordarson et al. 1996). Finally, although non- to poorly vesicular phreatomagmatic clasts from Laki have relatively high sulphur contents (1000 – 1300 ppm S), these values are 20-40% lower than the recorded melt inclusion (= initial sulphur content of 1810±100 ppm) sulphur values of the magma, suggesting that a significant portion of the magmatic gases escaped from the magma without leaving a mark in the form of vesicles (Thordarson et al. 1996). This evidence raises some interesting questions about the dynamics of the phreatomagmatic activity at Laki:

- Is the higher ash grade a reflection of more intense and powerful explosive activity during the phreatomagmatic phases?
- How significant a role does explosive magma-water interaction play in driving the intensity of the eruption?
- How much of the phreatomagmatic tephra was highly vesiculated at the time of magma to water interaction and what does that indicate for the level/depth within the conduit for the interaction?
- What principal mechanism governs magma vesiculation in the phreatomagmatic phases at Laki?

Before these questions are answered, this study will give an overview of the volcanic eruption history of Iceland (Chapter 2) and type of volcanic eruptions that typify the activity (Chapter 3). I then delve into the Grímsvötn volcanic system and in particular the course of events and the nature of the 1783-4 Laki event (Chapter 4). As well as, the methods used in this study and the theory behind using such methods (Chapter 5). Finally, the results of this study will be presented in Chapter 6 and discussed in Chapter 7. Concluding remarks (Chapter 8) tie everything together in a short and concise overview.
2 Geology of Iceland

2.1 Origin of Iceland

Iceland is located on the northern section of the Mid Atlantic Ridge (MAR) between Greenland and Northern Scandinavia and above a magmatic plume. The MAR is a divergent plate boundary that divides the Atlantic Ocean and is continuously producing new oceanic crust. Iceland is a product of the MAR and the mantle plume beneath it (Bjarnason 2008; Einarsson 2008). Hot spots (or plumes) are areas of anomalous volcanism that span millions to tens of millions of years. They often leave their mark on the associated tectonic plate, commonly as a trail of erupted material indicating their past locations. The Iceland-Greenland and Iceland-Faroe Ridges have been interpreted to be such markings that were formed in the period 54 to 25 Ma (Figure 2.1).

The Iceland mantle plume is thought to extend to at least 450 km depth in the mantle and at present, its centre is placed beneath Central Iceland at about 64 40’N and 18 10’W (Figure 2.1). The Iceland Basalt Plateau resides at an elevation of more than 3000 m asl (Thordarson & Hoskuldsson 2008) and has an unusually thick crust ranging from 19 km in West and West-Central Iceland to about 29-32 km in East and East-Central Iceland due to combined magma production from the MAR and mantle plume (Bjarnason & Schmeling 2009). Oceanic crust outside Iceland is, on average, about 7 km thick (Bjarnason 2008).

Using the hot spot trails and oceanic magnetic anomalies, the plume is thought to be as old as the opening of the Atlantic Ocean (~56 Ma) (Bjarnason 2008). The mantle plume is a narrow vertical flow of hot partially melted mantle material from the lower mantle. When the plume reaches the upper mantle then it spreads laterally due to the density change (Figure 2.2). We can detect mantle plumes using p and s seismic wave studies (Larsen & Eiriksson 2008; Einarsson 2008). As the waves move through the plume then they slow down due to the high temperature of the plume rocks and partially melted material. This area is called the low velocity zone (Bjarnason & Schmeling 2009).
Figure 2.1: Bathymetry map showing the mid-ocean ridge and the Icelandic hotspot. White shows areas above sea level, shallow oceans are shown by warm colours while deep oceans are indicated by dark colours. The Iceland basalt plateau is indicated. The red lines trace the hotspot trails connecting Iceland to Greenland (on the left) and the Faroe Islands (on the right). The red dot indicates the present day location of the hotspot and the grey dot is its position 40Ma. The black arrows show absolute plate motion of the North American (26 mm/yr) and European (15mm/yr) plate. (Modified from Bjarnason 2008).
Figure 2.2: Structure of the crust using seismic and dynamic modeling. a) shows the relative thickness of lithosphere and crust. b) shows the image of the plume, lithosphere and crust. The lithosphere has variable thickness (19 – 100 km) and is thickest in the Eastfjords and East Iceland Shelf. The upper part of the lithosphere is the rigid crust and the lower part is the cooled upper mantle. The plume rises up from the lower mantle and spreads at the base of the upper mantle. The low velocity anomaly detected using seismic measurements resides at 100-125 km below the surface. (Source: Bjarnason & Schmeling 2009).

2.2 Iceland Volcanology

Iceland is actively spreading apart and experiences an eruption every 3-5 years on average. The axial rift across Iceland and the plume beneath it interact and create areas of faulting, volcanism and deformation (Thordarson & Hoskuldsson 2008). The main geological formations and the neovolcanic zones in Iceland are shown in figure 2.3. The axial rift extends from the Reykjanes Peninsula in the SW to Axarfjörður in the NE and has a width ranging from 15-50 km. It is the continuation of the MAR across Iceland and the zone of
active spreading, comprising the main volcanic zones in Iceland: the West Volcanic Zone (WVZ) and North Volcanic Zone (NVZ). As on the MAR, the magmas of the axial rift are tholeiitic. The West and North Volcanic zones are connected by the Mid-Icelandic Belt and are linked to the MAR via the Reykjanes Volcanic Belt (RVB) in the southwest and the Tjörnes Fracture Zone (TFZ) in the northeast (Figure 2.3). The East Volcanic Zone (EVZ) is a relatively young (<3 myr old) and is a propagating to the southwest through older crust. The southern part of the EVZ is more alkali and the northern part is tholeiitic (Thordarson & Larsen 2007). At present, it is the most volcanically active region in Iceland that hosts the most productive volcanic systems in Iceland, namely the Hekla, Katla, Bárðabunga-Veiðivötn and Grímsvötn systems (Figure 2.3). The 1783-84 Laki eruption took place on the Grímsvötn volcanic system (Thordarson & Self 1993).

Figure 2.3: Geological map of Iceland showing the neovolcanic zones and the 30 active volcanic systems (see key for details). The large black dashed-line circle shows the apparent location of the mantle plume beneath Iceland. The abbreviations are as follows: RR – Reykjanes Ridge, RVB – Reykjanes Volcanic Belt, SVB – Snæfellsnes Volcanic Belt, WVZ – Western Volcanic Zone, MIB – Mid-Iceland Belt, SISZ – South Iceland Seismic Zone, EVZ – Eastern Volcanic Zone, OVB – Óræfi Volcanic Belt, NVZ – Northern Volcanic Zone, TFZ – Tjörnes Fracture Zone, and KR – Kolbeinsey Ridge. (From Thordarson & Hoskuldsson 2008).
The neovolcanic zones contain 30 identified volcanic systems and regions of active volcanism cover about 30,000 km² (Figure 2.3). On the axial rift, the plate motion is evident by discrete events, commonly referred to in Iceland as “eldar” (= Fires), typically confined to a single volcanic system and can last for years to decades. These events are associated with continuous ground deformation with alternating inflation and deflation episodes as well as periodic earthquake swarms and volcanic eruptions (Thordarson & Larsen 2007). The 1783-84 Laki eruption along with the 1783-85 eruption sequence at the Grímsvötn central volcano are a good example of such Fires (e.g. Thordarson & Self 1993). So are the 1975-84 Karfla Fires on the Krafla volcanic system in North Iceland (e.g. Björnsson 1985; Bragðsdóttir & Einarsson 1979).
3 Volcanoes

3.1 Volcanic Areas

It is impossible to know exactly how many active and inactive volcanoes there are on Earth. Finding all volcanoes on land is a heavy-handed task, but most of the volcanoes are said to reside on the sea floor (Carey 2005). Since 70% of the Earth is covered by water then we have very little information on just how many volcanoes inhabit the oceans. Carey (2005) describes that most of the volcanism on Earth occurs on the sea floor and more specifically along “belts or linear segments”, such as, along plate boundaries. The main area of volcanic activity and where, historically, the most severe and catastrophic eruptions have occurred is along the Pacific Ring of Fire (Figure 3.1). The Pacific Ring of Fire coincidentally consists of subduction zone boundaries. Subduction zones have been known to produce extremely violent eruptions, such as: Pinatubo, Toba, and Mount Churchill (Carey 2005).

![Figure 3.1: The Pacific Ring of Fire is the most active volcanic region on Earth. The sun icon indicates major active volcanoes. Most volcanic activity is located on the sea floor. (Source: Encyclopædia Britannica).](image-url)
More than 80% of volcanic output on Earth is basaltic (Parfitt 2004). Magmas will vary in composition based on the source rock melt composition, volatile concentration, whether the magma crystallizes during ascent or storage, magma mixing and/or contamination from the country rock during ascent (Carey 2005; Parfitt 2004). However, by studying the erupted products researchers can predict the processes this magma had to undertake to reach the surface.

3.2 Effusive Eruptions

Effusive eruptions are lower energy regimes and produce less violent eruptions where lava flows steadily out of the volcano. These eruptions are usually basaltic in composition and have a very low volatile concentration compared to silicic effusive and explosive eruptions (Carey 2005; Lejeune et al. 1999). Most basaltic eruptions, that are volatile poor, are effusive when they erupt because of their low viscosity inducing little resistance for magma flow (Alidibirov & Dingwell 1996). For an eruption to be effusive the magma needs to have a high enough permeability such that the gas is able to escape easily before the eruption (Carey 2005). The magma will then be able to flow out effortlessly after pre-eruptive degassing or outgassing. If the gases have difficulty leaving the magma then pressure builds up and the eruption becomes explosive (Kaminski & Jaupart 1997; Lejeune et al. 1999).

The magma, in effusive basaltic eruptions, maintains an equilibrium assemblage of melt and bubbles and is classified as a bubble suspension, which has typical vesicularities of 60-70%, whereas, an explosive eruption will occupy vesicular values of 75-98% (Mangan & Cashman 1996; Sparks 1978). Bubble suspension means that the magma volume has less than 74% of its gas phase in the form of bubbles (Mangan & Cashman 1996). Non-explosive eruptions typically have high permeability with increasing vesiculation, which allows for efficient open system degassing through bubble chains. Bubbles that are linked together in highly viscose magmas through the process of coalescence can induce efficient outgassing (Okumura et al. 2009). Effusive magma products tend to have a significantly lower vesicle number density ($N_{VM}$) compared to explosive products, often an order of three magnitudes lower (Mangan & Cashman 1996).
3.3 Explosive Eruptions

Explosive eruptions are violent eruptions that eject tephra, that are partly or solely made of disintegrated magma (Fisher & Schmincke 1984), anywhere from several hundred meters to tens of kilometers into the atmosphere. This debris can be anywhere from micron to meter sized (Kaminski & Jaupart 1997; Larsen & Eiriksson 2008). The explosivity of volcanic eruptions is largely governed by initial volatile content, rate at which magma degasses, and the speed at which it rises to the surface. Magma decompresses during ascension to the surface, which increases buoyancy of the magma and creates a driving force for further ascent (Lloyd 2014). Explosive magma are often volatile rich, with dissolved concentration of H$_2$O up to 4%, which leads to the rapid expansion of gases in the conduit and a large pressure build inducing a violent eruption (Figure 3.2; Carey 2005; Cashman 2004). At depth, if the magma is supersaturated in volatiles, such as water or carbon dioxide, then they will begin to degas from the magma to form bubbles (e.g. Lensky et al. 2004). At such a depth, bubble volume is reliant on bubble nucleation and gas element diffusion because the pressure exerted on each bubble is too intense to allow for bubble expansion so they must rise to lower pressures (e.g. Alidibirov & Dingwell 1996). The only way for bubbles to rise fast enough to lower pressures is to combine or coalesce with other bubbles on their way up to the surface (e.g. Lensky et al. 2004). Magma ascent rate can also affect the explosivity of an eruption. Magma that rises quickly to the surface can surpass the point at which bubbles start to degas causing a burst of nucleation. This magma is now incredibly buoyant and rises quickly, with the bubbles as one system, to the surface building up tremendous pressure inducing an explosive eruption (e.g. Cashman 2004; Carey 2005).

Viscosity of the magma can affect bubble growth and rise. Higher viscosity magmas prevent bubbles from rising faster than the magma so the bubble is then “locked” in place and their rise is coupled with the magma (Carey 2005; Lejeune et al. 1999; Parfitt 2004; Lensky et al. 2004). When water is separated from the melt as bubbles then the viscosity of the melt can increase significantly, even several orders of magnitude (Figure 3.3). Basaltic melts have lower viscosity, which allows the bubbles to rise faster than the
magma and, in turn, have less pressure exerted on each bubble (Kaminski & Jaupart 1997). Bubbles in basalt magmas tend to be significantly larger than in silicic magmas. These large bubbles accumulate into enormous amounts in the conduit causing magmatic foam to form at the surface (Carey 2005; Alidibirov & Dingwell 1996; Lensky et al. 2004). When the magma has transformed into a mixture of melt and mostly bubbles then the magma has fragmented. The fragmentation zone is when the magma is gas dominated and usually begins at about 1km depth in basaltic regimes (Alidibirov & Dingwell 1996). These gas-dominated regions are foam-like behaving like a solid desperately trying to keep the bubbles beneath it in the conduit as the pressure builds (Kaminski & Jaupart 1997). Explosive eruptions are induced when the foam structure collapses and the bubbles burst from the intense depressurization causing highly vesiculated pyroclastic material to eject at speeds over 400 m/s (Figure 3.2; Carey 2005).

Explosive eruptions that produce a great amount of ash tend to experience two or more fragmentation processes. The first, occurs in the conduit during rapid decompression of magma (Alidibirov & Dingwell 1996) followed by preferential (secondary) refragmentation (higher in the conduit) of larger fragments to produce ash (Girault et al. 2014). This kind of process can produce power law-like trends in the magma during degassing and fragmentation.

Parfitt (2004) explains that basaltic explosive eruptions tend to follow the rise speed dependent (RSD) model rather than the collapsing foam (CF) model. The CF model describes that bubbles will form in the magma and rise to the top of the conduit. They collect to form magmatic foam, which builds pressure and eventually collapses, due to the thin bubble walls, to form one giant air pocket. The RSD model is, by some, considered more common in natural systems (Parfitt 2004). The RSD model describes the difference between bubble rise in magmas with high and low rising speeds. When the rise speed of the magma is high then bubbles will rise with the magma as one homogenous system consisting of two phases acting as one. Bubbles will grow through diffusion and decompression. Bubbles will also nucleate during the rise eventually leading to fragmentation when the bubble volume fraction is large enough. The gas-liquid mixture will accelerate as it rises due to gas expansion increasing the buoyancy of the mixture.
Figure 3.2: Nucleation, growth and fragmentation of gas bubble population in the magma column. S: level of gas saturation where bubbles begin to evolve from the melt. N: level of bubble nucleation. E: level of bubble growth and exsolution. D: level of magma disruption and collapse prior to eruption. Stages of bubble growth are shown in figures (a) – (d). (a) Shows early stages of nucleation where bubbles grow inhibited. (b) Growth of bubbles continues and nucleation of new bubbles occurs. (c) Bubbles are closely packed together and growth ceases. (d) Bubbles build up pressure and burst when bubble walls become too thin inducing an explosive eruption. (Source: Cas & Wright 1987).

This coupled behaviour of bubbles and magma are thought to promote formation of subPlinian or Plinian eruptions. When the rise speed is low then bubbles rise quickly through the liquid and grow to large sizes due to decompression and coalescing. This causes very large bubbles to collect in the conduit or in an extreme case one large bubble. The large bubble mass will move through the conduit and collect at the surface under the cooled ‘skin’ of magma where it builds up pressure or explodes through after a certain amount of time depending on the strength of the magma ‘skin’. This process promotes pulsating strombolian-like eruptions typified by short-lived bursts and bubbles are able to
collect and repeat the explosive process many times (Parfitt 2004). Magma flow regimes are illustrated in figure 3.4.

![Figure 3.4: Magma viscosity as a function of water content. By decreasing the water content within the magma then you can increase the viscosity exponentially. (Source: Carey 2005).](image)

Magma composition and type of explosive eruption can be determined by observing the characteristics of tephra deposits. There are three main components of tephra: glass, crystals and lithics. Glass is juvenile melt, which has been super cooled by the surrounding air. The juvenile melt has expanded and fragmented to various degrees to form the glass. Crystals form from crystallization within the magma conduit or can be derived from the surrounding wall rock. Lithics are usually rock fragments eroded from the wall rock in the conduit (Larsen & Eiriksson 2008).
A special kind of glass pumice, which has been found in large quantities in this study, is called golden pumice. Golden pumice is incredibly fragile and bubble walls can easily shatter with a slight touch because they are so thin, which gives these clasts their characteristic gold colour. These clasts are the product of extreme fragment stretching during fountaining eruptions, which gives them a very thin texture that lets light pass through easily giving it a light colour. Golden pumice with high vesicularity (95-99%) is called reticulite, which is particularly frothy pumice (Stovall et al. 2012; Thordarson et al. 1996; Mangan & Cashman 1996). Reticulite is formed during fragmentation when mature bubbles are squished together forming very thin 6-sided bubble walls resembling honeycomb. The thin bubble walls are quenched extremely quickly from liquid to solid (Mangan and Cashman 1996). Stovall et al. (2012) explains that golden pumice grains or reticulite indicate times of increased eruption intensity and magma discharge. Therefore,
increased amounts of golden pumice could indicate an increase in eruption energy due to high magmatic volatile concentrations and/or magma discharge.

Houghton and Wilson (1989) present data showing that magma can reach vesiculation of 75-83% before the bubbles begin to deform and growth is constrained. This vesiculation can build up pressure causing violent eruptions. However, Mangan and Cashman put forth data illustrating that explosive eruptions can produce tephra (i.e. pumices) with higher vesicularities (75-98%) and represent true magmatic foam. This foam consists of tightly packed and sometimes deformed bubbles with liquid melt forming thin walls between adjacent bubbles. Surface tension makes free-floating bubbles spherical (i.e lowest energy state). In a monodispersed (uniformly sized) bubbles can remain spherical up to a packing equal to 74% vesicle volume and in case of polydispersed (i.e. range of sizes) systems the vesicularity can be as high as 85%. More efficient close-packing is required to reach higher vesicularity, which results in deformation of the bubbles, and consequently acquire polyhedral form (Figure 3.5; Mangan & Cashman 1996; Lensky et al. 2004; Llewellin & Manga 2005). The melt component in the foam is influenced by range of forces, which effect how it moves through the system. These forces are mainly controlled by the geometry of the gas phase (Figure 3.6). The melt flows along thin bubble walls towards the plateau borders and drains downward due to gravity rather than in multi-directions due to capillary suction (Mangan & Cashman 1996). Thinning of the bubble walls eventually leads to rupture and then the entire foam system can collapse. Bubble rupture influences the foam system in two ways. Firstly, if the melt viscosity is low then the ruptured film with retreat quickly. The plateau borders will then assimilate and cause adjacent bubbles to coalesced, which increases mean bubble size, reduces bubble number density and produces polydisperse size distribution. The second is if the viscosity of the liquid is high then the ruptured films will retract slowly and the time needed to coalesce may exceed the solidification time of the magma (Mangan & Cashman 1996). At lower pressures (i.e. near the surface) differences in pressure between bubbles of contrasting size will expel liquid melt from the bubble films and plateau borders leading to film thinning, rupture and coalescence.
Figure 3.5: Ideal 2D bubble structures. Bubbles take on a spherical shape due to surface tension but during fragmentation then bubbles deform and are packed together tightly to form more polyhedral structures. (Source: Mangan & Cashman 1996).

Figure 3.6: Mechanisms for interstitial liquid flow between bubble films and plateau borders. Arrows indicate the direction of flow. The force of gravity is usually the main driving force in the magma foam, \( \Delta P = \text{external pressure drop}, \Delta V = \text{total gas volume change}, R = \text{gas content}, T_a = \text{ambient temperature}, n_i = \text{total number of moles of the gas species}, \rho = \text{liquid density}, g = \text{gravitational constant}, \alpha = \text{angle between channel axis and vertical}, r_p = \text{typical length of a cell side}. \) Equations are shown in Mangan and Cashman 1996. (Source: Mangan & Cashman 1996).
3.4 Phreatomagmatic Eruptions

Phreatomagmatic eruptions are juvenile forming eruptions that result from the interaction of external water and magma (Thordarson & Larsen 2007). External water can be in the form of ground water, ice, snow, glaciers, lakes, steam, water saturated sediment and wall rock. These eruptions are very common in Iceland especially because of the high ground water table and many of the volcanoes reside beneath glaciers. Phreatomagmatic eruptions can be effusive, producing pillow lava basalts in subaqueous conditions, (Kokelaar & Durant 1983) or explosive, producing large concentrations of ash dominated units (Gudmundsson et al. 1997; Gudmundsson & Bjornsson 1991; Witham et al. 2007).

Characteristically, phreatomagmatic eruptions produce enormous black ash plumes that billow above the eruption site and deposit poorly sorted ash-rich units as well as lapilli-sized clasts. Large volumes of ash are able to deposit close to the eruption site due to ash aggregation. Ash aggregates form when electrostatic charges, generated during fragmentation of volcanic pumice, induce capillary forces between ash particles clumping them together into lapilli-sized clasts that are heavy enough to be deposited close to the eruption site. The forces will weaken and the aggregates disintegrate leaving large amounts of ash (James et al. 2003; Brazier et al. 1983).

Depending on the depth at which external water interacts with the magma, explosive phreatomagmatic eruptions are produced either by fuel coolant interaction (FCI or MFCI) processes (Wohletz 1986; Zimanowski et al. 1997) or by the expansion of magmatic gases (H₂O, CO₂, SO₂) in the conduit during magma ascent (Kokelaar 1986; Houghton et al. 1999; Houghton & Schmincke 1986).

There are three principle mechanisms producing explosive phreatomagmatic eruptions. The first, is the process of molten-fuel-coolant interactions (MFCI or FCI) labeled contact-surface steam explosivity in figure 3.7. Water (coolant) cools the magma (fuel) causing it to contract (roughly by 10%) and the water is heated conductively to expand rapidly (1000x volumetric expansion) and produces a thin film of steam residing on the contact surface between the magma and water. Intense thermal gradients are set up, making the thin film unstable and collapsing. When this thin film collapses then the hot magma is exposed to the cold water creating more and more vapour. The magma fragments,
however, a large part of the magma is still exposed to the cool water. This process continuously repeats itself until the collected vapour builds up enough pressure to explode. The timescale for this process is in a matter of seconds (Kokelaar 1986; Zimanowski et al. 1997; Wohletz 1983; Wohletz 1986; Büttner & Zimanowski 1998). The second mechanism is called bulk interaction steam explosivity and it is caused by the entrapment of external water within the magma or close to it. The water will expand rapidly as steam to eventually build up enough pressure to explode. This could occur if magma were to flow over a small pond or bog. The third mechanism is magmatic explosion paired with cooling contraction granulation, which is when water interacts with vesiculated clasts and melt to produce explosive eruptions (Kokelaar 1986).

![Diagram of Primary Processes](image.png)

Figure 3.7: Primary processes that generate volcano-clastic materials during magma-water interactions. These processes often occur together and potentially could enhance one another. (Source: Kokelaar 1986)

Cooling-contraction granulation is when water reaches the magma to rapidly quench it producing numerous fine cracks in the quenched magma or glass and these fractures may act as weakness causing further fragmentation when the glass disintegrates revealing the
molten substance from within. This process also produces a rich amount of fine grained when the quenched magma disintegrates and shatters. This process does not drive the eruption to become explosive because it will only induce clasts to granulate forming ash-sized particles (Kokelaar 1986; Sruoga et al. 2004). Lynch (2015) proposed that pumice clasts from the Grímsvötn 2011 eruption may have experienced this type of granulation due to late-stage quenching of already vesiculated clasts and melt. This process produced rich amounts of ash-sized particles in the deposits found at Grímsvötn.

Others researchers (Kokelaar 1986; Sruoga et al. 2004; Houghton et al. 1999; Houghton & Schmincke 1986) put forth the notion that not all phreatomagmatic eruptions are explosive and especially in magma that is fully expanded as a fragmented foam then water interaction has no effect on whether the eruption will proceed explosively or not. They advise that in these cases, explosive phreatomagmatic eruptions are induced by the expansion of gases in the conduit and not by the interaction of external water. Studies done by LaRue et al. (2013) describe that the addition of water to a highly vesiculated melt had no effect on the fragmentation process and only quenched it faster than the unaffected “dry” melt. Both melts had the same bubble number density at the end of the experiment, concluding that the main control on vesicle densities is the initial volatile content and not the addition of water.

Whether or not phreatomagmatic eruptions are unanimously more powerful/explosive than magmatic eruptions is still up for debate. Especially in Iceland, most eruptions tend to interact with external water close to or at the surface after the magma is fully fragmented and are explosive in nature. Are FCI processes occurring in environments like this or are these eruptions, with late-stage water interactions, predetermined to be explosive based on the expansion of magmatic gases? These are the kind of questions volcanologists ask themselves. In this study, we are on in favour of the later. Laki’s products have high vesicularities, which means that they were able to reside near the surface and fully fragment. We have seen in Lynch (2015) that pumice clasts from Grímsvötn 2011 eruption had very high vesicularities between 78.8-87.8% (very similar to Laki pumice from Thordarson et al. 1996 [60-96%] and our own study). Lynch (2015) concluded that the magma was able to fully vesiculate to foam before quenching occurred. Magmatic foam environments are unsuitable for FCI interactions to occur because the foam has high yield
strength and will not collapse to allow the external water to fragment the melt underneath. Therefore, in late-stage quenching situations when magma is highly vesiculated then external water will only quench the system and not drive explosive eruptions. Whether or not the eruption will be explosive is based on the expansion of gases in the system predetermined by the initial volatile content and expansion of bubbles in the conduit.

Houghton et al. (1999) studied the vesicularity and other parameters of erupted products from Crater Hill in New Zealand. The eruption varied between phreatomagmatic to magmatic phases and sometimes a mix. Phreatomagmatic clasts had lower vesiculation (20-50%) compared to dry eruptions with higher vesiculation (65-70%). The phreatomagmatic phases occurred at the beginning of the eruption when magma flux was higher. The phreatomagmatic products had a wide range (but low) of vesicularities because groundwater seeped in to quench the system at depths when the magma was beginning to vesiculate. As the eruption proceeded then groundwater no longer affected the eruption so the magma was able to fragment to shallower depths producing higher vesicularities. It was also said that during magma ascent, parts of the magma cooled by interacting with the surrounding wall rock, which caused the viscosity to increase prior to fragmentation and induced slower ascent rates.

Phreatomagmatic eruptions can produce clasts of various vesicularities because external water can enter into the magmatic systems at various depths quenching the magma. The top portion of the magma rising through the conduit has a higher vesicularity compared to the magma far below it. Moore and Schilling (1973) describes that submarine basalt phreatomagmatic eruptions along the Reykjanes Ridge have low vesicularities of 5-16% depending on the depth at which eruption is quenched in the vent. The Surtseyan eruption at Black Point in California had vesicularities ranging from 4-89% and deposits were ash and lapilli dominated (Murtagh & White 2013). The vesiculating magma interacted with external water near the surface. Murtagh and White (2013) explain that the range in vesicularity is due to magma heterogeneity similar to Mt. Etna 122BC. Moreland and Thordarson (2014) observed that Eldgjá phreatomagmatic vesicularities had a 40-50% range and a mean of 70-80%.

Phreatomagmatic eruptions cover a wide range of vesicularities, discharge rates, and intensities producing large amount of ash. These eruptions can become very explosive due
to FCI processes, which affect magma as great depths, as well as rootless eruptions and cones. However, not all phreatomagmatic eruptions are explosive and especially in Iceland when the magma only begins to interact with water at or near the surface then water has no effect on driving the eruption to become explosive. In this case, water will quench the system and leading to cooling-contraction granulation to form the ash rich layers in the eruptions and only the expansion of gases in the conduit is able to drive the eruption to become explosive or effusive.
4 Laki – Grímsvötn Volcanic System

4.1 Geology of the Volcanic System

The Laki fissure is located in South-Central Iceland in an upland area, which is bordered by a prominent scarp (remnants of an old sea cliff), and is part of the Grímsvötn volcanic system (Figure 4.1). The volcanic system consists of the Grímsvötn central volcano (located beneath Vatnajökull glacier) and a fissure swarm about 100 km long and 15 km wide running north-east to south-west from the Grímsvötn central volcano (Thordarson and Self 1993).

The Grímsvötn central volcano is the most active component of the system, periodically erupting every 10-15 years (e.g. Thordarson & Larsen 2007). However, the system was inactive for 38 years prior to the especially massive Laki fissure eruption of 1783-84. The name for the Laki fissure eruption is taken from Mt. Laki, an older hyaloclastite structure that is situated in the middle of the Laki vent system, which subdivides the cone row in SW and NE portions. Compositionally, the Grímsvötn volcanic system is tholeiitic and has erupted about 50-55 km³ of magma during the Holocene. Of this magma, 15.1 km³ was erupted from the Laki fissure (e.g. Thordarson et al. 1996; Thordarson & Hoskuldsson 2008). The Grímsvötn caldera is 20 km² in area and about 250-300 m deep and supports an active geothermal system. This volcanic system is thought to have a shallow magma chamber due to the active geothermal system forming a pool of melt water residing on top of the ice covered caldera (Alfaro et al. 2007).

The Laki vent system is composed of over 140 vents and 10 fissures covering a distance of 27km (Figure 4.1; Thordarson & Self 1993). It produced one of the largest basalt lava flows in historic times, devastated the populations inhabiting the Fire Districts and affected numerous countries in the Northern Hemisphere (Thordarson et al. 1996, Thordarson et al. 2003, Thordarson 2003).
Figure 4.1: Isopach map illustrating distal tephra thickness from the P1 phase. Total of 0.43 km$^3$ dense rock equivalent volume of tephra was produced from the Laki eruption. (Original isopach from Thordarson & Self 1993)
4.2 Laki Fissure Eruption 1783 -1784

The Laki fissure eruption was a combination of phreatomagmatic and Magmatic style erupting and lasted from June 8th, 1783 to February 7th, 1784 (Thordarson et al. 1996, 2003; Thordarson and Self 1993). The Laki eruption consisted of 10 separate episodes with phreatomagmatic and/or Magmatic characteristics. It has been described as the second largest flood basalt eruption in historic time (the first being the Eldgjá eruption of 934AD) producing 15.1 km³ of basalt. The volume of tephra produced was 0.4 km³ dense rock equivalent volume or 2.6% of the total erupted volume (Figure 4.1). The eruption occurred in the Síða highlands, which is an ice-free region south of the Laki fissure and between the sea scarp and Sídújökull. The highlands are bordered by the Skáftá and Hverfísfljót rivers and had built up a large outwash plain during the Holocene. The various settlements (Álfaver, Medalland, Skáftártunga, Landbrot, Síða and Fljótshverfi) living on this plain or scarp were known as the 'Fire Districts' and were most affected by the Laki eruption (Thordarson and Self 1993; Thordarson et al. 2003; Thordarson 2003).

Most of the erupted material (~13 km³ of solidified lava) was expelled during the 10 episodes in the first 5 months of the eruption (Figure 4.2). As demonstrated by Thordarson and Self (1993), each episode was marked by the onset of earthquake swarms with increasing intensity and the opening of a new fissure segment, followed by a violent Magmatic/subPlinian or phreatomagmatic phase (0.5-4 days) and then an effusive phase, which lasted days to weeks. The eruption was characterized by very episodic nature with a mean effusive rate of $5.1 \times 10^3$ m³/s during the first 5 episodes of the eruption and maximum rates of $8.5 \times 10^3$ and $8.7 \times 10^3$ m³/s from the first two fissures. The fluctuation of eruption rate and sequence of events shows more of an unsteady or non-uniform conduit flow and not due to separate inflation-deflation events within the Laki reservoir (Thordarson and Self 1993).

The Laki fissures erupted either Magmatic or phreatomagmatic tephra. Figure 4.3 displays the isopach map for Laki proximal tephra. The total volume of tephra produced from Laki was 0.4 km³ dense rock equivalent volume or 2.6% of the total erupted volume. The erupted Magmatic tephra is solidified from slightly more evolved magma compared to the phreatomagmatic magma and has higher concentrations of FeO (~13.7-14.6 wt%);
The magmatic tephra units consist of shiny black tephra and abundant Pele’s hair and contains an abundance of highly vesicular grains, including golden pumices, which are covered by shiny black glassy surface film that is 10-100 micrometer in thickness. This film is inferred to have formed by fusion of the outer surface of the foamy (pumiceous) grains as the extremely hot magmatic gas streamed around the grains at high velocity (Thordarson et al. 1996). These magmatic clasts are highly vesiculated because the magma was able to fragment before erupting, resulting in a high sulphur content within inclusions in the tephra and low sulphur concentrations in the ground mass (418 – 640 ppm) and lower FeO concentrations (~13.2-14.6 wt%; Thordarson et al. 1996). Phreatomagmatic eruptions produce juvenile tephra, which were quenched during the magma fragmentation process due to interaction with external water. The non- to poorly vesiculated phreatomagmatic clasts have a high sulphur content (1000 – 1260 ppm), where highest values approach the initial sulphur content of the magma. Highly vesiculated clasts (~60-80%) have similar sulphur content to that seen in magmatic clasts. The phreatomagmatic tephra and its clasts have a dull appearance and are greyish black to golden brown colour, which is a stark contrast to the shiny black tephra clasts of the magmatic (S) units. Simplest explanation for the absence of the shiny black glass film on the phreatomagmatic clasts is that the gas phase that carried them into the atmosphere was relatively cold (<500°C?). Phreatomagmatic clasts range from non- to highly vesicular (Thordarson et al. 1996). The highly vesiculated tephra are products of the magmatic foam. Figure 4.4 displays visual differences of the Laki Magmatic and phreatomagmatic tephra.

There are four main tephra units erupted from the SW row of the Laki fissure. The Magmatic units are represented as layers S1, S2, and S3 and the phreatomagmatic layer is P1. S1 Magmatic layer is restricted to the area around fissures 1 and 2. S2 is the erupted Magmatic tephra from fissure 3 and contains a very high abundance of Pele’s hair and tears. P1 is the erupted phreatomagmatic tephra from the explosive activity occurring at fissure 4, which created tuff cone 1 (TC1). Finally, S3 is an erupted Magmatic tephra, which corresponds to episode 5 but erupted from the entire SW-row (Figure 4.5) (Thordarson & Self 1993; Thordarson & Self 2003).
The main topic of this study is the phreatomagmatic phase of episode four, which was associated with the opening of fissure segment 4 on the SW-row of the Laki fissure. This eruption produced the tephra layer P1 and TC1 clearly visible along the Laki cone row (Figure 4.6). TC1 was the main sample collection site for this study. Studies performed by Thordarson and Self (1993) described at least 10 episodic events during the eruption. Each episode is linked to the opening of a new fissure segment. The main volcanic activity was
focused at the newly formed fissure segment starting in the SW and slowly propagating NE. The fourth episode was characterized by strong tremors felt from June 23\textsuperscript{rd} – 26\textsuperscript{th} and a large lava surge flowing down the Skæfjá Gorge on the 29\textsuperscript{th} ending the episode. There was also an increase in explosive activity between the 23\textsuperscript{rd} and the 25\textsuperscript{th}. Reverend Steingrimsson described an enormous black ash cloud rising up from the SW cone row on June 27\textsuperscript{th} and it twisted over the mountains and formed thick clouds resembling cumulous clouds (Thordarson et al. 2003; Thordarson 2003). This description fits with the characteristic display of a phreatomagmatic plume. The external water quenching the magma and causing the large black ash cloud to emerge is most likely from lake Lambavatn, which is located just NW of the tuff cone. Numerous faults propagate through Lambavatn and could potentially allow cool fluids to reach the magma before it reaches the surface (Thordarson & Self 1993).

\textbf{Figure 4.4:} Images show the key visual differences between the magmatic (left) and phreatomagmatic (right) clasts (-4.5φ) found at tuff cone 1. The magmatic clasts are from tephra unit S2, which is from episode 3. Magmatic clasts have a smooth dark surface that is somewhat shiny due to re-melting by hot gases.

The timeline of the fourth episode will be evaluated later in the Discussion Chapter based on the stratigraphy of the sample sites around TC1.

The Laki eruption is one of the best documented eruptions to have occurred prior to the 20\textsuperscript{th} century. Detailed first-hand descriptions of the eruption were made Reverend Steingrimsson, who served the farming community in the Sida district from 1778–1791 and resided at his farmhouse, Prestbakki, which was located 40 km south from the
fissures (Thordarson 2003; Thordarson et al. 2003). There are several other Icelanders who wrote about the eruption as it occurred but Steingrímsson’s were quite detailed and descriptive and he often wrote to the authorities about the progress of the eruption and its destruction to the settlements.

The following outline of the course of events of the Laki eruption is taken from Thordarson et al. (2003), which is constructed via detailed analysis of the contemporary accounts written by Steingrímsson coupled with field observations. An evaluation of the reliability of these historic accounts is given by Thordarson (2003).

In mid-May tremors were felt in the Fire districts, initially in Skaftártunga, which increased in intensity, especially in the first week of June until the eruption began at 9am on June 8th, 1783. This earthquake activity marked the onset of rifting within the Grímsvötn volcanic system and the emplacement of the Laki feeder dyke. In the early morning on June 8th, a black volcanic cloud rose up into the air above the Fire Districts. Tephra fell heavily during the day then easing in the late afternoon. The tephra covered the ground and was thick enough for footprints to be made. Pele’s hair scattered the ground around the districts and the rain was soaked with black ash and resembled black ink dripping off plants and homes. Numerous columns of fire were seen that day from the Síða and Skaftártunga districts. By the 10th the air was thick with the stench of sulphur causing dizziness and irritated the eyes and skin of the locals. The rain that fell burned through leaves and irritated the skin of newly shorn sheep by leaving red spots. The flow from the Skaftá River had begun to dwindle on the 9th and by the 10th was completely dried up. Lava quickly poured into the gorge at tremendous speed on June 12th. There were a total of five lava surges, corresponding to the first five episodes of the Laki eruption, that filled the Skaftá River Gorge occurring on the 12th, 15th, 18th, 29th and July 14th. There was a substantial amount of tephra falling in Síða on July 9th and 10th and there was increased activity on July 13th with thunder, lightning and earthquakes for four days after the 13th. Lava flowed into the gorge on July 13th and 14th. Ash covered the ground on July 18th with increasing activity and earthquakes. Fires were observed at the fissures till the end of July, fine ash layered the ground and a fire-coloured glow was seen on July 24th and 25th. Rumbling and boiling sounds were heard on July 29th from the NE cone row. On this day, a “huge and dreadful” black cloud rose up from this site raining down ash and pumice
over Fljótshverfi and eastern Sída. The water in the Hverfisfljót River began to heat up on August 3rd and was completely dried up the following day with lava flowing down the gorge August 7th. Intense rumbling and cracking sounds were heard from a new fissure between July 30th and August 7th. Lava flowed down the Hverfisfljót gorge till August 14th and ash rained down over Sída from August 17-23. The 23rd was accompanied by boiling sounds and earthquakes that were felt till the end of the month. The second lava surge flowed down the gorge on September 1st and was the most intense on September 7th. Steingrímsson recalled that all the rivers and streams on the Hverfisfljót sandur plain were dried up. The third lava flow was on September 10th. Water began to flow down the gorge till the 26th when strong earthquakes were felt and lava flows emerged drying up all the rivers again. Activity was said to have diminished till tremors on October 24th and a foul stench was in the air. A huge column of fire rose on the 25th and an intense lava surge flowed down the Hverfisfljót river gorge. After this lava flow, all subsequent flows were confined to the highlands. Rain was frequent during October and November and contained ash and sand. On November 24th a fire column was seen and strong earthquakes were also felt. During 1784, the air was thick with the smell of sulphur in the Fire districts and fires were seen on the fissures till January 14th 1784. Activity ceased on February 7th, 1784. The eruption at Grímsvötn continued till May 26th, 1785.

Grímsvötn was active during the Laki eruption and had an increase in activity during the autumn of 1783 and continued till May 1785. Grímsvötn and Laki are part of the same volcanic system and both eruptions overlapped each other, meaning they are from the same volcano-tectonic event (Thordarson et al. 2003). The composition of the products from each eruption show that Laki’s source is a large deep-seated reservoir located at crust-mantle boundary, whereas, Grímsvötn’s source is a shallow crustal magma chamber. This may indicate that as the eruption at Laki progressed then there was a gradual change in the flow of magma from a deep reservoir to a shallow chamber (Thordarson et al. 2003). This would also explain the episodic nature of the Laki eruption, which was mention earlier.
Figure 4.5: Stratigraphic section of the Laki tephra units illustrating the date of deposit and the proximal location of each unit as it corresponds to a particular eruption episode. (Source: Thordarson and Self 2003).

Figure 4.6: The Laki vent system showing the many volcanic cones and vents produced during the 10 eruption events from 1783 to 1784. The top picture shows the SW sections of the Laki cone row and the bottom picture is the NE section. Both tuff cones (TC-1 and TC-2) are labeled on the figure and correspond to the phreatomagmatic phases from fissure 4 and 6 (Source: Thordarson et al. 2003).
4.3 Laki Atmospheric Sulphur Loading

Laki was one of the most influential atmospheric pollution events in the past 250 years and affected most of the Northern Hemisphere, from parts of Europe to the Middle East and China (Stothers 1999; Thordarson et al. 1996). Laki’s degassing regime was determined using pre-eruptive inclusions of sulphur, chlorine and fluorine within the products at various stages of the eruption (Thordarson et al. 1996; Thordarson and Self 2003). Degassing had occurred in two ways: 1. During the explosive activity at the vents and 2. From the erupted lava flow. Thordarson et al. (1996) measured the sulphur within the inclusions and determined that about 122 megatons (Mt) of sulphur was released over the 8-month period of the eruption. This amount of sulphur is able to oxidize to produce about 250Mt of sulphate aerosols. Most of the sulphur was released within the first 1.5 months of the eruption, which was when Laki was most actively erupting magma (Figure 4.7). Thordarson et al. (1996) used this correlation to predict that in eruptions, the largest spells of gas are released at times of increased magma rate (usually at the beginning of an eruption). The sulphur and other volatiles were erupted to heights of 6 – 13km in the atmosphere, which allowed the sulphur to reside in the upper troposphere and lower stratosphere for 2 - 3 years after the Laki eruption ended (Thordarson et al. 1996).

First-hand observations of Laki’s influence, made by S.P van Swinden, described an unusual blue haze that crept over the hills on the morning of June 19th (Thordarson & Self 2001; Lindmand & Thordarson 2001). This blue haze resembled fog, however, it gave the sun a dimmed red glow, didn’t leave dew in the mornings and persisted though the sun was present. This phenomenon is known as dry fog because it doesn’t leave dew on plants, which is a main characteristic of typical wet fog. Dry fog was observed all over Europe starting around the 19th of June (1783) and was at its most intense by the 23rd. The dry fog dissipated around the 30th of June but often persisted till the autumn of 1783 or even January 1784 (Figure 4.8). Dry fog is made when sulphuric acid falls out from volcanic eruptions and was, therefore, a product of the Laki eruption (Thordarson and Self 2003). The dry fog lay mostly in the lower troposphere but with the fall of acid rain then this led researchers to believe the dry fog extended into the upper troposphere as well (Rampino et al. 1988). There were countless cases of skin irritation, headaches, and the number of
deaths during this period increased. Crop failure was common during this time due to the dry fog, which caused wide spread famine, especially in Iceland (Rampino et al. 1988; Thordarson & Self 2003; Thordarson & Self 1993).

Figure 4.7: Most of the sulphur emitted from Laki was within the first 1.5 months of the eruption. Figure shows the mass sulphur total released during the eruption. (Source: Thordarson and Self 2003)

The same blue haze was observed all over Iceland and at an even greater intensity than seen in Europe (Rampino et al. 1988). Laki caused the summer crops in Iceland to fail as well as 75% of the livestock and 24% of the population to perish. In Iceland, the dry fog was most intense in June and July of 1783 as Laki was at its most prevalent state but seemed to dissipate by December. The winter of December 1783 to February 1784 was especially cold and average temperatures had decreased significantly, with parts of Europe dropping as much as 4.5°C (Figure 4.9; Rampino et al. 1988). The following spring, autumn and winter had all shown decreased average temperatures as well (Grattan 2005). Grattan (2005) explains that it is the cooler summers, causing crop failure and famine, which lead to increased deaths and not so much the colder winter temperatures.
Figure 4.8: Historical accounts show that most of Europe was affected by dry fog just days after Laki began to erupt. (a) Locations and timing of the first appearances of the dry fog and haze from Laki. Locations of maps (b), (c), and (d) are shown in the dotted-line box in (a). (b)-(d) show locations (dots) and dates (numbers) of the observations made. (b) European Map showing locations of the very first observations made about the dry fog in June of 1783. Some dates are unknown but their locations are still given. (c) Dates and locations of when the thick tropospheric haze was first witnessed (black dots) and where there’s a sudden increase in opacity but the date isn’t specified (white dots). (d) Dates of the last occurrence of the haze. (Source: Thordarson and Self 2003).
Figure 4.9: The Laki eruption decreased mean temperatures all across Europe and especially in the winter months. The cooler temperatures lasted until the following winter of December 1784. (Source: Grattan 2005)
5 Methods

5.1 Grain Size Measurements

Grain size distribution is an important parameter used to determine initial fragmentation characteristics of the magma and the eruption style by comparing particle size to initial gas content and magma-water interaction processes (Bonadonna & Houghton 2005). Grain size measurements can also be used to determine plume dynamics and assess for hazard mitigation. The process of measuring and analyzing samples to produce a grain size distribution is outlined in the following sections.

5.1.1 Mechanical Sieving – Course Fraction

The first step for determining grain size distribution is to separate the tephra according to their grain size. This is done through dry sieving each sample using the Retsch (DIN ISO 3110/1) sieves, which are sized from -5.5Φ to 4.0Φ (or 45 mm to 63 μm) at intervals of 0.5Φ. Samples were dried in an oven at 40°C for 24 hours prior to sieving and transferred to individual bags and labeled. Contents and mass of each size interval in the sample were documented and graphed (See Results Chapter, section 6.2.1).

5.1.2 Sieving by Sedigraph – Fine Fraction

Each sample contained a large amount of fine particles (>3.0 g) that were less than 0.063 mm and therefore could not be completely sieved via the mechanical procedure. The fine particles are analyzed in a sedigraph to determine the grain size distribution of the fine fraction. Sedigraphs are ideal for analyzing samples that are in the mass range of 2.0-15 g, it is difficult to measure smaller or bigger samples than that. Before samples can be measured in a sedigraph, the density
of the ash fraction has to be determined by a pycnometer in order to calculate accurate settling rates for grains and to determine the appropriate amount of glycerol liquid to be added to the water column (Sartorius 1999). In order to acquire the desired precision at least three density measurements from three separate samples are required. This method is known for its precision, especially for materials with poor floatability, such as tephra and ash.

Formula (1) is used to calculate density of the solid material ($\rho_s$). However, the volume ($V_s$) and mass of the solid material ($m_s$) are unknowns and need to be calculated.

$$\rho_s = \frac{m_s}{V_s}$$

(1)

- $\rho_s$ = density of the solid material (g/cm$^3$)
- $m_s$ = mass of the solid material (g)
- $V_s$ = volume of the solid material (cm$^3$)

The volume of the solid is calculated by measuring the pycnometer’s volume when filled with fluid and solid sample and subtracting the volume of the fluid. This is shown in the formula below:

$$V_s = V_{ges} - V_{fl}$$

(2)

- $V_{ges}$ = volume of pycnometer filled with fluid and solid (ml)
- $V_{fl}$ = volume of the fluid (ml)

The pycnometer has a known volume and density and when filled with deionized water it can be used to find the solid’s volume. Deionized water is used in this technique because it has a known density (1.0 g/cm$^3$). As seen in formula (3):
\[ V_{ges} = \frac{m_{fl}}{\rho_{fl}} \]

(3)

- \( m_{fl} \) = mass of the fluid (g)
- \( \rho_{fl} \) = density of the fluid (g/cm\(^3\))

Next, the mass of the solid is calculated by using formula (4) and substituting it into formula (3). The sample solid and fluid are weighed in the pycnometer to determine the combined mass (\( m_{fl+s} \)) and subtracting the mass of the solid (\( m_s \))

\[ m_{fl} = m_{fl+s} - m_s \]

(4)

- \( m_{fl+s} \) = combined mass of fluid and solid (g)

Thus producing formula (5):

\[ V_{ges} = \frac{m_{fl+s} - m_s}{\rho_{fl}} \]

(5)

To summarize, first we measure the mass of sample solid in the pycnometer and subtract this value from the combined mass of the deionized water and solid in the pycnometer. The volume of the combined mixture is now known (\( V_{ges} \)) and is used to determine the solid volume. The pycnometer is of known volume and therefore, we know the volume of the deionized water (\( V_{fl} \)). The density of the solid is easily calculated now using the original formula (1). The previous formulas are used by Sartorius (1999) to determine sample density. A simpler technique was
used to measure sample density, though the original principles of Sartorius (1999) are still implemented (formula 6).

\[
\rho_s = \frac{\rho_{fl} \cdot m_s}{m_s - (m_{fl+p+s} - m_{p+fl})}
\]

(6)

- \( m_{p+fl} = \) mass of pycnometer and fluid (g)
- \( m_{fl+p+s} = \) mass of pycnometer, fluid and solid (g)

The mass of the pycnometer and fluid is 79.2249 g and the density of water is known to be 1.0 g/cm\(^3\). The mass of the solid is measured and is then placed in the pycnometer and filled with deionized water to be weighed again. The densities for sample L4-A002, L4-001, L4-006 and L4-012 are as follows: 2.674, 2.606, 2.573, and 2.503 g/cm\(^3\). The average density of 2.589 g/cm\(^3\) is used for sedigraph measurements.

### 5.1.3 Sedigraph Analysis

Once the density of the tephra is determined then each sample can be analyzed in the Sedigraph III 5120. The sedigraph can measure particles from <63μm down to 1μm as long as the user has at least 2.0-3.0 g of sample, but less than 15 g, available to analyze. The sedigraph determines particle size using Stoke’s Law, which is shown below. Stoke’s Law measures the rate at which particles fall through a liquid, of known properties, due to gravity (formula 7).

\[
D^2 = \frac{18 \nu \eta}{(\rho - \rho_o)g}
\]

(7)

- \( D = \) particles diameter (mm)
- \( \rho = \) density of the particle (g/mm\(^3\))
• $\rho_o = \text{density of the liquid (g/mm}^3\text{)}$
• $v = \text{terminal velocity (mm/s}^2\text{)}$
• $\eta = \text{viscosity of the liquid}$
• $g = \text{gravitational force constant (mm/s}^2\text{)}$

The sedigraph allows gravity to settle the particles as they fall through the analyzing cell in the glycerol mixture. While the particles are settling, the sedigraph sends out horizontal x-ray beams to measure the falling rates of the particles. The sedigraph then measures the mass of each size interval using the Beer-Lambert Law. The Beer-Lambert Law is a linear relationship between absorbance and concentration of an absorbing species. The mass concentration of the Laki tephra samples is based on their ability to absorb the low-powered x-rays. There are a few conditions that need to be fulfilled for Beer-Lambert Law to hold true. One being that the mixture must be as homogenous as possible and have no turbid flow or else the x-rays could scatter. Another condition is the x-rays must be parallel, monochromatic and travel the same length within the mixture (Jonsdottir 2015).

The liquid used in the mixture is a combination of glycerol and deionized water. The amount of glycerol can vary depending on how viscose the user needs the mixture to be. When the concentration of glycerol is high in the mixture then particles will take longer to settle out in the liquid and this leads to increased measuring time. The sedigraph program guides the operator to select the liquid mixture that is most suitable for their density measurements. Half of the samples, in this study, were analyzed using a 50% glycerol mixture and the other half used a 60% glycerol mixture. The sedigraph uses Reynold’s number to determine the liquid needed for analysis. Reynold’s number is a dimensionless quantity that measures the inertial or viscose forces during flow conditions. This quantity is important to insure that the Beer-Lambert Law holds true. When Reynold’s number is as close to 0.1 as possible and below 0.3 then the flow is laminar and viscose forces dominate, which leads to ideal conditions for testing (Sartorius 1999).

For analysis, the particles need to be separated and made into homogenous solid-liquid mixtures. This is accomplished using ultra sound. When the mixture is homogenized then x-rays are
projected through the cell to measure the settling particles. Values are collected and inserted into the grain size excel spreadsheet, which already has the sieved data recorded. The sedigraph data is rescaled to 100% because during sieving the fine particles $<3\Phi$ needed to be split in half. Half of the $<3\Phi$ sample is sieved till $4\Phi$ and the other half was analyzed in the sedigraph. After rescaling then a complete grain size record is displayed.

5.1.4 Gradistat

Gradistat is a program designed by Blott & Pye (2001) available for free online. This program is able to calculate many useful parameters. The values calculated using gradistat for this study were: mean, standard deviation, skewness, sorting, and kurtosis. Blott and Pye (2001) describe the formulas and theory behind how gradistat functions. Gradistat displays the values arithmetically (in microns), geometrically (in microns), and logarithmically (in phi). These values are displayed and discussed in the Results sections.

5.2 Bubble Imaging and Vesiculation

Vesicles in pyroclasts and lava flows display evidence of processes occurring in the magma conduit such as; gas exolution, expansion and degassing. These processes are known to drive most volcanic eruptions. Gas exolution in magma is controlled by magma decompression rate and initial concentrations of volatiles. Vesicle textures are controlled by nucleation and bubble growth, which are dependent on exolution, viscosity, magma rise rate, fragmentation and quenching. By examining and measuring density and vesiculation then we can gain an understanding of the magmatic processes that took place in the conduit during the Laki phreatomagmatic phase.
5.2.1 Density Measurements

To determine the density and vesiculation of the 12 density sample sets, I used the methods and sample preparation of Houghton & Wilson (1989) and Shea et al. (2010).

The samples were cleansed in a sonic bath for 15-20 minutes each. Samples were then drained and dried in the oven at 50°C for 24 hours. Any lithic clasts were taken out of the sample and 100 tephra clasts were chosen. Clasts were arranged from 1-100 or largest to smallest. Only clasts ranging from 16-32 mm were chosen for measurement because this size range best preserves original magma vesiculation at the time of the eruption. Clasts larger than 32mm are thought to experience thermal expansion after ejection from the vent and deposition on the surface leading to inaccurate vesicle sizes. Each individual clast was categorized by various componentry groups, which were based on vesicle size, shape, distribution and colour and were common to most samples. Clast densities are determined using Archimedes’ Principle, which states that the upward buoyant force exerted on an object submerged in water is equal to the weight of the fluid displaced by the object. By using formula (1) the density of each clast can be calculated. The mass of each clast is easily weighed on a scale and volumes of the samples are calculated by using the relationship that 1ml of water at 1atm and 4°C is equivalent to 1g.

Each clast was weighed on a scale and given a “dry” weight then wrapped in thin wax strips and labeled from 1-100 (Figure 5.1). The wax strips are of known size and weight and are easily added into any further calculations. The wax provides a necessary waterproof layer in order to keep the gas within the vesicles inside the clast and prevent the clast from filling with water and sinking. Next, each clast is placed into a container of water and held down with a ballast weight so it is completely submerged. A coiled wire in the shape of a soup ladle is attached to a ring stand and cradles the clast and ballast. The ring stand apparatus is placed on top of a scale so that the “wet” weight of each clast is recorded. The wax often doesn’t make the clasts completely airtight and will let bubbles out of the clast making the clast weight increase and fill with water. As long as the weight is noted before this occurs then there is no harm done. The steel ballast’s weight will be subtracted from the weight recorded during calculations.
To calculate the density of each clast, first we need to calculate the specific gravity. This is given by the following formula (8):

\[ SG_c = \frac{w_a}{w_a - (w_w - w_b + (w_f \times n_f))} \]

- \( SG_c \) = specific gravity of clast (g)
- \( w_a \) = weight of the clast in air (g)
- \( w_w \) = weight of clast in water with ballast (g)
- \( w_b \) = weight of ballast (g)
- \( w_f \) = weight of wax film (g)
- \( n_f \) = number of wax strips used
The density of each clast is equated by multiplying the specific gravity with the density of water (1.0 g/cm$^3$) (formula 9).

\[ \rho_c = SG_c \times \rho_w \]  

(9)

- $\rho_c$ = density of clast (g/cm$^3$)
- $\rho_w$ = density of water (g/cm$^3$)

Density is converted to vesicularity using the Dense Rock Equivalent (DRE) density (formula 10), which is effectively the melt density. The DRE density is used as a standard density for juvenile tephra clasts of basaltic origin. This density comes from measurements done on tephra with similar composition to that of Laki. Therefore, the DRE for this study is 2.75 g/cm$^3$.

\[ V = \frac{100(\rho_{DRE} - \rho_c)}{\rho_{DRE}} \]  

(10)

- $V$ = vesicularity (%)
- $\rho_{DRE}$ = Dense Rock Equivalent (g/cm$^3$)

The calculations were done on excel and clasts were organized into bins to determine the frequency of density measurements ranging from 0.0 to 1.4 g/cm$^3$. Density and vesicularity values were graphed against frequency and clasts were also plotted based on their componentry groups. All graphs and data are displayed in Results and Appendix.
5.2.2 Selecting Samples for SEM Imaging and VSD measurements

There are three sites in this study and each site has at least three density samples. To choose the most appropriate sample for SEM imaging and various vesicle distributions (VSD, VVD and NV\textsubscript{M}) we must consider the following criteria. Firstly, the samples must represent a time sequence within the stratigraphy section in order to compare between phases of the eruption. Secondly, each sample should represent a relatively short time interval during the eruption. Intervals during the eruption are separated by changes in stratigraphy. Lastly, each clast must be between 16-32 mm and be able to fit on a thin section slide without having significant pieces cut out. Sample site 3 has the best coverage, in terms of grain size and density measurements, out of all the sites. Therefore, two clasts were chosen from two separate samples. One clast was chosen from site 1 in order to correlate between the tephra deposited on both sides of the tuff cone.

The thin sections used for the SEM imaging and vesicle measurements were made at the University of Iceland by Atli Hjartarson.

The clasts were chosen from the modal peak in the density vs. frequency graphs (See Results Chapter 6.1). The mode gives the best representation of the entire sample. During the phreatomagmatic phase, external water quenched the system causing clasts with a wide range of vesicularities to be ejected during the eruption. By taking from the mode then we are able to represent the ejected material as a whole and this reduces error associated with measurements. Clasts were also selected based on the most abundant componentry group within the modal peak. The samples chosen to be made into thin sections were: Sample L4-004 clast 9 (L4-04-09), sample L4-008 clast 7 (L4-8-7) and 11 (L4-8-11), sample L4-010 clast 5 (L4-10-5) and 15 (L4-10-15), and sample L4-015 clast 8 (L4-15-8) and 20 (L4-15-20). However, only three of the samples were chosen for analysis: sample L4-04-09, L4-10-15 and L4-15-20. In addition to the three samples chosen, one sample (24-07-04-07(1)) from Benjamin Sellers’s study on the Laki phreatomagmatic phase 1 is analyzed and used in this study. One sample (L4-004) was chosen from site 1 to compare and contrast the results from both site 1 and site 3 because these areas are on opposite sides of the tuff cone. Each site on its own could represent different phases in the eruption.
Shea et al. (2010) discusses the inherent flaws associated with measuring vesicle size distributions in 2D. In most samples, the vesicles are polydispersed making it unlikely that each vesicle will be cut through its widest point. This can often underestimate the size of larger vesicles. Smaller vesicles are less likely to be intersected when the sample is cut compared to larger vesicles. Elongated vesicles are not properly represented either and samples that have experience any sort of shearing are not usually measured correctly. To correct for these problems, 2D vesicle area is converted to volumes and many models today are able to account for variations in shape due to intersection flaws. The model cannot account for samples that have experienced shear stresses but instead, assumes the samples have experienced the same deformation throughout the entire sample clast.

5.2.3 SEM Imaging

Image analysis methods were based on Shea et al. (2010) and were used to determine vesicularity distributions within the samples selected. The thin sections first needed to be scanned on the Cano Scan 5600F at 1200 dpi (dots per inch). This is to ensure there is an image of the entire sample so larger more prominent vesicles could be observed easily due to limitations with the SEM measuring at magnifications <30x. The scanned image is used for image analysis and is equivalent to 5x magnification. A scanned copy of the thin section is printed out so the locations of each image taken on the SEM can be recorded for future reference. Once the four thin sections had been scanned then they were coated in a thin gold layering using the Cressington Sputter Coater 108 Auto. Sputtering benefits samples by: reducing beam damage to the sample, increasing the thermal conduction between the sample and the SEM, and reduced sample charging. A single thin section is placed in the Sputter Coater and the surrounding air is removed to create a vacuum inside the chamber. This will ensure that the ejected argon atoms have a clear path to the thin section with no inferences or else atoms will be scattered and not effectively reach the sample. The sample is bombarded with argon gas particles, which causes the sputtering. Sputtering is a term used to describe when a glow discharge is formed by an anode and cathode using a gas to bombard a target material and erode it. The anode and the
cathode in this case are the argon gas and gold target, respectively. The resulting omni-directional deposition of sputtered atoms will form an even coating of gold on the thin section.

After the samples are coated then they are ready for the SEM. The Hitachi TM 3000 is the SEM used in this study. The sample is placed on a metal stand inside the chamber of the SEM and a small piece of carbon tape is placed on the thin section and the stage. This is to ensure appropriate conductivity between the thin section and the stage. The sample chamber is closed and the SEM evacuates the air to create a vacuum inside the chamber, which allows a clear path to the sample.

Figure 5.2: Sampling image strategy for samples L4-04-09, L4-10-15 and L4-15-20. Based off of Shea et al. 2010 15 image exponential nest. Four 50x images are taken adjacent to one another forming a large square image that mimics the amount of vesicles imaged by a 25x image. The 50x images are recognized as one image and three 100x images are taken from within the 50x image. Note: There are a total of 27 images taken of the three samples. Another set of images is taken within the same sample labeled: 50ba, 50bb, 50bc, 50bd etc. This second set of images follows the same image strategy as the one shown in the figure above.
Figure 5.3: Sampling image strategy for Sample 24-07-04-07 (1) based off of Shea et al. 2010 15 image exponential nest. Imaging is similar to the technique shown in figure 5.8, however, three 250x images are taken from each 100x instead of two and the 50x images overlap partially over one another. Note: There are a total of 33 images taken of 24-07-04-07(1). Another set of images is taken within the same sample labeled: 50ba, 50bb, 50bc, 50bd etc. This second set of images follows the same image strategy as the one shown in the figure above.

The Hitachi TM3000 image program displays the image on the computer screen at the selected magnification and allows for accurate control over image selection. Each image is made binary so the bubbles are black and the melt is white. For samples L4-04-09, L4-10-15 and L4-15-20 the image system in figure 5.2 was used. For Benjamin Sellers sample 24-07-04-07(1) the image system in figure 5.3 was used. The SEM located at the University of Iceland doesn’t have the ability to image samples under 40x magnification. Therefore, four 50x images are taken adjacent to each other to form one large image. This is to insure that the amount of vesicles analyzed in a 25x image is the same or less than the vesicles analyzed in the 50x-compiled image. The image scheme displayed in figure 5.2 and 5.3 are repeated twice.

Shea et al. (2010) explains the differences between imaging techniques and which method gives the most accurate results based on the amount of images analyzed. Figure 5.4 shows the differences between a 73 and 25 image grid system. The results for each imaging technique are
almost exactly the same. The 73 image grid system takes a longer time when analyzing each sample, however, it covers a much larger surface area on the sample compared to the 25 image system. Figure 5.4 also shows that the 15 image exponential nest system produces the same level of accuracy as the grid systems and requires much less time and effort to produce the images for each sample. Shea et al. (2010) discusses the appeal of the grid system as it avoids user bias. If the sample has heterogeneities then the user will often not pick these areas for analysis and instead choose areas that are more homogenous in character. The grid systems analyze larger areas and lead to less user bias.

Figure 5.4: Image on the left is the 3 image strategies tested by Shea et al. 2010; (a) 73-image grid nest where one image is chosen randomly within the previous image, (b) 25-image grid nest, which has a similar strategy to (a) but less images are taken, and (c) 15-image exponential nest where each magnification has twice as many images as the previous magnification. Image on the right is the number of objects per melt area for all three imaging strategies are compared. All three strategies give similar results meaning that a lesser number of images may be taken and still obtain accurate results. (Source: Shea et al. 2010).
For this study, an exponential nest system was used with similar traits to the 15-image nest system but with slight variation (Figures 5.2 and 5.3). The location of the images taken for Benjamin Sellers sample is displayed in figure 5.5.

Figure 5.5: SEM (relative) image locations displayed on the scanned image of sample 24-07-04-07(1). The black outline represents the compiled 50x images, which are shown overlapping each other. The black square on the left is 50 A series and the one on the right is the 50 B series. 50aa is red, 50ab is orange, 50ac is green and 50ad is blue. The 100x images are shown in purple. Three 250x images are taken within each 100x images but it is too small to display on the image above. The 50 B series images follows the same colouring as the A series.

5.2.4 Imaging in GIMP

After images were taken using the SEM then they were rectified using the GNU Image Manipulation Program (GIMP), which is available online as public domain. This was used to prepare each image for vesicle analysis in ImageJ.

The scanned image is rectified first by duplicating the original image and converting it to binary by using the threshold tool. It is rare for the black bubbles to be completely black and vice versa for the white melt. By using the Shrink 1 Grow 1 technique then areas that are only a couple
pixels wide will be eliminated from the image. The user then selects the melt and bubbles by colour and uses the paint bucket to fill in the areas to cover any stray pixels left over. Next, the pencil tool is used to clean up the image to look identical to the original SEM image. It is impossible and unnecessary to draw in every single vesicle in the 5x image. The smallest vesicles can be disregarded because it is proven that their disappearance on the 5x image will have very little to no effect on the vesicle distribution. Once the image has been cleaned up then it must be decoalesced. A decoalesced image depicts the appearance of the bubble population prior bubble coalescence and collapse of the magma foam (Figure 5.6). Therefore, original bubble walls need to be drawn in with the 3-pixel wide pencil tool. The walls need to be 3 pixels wide in order for ImageJ to process each bubble effectively and separately. After the image is decoalesced then a similar technique is done to display any crystals that may be in the images. Following the same steps, the user creates another binary layer for crystals so they can be analyzed separately from the melt and vesicles. The final images are saved as jpegs to be later uploaded to ImageJ. There are three separate saved images: 1. Bubbles and melt, 2. Crystals only and 3. Bubbles, melt and crystals.

Manually decoalescing the bubbles can lead to uncertainty and error in the data. Despite good knowledge on the interactions and coalescence of bubbles of varying sizes, the manual process of decoalescence can lead to error because the process is judgement dependent. In extreme cases this can affect the overall outcome to a significant degree, by increasing the bubble number densities as well as the weight of the smaller bubble sizes. Most of the time the reconstruction is within acceptable limits.
5.2.5 Image Adjustment via ImageJ

ImageJ is a program run by Java and is available online as public domain. It was initially developed by Wayne Rasband at the National Institute of Mental Health and was mainly used for Health Sciences (Rasband 1997). Recently, Bruce Houghton and colleagues at the University of Hawai’i adopted it as an imaging technique for vesicle size distribution analysis.

Images, decoalesced by GIMP, are uploaded to ImageJ as jpegs. These images need to be adjusted before any analysis can take place. First, the user must make the image binary then invert and save it as a separate binary image. Images are saved as separate files in order to retrace steps easily in case a mistake is made in the calculations. The image must be “smoothed”, which will get rid of any stray pixels as long as the four adjacent pixels are a different colour. Therefore, the user must set the count to 4 and iterations to 1 so the process occurs only once. Finally, the image is “opened” and “smoothing” can begin. After “smoothing” then the image is ready to be measured to determine the mean grey scale value ($G_e$) for the vesicles in the image. This value is recorded and used in later calculations. The mean grey scale value will lie between 0 and 255, which are the values for pure white and black pixels, because the mean value is an

Figure 5.6: Manually decoalescing images using GIMP. Coalesced image is shown on the left and decoalesced image is shown on the right.
indication of the ratio of bubbles to melt or black to white pixels. The mean value and the image are saved.

Next step is to select all the vesicles that are touching the borders of the image. Since these bubbles are not whole then they must be accounted for and excluded during analysis of the whole image. Instead, these partial vesicles are saved as a “mask” image and analyzed separately. The mean grey scale of the mask is recorded and saved. The working image will now have a yellow outline separating the mask from the rest of the whole vesicles. When the image is analyzed then ImageJ will know to exclude the mask vesicles. This image is inverted and saved as “analyzed image” and used for further image analysis.

To measure the mean grey scale value for the crystals the user only needs to make the crystal image binary, then invert it and measure the mean.

### 5.2.6 Image Analysis via ImageJ

The analyzed image is selected and opened in ImageJ. The user must set a scale for the image by setting “distance in pixels” to the magnification scale ie. 48 for 5x scan, 337.5 for 50x, 676 for 100x and 1693.33 for 250x. Then set the known distance to 1, pixel ratio to 1, unit length is in mm and the global box needs to be selected for the scale to be set correctly. After the scale is set the parameters for each measurement is selected: area, mean grey scale, min and max grey scale, perimeter, fit, ellipse, shape descriptors, Feret’s diameter, and area fraction. An image is created where all of the vesicles are represented as separate ellipses. This image is also accompanied by a results and summary spreadsheet and all of these items are saved separately. A second analysis is performed producing an outline image, which is an image counting every single bubble except for the mask bubbles. The outline image is saved in a separate file. The third image produced is a mask image showing all the bubbles minus the outside bubbles that touch the borders of the image. The only items that are actually used for analysis from ImageJ are the mean value for the whole image and mask, and the results excel sheet. The rest of the images produced are not used.
5.2.7 Calculating Vesiculation Size and Volume Distribution

An Excel spreadsheet, created by Bruce Houghton and colleagues at the University of Hawaii and later modified by William M. Moreland at the University of Iceland, is used to organize and analyse the data acquired from ImageJ. It is used to produce values for NA (number of vesicles per unit area), NV (number of vesicles per unit volume) and NV\textsubscript{M} (density of vesicles with respect to melt). This spreadsheet also produces graphical representations of the NA, NV, and NV\textsubscript{M} values.

Each image from ImageJ analysis produced an excel sheet with the area values (results excel sheet) as well as a number of other measurements. The column labeled area is copied from each of the ImageJ results and pasted into the new vesiculation spreadsheet and grouped together based on magnification. The area values represent each bubble’s area within the image in mm\textsuperscript{2} and in turn show the amount of bubbles in each image. The vesiculation spreadsheet performs a number of calculations to produce the vesicle data displayed in the Results section. The calculations are as follows:

Image Area

The total image area (pixels\textsuperscript{2}) is calculated by multiplying the dimensions of the image produced by the SEM (i.e. 1280x960) (formula 11). The scale factor is the “distance in pixels” value used during ImageJ analysis. The total image area (mm\textsuperscript{2}) is used to calculate the reference area.

\[
A_T = \frac{A_P}{SF^2}
\]

(11)

- \(A_T\) = Total image area (mm\textsuperscript{2})
- \(A_P\) = Total image area (pixels\textsuperscript{2})
- \(SF\) = Scale factor (pixels/mm)
Reference Area

The total image area includes the mask area and the reference area excludes the mask area. Formula (12) is used to subtract the mask area from the total area to calculate the reference area. The number “255” in the equation represents the mean grey scale value of an area completely covered by black pixels.

\[ A_R = A_T - A_T \left( \frac{G_e}{255} \right) \]  

(12)

- \( A_R \) = Reference Area (mm\(^2\))
- \( G_e \) = Mean grey scale value

Vesicle Fraction

The vesicle fraction is calculated dividing the mean grey scale value for vesicles by 255 (formula 13). The percent vesicle fraction is calculated by multiplying the results by 100. However, the value includes the edge vesicles and also very small vesicles, which should be disregarded from the calculation.

\[ V_f = \frac{G_v}{255} \]  

(13)

- \( V_f \) = Vesicle fraction
- \( G_v \) = Mean grey scale value for all vesicles
Crystal Fraction

The crystal fraction is calculated in the same way as the vesicle fraction (formula 14). This value is for phenocrysts only and can be represented as a percent fraction when multiplied by 100.

\[ P_f = \frac{G_c}{255} \]

(14)

- \( P_f \) = Phenocryst fraction
- \( G_c \) = Grey scale value for phenocrysts

Number of Vesicles

This section describes how to calculate NA and NV. NA is the number of vesicles per unit area (mm\(^2\)). NV is the number of vesicles per unit volume (mm\(^3\)). To calculate NA, the values from the previous calculations must be used to find the equivalent diameter (EqD). The equivalent diameter is the diameter of a sphere that has the same volume as the irregular shaped vesicle. To calculate the EqD, first the user must combine all the areas produced in ImageJ and organize them by their magnification. To calculate EqD formula (15) is used:

\[ EqD = \sqrt{\frac{4 \times A_v}{\pi}} \]

(15)

- EqD = Equivalent diameter (mm)
- \( A_v \) = Area of the vesicle (mm\(^2\))

Formula (15) is the combined area of a sphere and the area of each vesicle from ImageJ. The equivalent diameters are binned according to their value and each bin is 100.1 times larger than
The minimum geo bin size used in this study is equivalent to the minimum vesicle size in mm or 15 pixels wide.

The vesicles included ($V_I$) in this study are ones that are counted in the designated magnification bins for any further calculations. The designated bins are selected so there is a steady decrease of NA with increasing bin size. This is because each sample contains large amounts of small vesicles vs. larger ones. The bins selected from each magnification are also chosen so there is a small amount of overlap but no gaps in the data and empty bins are disregarded. Bins with a frequency of 10 vesicles or less, in that specific magnification, are not chosen and that bin will be chosen in the next lower magnification. The graphs produced by this technique are shown in Results. The Bulk NA ($NA_B$) is used to describe how NA is distributed among the bin sizes (formula 16).

$$NA_B = \frac{V_I}{A_{RM}}$$

- $NA_B$ = Bulk number of vesicles per unit area (mm$^2$)
- $V_I$ = Vesicles included in the reference area
- $A_{RM}$ = Reference area for each magnification (mm$^2$)

When calculating the $A_R$ of an image in a specific magnification (seen previously in “Reference Area” section) then all images from the same magnification are added together to get the total reference area. For example with 100x magnification:

$$A_{RMT} = A_{100aa} + A_{100ab} + A_{100ac} + A_{100ba} + A_{100bb} + A_{100bc}$$

The $A_{RMT}$ is used to calculate the NA for each bin (formula 17).
\[ NA_{\text{bin}} = \frac{f_{\text{bin}}}{A_{\text{RMT}}} \]  

(17)

- \( NA_{\text{bin}} \) = Number of vesicles per unit area in each bin (mm\(^2\))
- \( f_{\text{bin}} \) = Frequency of vesicles in each bin
- \( A_{\text{RMT}} \) = Total reference area (mm\(^2\))

These values were collected and used to plot NA vs. vesicle size (EqD).

Some of the images show a couple phenocrysts within the sample but not enough to actually affect the various vesicle measurements done in this study. Nevertheless, the number of phenocrysts per unit area (\( NA_P \)) can be calculated using the phenocryst fraction (\( P_f \)) and the \( NA_{\text{bin}} \). The \( NA_P \) is the value for NA but adjusted for the presence of phenocrysts (formula 18).

\[ (NA)_P = \frac{NA_{\text{bin}}}{1 - P_f} \]  

(18)

- \( NA_P \) = Number of phenocrysts per unit area (mm\(^2\))

Finally, the NA total is calculated by summing all the NA values in each bin (formula 19).

\[ NA_T = \sum_{bin=1}^{32} NA_{\text{bin}} \]  

(19)

- \( NA_T \) = Total vesicles per unit area (mm\(^2\))
Vesicle Size Distribution

Vesicle size distribution (VSD) is plotted as vesicle size vs. \( \ln(n) \) (mm\(^{-3}\)) where \( n \) is the number of vesicles (NV) divided by the lower size limit (lsl). Vesicle size distribution (VSD) is used to infer the kinematics of growth and nucleation rate of bubbles in the melt prior to erupting. Shea et al. (2010) explains that the final number density of a sample depends on the available volatile concentration in the melt, the rise rate and the ability for the bubbles to expand. Figure 5.7 (c) displays the visual differences between VSD graphs depending on the processes occurring in the melt. If there is one nucleation event where the entire system degasses at once in a single pulse then all the bubbles will be of the same size and the trend line on a VSD graph will have a very narrow size range and all bubbles will occupy this size range. A steady state nucleation or power law trend will be a negative linear sloping line. There is no coalescence or ripening during steady state processes. The bubbles grow during steady state by rising through the magma. Blower et al. (2002) describes that power law trends arise in nature when there are at least 5 nucleation events occurring in the magma. Whereas, exponential trends occur after 3 nucleation events and unimodal trends occur after just 1 event. A system that experiences coalescence will have a relatively linear trend line but with a curled up tail on the end (Shea et al. 2010) or a kinked trend with two separate slopes (Mangan and Cashman 1996). However, in Shea et al. (2010) they explain that the kink trend is seen in samples that experience multiple stages of nucleation. Coalescence and ripening will increase the dominant bubble diameter and decrease the number of bubbles in the melt, compared to steady state growth.
Figure 5.7: Images from Shea et al. 2010 displaying the trends observed in pumice clasts. (a) Vesicle volume distribution (VVD). (b) Cumulative vesicle volume distribution (CVVD). (c) Vesicle size distribution (VSD). (d) Cumulative vesicle size distribution (CVSD). Source: Shea et al. 2010.

The values for n are calculated by dividing NV by the lower size limit (lsl) in mm for each class number. The vesicle size values are also known as the middle size range for the vesicles and are calculated by taking the average of the upper and lower size limits (formula 20 and 21). For this study, class 26 is the minimum size (mm) and any class greater than 26 is equal to:

\[
\text{lsl for class } X = \frac{(\text{lsl class } x - 1)}{10^{0.1}}
\]

For class numbers less than 26 this relationship is used:

\[
\text{lsl for class } X = (\text{lsl class } x + 1)(10^{0.1})
\]

(20)

(21)
The same relationship can be used for upper size limits.

Vesicle Volume Distribution

To calculate vesicle size distribution (VVD), first step is to set each geo bin value to the power of three to calculate the volume. In order to proceed, the alpha values devised by Sahagian and Proussevitch (1998) must be recorded. These alpha values are used as a simple way to incorporate probability into the calculations. Though Sahagian and Proussevitch (1998) have only developed alpha values for 12 classes, Kathy Cashman has calculated class sizes for all 32 classes (Bruce Houghton, personal correspondence, 2015).

Hbar is the mean projected height of a shadow cast by a particle spinning about an axis. In this study, Hbar represents a spherically shaped particle, which is the diameter of the class size. Each geometric bin is described as a spherical diameter (Hbar) that represents the features in that bin size range. In order to do this, we use the mid-point in a volume linear scale. The bin diameters are cubed to convert these values to volume and then the arithmetic mean is taken to calculate the mid-point. Since Hbar is a value in mm then the mean is set to the power of $1/3$ to transform the volume to diameter.

The Hbar and alpha values are used to calculate the number of vesicles per unit volume (NV in mm$^{-3}$) for each geometric bin. Sahagian & Proussevitch (1998) used formula 22 to calculate NV:

$$NV_i = \frac{1}{H_i} (\alpha_1 NA_i - \sum_{j=1}^{i-1} \alpha_{(j+1)} + NA_{(i-j)})$$

(22)

- $H_i = $ Hbar (mm)
This formula calculates NV values from 2D observations to 3D. This discounts any larger vesicles that were cut at less than their maximum diameter; otherwise these vesicles would be placed into smaller class sizes and thus skewing results.

To calculate the volume fraction, we need to determine the volume of a sphere particle that has an equivalent diameter to Hbar (formula 23). Hbar is halved and substituted for the radius in a standard volume calculation of a sphere.

\[ V_s = \frac{\pi \times Hbar^3}{6} \]  

(23)

- \( V_s \) = Volume of a sphere with the same diameter as Hbar (mm³)

Then the volume fraction is calculated by multiplying NV and \( V_s \) (formula 24).

\[ F_v = NV \times V_s \]  

(24)

- \( F_v \) = Volume fraction

The volume fraction is adjusted by applying a correction factor, which is equal to the density-derived vesicularity divided by the sum of the volume fractions. This correction factor is used when the density-derived vesicularity and volume fractions sum have disagreeing values. If both values were the same then the correction factor would be 1 and the volume fraction would need no correction.

Figure 5.7 (a) displays the visual differences between various bubble growth processes seen in VVD graphs. In general, each mode in the VVD trend is interpreted to represent a pulse of...
nucleation and growth. Coalescence usually skews the data positively and can produce a coarse mode (Shea et al. 2010; Constantini et al. 2010).

Density of Vesicles with Respect to Melt

The vesicle number density with respect to the melt ($N\nu_M$) is an important parameter that is considered to correlate to the explosivity of an eruption (Stovall et al. 2012; Shea et al. 2010). $N\nu_M$ depends on decompression rate, diffusivity, viscosity and surface tension in the melt. If the viscosity and surface tension are high in a melt then bubbles are restricted from growing. These bubbles will then build up pressure and more bubbles will nucleated from the melt eventually exploding at the surface. The clasts produced will most likely have high number of bubbles but they will cover a finer vesicle size range because the bubbles are so small. This system will have a very high $N\nu_M$ because there is an increasingly large amount of bubbles in a given volume of melt (Stovall et al. 2012; Sable et al. 2006). A system that has a high magma ascent rate will cause the system to quickly degas at once because the magma rises so fast that it passes the liquidous point rapidly. The system doesn’t register that the volatiles need to degas at this point so it hurriedly degases all at once. This pulse of nucleation can have the same effect that the increased viscosity has on a system.

The number of vesicles per unit volume with respect to the melt is calculated using formula 25:

$$N\nu_M = \frac{NV \times 100}{100 - V_\rho}$$

(25)

- $N\nu_M =$ Number of Vesicles per volume corrected for the melt ($\text{mm}^{-3}$)
- $V_\rho =$ Density derived clast vesicularity (%)
6 Results

6.1 Field Sampling

The sampling sites for this study were situated at three separate locations in close proximity to tephra cone 1 (Figure 6.1), which is the location and source of the phreatomagmatic phase during the fourth episode. Dotted lines in the figure show the outline of large channels that cut into the tephra deposits. Sample site 1 is located in a wide channel at N64.05436 and W018.29416 less than 350 meters NW from the cone rim. Catherine Gallagher and Sarah Tapscott collected the samples from site 1. Sample site 2 is located in a wide channel running SW to NE almost parallel to the Laki fissures. It is located at N64.05316 and W018.29029 about 150 meters NW from the cone rim. Paavo Nikkola and Louis Steigerwald collected the samples from site 2. Sample site 3 is located on the opposite side (SE side) of the Laki tephra cone at N64.05106 and W018.27942 less than 200 meters SE of the cone rim. Catherine Gallagher, Sarah Tapscott, Morgan Haldeman, Megan Grainger, Thorvaldur Thordarson, and myself collected the samples from site 3.

A total of 38 samples (26 bulk grain size samples and 12 density sample sets) were collected in the lapilli and ash dominated units (S2 and P1) within the Laki eruption sequences. The Laki eruption consisted of a phreatomagmatic episode (P1) interrupting magmatic episodes (S2 and S3). In the stratigraphic sequence, we could observe both magmatic and phreatomagmatic phases. The uppermost unit of each sequence was the magmatic eruption (S3) with shiny black tephra and abundant Pele’s hair. Beneath this unit is the phreatomagmatic tephra (P1), the main target of our sampling campaign. Unit P1 consists of multiple layers of ash, pumice and various lapilli sizes. The deepest layer observed in this study is another magmatic tephra unit, S2, which has similar characteristics to unit S3, although significantly greater abundance of Pele’s hair. Grain size and density samples were all collected from the P1 sequence, except for sample L4-006 (and L4-A001?), which came from unit S2. Samples were collected when changes within the tephra sequence were observed indicating potential changes in eruption behaviour.
Figure 6.1: Map area taken from Google Earth showing the locations of the three sample sites. Dotted black lines indicate channels in the stratigraphy and black circles show the location of the sites. Top of the figure is oriented north.

6.1.1 Site Descriptions

Sites 1-3 have the layers within its stratigraphic column marked from A-Z. However, the correlation between sites is not straightforward and will be explained later on.

Site 1 is the farthest sampling area from the tephra cone out of the three visited during this study. It is located in a large channel running NW, almost perpendicular to the Laki fissure. The stratigraphic section at site 1 is shown in figure 6.2. The top layer (A) in site 1 is redeposited tephra, mostly derived from reworking of unit S3. Layer B consists of black shiny coarse lapilli tephra with Pele’s hair, particularly in the lowest 5 cm of layer and is identified as unit S3 (Thordarson and Self, 1993). The phreatomagmatic tephra unit, P1, is represented here by layers C-O, that extend from 25 cm to 95 cm below the surface of the pit dug for the section and thus is about 70 cm thick. The P1 layers are relatively ash rich with fine to course dull lapilli fragments. The phreatomagmatic section also consists of course lapilli layers dominated by ash interbedded with thinner dark black and brown ash layers (F-O), which is about 40 cm thick.
Figure 6.2: Stratigraphic section at Site 1. Unit A is redeposited tephra from the S3 unit and B is also part of S3. Layers C to O are P1 tephra. Layer P is S2 tephra. Samples marked in red are density samples and black ones are grain size samples.

Beneath the phreatomagmatic layer at about 95cm below the top of the section is S2 tephra layer (layer P). The S2 layer is magmatic in origin and is dominated by course shiny black lapilli and Pele’s hair. It is unknown, at this site, how far the S2 layer spans beneath P1.
Figure 6.3: Stratigraphic section of Site 2 showing sections A, B and C. Section A is lowest in the stratigraphic column and the measurements and sampling begins there and ends at the top layer of section C. Section A is 55 cm thick (0 – 55 cm), Section B is 65 cm thick (135 – 200 cm) and section C is 80 cm thick (260 – 340 cm). Red samples are density and black are grain size samples.

Site 2 is located in a wide deep channel that runs almost parallel to the Laki fissure and is located about 200 meters closer to the cone rim than site 1. The stratigraphic section of site 2 is displayed in figure 6.3. Figure 6.4 shows a photo of the site and where each of the three sections (A, B, and C) are located. There were 12 grain size and 3 density sample sets collected from this site. Layer A is potentially part of the S2 tephra unit and is poorly sorted with an overall very large grain size (lapilli to block sized). Layers B-T are characteristic of P1 tephra and cycle between poorly sorted course lapilli to bomb-sized layers (C, E, F, H, J, K, L, M, N, P, R and T) to fine-grained, homogeneous, relatively well-sorted layers (B, D, G, I, O, Q, S). The bottom section sampled in site 2 is section A. It is approximately 55 cm thick and has four noticeable layers within it (A, B, C and D). Section B begins about 80 cm above the top of section A and includes layers E to J. Four samples plus one density sample were collected in section B, which
is about 65 cm thick. Section C is located 60cm above the top of section B and consists of layers K to T. Four samples plus one density sample were collected in section C. This section is about 80cm thick.

Site 3 is located on the SE side of the Laki tephra cone in a wide channel running SE to NW. The entire section is more than 300 cm thick and consists of layers A-O. Ten grain size and five density sample sets were collected from this site and two of the four thin sections used for bubble analysis are representative of this area. Site 3 has a thick basalt flow covering and preserving the outcrop. The first 20cm of the stratigraphy column or layer A (figure 6.5(a)) is a thick basalt flow covering the section. Layer B is the S3 tephra unit, which is about 25cm thick and consists of Pele’s hair and course lapilli. Interbedded gold pumice and fine layers begin to occur in layer F to H. These interbedded layers pinch out about 10m laterally down the channel, but the course pumice unit (layer J) found at 180cm is still present. We moved to the south side of the channel and layer J is noted on the opposite side of the channel where sampling resumes (Figure 6.5(b)). A second pumice layer is shown in layer L and has similar size and properties to layer J. The layers K, M and N are medium lapilli dominated by ash layers that seem to fine as the eruption proceeded. Layer O is a course lapilli layer rich in ash with finer lapilli than layer N.

Figure 6.4: Site 2 outcrop. Sections 1, 2 and 3 are labeled. Photo provided by Paavo Nikkola.
Figure 6.5: a) Stratigraphic section of Site 3 on the northern side of the channel. Red samples indicate density measurements and black are grain size. b) Stratigraphic section of Site 3 on the southern side of the channel. Red samples indicate density measurements and black are grain size.


6.2 Grain Size

Each grain size (Table 6.1) sample was measured in 0.5φ intervals by sieving and sedigraph techniques described in the Methods section. Samples are graphed based on their grain size (φ) vs. weight (%) and all graphs each sample are presented in Appendix A. Example graphs from each site are displayed in figure 6.6, 6.7 and 6.8 for the purpose of exhibiting the main changes in grain size with eruption time. Samples are, also, described by their sorting using gradistat (Table 6.2; Blott & Pye 2001). Table 6.3 organizes the grain size data derived from the samples that were collected from sites 1, 2 and 3 and the locations of each sample are displayed in the stratigraphic sections in figures 6.2, 6.3, and 6.5.

Table 6.1: Volcanic particle description based on size

<table>
<thead>
<tr>
<th>Particle Name</th>
<th>Particle Size (mm)</th>
</tr>
</thead>
<tbody>
<tr>
<td>Bombs</td>
<td>&gt;64mm</td>
</tr>
<tr>
<td>Course Lapilli</td>
<td>32-64mm</td>
</tr>
<tr>
<td>Medium Lapilli</td>
<td>8-32mm</td>
</tr>
<tr>
<td>Fine lapilli</td>
<td>2-8mm</td>
</tr>
<tr>
<td>Very Course Ash</td>
<td>1-2mm</td>
</tr>
<tr>
<td>Course Ash</td>
<td>0.5-1mm</td>
</tr>
<tr>
<td>Medium Ash</td>
<td>0.25-0.5mm</td>
</tr>
<tr>
<td>Fine Ash</td>
<td>0.125-0.25mm</td>
</tr>
<tr>
<td>Very Fine Ash</td>
<td>&lt;125mm</td>
</tr>
</tbody>
</table>

6.2.1 Grain Size Sample Trends

Sample L4-006 is from Unit S2 and magmatic phase of episode 3. Its grain size distribution is weakly bimodal with the main mode of fine lapilli (Mdø = 5.66 mm) and it is poorly sorted (φ = 2.26) with a second mode at fine ash (Mdø = 0.125 mm) and relatively small (19%) ash-grade component.
<table>
<thead>
<tr>
<th>Sample #</th>
<th>Mean (φ)</th>
<th>Sorting (ϕ)</th>
<th>Skewness</th>
<th>Kurtosis</th>
</tr>
</thead>
<tbody>
<tr>
<td>L4-A002</td>
<td>0.187</td>
<td>2.237</td>
<td>0.438</td>
<td>2.618</td>
</tr>
<tr>
<td>L4-A003</td>
<td>-0.921</td>
<td>2.182</td>
<td>0.314</td>
<td>2.201</td>
</tr>
<tr>
<td>L4-A004</td>
<td>-0.742</td>
<td>2.519</td>
<td>0.519</td>
<td>2.856</td>
</tr>
<tr>
<td>L4-B005</td>
<td>0.052</td>
<td>2.108</td>
<td>0.629</td>
<td>3.06</td>
</tr>
<tr>
<td>L4-B006</td>
<td>-0.108</td>
<td>2.254</td>
<td>0.382</td>
<td>2.884</td>
</tr>
<tr>
<td>L4-B007</td>
<td>1.032</td>
<td>2.059</td>
<td>0.369</td>
<td>3.148</td>
</tr>
<tr>
<td>L4-B008</td>
<td>-0.463</td>
<td>2.439</td>
<td>0.167</td>
<td>2.482</td>
</tr>
<tr>
<td>L4-C009</td>
<td>0.208</td>
<td>2.193</td>
<td>0.387</td>
<td>3.118</td>
</tr>
<tr>
<td>L4-C010</td>
<td>-0.676</td>
<td>2.203</td>
<td>0.702</td>
<td>3.335</td>
</tr>
<tr>
<td>L4-C011</td>
<td>-0.618</td>
<td>2.446</td>
<td>0.475</td>
<td>2.789</td>
</tr>
<tr>
<td>L4-C012</td>
<td>1.26</td>
<td>2.065</td>
<td>0.326</td>
<td>2.861</td>
</tr>
<tr>
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<td>2.538</td>
<td>0.817</td>
<td>3.33</td>
</tr>
<tr>
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<td>2.694</td>
<td>0.372</td>
<td>2.539</td>
</tr>
<tr>
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<td>2.371</td>
<td>0.225</td>
<td>2.396</td>
</tr>
<tr>
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<td>2.263</td>
<td>1.69</td>
<td>5.304</td>
</tr>
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</tr>
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</tr>
<tr>
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<td>2.904</td>
<td>0.158</td>
<td>1.981</td>
</tr>
<tr>
<td>L4-013</td>
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<td>2.862</td>
<td>0.142</td>
<td>1.888</td>
</tr>
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<td>L4-014</td>
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<td>2.795</td>
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<td>L4-015</td>
<td>0.654</td>
<td>3.229</td>
<td>0.164</td>
<td>2.038</td>
</tr>
</tbody>
</table>

Sample L4-004B is a P1 phreatomagmatic sample from fine-grained bottom section of layer L in site 1. Its grain size distribution is polymodal with the main mode of medium ash (Mdø = 0.35 mm) and it is poorly sorted (φ = 2.37) with additional modes at fine and very fine ash (Mdø = 0.125, 0.063 and 0.031 mm) and very high (94%) ash-grade component.

Sample L4-004 is a P1 phreatomagmatic sample from the course middle of layer L in site 1. Its grain size distribution is polymodal with the main mode of course to very course ash (Mdø =
and fine ash (Mdø = 11.0, 4.00 and 0.125 mm) and moderate (56%) ash-grade component.

Sample L4-001 is a P1 phreatomagmatic sample from layer C in site 1. Its grain size distribution is polymodal with the main mode of fine lapilli (Mdø = 2.80 mm) and it is poorly sorted (Φo = 2.54) with additional modes at medium and fine lapilli and fine ash (Mdø = 22.6, 11.0 and 0.125 mm) and moderate (45%) ash-grade component.

Samples from site 2 are graphed and displayed in figure 6.2. Sample L4-A002 is a P1 phreatomagmatic sample from layer B in section A in site 2. Its grain size distribution is bimodal with the main mode of fine lapilli (Mdø = 1.4 mm) and it is poorly sorted (Φo = 2.24) with an additional mode at fine ash (Mdø = 0.125 mm) and moderate (63.6%) ash-grade component.

Sample L4-A003 is a P1 phreatomagmatic sample from layer C in section A in site 2. Its grain size distribution is polymodal with the main mode of medium lapilli (Mdø = 22.6 mm) and it is poorly sorted (Φo = 2.182) with additional modes at very course and fine ash (Mdø = 1.00 and 0.125 mm) and moderate (48%) ash-grade component.

Sample L4-A004 is a P1 phreatomagmatic sample from layer D in section A in site 2. Its grain size distribution is polymodal with the main mode of fine lapilli (Mdø = 5.70 mm) and it is poorly sorted (Φo = 2.52) with additional modes at medium lapilli and very course and fine ash (Mdø = 16.0, 1.40 and 0.125 mm) and moderate (48.7%) ash-grade component.

Sample L4-B005 is a P1 phreatomagmatic sample from layer E in section B in site 2. Its grain size distribution is bimodal with the main mode of very course ash (Mdø = 1.4 mm) and it is poorly sorted (Φo = 2.11) with an additional mode at fine ash (Mdø = 0.125 mm) and moderate (62.3%) ash-grade component.

Sample L4-B006 is a P1 phreatomagmatic sample from layer G in section B in site 2. Its grain size distribution is polymodal with the main mode of very course ash (Mdø = 1.4 mm) and it is poorly sorted (Φo = 2.25) with additional modes at medium lapilli and medium and fine ash (Mdø = 16.0, 0.35 and 0.125 mm) and moderate (48.7%) ash-grade component.
Sample L4-B007 is a P1 phreatomagmatic sample from layer I in section B in site 2. Its grain size distribution is bimodal with the main mode of very course ash (Mdø = 1.00 mm) and it is poorly sorted (□ø = 2.06) with an additional mode at fine ash (Mdø = 0.125 mm) and high (83.4%) ash-grade component.

Sample L4-B008 is a P1 phreatomagmatic sample from layer J in section B in site 2. Its grain size distribution is polymodal with the main mode of very course ash (Mdø = 1.4 mm) and it is poorly sorted (□ø = 2.44) with additional modes at medium lapilli and fine ash (Mdø = 22.6 and 0.125 mm) and moderate (56%) ash-grade component.

Sample L4-C009 is a P1 phreatomagmatic sample from layer K in section C in site 2. Its grain size distribution is polymodal with the main mode of very course ash (Mdø = 1.4 mm) and it is poorly sorted (□ø = 2.19) with additional modes at medium lapilli and fine and very fine ash (Mdø = 22.6, 0.125 and 0.044 mm) and moderate (68%) ash-grade component.

Sample L4-C010 is a P1 phreatomagmatic sample from layer M in section C in site 2. Its grain size distribution is bimodal with the main mode of very course ash (Mdø = 1.4 mm) and it is poorly sorted (□ø = 2.20) with additional modes at fine ash (Mdø = 0.125 mm) and moderate (67.4%) ash-grade component.

Sample L4-C011 is a P1 phreatomagmatic sample from layer R in section C in site 2. Its grain size distribution is polymodal with the main mode of fine lapilli (Mdø = 2.0 mm) and it is poorly sorted (□ø = 2.45) with additional modes at medium lapilli and fine ash (Mdø = 8.0 and 0.125 mm) and moderate (50.7%) ash-grade component.

Sample L4-C012 is P1 phreatomagmatic sample from layer S in section C in site 2. Its grain size distribution is polymodal with the main mode of very course ash (Mdø = 1.0 mm) and it is poorly sorted (□ø = 2.07) with additional modes at fine and very fine ash (Mdø = 0.125 and 0.031 mm) and very high (85.9%) ash-grade component.

Samples from site 3 are graphed and displayed in figure 6.3. Sample L4-007 is a P1 phreatomagmatic sample from layer D on the north side of the channel in site 3. Its grain size
distribution is polymodal with the main mode of very course ash (Mdø = 1.0 mm) and it is poorly sorted (\(\bar{\phi} = 3.11\)) with additional modes at medium lapilli and fine and very fine ash (Mdø = 16.0, 0.125, 0.063 and 0.022 mm) and high (79.8%) ash-grade component.

Sample L4-008 is a P1 phreatomagmatic sample from the top portion of layer E on the north side of the channel in site 3. Its grain size distribution is polymodal with the main mode of fine ash (Mdø = 11.0, 1.00 and 0.063 mm) and high (80%) ash-grade component.

Sample L4-009A is a P1 phreatomagmatic sample from the bottom portion of layer E on the north side of the channel in site 3. Its grain size distribution is weakly bimodal with the main mode of very fine ash (Mdø = 0.044 mm) and it is poorly sorted (\(\bar{\phi} = 2.85\)) with additional modes at medium lapilli and fine and very fine ash (Mdø = 1.4 and 0.0125 mm) and high (79.8%) ash-grade component.

Sample L4-009B is a P1 phreatomagmatic sample from layer F on the north side of the channel in site 3. Its grain size distribution is bimodal with the main mode of very course ash (Mdø = 1.4 mm) and it is poorly sorted (\(\bar{\phi} = 2.50\)) with an additional mode at fine ash (Mdø = 0.125 mm) and high (75.9%) ash-grade component.

Sample L4-010 is a P1 phreatomagmatic sample from layer J on the north side of the channel in site 3. Its grain size distribution is polymodal with the main mode of very fine ash (Mdø = 0.063 mm) and it is poorly sorted (\(\bar{\phi} = 2.65\)) with additional modes at very course, fine and very fine ash (Mdø = 1.00, 0.125 and 0.022 mm) and very high (89%) ash-grade component.

Sample L4-011 is a P1 phreatomagmatic sample from layer J (10m east along layer J) on the north side of the channel in site 3. Its grain size distribution is polymodal with the main mode of fine ash (Mdø = 0.125 mm) and it is poorly sorted (\(\bar{\phi} = 2.59\)) with additional modes at fine lapilli and very course and very fine ash (Mdø = 5.66, 1.00 and 0.044 mm) and very high (87.7%) ash-grade component.
Table 6.3: Organized grain size data

<table>
<thead>
<tr>
<th>Unit</th>
<th>Sample #</th>
<th>Layer</th>
<th>Mode (mm)</th>
<th>Secondary mode(s) (mm)</th>
<th>Sorting</th>
<th>Ash Comp. (%)</th>
<th>Uni, Bi or Poly</th>
</tr>
</thead>
<tbody>
<tr>
<td>Site 1 P1</td>
<td>L4-001</td>
<td>C</td>
<td>2.8</td>
<td>22.6, 11.0 and 0.125</td>
<td>2.54</td>
<td>45.0%</td>
<td>P</td>
</tr>
<tr>
<td>P1</td>
<td>L4-004</td>
<td>L</td>
<td>1.0</td>
<td>11.0, 4.0 and 0.125</td>
<td>2.69</td>
<td>56.0%</td>
<td>P</td>
</tr>
<tr>
<td>P1</td>
<td>L4-004B</td>
<td>L</td>
<td>0.35</td>
<td>0.125, 0.063 and 0.031</td>
<td>2.37</td>
<td>94.0%</td>
<td>P</td>
</tr>
<tr>
<td>S2</td>
<td>L4-006</td>
<td>P</td>
<td>5.66</td>
<td>0.125</td>
<td>2.26</td>
<td>19.0%</td>
<td>B</td>
</tr>
</tbody>
</table>

| Site 2 P1  | L4-C12   | S     | 1.0       | 0.125 and 0.031        | 2.07    | 85.9%         | P               |
| P1         | L4-C11   | R     | 2.0       | 8.0 and 0.125          | 2.45    | 50.7%         | P               |
| P1         | L4-C10   | M     | 1.4       | 0.125                  | 2.2     | 67.4%         | B               |
| P1         | L4-C09   | K     | 1.4       | 16.0, 0.125 and 0.044  | 2.19    | 68.0%         | P               |
| P1         | L4-B08   | J     | 1.4       | 22.6 and 0.125         | 2.44    | 56.0%         | P               |
| P1         | L4-B07   | I     | 1.0       | 0.125 and 0.031        | 2.06    | 83.4%         | B               |
| P1         | L4-B06   | G     | 1.4       | 16.0, 0.35 and 0.125   | 2.25    | 61.4%         | P               |
| P1         | L4-B05   | E     | 1.4       | 0.125                  | 2.11    | 62.3%         | B               |
| P1         | L4-A04   | D     | 5.7       | 16.0, 1.4 and 0.125    | 2.52    | 48.7%         | P               |
| P1         | L4-A03   | C     | 22.6      | 1.0 and 0.125          | 2.18    | 48.0%         | P               |
| P1         | L4-A02   | B     | 1.4       | 0.125                  | 2.24    | 63.6%         | B               |

| Site 3 P1  | L4-007   | D     | 1.0       | 16.0, 0.125, 0.063 and 0.022 | 3.11 | 79.8% | P |
| P1         | L4-008   | Top E | 0.125     | 11.0, 1.0 and 0.063      | 3.17  | 80.0% | P |
| P1         | L4-009A  | Bottom E | 0.044     | 1.4 and 0.125           | 2.85  | 79.8% | B |
| P1         | L4-009B  | F     | 1.4       | 0.125                   | 2.5   | 75.9% | B |
| P1         | L4-010   | J     | 0.063     | 1.0, 0.125 and 0.022    | 2.65  | 89.0% | P |
| P1         | L4-011   | J     | 0.125     | 5.66, 1.00 and 0.044    | 2.59  | 87.7% | P |
| P1         | L4-012   | K     | 1.4       | 11.0, 0.125 and 0.063   | 2.9   | 76.7% | P |
| P1         | L4-013   | K     | 1.4       | 5.66, 0.125 and 0.044   | 2.86  | 76.1% | P |
| P1         | L4-014   | M     | 0.044     | 8.00, 1.00 and 0.125    | 2.8   | 82.2% | P |
| P1         | L4-015   | N     | 1.4       | 16.0, 0.125 and 0.063   | 3.23  | 63.6% | P |

Sample L4-012 is a P1 phreatomagmatic sample from layer K on the north side of the channel in site 3. Its grain size distribution is polymodal with the main mode of very coarse ash (Mdø = 1.4 mm) and it is poorly sorted (dø = 2.90) with additional modes at medium lapilli and fine and very fine ash (Mdø = 11.0, 0.125 and 0.063 mm) and very high (89%) ash-grade component.
Sample L4-013 is a P1 phreatomagmatic sample from layer K on the south side of the channel in site 3. Its grain size distribution is polymodal with the main mode of very coarse ash ($\text{Md}_\phi = 1.4$ mm) and it is poorly sorted ($\phi = 2.86$) with additional modes at fine lapilli and fine and very fine ash ($\text{Md}_\phi = 5.66, 0.125$ and $0.044$ mm) and very high (89%) ash-grade component.

Sample L4-014 is a P1 phreatomagmatic sample from layer M on the south side of the channel in site 3. Its grain size distribution is polymodal with the main mode of very coarse ash ($\text{Md}_\phi = 0.044$ mm) and it is poorly sorted ($\phi = 2.80$) with additional modes at medium lapilli and very coarse and fine ash ($\text{Md}_\phi = 8.00, 1.00$ and $0.125$ mm) and high (82.2%) ash-grade component.

Sample L4-015 is a P1 phreatomagmatic sample from layer N on the south side of the channel in site 3. Its grain size distribution is polymodal with the main mode of very coarse ash ($\text{Md}_\phi = 1.4$ mm) and it is poorly sorted ($\phi = 3.23$) with additional modes at medium lapilli and fine and very fine ash ($\text{Md}_\phi = 16.0, 0.125$ and $0.063$ mm) and moderate (63.6%) ash-grade component.
Figure 6.6: Grain size distribution graphs of site 1 oriented in stratigraphic order. Sample L4-006 is a sample from the magmatic unit S2. P1 units coarsen with decreasing stratigraphic depth. The rest of the samples are displayed in Appendix A.
Figure 6.7: Grain size distributions of samples from site 2 sections B and C. Graphs show fine lapilli and coarse ash dominated trends. The rest of the samples are displayed in Appendix A.
Figure 6.8: Grain size distributions of samples from site 3. Samples from site 3 have relatively wide grain size distribution and some graphs (i.e. sample L4-013) show bimodal trends. Graphs are oriented in stratigraphic order. The rest of the samples are displayed in Appendix A.
To summarize, site 3 samples are very fine to fine ash dominant with modes between 0.022 and 2.00 mm. Site 2 samples are very coarse ash to fine lapilli rich with dominant modes between 1.4 and 22.6 mm. Site 1 has samples near the top of the stratigraphy with modes similar to site 2 and samples near the bottom with modes similar to site 3. Figure 6.9 (a and b) shows the visual correlation between site 1, 2 and 3.

![Figure 6.9: Correlation between the site 1, 2 and 3. Samples L4-001 and L4-004, located at the middle and top of site 1, show strong coarse ash and fine lapilli dominant trends similar to those seen in site 2 samples. Sample L4-004B, located at the bottom of site 1, shows strong fine ash trends similar to samples from site 3.](image-url)
6.2.2 Half Distance and Tephra Volumes

We were able to calculate thinning half distance of the phreatomagmatic phase 1 tephra using the isopach map (Figure 6.10). The isopach map was made using arcmap and superimposing an existing isopach map of the P1 tephra created by Thorvaldue Thordarson. Thickness of each isopach (mm) is plotted against thinning half distance and compared to previous data by Bruce Houghton on the Laki tephra (Figure 6.11). Tephra fall data is important in assessing eruption intensity and for future work on total grain size distribution. For perspective on the intensity of the P1 phase, Mount St. Helens half distance spans from the volcano (0 km) to more than 350 km and Taupo spans well past 120 km (Pyle 1989). Both Taupo and Mount St. Helens were more silicic in composition compared to Laki. The total volume of the Laki P1 tephra is about 0.008 km$^3$.

![Figure 6.10: Isopach map of the P1 tephra unit. Thickness of each isopach is displayed in cm. Laki TC1 is displayed in the center of the 50 cm isopach line.](image)
Figure 6.11: Pyroclastograph illustrating more accurate data from more recent tephra thickness calculations labeled as Laki 1783 P1. Other trends by Bruce are previous studies done on the Laki tephra. The trend lines labeled M1 and M2 correspond to the S1 and S2 magmatic tephra layers.

Table 6.4: Isopach thickness, thinning half distance (Area $\frac{1}{2}$) and volume of Laki P1 tephra

<table>
<thead>
<tr>
<th>Thickness (mm)</th>
<th>Area (pixels$^2$)</th>
<th>Area (m$^2$)</th>
<th>Area (km$^2$)</th>
<th>Area$\frac{1}{2}$</th>
<th>Volume (km$^3$)</th>
</tr>
</thead>
<tbody>
<tr>
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<td>7.23E+05</td>
<td>7.23E-01</td>
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<td>8.15E-04</td>
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<td>2.28E+01</td>
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<td>9.13E-04</td>
</tr>
<tr>
<td>30</td>
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<td>5.06E+07</td>
<td>5.06E+01</td>
<td>7.112</td>
<td>1.52E-03</td>
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<tr>
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<td>1.05E+23</td>
<td>1.05E+08</td>
<td>1.05E+02</td>
<td>10.239</td>
<td>2.10E-03</td>
</tr>
<tr>
<td><strong>Total</strong></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td><strong>7.89E-03</strong></td>
</tr>
</tbody>
</table>

Volume
6.3 Bubble size, number, and vesicularity

Each density sample was measured based on the techniques discussed in Methods Chapter. Each sample is described based on its graphical distribution in relation to other samples and other parameters such as colour and vesicle size. When samples were being initially measured for density and vesicularity and wrapped in wax then they were categorized based on a number of components. The juvenile tephra clasts are first placed in a main component group (A, B or C, see below) and then into a sub-component group (G-1, G-2, G-3 or G-4). Groups A, B and C describe the vesicle size and degree of roundness. Sub-components G-1, G-2, G-3 and G-4 describe variations of colour.

Componentry group A classifies clasts that have large angular vesicles (~1-5 mm). The vesicles are often non-circular and are much larger compared to the ones seen in B or C clasts. These clasts also contain small round vesicles, but the main criteria for group A is that the larger angular bubbles are present.

Componentry group B clasts have uniform sized vesicles that are well rounded, small (0.2-2 mm) and usually abundant throughout the clast. These clasts can be more dense (>0.2 g/cm³) than the A group clasts, except golden pumice clasts (BG-2).

Componentry group C clasts have little to no vesicles. They are not lithic fragments and still have relatively high vesiculation values (greater than 50%). These clasts are usually denser (0.3-1.0 g/cm³) than B and A clasts.

Each component can be described further using the sub-components: G-1, G-2, G-3 or G-4.

Sub-component G-1 describes a clast that has a mixture of gold and metallic (blue-silver) colouring. Any combination of this colouring is classified as G-1. Example of an A G-1 clast is shown in figure 6.12

Sub-component G-2 is a clast that is pure gold colouring. There is extremely little to no trace of metallic colour on this clast. These clasts are very delicate and very low density
 (~0.1-0.3 g/cm³). Clasts are usually small and uniform in size and are usually classified as B group clasts. Example of a B G-2 clast (golden pumice) is shown figure 6.12.

**Sub-component G-3** clasts are dull brown and have relatively no metallic colour. Often this sub-group is paired with C clasts but also have occurred in A and B clasts.

**Sub-component G-4** clasts are dull brown and only show metallic and gold colouring inside the vesicles. This sub-component does not show up very much in these samples.

![Figure 6.12: Image on the left is an A G-1 clast with larger more angular vesicles and a slight metallic sheen. Image on the right is a B G-2 golden pumice clast with smaller more uniform sized bubbles and purely gold colouring.](image)

### 6.3.1 Clast Density and Vesicularity

The density and vesicularity of the 1180 sample clasts from 12 different samples were measured and graphed. Each sample’s graphical distribution, mean density and vesicularity will be discussed in relation to the site it resides in. The graphed results are shown in figure 6.13 and the mean values are displayed in table 6.4. For reference, a dotted line is shown in the vesiculation and density graphs at 75% vesicularity. The location of each density sample is shown in the stratigraphic sections of each site in figures 6.1, 6.2, 6.3 and 6.5. Componentry groups for each sample are graphed and displayed in the Appendix.
The mean density values of the P1 tephra range from 0.31-0.44 g/cm$^3$ (Table 6.5). The mean vesicularity values range from 84.5-88.1%, which basically categorises these phreatomagmatic clasts as “basaltic pumice” (Mangan and Cashman 1996).

Sample L4-006, the magmatic S2 tephra unit has a mean density of 0.6 g/cm$^3$, which corresponds to approximately 80% vesicularity. This is at the lower end, yet compares favourably with clast density and vesicularity from other magmatic (S) units from Laki, which range from about 60-90% vesicularity, respectively (Thordarson et al. 1996).

**Site 1**

Samples L4-002, L4-003, L4-004 and L4-006 are from site 1, listed here in reverse stratigraphical order (See figure 6.2 in Section 6.1). The magmatic S2 sample L4-006 is typified by a unimodal Gaussian distribution with a mode at 0.6 g/cm$^3$ (Figure 6.13), which corresponds to vesicularity of 79.8%. It has the widest density distribution of the Site 1 samples, ranging from 0.23-1.00 g/cm$^3$ and a standard deviation of ±0.06 g/cm$^3$. This sample is dominated by AG-1 lapilli (82%) but also contains BG-2 gold pumice (6%).

Sample L4-004 is from layer L in the P1 unit at site 1 and has a mean density of 0.31 g/cm$^3$. It has a tight density distribution, with a mode of 0.2-0.3 g/cm$^3$ and a range from 0.14-0.68 g/cm$^3$. The standard deviation is ±0.04 g/cm$^3$. More than 50% of the clasts have densities less than 0.30 g/cm$^3$. The corresponding vesicularities are 74-95%, with a mean of 88%. This sample has an equal amount of A and B clasts, 48% each, and 4% group C clasts. The lowest densities are dominated by A-type clasts and higher densities by B and C-type clasts (0.30-0.68 g/cm$^3$). A portion golden pumice clasts (BG-2; 7%) have vesicularities > 90%, suggesting that some of these clasts might be classified as reticulite.

Sample L4-003 is from layer J, which is a course lapilli layer within a sequence of alternating ash and course lapilli layers. Sample L4-003 has a less narrow density range, compared to L4-004, from 0.18-0.7 g/cm$^3$. Its main modal peak is at 0.27 and 0.33 g/cm$^3$ and has a mean density of 0.34 g/cm$^3$. This sample has greater density than L4-004, which occurred earlier in the phreatomagmatic phase. Vesiculation ranges from 74-93% and has a mean of 87.2%. The standard deviation is ±0.052 g/cm$^3$. This sample is dominated by both
A (43% of the sample) and B (50% of the sample). Only 2% of the sample consists of golden pumice. One of the golden pumice clasts has a vesiculation of 92%. Indicating a potential reticulite sample. Most of the B componentry clasts in this sample are BG-1 (47%).

Sample L4-002 is the youngest sample in terms of eruption timing and is found in layer F. It has a density range of 0.10-0.59 g/cm³ and potential outliers at 0.63, 0.73 and 0.89 g/cm³. It has a mean density of 0.32 g/cm³ and a mean vesiculation of 88.5%. The standard deviation is ±0.038 g/cm³. The vesiculation ranges from 79-96% and has very similar mean density and vesiculation values to L4-004 but a very similar graphical distribution to L4-003. Its main modal peak is at ~0.23 g/cm³ but with a significant peak at ~0.33 g/cm³. L4-002 is made up of 69% A componentry clasts and 31% B componentry clasts. The sample is dominated by AG-1 clasts (69%) and consists of 7 golden pumice clasts that have vesicularities above 90%, which is the same amount as in sample L4-004.

**Site 2**

Site 2 consists of samples L4-A, L4-B and L4-C. Sample L4-A was deposited first in the eruption and L4-C was last out of the three.

Sample L4-A density sample is from layer C in section A (Figure 6.3 Section 6.1). It has a mean density of 0.39 g/cm³ with a range of 0.16 to 0.79 g/cm³ and a potential outlier at 1.3 g/cm³. The modal density peak is at ~0.27 g/cm³. This sample has a mean vesiculation value of 85.2% and a standard deviation of ±0.05 g/cm³. Group B clasts make up 52% of the sample, group A are 42% and group C are 6% of the sample. Sample L4-A contains 19% BG-2 clasts and 8 BG-2 clasts have vesiculation values above 90%. The lower density size bins are dominated by A clasts and higher densities (>0.4 g/cm³) are dominated by B clasts. Most of the sample (77%) is classified as G-1 (AG-1: 42% and BG-1: 32%). It has a slightly higher density than the samples found at site 1.

Sample L4-B is from layer J in section B. This density sample has a mean density of 0.41 g/cm³ and a mean vesiculation of 83.5%. The standard deviation is ±0.05 g/cm³. The
density ranges from 0.21 – 0.84 g/cm³ with three outliers at 0.94, 1.31 and 1.42 g/cm³. The modal density peak is at ~0.38 g/cm³. This density sample has an overall higher density than section A, but both have very similar data ranges. About 4% of the sample consists of golden pumice with vesicularities above 90%, which is a decrease from sample A. Sample L4-B has a relative equal spread of A (43%) and B (40%) clasts within the sample. However, A clasts dominate the lower densities (<0.3 g/cm³). There is a significant amount of C clasts (17%) found at the higher density bins (>0.6 g/cm³).

Sample L4-C is from layer T in section C. This density sample has a mean density of 0.36 g/cm³ and a mean vesiculation of 86.4%. The standard deviation is ±0.039 g/cm³. The density ranges from 0.08 to 0.62 g/cm³ and potential outliers at 0.95, 1.03 and 1.11 g/cm³. The modal density peak is ~0.37 g/cm³. Sample L4-C has the lowest mean density out of the three sections at site 2. This sample is made up of 47% A clasts and 46% B clasts and have a relative equal spread of A and B clasts throughout the density bins. The sample is made up of 26% BG-2 clasts. About 6% of this sample is golden pumice with vesicularities above 90% and one of those clasts has a vesiculation of 97%.

**Site 3**

Samples L4-008, L4-010, L4-011A, L4-011B and L4-015 were collected from site 3. Sample L4-015 was deposited earliest in the eruption and L4-008 was deposited latest.

Sample L4-008 is from the top area of layer E. It has a mean density of 0.37 g/cm³ and a density range of 0.18-0.77 g/cm³, with potential outliers at 0.89, 0.94 and 1.14 g/cm³. It has a mean vesiculation of 86% and a modal density peak at ~0.25 g/cm³. It has a standard deviation of ±0.052 g/cm³. This sample contains 80 clasts and is the only sample that measures less than 100 clasts. Sample L4-008 is made up of 26% A group clasts, 66% B group clasts and 8% C clasts. About 14% of this sample is made up of golden pumice with vesiculation above 90%. Golden pumice, regardless of the vesiculation, make up more than 50% of the sample. Group A ranges mostly in the lower density bins (<0.2 g/cm³) and B groups increase as the density increases.
Sample L4-010 was collected in the golden pumice layer J and has a mean density of 0.35 g/cm³. Its density ranges from 0.19-0.70 g/cm³ (0.76 and 0.9 g/cm³ are potential outliers) and has a modal density peak at ~0.27 g/cm³. This sample has a mean vesiculation of 86.7% and ranges from 74-93%. The standard deviation is ±0.049 g/cm³. Sample L4-010 consists of 26% A group clasts, 70% B group clasts and 4% C group clasts. Golden pumice makes up 69% of the sample and 24% of the sample consists of gold pumice with vesiculation values above 90%.

Sample L4-011A was collected in the same pumice unit as L4-010 but 10m down south-east along the channel after the upper units (F, G, H, and I) had pinched out. This sample has a mean density of 0.32 g/cm³ and it ranges from 0.21-0.51 g/cm³. This is a much narrower data range compared to L4-010 and has a lower average density. The sample has a modal density peak at ~0.3 g/cm³. Its mean vesiculation is 87.9% and ranges from 81-92%. It has a standard deviation of ±0.035 g/cm³. L4-011A consists of 56% A clasts, 41% B clasts and 3% C clasts. It is made up of 39% gold pumice and only 7% of the sample is gold pumice with vesiculation greater than 90%. This sample has significantly less gold pumice compared to sample L4-010.

Sample L4-011B was collected in the same pumice unit (layer J) as L4-011A and L4-010 but it is on the opposite side (south side) of the channel and is adjacent to where L4-010 was collected. This sample has a mean density value of 0.36 g/cm³ and density values range from 0.18-0.55 g/cm³. Its modal density peak is at ~0.33 g/cm³ where more than 50% of the sample has density values between 0.30 and 0.39 g/cm³. This sample has a mean vesiculation of 86.6% and it ranges from 79-93%. It has a standard deviation of ±0.030g/cm³. L4-011B consists of 68% A clasts and 32% B clasts. Only 4% of gold pumice has a vesiculation above 90% and regardless of vesiculation, this sample consists of 16% gold pumice.

Sample L4-015 was collected in layer N on the south side of the channel. This sample has a mean density of 0.4 g/cm³ and a density range of 0.24-0.62 g/cm³. Its modal density peak is at ~0.37 g/cm³ and very little of the sample has densities between 0.2 and 0.29 g/cm³.
L4-015 has a mean vesiculation of 85.0% and the vesiculation ranges from 77-91%. More than 80% of the sample has vesicularities between 80-89%. It has a standard deviation of ±0.035 g/cm³. This sample is composed of 36% A clasts, 63% B clasts and 1% C clasts. L4-015 is quite rich in gold pumice with 47% of the sample composed of gold pumice. However, only a couple gold pumice clasts have vesicularities above 90%.

Various sizes of tephra particles (16.0, 1.00 and 0.125 mm) were observed under the binocular microscope. The bubbles spacing was relatively identical in these various fragment sizes. Indicating that the ash-sized clasts have the same or similar vesicularity to the lapilli sized clasts analyzed above.

<table>
<thead>
<tr>
<th>Sample #</th>
<th>ρM (g/cm³)</th>
<th>VM (%)</th>
<th>≥75% Ves.</th>
<th>≥85% Ves.</th>
</tr>
</thead>
<tbody>
<tr>
<td>L4-002</td>
<td>0.32</td>
<td>88.0%</td>
<td>98%</td>
<td>90%</td>
</tr>
<tr>
<td>L4-003</td>
<td>0.36</td>
<td>86.4%</td>
<td>95%</td>
<td>72%</td>
</tr>
<tr>
<td>L4-004</td>
<td>0.31</td>
<td>88.1%</td>
<td>99%</td>
<td>82%</td>
</tr>
<tr>
<td>L4-006</td>
<td>0.6</td>
<td>79.8%</td>
<td>70%</td>
<td>17%</td>
</tr>
<tr>
<td>L4-A</td>
<td>0.4</td>
<td>84.8%</td>
<td>96%</td>
<td>58%</td>
</tr>
<tr>
<td>L4-B</td>
<td>0.44</td>
<td>83.5%</td>
<td>93%</td>
<td>54%</td>
</tr>
<tr>
<td>L4-C</td>
<td>0.38</td>
<td>85.6%</td>
<td>97%</td>
<td>66%</td>
</tr>
<tr>
<td>L4-008</td>
<td>0.39</td>
<td>85.2%</td>
<td>93%</td>
<td>61%</td>
</tr>
<tr>
<td>L4-010</td>
<td>0.36</td>
<td>86.3%</td>
<td>97%</td>
<td>71%</td>
</tr>
<tr>
<td>L4-011A</td>
<td>0.33</td>
<td>87.4%</td>
<td>99%</td>
<td>88%</td>
</tr>
<tr>
<td>L4-011B</td>
<td>0.36</td>
<td>86.5%</td>
<td>100%</td>
<td>77%</td>
</tr>
<tr>
<td>L4-015</td>
<td>0.41</td>
<td>84.6%</td>
<td>97%</td>
<td>64%</td>
</tr>
</tbody>
</table>
**Site 1**

L4-002  
$\phi = 88.5\%$

L4-003  
$\phi = 87.2\%$
L4-004
\( \phi = 88.1\% \)

L4-006
\( \phi = 77.5\% \)
Site 2

L4-C
$\phi = 86.4\%$

L4-B
$\phi = 84.5\%$
**Site 3**

**L4-A**

$\phi = 85.2\%$

**L4-008**

$\phi = 86.0\%$
L4-010
φ = 86.7%

L4-011A
φ = 87.9%
Figure 6.13: Mean density and vesicularity measurements of samples from site 1, 2 and 3. Outlier clasts are indicated with black dots. SEM thin section samples (L4-04-09, L4-10-15, L4-15-20) are displayed on the graphs with a star.
6.3.2 SEM Image Analysis

There were 7 clasts chosen from 4 of the samples from site 1 and 3. These 7 clasts were made into thin sections, however due to time issues only 3 of the samples were analyzed using the SEM. The 3 thin sections represent clast 9 from sample L4-004 (L4-04-09), clast 15 from sample L4-010 (L4-10-15), and clast 20 from sample L4-015 (L4-15-20). Samples were chosen from site 1 and 3 because site 3 has the overall most detailed coverage in terms of grain and density measurements and site 1 and 3 need to be compared to determine what the correlation is for the NW and SE sides of the tephra cone. The clasts chosen are from the modal peak and most abundant componentry group in each sample. So L4-04-09 clast is an AG-1 and L4-10-15 and L4-15-20 are BG-2 clasts. Therefore, each clast gives a good representation of the particular sample as a whole. During the phreatomagmatic portion of the eruption, water rushed in to quench the system and tephra with a various range of vesicularities were ejected and deposited. Analyzing clasts from the modal peak of a sample give a better representation of the sample as a whole. Each sample was analyzed in ImageJ and presented in figure 6.14.

The thin sections were first observed through an imaging microscope to see any microscopic features and if any phenocrysts are present. Sample L4-04-09 is an AG-1 clast with a vesicularity of 89%. This clast is taken from L4-004 (from layer L in site 1), which has a mean vesicularity of 88%. Group A clasts have large (1-5 mm) angular vesicles and can have a broad range of bubble sizes. The brown/dark area represents the glass or magmatic melt and the white areas represent the bubbles or vesicles (Figure 6.15). L4-04-09 has large (<2.0 mm) round bubbles that are packed closely against other bubbles. Some bubbles take on a more polygonal shape due to the close proximity of the other bubbles. This kind of bubble packing resembles that seen in magma foam. Clast L4-04-09 also contains long chains of partly coalesced bubbles. These chains illustrate onset of coalescence, where bubble walls (= films) have partly retreated, illustrating that the relaxation time needed to complete coalescence exceeded the time needed for the glass transition (Costantini et al. 2010). Therefore, the rupture bubble film only retreated partially before the melt was quenched, which explains why the bubble chains still feature relatively round or indistinctly polygonal bubbles pressed together. This clast has small clusters of crystals within the melt. The crystals are mostly plagioclase and pyroxene.
Figure 6.14: 50x, 100x, and 250x sample images produced from ImageJ. L4-04-09 has a vesiculation of 89%. L4-10-15 has a vesiculation of 92%. L4-15-20 has a vesiculation of 86.7%. 24-07-04-07(1) has a vesiculation of 84.2%
Figure 6.15: Sample L4-04-09 showing polygonal bubbles. Some bubbles have coalesced but others still have their bubble walls intact. Brown and darker areas are glass (i.e. melt), and white areas are bubbles (or vesicles). The object with the speckled domain (top left) is a plagioclase crystal. Sample has a vesiculation of 89%.

Sample L4-10-15 is a golden pumice (BG-2) clast with a vesicularity of 92%. This clast was taken from sample L4-010, which is from layer J in site 3 and has a mean vesicularity of 86.7%. B group clasts have smaller more uniformly shaped bubbles. Golden pumice clasts have a high abundance of small (0.5-2 mm) closely packed vesicles. Golden pumice clasts are often smaller in size (<2.5 cm) and well rounded. This sample has an abundance of polygonal bubbles (<0.75 mm) that are closely packed together (Figure 6.16). With a vesiculation of 92% then this sample is very mature and represents more “foam-like” conditions in the melt. Clast L4-10-15 has an abundant amount of bubble chains throughout the sample and in between these bubble chains there are smaller bubbles that are squished and have an elongated flattened appearance (Figure 6.17). This sample does not have many crystals within the melt.
Figure 6.16: Sample L4-10-15 has smaller bubbles compared to sample L4-04-09. Bubbles are polygonal in shape and are packed tightly due to its high vesiculation of 92%.

Figure 6.17: Sample L4-10-15 has abundant bubble chains within the entire sample.

Sample L4-15-20 is a golden pumice clast (BG-2) with a vesicularity of 86.7%. This clast is from sample L4-015, which is from layer N in site 3 and has a mean vesiculation of 85%. Bubbles are densely packed together giving most of them polygonal or flattened appearances. Most bubbles are quite small (<1.0 mm) and there are only a few large bubbles (>2.0 mm) Some bubbles have completely coalesced with only faint indications of a former bubble film left over (Figure 6.18). Bubbles are much smaller and bubble chains are less frequent than in the other samples. The bubble chains occur randomly throughout
the sample and do not have a preferred orientation. Clast L4-15-20 contains some crystals but not a significant amount to distort the entire clast (Figure 6.19). The large plagioclase phenocryst featured in figure 6.19 seems to distort the bubbles around it slightly due to the angular appearance of some of the bubbles and “pinching” of the bubble walls.

**Figure 6.18:** Sample L4-15-20 showing elongated bubble chains and polyhedral bubbles. Bubbles are closely packed together giving a rounded polygonal shape. Sample has a vesicularity of 86.7%.

**Figure 6.19:** Sample L4-15-20 contains larger crystal phenocrysts. Bubbles seem slightly distorted around the large plagioclase phenocryst.

Clast 24-07-04-07(1) is a sample from the P1 tephra unit and has a vesicularity of 84.2%. This clast is from a sample with a mean vesicularity of 86% and a range of 64-94%. Clast
24-07-04-07(1) contains a small amount of crystals and the bubbles within the sample are significantly less round compared the other samples. The bubbles have “pinched” edges from being distorted somehow and most of the bubbles have coalesced to make very long and irregularly shaped bubbles. Most bubbles are less than 1.5 mm and are not close to one another so the amount of melt exposed in this sample is greater than the other samples.

Vesicle Size Distribution

The VSD graphs are displayed in figure 6.20 (a-d). Sample 24-07-04-07(1) presents a relatively smooth curved trend with a negative logarithmic slope (Figure 6.20a). There are a couple small ‘pick-ups’ and a more obvious kink in the curve leading to the conclusion that this curve most likely represents multiple nucleation events or power-law trend (=continuous growth and nucleation). The term ‘pick-ups’ is used to describe small perturbations in the trend line where there is a sudden increase in the amount of bubbles then immediately following is a decrease back to the general trend of the line. The multiple nucleation events are evident in the trend line because until about a vesicle size of 0.25mm the slope is quite steep and then after 0.25mm the slope decreases dramatically, giving the trend line a ‘kinked’ appearance. This sample has a large population of bubbles in the small bubble size bins (<0.018 mm).

Sample L4-04-09 has a relatively similar trend to sample 24-07-04-07(1) with an obvious kink in the trend line at 0.5mm hinting at multiple nucleation and growth events super imposed on each other or power-law trend (=continuous growth and nucleation; Figure 6.20b). Sample L4-04-09 has a smaller amount of bubbles, in general, compared to the other clasts, and therefore a lower bubble size distribution.

Clast L4-10-15 features a flatter distribution with a shallower slope (Figure 6.20c). It has bubble number densities rivalling that of sample 24-07-04-07(1). Clast L4-10-15 also features a very steep vertical slope in the smallest size bins. This means that it has a large frequency of small bubbles (<0.018 mm) compared to any other size, at the time the clast was quenched. Clast L4-10-15 displays the most linear slope out of all the samples resembling either a single nucleation event trend or multiple nucleation events. It is not a
perfect “single nucleation event” slope, however, this may show how influential the abundant smallest bubble population is on the trend line.

Figure 6.20: VSD trends indicating kinetics of nucleation, growth and coalescence within the magmatic system. (a) Sample 24-07-04-07. (b) Sample L4-04-09. (c) Sample L4-10-15. (d) Sample L4-15-20.

Sample L4-15-20 has a slightly jagged trend line with a few ‘pick ups’ and kinks (Figure 6.20d). This kind of a trend is indicative of various stages of nucleation and growth or its relatively curved appearance mimics that seen in power-law (=continuous growth and nucleation). This sample has a large population of bubbles in the small bubble size bins (<0.018 mm).
**Vesicle Volume Distribution**

Vesicle volume distribution (VVD) is commonly used to describe the nature of nucleation and/or coalescence events during bubble formation in the melt (Shea et al. 2010). VVD graphs volume fraction and volume fraction adjusted vs. vesicle size (mm). Figure 6.21 (a-d) show the VVD graphs of the four samples studied.

Clast 24-07-04-07(1) features, to the first order, a bimodal VVD, with clear modes at 0.14 mm and between 0.7, 1.12 and 1.8 mm. However, the latter mode does not feature a distinct peak, rather rhythmic alternation of higher versus lower vesicle volumes across six bins (Figure 6.21a). Hence, the distribution can be viewed as polymodal, three additional modes at 0.7, 1.12 and 1.8 mm. This sample’s trend exhibits evidence of multiple nucleation and growth events due to its wide distribution and multiple peaks. Also, the VVD in clast 24-07-04-07(1) is broad, yet dominated by rather large vesicles as demonstrated by the prominent 0.7 to 1.8 mm mode. These vesicle sizes are significantly larger than those that typify silicic pumices (e.g. Carey et al. 2009), but similar to that observed for golden pumices from the 1959 Kilauea Iki eruption (Stovall et al. 2011).

Clast L4-04-09 features a bimodal VVD, with clear modes between 0.28 to 0.44 mm and 1.8 mm. However, similar to the previous sample, the first mode in L4-04-09 does not feature a distinct peak, rather rhythmic alternation of higher versus lower vesicle volumes across four bins (Figure 6.21b). Therefore, the distribution can be viewed as polymodal, with modes at 0.28, 0.44 and 1.8 mm. This clast features a narrower vesicle size range and most of its vesicle size population is clustered around 0.14-0.7 mm range. This sample does have some vesicle populations around the larger sizes (~3.5 mm), but has very little amount of vesicles with sizes less than 0.06 mm. Clast L4-04-09 exhibits evidence of multiple nucleation and growth events due to its wide distribution and multiple peaks, as well as, weakly resembling power law (=continuous growth and nucleation events) due to the concentrated distribution of vesicle sizes around a common size bin giving the bar graph a narrow distribution. These vesicle sizes are very similar to those that typify silicic pumices (e.g. Carey et al. 2009).

Clast L4-10-15 features a polymodal VVD with clear modes at 0.35, 0.7 and 1.12 mm. However, the first two modes feature a relatively indistinct trough between them leading
this distribution to be considered as weakly bimodal (Figure 6.21c). This clast features a narrower vesicle size range (compared to the other samples) and most of its vesicle size population is clustered around 0.28-0.7 mm range and a maximum vesicle size of 1.8 mm and very little amount of vesicles with sizes less than 0.06 mm. Clast L4-10-15 exhibits evidence of multiple nucleation and growth events due to its multiple peaks, as well as, resembling power law (=continuous growth and nucleation events) due to the concentrated distribution of vesicle sizes around a common size bin giving the bar graph a narrow distribution. These vesicle sizes are very similar to those that typify silicic pumices (e.g. Carey et al. 2009).

Figure 6.21: VVD trends displaying volume fraction in black and volume fraction adjusted in red. (a) Sample 24-07-04-07. (b) Sample L4-04-09. (c) Sample L4-10-15. (d) Sample L4-15-20.

Clast L4-15-20 features a polymodal VVD, with clear modes at 0.13, 0.22 to 0.35, 0.7, 1.12 and 1.8 mm. The second mode is a very broad peak covering three size bins and the peaks of the higher vesicle sizes are quite small (Figure 6.21d). Therefore, the distribution is polymodal with the most prominent peaks at 0.13, 0.22 to 0.35 and 0.7 mm. This clast features a narrower vesicle size range (compared to the previous samples minus L4-10-15) and most of its vesicle size population is clustered around 0.089-0.7 mm range and a maximum vesicle size of 1.8 mm and very little amount of vesicles with sizes less than
0.035 mm. Clast L4-15-20 exhibits evidence of multiple nucleation and growth events due to its multiple peaks, as well as, weakly resembling power law (=continuous growth and nucleation events) due to the concentrated distribution of vesicle sizes around a common size bin range giving the bar graph a narrow distribution. These vesicle sizes are very similar to those that typify silicic pumices (e.g. Carey et al. 2009).

**Vesicle Number Density**

Vesicle number density ($NV_m$) values are displayed in table 6.6. The sample clasts from P1 Laki are graphed in figure 6.22 along side other famous eruptions; Tarawera, Kilauea Iki and Etna. Laki has an average effusive rate of about $1.4e^7$ kg/s during the first five episodes (Thordarson and Self 1993), which falls between effusive rates from Kilauea Iki and Etna but closer to Etna. Etna 122BC was a basaltic Plinian eruption featuring high bubble number densities and fast magma ascent rate (Sable et al. 2006). Bubble density values of Laki P1 phase are also much closer to those seen in Etna. Grímsvötn 2011 eruption (Lynch 2015) has similar effusive rates to Laki and Etna ($\sim10^7$ kg/s) and slightly higher vesicle number densities ($\sim10^7-10^8$ cm$^3$) and is also classified as subPlinian to Plinian. Tarawera is the youngest known Plinian eruption of high intensity and high magma effusion rates (Houghton et al. 2004).

**Table 6.6: Vesicle number densities**

<table>
<thead>
<tr>
<th>Sample #</th>
<th>$NV_m$ (cm$^3$)</th>
</tr>
</thead>
<tbody>
<tr>
<td>L4-04-09</td>
<td>8.06E+06</td>
</tr>
<tr>
<td>L4-10-15</td>
<td>2.14E+07</td>
</tr>
<tr>
<td>L4-15-20</td>
<td>3.30E+07</td>
</tr>
<tr>
<td>24-07-04-07(1)</td>
<td>6.55E+07</td>
</tr>
</tbody>
</table>
Figure 6.22: Vesicle number density and mass discharge rate of various eruptions. (Source: Houghton & Gonnermann 2008).
7 Discussion

There is no mistaking that the fourth episode during Laki eruption was phreatomagmatic. The deposits are rich in ash and first hand accounts report seeing a large black billowing plume rise up on June 27\textsuperscript{th} 1783 during the fourth episode. The external water quenching the magma is most likely from lake Lambavatn, which is located just NW of the tuff cone. Numerous faults propagate through Lambavatn and could potentially allow cool fluids to reach the magma before it reaches the surface.

7.1 Density and Depth of Quenching

The density samples from the phreatomagmatic phase have a very high and narrow range of vesicularities (~70-95\%) and mean vesicularities (~84-88\%). Usually phreatomagmatic eruptions have a wide range of vesicularities (~40-50\%) and with much smaller mean values (<70\%; Moreland & Thordarson 2014) because water can quench the magma at a wide range of depths and before it has fully vesiculated. The P1 samples are similar to magmatic tephra (i.e. 72-86\% Mount St. Helens; Klug and Cashman 1994) and the Magmatic tephra from Laki (Sample L4-006 ~77\% ves.). Therefore, P1 tephra has high vesicularities because it was able to reach shallow conditions (~1 atm) similar to magmatic systems and fully fragment before quenching occurred. High vesicularity values like this are representative or true magmatic foam. Since foam has such a high yield strength then it is not a suitable environment for FCI processes to occur (Lynch 2015). Therefore, water had no effect on the explosivity of the phreatomagmatic phase from fissure four. The water only quenched the magma and most likely lead to cooling-contraction granulation of fragile clasts (i.e. golden pumice) leading to additional fragmentation and increasing ash content in the deposits. FCI processes will lead to explosive eruptions when external water interacts with magma at depth. We have seen from the vesicularity measurements from Laki that was external water interacts with fully vesiculated magma near the surface then the water does little to aid in whether or not the eruption will be explosive and instead leads to the ash rich component in the deposits.
Using the binocular microscope, it was evident that the bubbles spacing in grains of various sizes (16, 1.0 and 0.125 mm) were nearly identical. Indicating that not only fine lapilli clasts, but also the ash particles all have the same vesicularity and therefore, are all part of the same bulk area in the magma that was quenched. To produce this and the very high vesiculation then the magma near the surface was the only part of the magma in the conduit to be quenched, most likely from surface water or by Lake Lambavatn rather than ground water, which likely would have quenched the magma over a wider range.

7.2 Stratigraphy and Course of Events

The stratigraphy of the Laki deposits, in general, consist of alternating ash rich (<1.4 mm) with dominant modes in the very fine ash region (0.063 mm) and lapilli/ash rich (<22.6 mm) with dominant modes in the course ash (1.0 mm) and fine lapilli region (2.0 mm).

Site 1, which is located on the NW side of tuff cone 1, consists of alternating layers of ash (course to fine) and lapilli/ash (course lapilli to fine ash) rich layers. The entire P1 section at this location is about 0.7 m, which is the thinnest P1 unit out of the three sites in this study and it is also the farthest from the tuff cone (~350 m). The tephra layers in site 1 are on average about 2-10 cm thick. Site 1 samples show a coarsening upwards pattern for the main modes (0.35 to 1.0 to 2.8 mm) as well as the secondary modes and decreasing of ash content (94 to 56 to 45 %). All samples have a secondary mode at 0.125 mm.

Site 2 is located about 150 m NW of the Laki tuff cone. It consists of thick (5-20 cm) alternating ash (course to fine) rich and lapilli/ash (course lapilli to fine ash) rich layers. The entire P1 section at this location is at least 3.5 m because we don’t know exactly how far this unit reaches and sample L4-A001 grain size trend is most likely due to sampling error. Site 2 samples exhibit a cyclical pattern consisting of alternating layers of; 1. Very course ash (1.0 to 1.4 mm) mode and a fine to very fine ash (0.031-0.125 mm) secondary mode and are usually bimodal samples with ash content between 60-85% and 2. Very course ash to medium lapilli (1.4-22.6 mm) and a very fine ash to medium lapilli (0.044-22.6 mm) secondary modes and are polymodal with ash contents between 48-70%. All samples have a mode at 0.125 mm.
Site 3, which is located about 200 m SE of tuff cone 1, consists of very similar alternating layers to the previous two sites, however, this site is overall more ash (course to very fine) rich and most samples show a relatively bimodal trend with very wide distributions (0.022-16.0 mm). The layers within the stratigraphy are about 10-50cm thick and the entire P1 unit is more than 3.5 m thick. Site 3 samples are the most fine-grained samples with rich abundance of ash within each layer. The main modes are course to very fine ash (0.044-1.4 mm). Most (but not all) of the samples have a fine to medium lapilli (5.66-16 mm) secondary mode. All samples have a mode at 0.125 mm. The samples within this site do seem to experience cycles from courser to very fine secondary ash modes. The samples here have the largest amount of ash compared to the other sites. The only way for ash to be deposited so close is by ash aggregation. Ash aggregates are made from fragmentation in the magma. We will talk about its significance to the eruption later.

In site 1, sample L4-004B seems out of place among the courser rich samples in site 1 and 2. This fine ash rich sample has a 94% ash content and similar ash-rich grain size trend to those seen in site 3. The grain size trends, when comparing the NW and SE sides of the tuff cone have very little visual similarities except for the beginning few layers of site 1 (L4-004B), which has a very similar trend and ash content to the SE side of the tuff cone. I propose that site 3 deposits were deposited first at the beginning of the eruption when the wind was blowing to the east on June 27th 1783 and then the wind shifted on June 30th and blew towards the west to deposit site 1 and 2. Sample L4-004B exhibits the last fine-grained pumice phase and then the eruption became more lapilli rich.

Overall, the erupted deposits (SE then NW deposits) reveal a coarsening with time pattern. There are two reasons to describe why this process could be occurring. One, there was increased fragmentation at the beginning of the eruption leading to the disintegration of the magma into more ash rich particles. A more vigorous eruption (i.e. beginning of an eruption) would have increased gas available to perform secondary re-fragmentation of the larger pumice clasts to produce deposits rich in ash (Girault et al. 2014). Additional fragmentation also favours ash aggregation, which would’ve allowed for the fine ash to be deposited so close the tuff cone at such high concentrations. A second reason could be that external water is quenching the magma and causing larger more fragile clasts (i.e. golden pumice) to disintegrate via cooling-contraction granulation (Kokelaar 1986).
Overall, the P1 samples have an almost identical vesicularity spread to the magmatic samples and the grain size sample trends are also relatively the same. Magmatic grain size trends (i.e. sample L4-006) have a mode of medium lapilli (5.66 mm) and a tail end with very little ash. Most of the grain size graphs from the P1 samples have the same course fraction but additional modes in the ash range (<2.00 mm). This leads us to conclude that since the only difference between the magmatic and phreatomagmatic samples is the addition of water then this fine fraction must have been the result of cooling contraction granulation via external water interaction and secondary fragmentation of the vesiculated clasts and melt near the surface.

7.3 Vesicle Distribution and Magma Ascent

VSD trends reveal that the four samples (24-07-04-07(1), L4-04-09, L4-10-15 and L4-15-20) all have increasingly high abundance of smaller (~0.009-0.018 mm) bubbles and trends show evidence for either multiple nucleation events or power law. The high vesicularities and high abundance of bubbles residing in the smallest bubble sizes (0.009-0.018 mm) could indicate possible burst nucleation from high magma ascent rates leading to little time for bubbles to grow and expand in the melt.

VVD trends display that all clasts are relatively bimodal (less so L4-15-20) with dominant mode in the lower bins sizes (0.14-0.6 mm) and then a secondary mode at about 1.4 mm. Bubbles number densities from these samples range between $10^6$ and $10^7$ cm$^3$ and are very similar to those found at Etna 122BC eruption and Tarawera. Etna 122BC VVD trends revealed bimodal distributions due to early degassing of CO$_2$ and burst nucleation of H$_2$O. Also, Etna revealed high vesicle number densities and small bubble sizes likely indicating fast magma ascent rates, most likely due to the early degassing of CO$_2$ (Sable et al. 2006). Phreatomagmatic phase 1 at Laki has relatively high bubble number densities and most of the bubbles are quite small in size (~0.009-0.018 mm) and display bimodal trends. High bubbles number densities indicate rapid bubble nucleation and, in addition, small bubble sizes means the bubbles had limited opportunity to grow. This can occur during high magma ascent rate, similar to that seen in Etna 122BC. I propose that the larger bubble mode ~1.4 mm is from early degassing of CO$_2$ causing the magma to accelerate and
induce burst nucleation from H$_2$O saturation making the magma accelerate further. This could also explain the intense ‘jumps’ in the bubbles mode populations (in the VVD graphs) instead of gradually increasing the number of bubbles over a few bubble size bins.

High bubble number densities are also indicative of increased eruption intensity. The values for $NV_M$ decrease going from L4-15-20 to L4-10-15 to L4-04-09, which reiterates that clast L4-15-20 or the first deposits in site 3 are representative of a more energetic part of the eruption (i.e. the beginning). High bubbles number density paired with the high effusive flow ($10^7$ kg/s) gives very similar values to Etna 122BC, Tarawera and Grímsvötn 2011. Leading us to conclude that the phreatomagmatic phase and the Laki eruption, in general, was of more subPlinian-Plinian intensity.

### 7.4 Summary of Processes

Using the data from this study we have determined a timeline for the phreatomagmatic phase 1 of episode four during the Laki eruption. CO$_2$ began degassing prior to 1km depth, causing the magma to accelerate inducing burst nucleation of H$_2$O from the melt. Bubble sizes are weakly bimodal with larger CO$_2$ bubbles and various sizes of H$_2$O bubbles (~0.009-0.018 mm) collecting near the surface (~1 atm). Bubbles grow and begin to deform against one another and becoming foam-like and fully vesiculated. Water quenches the vesiculated clasts and melt, causing cooling-contraction granulation of more fragile medium to fine lapilli clasts (golden pumice) to form a high concentration of ash. Earthquakes are felt, fissure four opens up and the phreatomagmatic phase 1 begins erupting on June 27$^{th}$ 1783 with the wind blowing to the east. An enormous black ash plume rose up high, billowing to the east. Fragments from the eruption were blown and deposited on the SE side of the tuff cone. The additional fragmentation formed ash aggregates, which were able to deposit in close proximity to the tuff cone. Due to the increased intensity of the beginning of the eruption then the deposits were ash rich and with high concentrations of golden pumice. Lava flowed out from the Skáfta River Gorge on June 29$^{th}$. The wind changed directions on June 30$^{th}$ towards the west and deposited the ash rich fragments on the NW side of the tuff cone. The supply of water began to dwindle causing the fine ash deposits to decrease within the layers on the NW side of the tuff cone.
8 Conclusion

8.1 Summary

The Laki fissure eruption lasted from June 8\textsuperscript{th} 1783 to February 7\textsuperscript{th} 1784 and was one the most deadly and well documented eruptions in history. Ash and lower temperatures were observed all over Iceland, as well as, various countries in the northern hemisphere including parts of Europe, China and Canada. The Laki eruption featured 10 separate events marked by tremors and the opening of a new fissure.

In this study we sought out to answer the following questions about the phreatomagmatic (P1) eruption from fissure 4:

- Is the higher ash grade a reflection of more intense and powerful explosive activity during the phreatomagmatic phases?
- How significant a role does explosive magma-water interaction play in driving the intensity of the eruption?
- How much of the phreatomagmatic tephra was highly vesiculated at the time of magma to water interaction and what does that indicate for the level/depth within the conduit for the interaction?
- What principal mechanism governs magma vesiculation in the phreatomagmatic phases at Laki?

We determined that due to the contrasting stratigraphy, the phreatomagmatic phase began on June 27\textsuperscript{th} 1783 when the enormous black plume was seen. The wind was blowing to the east depositing samples on the SE side of tuff cone 1 (TC1) first. Then the wind changed on June 30\textsuperscript{th} to deposit samples on the NW side of TC1. The grain size samples show that the eruption deposits coarsened through time due to decreasing availability of external water leading to cooling-contraction granulation.

Higher ash content is not a reflection of more intense activity. The ash content was a product of late stage quenching due to external water causing cooling-contraction granulation of the already vesiculated tephra clasts.
Magma-water interactions played an insignificant role with regards to the P1 phase because the magma was already highly vesiculated near the surface when the external water quenched the system. The eruption was already going to be explosive due to the expansion of gases, burst nucleation and high magma ascent rates.

The phreatomagmatic tephra was able to fully vesiculate near the surface and most of the clasts within each sample set (>93%) had vesicularities above 75%. Ash particles also display evidence for having high vesicularities similar to those found in the lapilli clasts. This indicates that the only magma that was quenched was the bulk of the magma that was residing very near to the surface and was able to fully vesiculate. This magma was likely quenched by surface water or Lake Lambavatn rather than ground water.

The principle mechanism that governs magma vesicularity is the magmatic expansion of gases at depth in the conduit. External water had very little to do with the vesiculation of the magma because the magma-water interaction was occurring at such shallow conditions.

Therefore, the effect that external water has on the intensity of an eruption depends on the depth of magma-water interactions. We have seen that FCI processes do not occur when water interacts with magma that has fully vesiculated near the surface (~1 atm). Instead, late stage quenching takes on a more passive role leading to the increased ash content by cooling-contraction granulation.

8.2 Concluding Remarks

The next steps for this study are to make a detailed total grain size distribution of the phreatomagmatic tephra from fissure four by using all the samples collected by Thorvaldur Thordarson and myself, in order to fully understand the fragmentation processes in the melt. As well as, collecting more data on bubble distribution in the melt.

I hope that this study leads to redifining the nature of magma-water interactions in phreatomagmatic eruptions and understanding the complexity of the Laki fissure eruption.
References


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Appendix A

Site 1

L4-001

L4-004
Site 2
Site 3
Appendix B

Component Subgroups of A, B, and C L4-002

Component Subgroups of A, B, and C L4-003
Component Subgroups of A, B, and C L4-015

Density (g/cm³) vs. Frequency

- AG-1
- AG-4
- BG-1
- BG-2
- BG-3
- CG-3