Ash Generation in the 2012 Eruption of Havre Volcano, Kermadec Arc: The Largest Deep Subaqueous Eruption of the Last Century

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Abstract

The 2012 silicic eruption at Havre volcano, in the Kermadec Arc, was the largest deep subaqueous eruption observed in the last century. A data set of unprecedented richness was collected during a dedicated research cruise in 2015, including detailed bathymetric mapping and systematic in situ sampling of seafloor clastic and effusive products. This thesis characterizes the seafloor Ash and Lapilli (AL) Unit produced during the 2012 Havre eruption, with the aim of determining the effect of the water column on ascent, fragmentation, and dispersal of ash during a deep silicic subaqueous eruption. To this end, sample grain-size distributions, sample componentry, ash shape and microtextural data, major-element chemistry, and groundmass volatile contents were acquired. Results and interpretations from the AL Unit support inferences on eruption processes.

It is demonstrated that the AL Unit is a composite deposit composed of four subunits; from base to top these are Subunit 1 (S1), 2 (S2), 3 (S3), 4a and 4b (S4). Each of the subunits in the AL Unit shows distinctive grain size or componentry characteristics, different mapped dispersal limits, and specific stratigraphic relationship with the other seafloor products of the 2012 Havre eruption. Using results of subunit depositional characteristics and particle microtextural features mechanisms are inferred to explain the generation of each subunits of the AL Unit.

Subunit 1 directly overlies the Giant Pumice Unit, draping the entire study area and fining towards the NW. This deposit is composed of an average 125 to 800 µm glassy vesicular ash showing dominantly curvi-planar morphologies, in addition to lesser amounts of angular and fluidal particles. Subunit 1 is therefore inferred to have been deposited by fallout following dispersal in an eruptive plume. The plume was driven by an eruption defined by energetic fragmentation with a large component of magma water interaction, however also apparently showing a range of other fragmentation processes.

Subunit 2 overlying S1 across a gradational contact shows a deposit boundary along the northern caldera wall. To the south of the boundary S2 is heavily thickened in the caldera showing a consistent
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grain size. Subunit 2 is composed of 16 to 32 µm glassy vesicular ash showing dominantly curvi-
planar morphologies, in addition to lesser amounts of angular and fluidal particles. This subunit is inferred to have been deposited from dilute density currents that ponded in the Havre caldera. The similarity in microtextural features to S1 and their gradational contact suggest these two subunits were generated from the same event. With density currents potentially generated off a larger eruption column. The microtextural similarity of these deposits to the GP Unit and ALB Unit suggests their eruption from the dome OP vent, while the presence of fluidal particles and energetic fragmentation indicates and explosive eruption.

Subunit 3 drapes topography in a NE-SW trend across the caldera thinning and fining towards a lava flow on the southern caldera rim. The morphological and microtextural similarity of the ash the S3 is composed of to the pumiceous carapace of the said lava suggests this was its source. By modelling the thermal plume required to generated S3 however it is shown that weakly pyroclastic activity is required to produce the wide dispersal, likely occurring synchronously with lava effusion.

Subunit 4 is composed of microcrystalline ash, the low vesicularity and high crystallinity of which suggests fragmentation from the lava flows. Subunit 4a dispersed in a NE-SW trend across the caldera is inferred to have been generated during a caldera wall collapse near the source vents of 3 lavas produced during the 2012 eruption. Subunit 4b dispersed around Dome OP in inferred to have been generated by quenching and brecciation of the lava as it was extruded.

The results presented in this thesis show that the 2012 deep subaqueous eruption of Havre volcano was a complex event, with both explosive and effusive activity occurring over several phases. The eruptive processes were significantly influenced by the water column, which affected magma rheology, magma fragmentation to produce fine ash, initial particle dispersal, and final deposition.
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Thesis Introduction
Chapter 1: Introduction

1.1. Overview

The thesis introduced here presents a study of volcanic ash collected from the seafloor, produced during the 2012 Havre submarine eruption. The 2012 eruption of Havre volcano was the largest deep subaqueous silicic eruption of the last century (Carey et al. 2014, 2018). I participated in a focused follow-up cruise, conducted in 2015, which produced a detailed sample set of both effusive and clastic seafloor products, along with in situ observations and images (Carey et al. 2018). The seafloor ash samples from the 2015 cruise are the first collected following a submarine silicic eruption at a known time, from a known location, and the first sampled in situ by a submersible in a stratigraphic context. The Havre sample set is globally significant in its completeness, how fresh it is, the fact that all samples were taken in situ, and the known eruptive source.

The abundance of ash discovered on the seafloor was unexpected. The paradigm of submarine eruptions is that explosivity will be suppressed at depth where there is high confining water pressure (Fisher and Schmincke 1984) resulting in little fine ash been produced (Wohletz and Sheridan 1983). This is backed up by observations of ‘fines flushed’ uplifted subaqueous volcanic successions (Kano et al. 1996; Allen and McPhie 2000, 2009; Kano 2003). However, the extremely fine grained ash observed at Havre indicates intensive fragmentation (Walker 1981; Zimanowski et al. 2003; Rose and Durant 2009). In order to learn about the 2012 Havre eruption, I have used the ash samples to establish a stratigraphic framework. In addition, to understand why the eruption produced this abundance of fine ash, in this study I have subsampled the ash, identified different morphological populations using various techniques, made comparisons with other volcanic ash, and produced new ash by re-melting and fragmenting Havre rocks in a lab in Germany. My conclusions, as supported in the following chapters, are that the ash formed in several phases with different processes operating in each. The initial phase was a high mass flux explosive event during which primary fragmentation was driven by both magmatic and hydromagmatic fragmentation. Deposition occurred both through fallout of ash through the water column, and by deposition from dilute density currents. The second phase occurred concurrently with lava extrusion, initially with weakly pyroclastic syn-effusive ash.
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venting driving a high convective plume from which a fallout deposit was produced. Following this, fragmentation was driven by mechanical break up and magma water interaction generated by the collapse of the caldera wall under three lavas.

One of the major remaining uncertainties is the processes by which a population of fluidal silicic ash grains were produced. Some initial hypothesis addressing this are presented, however further work involving examination of the volatile contents of the fluidal particles is required to answer the questions posed by their presence. In addition, all the ash deposits so far identified at Havre extend beyond the study area, so the total volume of material produced during the eruption is highly uncertain. Additional in situ seafloor sampling would be required to fully constrain the full extent and thickness of these deposits, the results of which would be required for modelling of eruption processes.

1.2 Submarine volcanism

Submarine volcanism accounts for approximately 75% of all volcanism occurring on the Earth today and throughout time since the early formation of the continents (Crisp 1984; Mueller and White 1992; White et al. 2003; Furnes et al. 2015). Our understanding of the dynamics of submarine eruptions lags compared to that of subaerial volcanism. Our comparatively limited understanding is a consequence of the lack of direct observations and the high costs associated with in situ sampling and ‘field work’. However, in recent decades significant contributions have been made to the study of subaqueous volcanism, by examining uplifted successions (Kokelaar 1986; Kano 2003; Allen and McPhie 2009) and seafloor deposits (Allen et al. 2010; Schipper et al. 2010; Rotella et al. 2013), along with numerical (Head and Wilson 2003) and experimental studies (Manville and Wilson 2004; Schmid et al. 2010; Verolino et al. 2017). Recent observations of subaqueous volcanism at NW-Rota and W-Mata at depths of ~550 mbsl (Chadwick et al. 2008a) and 1200 mbsl (Resing et al. 2011) have significantly advanced the understanding of such events, and are corner stones in our understanding of subaqueous eruption dynamics.
Table. 1.1. A comparison of the major thermo-physical properties of air, water, and steam from (White et al. 2003, 2015). Large differences can be observed between water and air/steam that lead to contrasting eruption behaviour.

One important component of all subaqueous volcanic eruptions is the interaction with ambient water that occurs to a greater or lesser extent. The effect of water on the magma can occur indirectly due to the large differences in viscosity and density of water relative to air, as well as during direct contact as a result of differences in water’s thermal properties (White et al. 2003, 2015; Cas and Giordano 2014) (Table. 1.1.). The change in the thermo-physical properties of the ambient environment can impact fundamentally on eruption processes in the shallow conduit, fragmentation mechanisms, along with transport and dispersal processes (Wohletz 1983; Cashman and Fiske 1991;
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Zimanowski et al. 1997b; Head and Wilson 2003; Murtagh and White 2013; Rotella et al. 2015). Here, I provide a brief outline of water’s effects on the dynamics of volcanism and highlight areas and questions that remain unanswered in the study of subaqueous volcanism.

1.2.1 Impact of the water column on submarine volcanism

Compared to air, water is a vastly different physical medium (Table. 1.1.). Water has a density and viscosity 1000 and 100 times larger than that of air, respectively (White et al. 2003, 2015; Cas and Giordano 2014). Water also has a specific heat capacity and thermal conductivity four and 30 times larger than that of air respectively, at similar conditions (White et al. 2003, 2015; Cas and Giordano 2014). In addition, at expected eruption pressure-temperature (PT) conditions, water can undergo a phase change to steam, which involves a volumetric expansion of up to 1700 times at shallow depths (White et al. 2003, 2015; Cas and Giordano 2014).

Direct magma water interaction is the result of contact between hot magma and liquid water (Kokelaar 1986; Wohletz 1986; Zimanowski et al. 1997b; Büttner et al. 1999, 2002; Austin-Erickson et al. 2008; Liu et al. 2015). This drives a range of fragmentation mechanisms unique to this process that result from the rapid transferal of thermal energy (Wohletz and Sheridan 1983; Kokelaar 1986). The high heat fluxes result in rapid volumetric change of both the magma and water (Wohletz 1983; Kokelaar 1986; Zimanowski et al. 1997a; Mastin et al. 2004; van Otterloo et al. 2015). Cooling of the magma on direct contact with water can generate high thermal stress in the melt due to rapid volumetric changes across the glass transition (Wohletz 1983; Giordano et al. 2005; Patel et al. 2013; van Otterloo et al. 2015). If the thermal stress exceeds the fracture toughness of the glass, fracturing and fragmentation is induced (e.g. Irwin, 1957). Energetic fragmentation can be generated by magma water contact by the sudden collapse of insulating steam films (Wohletz 1983; Zimanowski et al. 1997b; Austin-Erickson et al. 2008). Rapid contraction and fracturing of the melt, generates and drives seismic waves deeper into the magma body, behind which fractures propagate (Wohletz 1983; Zimanowski et al. 1997b; Austin-Erickson et al. 2008). Water infiltration into the cracks generates
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further fracturing as more surface area is exposed, resulting in a runaway fragmentation process driven entirely by the interaction of magma and external water (Wohletz 1983; Zimanowski et al. 1997b; Austin-Erickson et al. 2008). Water vaporisation during this process then leads to expansion of the system and particle dispersal producing explosive magma water interaction, also referred to as molten fuel coolant interaction (MFCI) (Wohletz 1983, 2003, Zimanowski et al. 1997a, b; Austin-Erickson et al. 2008). Propagation of shock waves, and system expansion upon vaporisation can cause fragmentation of a much larger volume of melt than that in direct contact with water (Zimanowski et al. 1997b; Austin-Erickson et al. 2008). Magma water contact may not result in explosive fragmentation. Quench fragmentation is a more passive magma water interaction process, that results from the propagation of cracks into the melt because of volumetric changes across the glass transition, fracturing however does not generate seismic waves meaning the process is not explosive (Carlisle 1963; Kokelaar 1986; Patel et al. 2013; Gonnermann 2015; van Otterloo et al. 2015). Magma water interaction may drive fragmentation or assist it in a primarily magmatic eruption (Wohletz and Sheridan 1983; Mastin et al. 2004; Liu et al. 2015).

The dynamics of magma water interaction processes can be studied experimentally with simple starting geometries (Zimanowski et al. 1997a, b; Mastin et al. 2004; Austin-Erickson et al. 2008). Although applicable on small scales, these small and simple experiments cannot represent large sustained explosive eruptions. Phreatoplinian eruptions show clear evidence for extensive magma water interaction (Wilson 2001; Houghton et al. 2010, 2015; Van Eaton and Wilson 2013), but the dynamics of this interaction remain somewhat unclear.

The indirect impact of the water column on eruption processes is the result of the differences in the physical properties of water compared with air (Table. 1.1.). Volatile exsolution and expansion during magma ascent is predominantly controlled by pressure (Verhoogen 1951; Sparks 1978; Proussevitch and Sahagian 1998). The higher density of water means that subaqueous eruptions take place under an applied hydrostatic pressure from the environment. Pressure increases by approximately 1 MPa per 100 m of depth (White et al. 2003, 2015; Cas and Giordano 2014). The hydrostatic pressure
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reduces volatile exsolution and expansion at equivalent shallow conduit levels compared to a similar subaerial eruption. The reduction in volatile exsolution and expansion lowers the driving force of the eruption, reducing explosivity (Fisher 1984; Yamagishi and Dimroth 1985; Mueller and White 1992). Water’s higher viscosity also means that rapid expansion requires more energy, thus further lowering explosivity (Kokelaar 1986; White 1996).

Supercritical fluids have properties between a liquid and a gas, and a phase boundary between the gas and liquid phases does not exist. Beyond its critical point, a fluid will no longer undergo rapid volumetric expansion, and will thus be unable to drive explosive fragmentation (White et al. 2003, 2015; Cas and Giordano 2014). Water is the dominant volatile component of silicic magmas, and has a critical point \( (C_p) \) at 22.06 MPa and 373.96 °C (Fig. 1.1) (White et al. 2003), for seawater the \( C_p \) is approximately 36 MPa and 403-406 °C (Bischoff and Rosenbauer 1984). Water’s critical temperature is well exceeded during MWI, while its critical depth equates to roughly 2200 mbsl (fresh water) (White et al. 2003) 3600 mbsl (seawater) (Bischoff and Rosenbauer 1984). The physics of MWI at conditions above the critical point are not well understood, the large variations in thermo-physical properties of the \( H_2O \) system from small changes in temperature pressure conditions above \( C_p \) mean this is a highly complex system to attempt to model (White et al. 2003).

The reduction in volatile exsolution results in reduced magma viscosity, relative to a similar subaerial eruption, since more water remains dissolved in the melt (Yamagishi and Dimroth 1985; Cas et al. 2003; Busby 2005; Giordano et al. 2008) (Fig. 1.2). For a silicic magma, viscosity may be reduced by several orders of magnitude depending on the eruption depth, and original water content (Fig. 1.2) (Yamagishi and Dimroth 1985; Cas et al. 2003; Busby 2005; Giordano et al. 2008). The reduction in viscosity has large implications for magma vesiculation, crystallisation, conduit strain and fragmentation.

The direct and indirect impact of the water column on subaqueous silicic volcanism fundamentally alters eruption dynamics compared with a similar system on land. These impacts have long been known (Reynolds and Best 1957; Honnorez and Kirst 1975; Kokelaar 1986; Kato 1987; Kano 2003).
Silicic magma water interaction processes and effects have been documented and studied from examination of ancient uplifted successions (Kano et al. 1996; Allen and McPhie 2000, 2009), seafloor deposits (Allen et al. 2010; Rotella et al. 2013, 2015), and from the few partly observed eruptions (Reynolds and Best 1957; Fiske et al. 1998; Kano 2003). However, studies of subaqueous volcanism are generally restricted, due to only being able to examine part of the eruption products, commonly from unknown vent environmental conditions. The inferred effects of the water column on subaqueous eruption dynamics have been included in published several eruption models (Kokelaar and Busby 1992; Kano 2003; Allen and McPhie 2009; Schipper et al. 2010; Rotella et al. 2013). However, the exact dynamics have never been demonstrated through physical observation and examination of all the products of a subaqueous silicic eruption.

**Fig. 1.1.** Schematic pressure temperature diagram for fresh water showing the phase at each point. The line in red marks the pressure at sea level, and the phase boundaries at that point. At conditions beyond the critical point, typical phase boundaries do not exist. For seawater the critical point is at approximately 36 MPa and 403-406 °C (Bischoff and Rosenbauer 1984).
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Fig. 1.2. Viscosity vs temperature graph for rhyolitic glass produced during the 2012 subaqueous eruption at Havre volcano. Viscosity was calculated for a range of eruption temperatures using the methods of Giordano et al. (2008), where chemistry was determined using EPMA of the glass (Chpt. 5.5). Viscosity temperature profiles for different groundmass water contents were also calculated. The decrease in viscosity with increasing water content is associated with the depolymerisation of silica tetrahedra chains through their reaction with water molecules to form hydroxyl molecules.

1.2.2. The effect of the water column on clast dispersal and transport processes

In subaqueous volcanism the water column has large impacts on the mechanisms by which the generated clasts are dispersed. Water’s different thermo-physical properties alter some processes, prevent others, and produce entirely different mechanisms unique to the subaqueous environment (Reynolds and Best 1957; Cashman and Fiske 1991; Fiske et al. 1998; Kano 2003; Allen et al. 2008, 2012; Jutzeler et al. 2016; von Lichtan et al. 2016). In the subaqueous environment steam is readily produced around hot magma/lava, however is highly transient due to its high buoyancy and the rapidity with which it is condensed once the heat source is removed (White et al. 2003, 2015; Cas and Giordano 2014). Steam’s brevity impacts on dispersal and transport in subaqueous volcanism.
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The higher density and viscosity of water relative to air has several impacts on dispersal and transport processes. Water’s high viscosity means ballistic transport of particles does not occur (Kokelaar 1986; White 1996). The higher density and viscosity of water compared to air also means that convection driven by heating of the water, due to the underlying volcanism, is more easily able to entrain fine particles (Ferguson and Church 2004; White et al. 2009). A convective cell driven by heating of the overlying water column by effusive volcanism could entrain particles to produce plumes that form a fallout style suspension deposit without explosive volcanism. Due to the low contrast in density between particles and the ambient water, settling rates are also comparatively reduced relative to the subaerial environment. The extremely slow settling rates of very fine ash particles in water could lead to the extremely long-distance transport of a significant proportion of the ash component (Cashman and Fiske 1991; Ferguson and Church 2004; White et al. 2009). In the case of transport of fine clasts, the effect of the water column on individual particles is relatively well understood. Its impact on an eruptions deposit, however, is not. An important consideration when interpreting subaqueous clastic stratigraphy is that the viscosity of water amplifies the settling times between large and small particles produced at the same time. Additionally, if particles arrive at the water-air interface their transport dynamics will be modified by wave and wind currents and may be completely missing from the stratigraphic record. Therefore, although the relative timing of deposition is clearly shown in stratigraphy the large effect of the water means that order of deposition may not reflect the times at which particles were produced. Determining the primary total grain size distribution (TGSD) of a deposit is always difficult (Bonadonna and Houghton 2005), and for submarine deposits it may not yet be possible because so many unknowns remain about the range of impacts from particle transport.

The formation of steam can insulate clasts from contact with water (Allen et al. 2008; Fauria et al. 2017). Steam insulation can allow transportation of molten clasts, allowing fluidal behaviour (Mueller and White 1992). Steam charging can also lower particles’ density sufficiently for them to remain buoyant within the water column (Allen et al. 2008; Fauria et al. 2017). Since the production of steam
from enclosing water or the maintenance of a magmatic steam film requires a constant heat source, constant steam films are typically restricted to larger particles (White et al. 2003, 2015; Allen et al. 2008; Cas and Giordano 2014; Fauria et al. 2017). The high surface area to volume ratios of ash particles means they become rapidly quenched (van Otterloo et al. 2015). During eruptions at Tulumau 1953-57 (Reynolds and Best 1957), Shin-Iwojima 1934-35 (Kano 2003), and Irimote Island 1924 (Kano 2003) sea surface pumice blocks were inferred to have buoyantly ascended several hundred meters through the water column after detaching from lava. Although the effect of steam is clearly demonstrated by some depositional processes (i.e. buoyant pumice blocks), it is unlikely to represent its full impact on the eruption processes. There is thus far a lack of direct observational evidence to fully quantify the impact on steam on subaqueous eruption processes.

A high sustained heat and material flux could generate a sustained jet, where entrained water is rapidly vaporised (Head and Wilson 2003). The condensation of generated steam and magmatic volatiles in explosive jets can rapidly stall their ascent (Deardorff et al. 2011). For this reason, subaqueous jets are inferred to ‘pinch out’ quickly with height as heat is dissipated through the water (Head and Wilson 2003). The dynamics of subaqueous eruption jets however are entirely inferences from experiments (Verolino et al. 2017), numerical modelling (Head and Wilson 2003), and microtextural studies (Allen and McPhie 2009; Rotella et al. 2015).

The main impedance on the understanding of silicic subaqueous volcanism is the lack of direct observations of an eruption. Even examining all the fresh in-situ products of an eruption from a known source is no replacement, and in addition is very expensive. Replication on the seafloor of a typical subaerial post-eruptive study at the same resolution would be prohibitively expensive. Therefore, the understanding of eruption dynamics has historically been derived from studies of uplifted successions (e.g., Allen and McPhie, 2009; Kano, 2003; Kokelaar, 1986). However, ancient subaqueous successions are generally restricted by the lack of detail in the eruptive and environmental conditions. Observations of subaqueous volcanism have been conducted at the sea surface (Reynolds and Best 1957; Fiske et al. 1998; Kano 2003), however these are unlikely to be
representsative of seafloor eruption processes (e.g. Carey et al., 2018). The study of subaqueous volcanism is currently missing a contextualised, integrated examination of all products from several events of varying eruptive size and depth. The results of such studies would underpin our understanding of volcanism underwater and provide context to studies examination the geological record.

1.3. 2012 Havre eruption

In 2012 a large silicic subaqueous eruption occurred from a vent depth of approximately 900 mbsl at Havre volcano (Carey et al. 2014, 2018; Jutzeler et al. 2014). A comparison of pre- and post-eruption bathymetric maps of the volcano illustrated many new features on the top of the Havre volcanic edifice, in addition to a ~400 km$^2$ pumice raft produced in ~21.5 hours. The before and after maps, coupled with volumetric and temporal constraints on the pumice raft offer an unprecedented opportunity for insight into subaqueous eruption processes.

1.3.1 Geological setting - Kermadec Arc

Havre volcano is located on the Tonga-Kermadec arc, a region reasonably well-known in geological terms. The Tonga-Kermadec arc-back-arc system is predominantly a submarine subduction system, that is the result of the ocean-ocean collision between the Pacific and Australasian plates (Oliver and Isacks 1967; Isacks et al. 1968; Smith and Price 2006; Meffre et al. 2012; Bassett et al. 2016) (Fig. 1.3.). Westward subduction of the Pacific plate has given rise to the north-northwest trending arc that extends from the North Island of New Zealand and transitions in to the Fiji Fracture Zone at its northern end (Smith and Price 2006; Wysoczanski et al. 2010; Meffre et al. 2012; Bassett et al. 2016) (Fig. 1.3.).

The Tonga-Kermadec Arc is a complex system with northern and southern segments defined by the morphology of the ridge front, style and speed of back arc spreading, speed of plate convergence,
Fig. 1.3. A regional bathymetric model of the Tonga-Kermadec subduction system. Arc front subaerial (green) and subaqueous (blue) volcanic centres are shown (data from Wright et al. 2006; Global Volcanism Program, 2013), with the location of Havre volcano marked in bold.
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and the morphology of the volcanoes among other features (Smith and Price 2006; Wysoczanski et al. 2010; Meffre et al. 2012; Bassett et al. 2016) (Fig. 1.3.). The north south split occurs at roughly 25 °S, where the Louisville seamount chain intercepts the subduction front (Smith and Price 2006) (Fig. 1.3.). The northern segment is the Tonga Arc, and includes the Tonga Ridge, Lau Basin and remnant Lau Arc (Smith and Price 2006). The southern segment is composed of the Kermadec Ridge, Havre Trough and remnant Colville Arc, and is referred to as the Kermadec Arc (Smith and Price 2006). Subduction rates increase from approximately 6 cm yr⁻¹ in the south to 24 cm yr⁻¹ in the north (Smith and Price 2006). An equivalent trend in the back-arc spreading rate is observed, with a high in the north of 80 mm yr⁻¹, associated with active seafloor spreading and oceanic crust generation in the back-arc Lau Basin (Pelletier et al. 1998; Zellmer and Taylor 2001; Smith and Price 2006). Towards the south, the spreading rate gradually drops to about 8 mmyr⁻¹, and is dominantly controlled by rifting and graben formation (Parson and Wright 1996; Smith and Price 2006).

A continuous volcanic chain extends for the whole length of the Tonga-Kermadec arc (Fig. 1.3.). The volcanoes themselves are restricted to an approximately 40 km wide zone, offset to the west of the of the arc front ridge (Smith and Price 2006; Wright et al. 2006). In the north, the Tonga ridge front forms a shallow broad terrace (Smith and Price 2006). In the south the ridge front is narrower and deeper, disappearing into seafloor sediments towards the south (Smith and Price 2006; Wysoczanski et al. 2010). The volcanoes are dominantly submarine, with only a few edifices breaching the sea surface. In the central segment of the Tonga Arc, the volcanoes are typically large with shallow summits, occasionally generating ephemeral pyroclastic islands (Siebert et al., 2010). Most of the knowledge regarding volcanism on the Tonga-Kermadec Arc comes from this segment due to its higher proportion of volcanos exposed above the sea surface (Smith and Price 2006). In the southern segment, volcanoes occur as both stratocones and calderas, with larger calderas typically restricted to shallower areas of the Kermadec Arc (Wright et al. 2006).

Magma from the Tonga-Kermadec Arc varies in composition from basalts through to rhyolite (Ewart et al. 1977, 1998; Ewart and Hawkesworth 1987; Wright et al. 2006; Barker et al. 2013) (Fig. 1.4.).
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Chemistry in the Kermadec Arc as a whole, however, shows more variability when compared to that of the northern segment. Smith et al. (2003) have shown that eruptions in the Holocene on the Kermadec Arc have been dominantly silicic in nature, producing dacites and rhyolites.

Fig. 1.4. A total Alkali vs Silica (TAS) diagram of whole rock composition for a range of volcanic centers along the Kermadec arc, data from (Wright et al. 1996, 2006; Wright and Gamble 1999; Barker et al. 2013). The key denotes the name of the volcanic center each sample is from.

Bathymetric work has also shown the presence of many stratocones and calderas located along the Kermadec ridge, which commonly show features indicative of complex eruptive histories (Wright et al. 2006). There is extensive evidence for a long history of volcanism along the Kermadec Arc, both subaerial and subaqueous. Observations back to 1814 (e.g. Smith 1888), show evidence for a range of pumice forming and effusive eruptions. In addition, recent studies have shown evidence of active subaqueous volcanism occurring recently, including hydrophone recordings, pumice rafts, and observation of discoloured water (Latter et al 1992; Global Volcanism Program, 2013; Rotella 2013). This occurs most notably at Monowai (Chadwick et al. 2008b; Watts et al. 2012; Metz et al. 2016).
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The largest recent eruption occurred at Havre Volcano in 2012 with the generation of a large pumice raft and significant seafloor changes (Carey et al. 2014, 2018; Jutzeler et al. 2014; Rotella et al. 2015).

1.3.2. Havre Volcano

Havre Volcano is fully submerged located at 31° 05’S 179°5’W (-31.103511, -179.035231) along the Kermadec arc just west of the Kermadec ridge (Wright et al. 2006) (Fig. 1.5.). The volcano rises from the seafloor at 1500-2000 meters below sea level (mbsl) to a summit of ~650 mbsl. Havre is truncated at 900 mbsl by an oval shaped caldera, four by six kilometres, elongate northwest southeast (Fig. 1.5.). The caldera floor is relatively flat at approximately 1500 mbsl. The caldera walls are typically steep, commonly forming vertical cliffs. Along the south rim however a relatively flat plain occurs at 900 mbsl. Havre summit is located on the south west corner of the caldera and was formed an eruption in 2012. The volcanoes flanks fall away gently towards the north and more steeply towards the south and are mantled by many dome-like features. A rubbly uneven surface marks the inferred base of the volcano also showing a northwest southeast elongated oval shape 20 km by 15 km (Fig. 1.5.). Towards the south west and north a series of concentric ridges can be seen that drop-in height away from the volcano (Fig. 1.5.).

Little is known of the eruptive history of Havre volcano prior to its eruption in 2012. The central caldera has been inferred to be have been formed as a result of syn-eruptive collapse, associated with high-mass-discharge pyroclastic eruptions (Wright et al. 2006). Wright et al. (2006) also indicated that the outer flanks of the volcano were mantled by weathered pumice when observed (2002), and that there are some dacite and basalt lava flows, which were suggested to be of pre-caldera age. Recent activity appears to have centred on the southern caldera rim plain where an aligned dome and crater complex can be observed. All these features were present prior to the 2012 eruptive episode (Wright et al. 2006).
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Whole rock geochemistry conducted on five samples collected by Wright et al. (2006) at Havre shows four silicic samples of dacite to rhyolite composition and a single basaltic one. The caldera wall exposes a pre-caldera lava and intrusive stack of basaltic andesite, dacite, gabbroic and diorite compositions (Wright et al. 2006).

1.3.3. The 2012 eruption of Havre Volcano

At 1050 on the 18th July in 2012 universal standard time (UTC) (all following times are in UTC), an atmospheric plume and hot spot were captured in a satellite image taken at night, and emanating from a point source above Havre (Carey et al. 2014; Jutzeler et al. 2014). An image taken in daylight at 2151 on the same date showed the same atmospheric plume in addition to an extensive pumice raft (Carey et al. 2014; Jutzeler et al. 2014) (Fig. 1.6.). Satellite imagery indicates that eruptive activity producing the pumice raft continued for 21.5 hours (Fig. 1.6.). An image taken at 0209 on the 20th July in 2012 (UST) shows the pumice raft had detached from its source and the atmospheric plume had ended, indicating the eruption had either ended, or was no longer powerful enough to produce effects at the sea surface (Carey et al. 2014, 2018; Jutzeler et al. 2014) (Fig. 1.6.). Over the course of the eruption a pumice raft approximately 400 km$^2$ in size was produced, prior to detachment (Carey et al. 2014, 2018; Jutzeler et al. 2014) (Fig. 1.6.). From 17th to 21st July frequent earthquakes of magnitude three to five were also recorded from Havre (Carey et al. 2014; Jutzeler et al. 2014). After 21st July 2012 there was no detectable evidence of volcanism at Havre in satellite imagery or seismometers.
Fig. 1.6. A series of Aqua and Terra MODIS images taken from 0045 18th July 2012 to 0110 20th July 2012 UTC (all following times are in UTC) showing the development of the 2012 Havre eruption. The generation of a pumice raft, a discoloured water plume, and an atmospheric steam plume extending from a point source, located above Havre, can be seen in true colour images between 0045 18th July to 0126 19th July 2012. An image of band 22; 3.959 μm spectrum, which shows temperature, taken 1050 18th July 2012 displays the presence of an ocean hot spot and indicates the atmospheric plume was cold. An image taken 2230 19th July 2012 shows the pumice raft has disconnected from the point source indicating an end to the subaerial component of the eruption.
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A comparison of bathymetry surveys conducted in 2002 (Wright et al. 2006) and after the eruption, on 26th October 2012, shows that there had been large scale topographic changes on the volcano summit (Carey et al. 2014) (Fig. 1.7.). Several cone shaped features had been produced on the southern caldera rim, initially inferred to be pyroclastic cones resulting from a pyroclastic eruption (Carey et al. 2014). The seafloor products of the Havre eruption add to a total volume of ~380x10^6 m^3, the addition of the raft volume of ~1,200x10^6 m^3 showed that the 2012 Havre eruption was the largest silicic deep water eruption of the last century (Carey et al. 2014, 2018).

In 2015, the seafloor products of the Havre eruption were subsequently mapped at high resolution (1 m) by autonomous underwater vehicle (AUV) Sentry, and observed and sampled using the remotely operated vehicle (ROV) Jason (Fig. 1.8.) (Carey et al. 2018). The cone shapes on the caldera rim were mapped and visited by the ROV and shown to be lavas and domes erupted on the southern caldera rim from vents at depths of 1050 – 900 mbsl. The bulge was mapped and visited and consists of five adjacent lava flows erupted from five vents at depths of 1220 mbsl and 1140 mbsl on the southwest caldera wall. In total 14 lavas were erupted along the southern and southwestern caldera rim, following structural lineaments which are likely faults (Fig. 1.8.) (Carey et al. 2018).

In ROV observations also revealed three mappable clastic units (Fig. 1.8.). (1) A deposit composed of giant pumice clasts >1 m in diameter (Giant Pumice Unit (GP Unit)) is widespread extending off the edifice to the NW, and encloses Dome OP in the southwest (Fig. 1.8.) (Carey et al. 2018). The GP Unit has been deposited over approximately 35 km^2 within the study area. Clasts in the GP Unit are commonly stacked on top of one another or precariously balanced. Assuming an average deposit thickness of 5 m, and a packing efficiency of 60% a volume of 0.1 km^3 was calculated for the volume of the GP unit within the study area. The basal contact of the GP Unit with pre-2012 deposits was not directly observed during the cruise. The giant pumices themselves are moderately to highly vesicular, are commonly bounded by curvi-planar fractures, and commonly show normal jointing. No lithic component has been identified in any of the clastic deposits from the 2012 Havre eruption. (2) A deposit composed of ash to 1 m blocks sized pumice clasts (Ash, Lapilli and Block Unit (ALB Unit)).
Fig. 1.7. 1 m resolution bathymetric model of Havre caldera collected by *AUV Sentry* (Carey et al. 2018). The bathymetric change between 2002 and 2012 is plotted where increase is shown in red and decrease in blue.
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Fig. 1.8. 1 m resolution bathymetric model collected by AUV Sentry of the caldera and proximal rim of Havre Volcano (Carey et al. 2018). The boundaries of the seafloor products of the 2012 Havre eruption are marked; GP Unit and ALB Unit in white, the lavas in green. The AL Unit is deposited over the whole study area with no boundary observed.
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The ALB Unit is locally dispersed around Dome OP with several lobes extending from Dome OP into the caldera and laterally across the caldera rim, as well as down the outer flanks of Havre (Carey et al. 2018). Approximating the ALB Unit as a cone with a peak of 2 m and a radius of 2 km its volume is estimated at 0.005 km$^3$ for a packing efficiency of 60% (Carey et al. 2018). (3) The final Unit is the Ash and Lapilli Unit (AL Unit), composed dominantly of ash with only minor amounts of lapilli. The AL Unit has been dispersed over the whole study area and found in every clastic sample taken at Havre. The thickness of the AL Unit is somewhat uncertain since its stratigraphic base was rarely observed. Using an inferred thickness of 20 cm, and a packing efficiency of 90% a volume of 0.063 km$^3$ is estimated for the AL Unit over the 35 km$^2$ study area. The AL Unit is composed of four distinct subunits, these are the focus of the present study. The spatial extent of the GP and AL Units remain largely unknown with GP Unit continuing off map to the northwest, while the AL Unit extends beyond the mapped area in every direction (Fig. 1.8.). Bulk rock major element analyses of all the sampled clastic and effusive materials from the 2012 eruption are rhyolitic with a silica composition of 70-72 wt% SiO$_2$ (Carey et al. 2018) (Fig. 1.9.).
The 2012 Havre eruption is temporally constrained by the timing of the pumice raft on 18-19<sup>th</sup> July (2012), and a multibeam survey conducted by the <i>R/V Tangaroa</i> on 17<sup>th</sup> Oct 2012 (Fig. 1.10.). Comparison of the <i>R/V Tangaroa</i> multibeam with that undertaken by <i>R/V Revelle</i> in 2015 shows no differences indicating that all the deposits were fully emplaced by 17<sup>th</sup> Oct 2012. There is no direct link between the seafloor deposits and the pumice raft (Fig. 1.10.). Based on similarities in the model densities of the GP and pumice raft along with, similar banding features, phenocryst and microlite composition and texture, and dispersion azimuths (Carey et al. 2018) inferred the GP and pumice raft to have been formed at the same time (Fig. 1.10.). Stratigraphically the GP Unit is overlain by both the ALB Unit and the AL Unit (Fig. 1.11.). The GP Unit also occurs under lavas H-P (Fig. 1.11.). The implication of this is that lavas H-P, the ALB unit, and AL Unit were emplaced after 19<sup>th</sup> July 2012 but prior to 17<sup>th</sup> Oct 2012 (Fig. 1.11.). For the largest dome at Havre, Dome OP, this gives a time averaged effusion rate of > 14 m<sup>3</sup>s<sup>-1</sup>.

**Fixed points**

2002 bathymetry survey (Fig. 1.5a.)

0733 17 July 2012 to 0300 18 July 2012
157 hydroacoustic events near Havre

0045 18 July 2012
seafloor pumice raft first appears (Fig. 1.6.)

2230 19 July 2012
seafloor pumice raft detached from point source (Fig. 1.6.)

No seafloor activity

17 Oct 2012 bathymetry survey (Fig. 1.7.)

Seafloor products fully emplace

**Broad seafloor activity**

Lavas A to E emplaced

GP Unit emplaced

AL Unit emplaced

Lavas F to P emplaced

Fig. 1.10. Time line of Havre eruption showing recorded eruption events (above). Below is show the broad inferences on eruptive activity based on seafloor features.
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During the 2015 Havre cruise 290 different samples were collected from various locations across the caldera floor and around the caldera rim on 12 different dives with ROV Jason. Gravity coring was also undertaken with successful recovery of material occurring at 19 locations. Of the ROV samples 98 were taken of clastic material and were used in this thesis. The sampling methods are described in Chapter 2. Methods.

Fig. 1.11. The broad inferred stratigraphy for the seafloor deposit of the 2012 Havre stratigraphy, including the relative timings of deposition all the subunits of the AL Unit. The trace of Fig. 1.11 is shown in Fig. 8.

The 2012 eruption of Havre volcano was volumetrically the largest deep subaqueous silicic eruption in the past century (Carey et al., 2018, 2014). The satellite imagery provides constraints on pumice raft volume, timing and mass eruption rates. This fact alone would make the 2012 Havre eruptive episode a globally significant eruption. The focused follow-up cruise, conducted in 2015 however, produced a detailed sample set of both effusive and clastic seafloor products (Carey et al. 2018). The Havre sample set is globally significant in its completeness, how fresh it is, the fact that all samples were taken in situ, and the known eruptive source. Specific to the ash component, samples from Havre show ash generation via a range of potential mechanisms. The fact samples were acquired along with in situ site observations allows us to look in detail at specific processes, normally beyond studies where the vent location/conditions are unknown, or samples were acquired en masse by dredging of unobserved seafloor. There are many yet unanswered questions regarding the dynamics
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of subaqueous volcanism. The Havre observational data and sample set provide an excellent opportunity for insight into questions regarding subaqueous ascent, fragmentation, effusion, and dispersal. Its global significance and uniqueness means that it is likely to become a corner stone for future studies looking at subaqueous silicic eruption processes.

1.4. Thesis aims within context of the overall Havre research project

My project is part of a larger undertaking, the goal of which is to fully quantify the eruption dynamics of 2012 Havre eruption, and place it in the context of the study of deep subaqueous volcanism. Several studies on the seafloor lavas, the seafloor and sea surface pumiceous products of lapilli size and larger of the 2012 eruption, and the chemistry of the 2012 eruption and the Havre edifice are being undertaken in tandem. The aim of the research presented here is to investigate specifically the processes involved in ash generation, dispersal and deposition during the 2012 silicic subaqueous eruption of Havre volcano, New Zealand. My additional aim is to assess the influence of ambient water both indirectly and directly on these processes.

Specific aims of this thesis:

- To constrain the relative timing of deposition from the 2012 Havre eruption, by establishing a detailed ash stratigraphy
- To establish the eruption, fragmentation, and depositional mechanisms by each ash subunit in the AL Unit was produced
- To investigate a population of fluidal grains observed in two subunits at Havre, and establish the mechanisms by which they formed
- To assess the dispersal mechanism associated with a population of mafic grains found in Havre ash
- To establish a unified eruption model for the 2012 Havre eruption
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In this thesis, chapters 3 to 4 are presented as stand-alone papers to be submitted for publication. The methods for each chapter, however, has been removed and combined into a single section (Chapter 2. Methods) to avoid unnecessary repetition. All the methods applied here are well established and used frequently in the study of pyroclastic deposits in both the subaerial and subaqueous realm.

Overall this thesis comprises a combination of manuscripts for publication and several short chapters that convey focused results and interpretations from specific investigations or methods. Draft manuscripts have been reformatted for consistency through the thesis. The contents of the chapters are as follows:

Chapter 3. Characteristics and stratigraphy of submarine-erupted silicic volcaniclastic deposits, Havre volcano, Kermadec Arc, New Zealand: outlines the clastic stratigraphy of the 2012 Havre eruption, constraining the spatial extent and relative timings of four subunits defined from dominantly mixed samples of the seafloor AL Unit. The reconstructed stratigraphy is used to infer the relative timings of effusive and pyroclastic phases in the 2012 eruptive episode. As part of the analysis detailed granulometry, componentry, and SEM morphological data are presented for all subunits. These results are then further used to infer the eruption, fragmentation, and dispersal processes for each of the subunits, allowing a broad eruption model to be built.

Chapter 4. Explosive and passive mechanisms associated with submarine lava effusion, Havre volcano, Kermadec Arc: examines in detail the formation and dispersal mechanisms associated with Subunit 3, a clastic deposit closely associated with Lava G. Here I model particle dispersal in thermal plumes driven both by heating from a lava flow surface and from low volume pyroclastic activity. It is suggested that fragmentation driven by gas flow through heated cracks can explain the formation of elongate fluidal grains from a rhyolite dome or lava.

Chapter 5: Unusual fluid behaviour of a silicic magma during fragmentation in a deep subaqueous eruption, Havre 2012: examines a population of ash grains at Havre that show fluidal surfaces. Here I
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suggest how these particles formed despite their high viscosity. None of the models to date fit all the
observations, and thus the formation mechanism of these grains remains an open question.

Chapter 6. Foreign grains: is a short chapter that examines a population of relatively fresh mafic
grains found in the 2012 Havre deposits. Glass chemistry shows the mafic grains to be have a range
of compositions from basaltic through to dacite, and they do not appear to be primary products of
the 2012 Havre eruption. The source of these grains is considered in light of potential source
volcanoes and transport of particles.

Chapter 7. SR-FTIR results: is a short chapter that examines specifically results of synchrotron
radiation Fourier transform infrared spectroscopy (SR-FTIR) analysis to determine abundance of
dissolved volatiles in the rhyolitic ash grains. From this I show that ash particles in the 2012 eruption
were quenched high in the water column.

Chapter 8. Discussion: is the synthesis and conclusion chapter. It presents an integrated
interpretation of results from the work conducted as part of this thesis.
Chapter 2

Methods
Chapter 2: Methods

2.1. 2015 Havre cruise

Sampling was undertaken at Havre volcano by remote operated vehicle (ROV) Jason. In total 290 seafloor samples were collected from various sites around Havre caldera, of which 92 were clastic samples of the AL Unit used in this study. The locations of the samples examined in this study, the methods by which they were collected, and the analysis undertaken on each are summarised in Appendix 1. Each sample has been given a code where HVR denotes its collection from Havre volcano and the following number indicates the relative timing of its collection on the 2015 cruise.

All the clastic samples taken at Havre caldera were collected in an approximately 12 km² area centred on the summit caldera. As such no sample was taken more than 4 km away from any of the vents inferred to be active during the 2012 Havre eruption. In addition, sampling was not conducted on the outer flanks of Havre volcano or the surrounding seafloor.

Due to the restrictions that the deep subaqueous environment places on field work, seafloor imaging and sampling can only occur in ~5 m wide ROV Jason dive tracts (Fig. 2.1.), which are also time restricted. This potentially introduces some bias into the data, never been able to directly observe the ‘big picture’. Despite this sampling was conducted over the whole study area giving a good spatial range (Fig. 2.1.). In addition, the high resolution bathymetric model of Havre caldera collected by AUV Sentry (Carey et al. 2018) gives good topographic constraint on each sampling location. Therefore, sample quality can be assessed and any potential influence of secondary processes (e.g. reworking, etc.).

Samples were collected from the seafloor at Havre volcano using ROV Jason and employing a range of techniques. Coherent pumice and dense rock could be sampling using a robotic manipulator. The manipulator however, cannot collect fine-grained clastic deposits or weak aggregates. To collect such samples, three other devices were employed; push cores, scoops, and vacuum sampling.
Fig. 2.1. *ROV Jason* dive tracks and sampling sites from the 2015 Havre cruise plotted over the 1 m scale resolution *AUV Sentry* bathymetric model of Havre caldera (Carey et al. 2018). The sampling method is showed at each site.
2.2. Sampling methods

A push core is a 3.5” diameter plastic tube with a handle at one end and open at the other; a core catcher supplemented with stretch-nylon fabric is mounted on the inside near the opening (Fig. 2.2.). This tube is designed to be inserted into sediment until either full or until the tube can be inserted no further, then retrieved with the sampled sediment in the tube (Fig. 2.2.). For many Havre samples, unfortunately, the tube was either inserted repeatedly and/or rotated to improve penetration. The rotation and multiple plunging during sampling via push cores generally destroyed any deposit layering present. Only two cores returned samples with preserved stratigraphy.

Fig. 2.2. Outlines push core sampling (a) Shows a full assembled push core, where the right-hand end is inserted into the sediment, which is caught by the netting and orange sediment catcher when the core is withdrawn. (b) Seafloor sampling using a push core (sample HVR004). When sampling the loose seafloor clastics of the 2012 Havre eruption push core were typically inserted vertically. (c) Showing a recovered push sample (HVR134) that has been cut open to expose the preserved stratigraphy.
A scoop is a frame holding two layers of netting, one fine (200 µm) and one coarse (1 mm), on a metal rod (Fig. 2.3.). The scoop is dragged through the sediment and then placed into a sample box mounted on Jason. ‘Scooping’ generally destroys any deposit layering present, but one sample did preserve layering in a sediment clump (Fig. 2.3.).

Fig. 2.3. Outlining scoop sampling. (a) Showing a range of scoop shapes in various stages of readiness. (b) Showing the collection of sample HVR042, scoops were dragged through the sediment
collecting a mixture of all deposits present at that location. (c) Scoop samples typically produced large volume mixed samples.

Vacuum sampling uses a pump to draw seawater and sediment into a tube that exits into one of several water-filled sampling canisters where particles are collected (Fig. 2.4.); the excess water is released through two filters; a >1 mm mesh and a 1000-200 µm fabric filter. Vacuum sampling disaggregates sampled material, destroying any fabric or layering present. The vacuum sampler was effective in deposits with particles up to 3 cm in diameter.

Fig. 2.4. Outlines vacuum sampling. (a) Shows the sample canister chamber for the vacuum pump. The canisters could be rotated allowing up to four samples to be taken before they need to be exchanged. (b) The vacuum sampler works much the way one would expect and was extremely effective at exposing in situ stratigraphy on the seafloor. (c) Samples taken using the vacuum method preserved no stratigraphic evidence.
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Despite the differences in collection methods there are no discernible differences in the grain size characteristics among samples taken by different mechanism (push core, scoop and vacuum). All sampling methods generally produced mixed samples of material of the clastic deposits collected. As such only two push core samples were recovered that preserved stratigraphy.

2.3. Granulometry

Prior to granulometry all samples were dried either in an oven at 90 °C or under an array of heat lamps for at least 8 hrs. Whole samples were later hand sieved from -4 φ to 4 φ (from 16 mm to 0.063 mm) in ½ φ steps (Fig. 2.5.). The fraction remaining in each sieve was weighed on an electronic scale with 0.01 g resolution. The fraction in each sieve was examined for aggregates that formed during the drying of the sample, and any found were crushed gently by hand. Once the aggregates were crushed the sample was again sieved to ensure all material fell to its appropriate size. When transferring the pan fraction (smaller than .063 mm; >4 φ) to the scales, a small volume would occasionally become airborne and drift away. Care was taken to minimise such losses of fine particles, but very small masses of some extremely fine particles were still lost.

A Mastersizer 2000® laser particle analyser at the University of Otago was used to measure the size distribution, in vol.%, of particles smaller than 0 φ (1 mm). Grain-sizes from 0 to 20 φ (1 mm to 95 nanometres) were measured in bins of approximately 0.2 φ. Laser particle size analysers operate by passing particles, suspended in water as a carrier liquid, between a laser and series of detectors; the amount of diffraction as the laser light passes through particles is proportional to their sizes, and the Mastersizer® calculates the size of each grain from the measured diffraction. To undertake Mastersizer® analysis the hand sieved fractions smaller than 0 φ (1 mm) were recombined, then the combined sample split using a mechanical splitter to get ~2 g of each sample for use in the Mastersizer 2000®. Three Mastersizer runs were conducted per sample and the average of these runs used for analyses. Mastersizer® results were merged into bins of 0.5 φ, for consistency with the
Chapter 2: Methods

results from hand sieving, assuming a linear distribution of particle sizes within each bin (Fig. 2.5.). Mastersizer® results were converted from vol.% to wt.% \( (X_f) \) prior to merger using the equation:

\[
X_f = \frac{(x_f \times X_{ct})}{100}
\]

Where \( x_f \) is the vol.% of a specific grain size fraction and \( X_{ct} \) is the cumulative wt.% of the analysed grain-size fraction, from (Eychenne et al. 2012). The equation presented here assumes relatively homogeneous grain density in the size fraction analysed by Mastersizer®, and that Mastersizer® results are collected from a representative volume of grains (Eychenne et al. 2012).

The chosen overlap point was generally around the 0.5 \( \Phi \) (Fig. 2.5.). Merging was undertaken by scaling the rebinned Mastersizer® data to the fraction of the sieved sample below the chosen point (i.e. the

\[\text{Fig. 2.5.}
\]

Showing a representative grain size distribution graph with the sieved (green), Mastersizer (red), and combined (dashed black) trends. In this sample the point of overlap occurs at 1 \( \Phi \).
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Mastersizer® data was rescaled so that its total wt% was reflective of the percentage of the sieved sample smaller than the chosen overlap point). The chosen point ideally produced the lowest difference from 100 wt% when adding the sieving and Mastersizer® data (Fig. 2.5.).

2.4. Componentry

"Componentry" is the quantification of different particle types (components) in a sample or deposit. Sample componentry was done for two size ranges. For larger particles, -2 φ to 0 φ (4 to 1 mm), categorization and identification were done with the naked eye and for 1 φ and 2 φ (500 to 250 µm) fractions, by binocular microscope (Optical componentry). For smaller particles, I produced scanning electron microscope (SEM) secondary electron images maps of ash size fractions then categorised and counted grains imaged (SEM componentry).

Optical componentry was conducted in 1 φ (500 µm) steps on material from -2 φ to 2 φ (4 mm to 250 µm). After the ½ φ sieving, particles coarser than 0 φ (1 mm) were combined into 1 φ (500 µm) fractions (e.g. the 0 φ fraction for componentry included material from the 0 φ and -0.5 φ sieves). All particles smaller than 0 φ (1 mm) were also sieved into 1 φ steps. For each size fraction, at least 300 grains, isolated using a mechanical splitter were analysed, or the entire sample if there were fewer than 300 grains. Particles were counted into three distinctive first-order juvenile component groups defined based on grain colour and vesicle patterns:

- **Glassy Vesicular grains:** Light coloured particles that appear generally glassy and moderately to highly vesicular
- **Microcrystalline grains:** Dark coloured grains that appear moderately to non-vesicular
- **Elongate Tube-Vesicle grains:** Light coloured grains that are often extremely elongate defined by tube vesicles the run parallel to the particle elongation direction.
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These first-order componentry classes were then subdivided into several second-order subclasses defined by surface morphology and texture. Glassy Vesicular and Microcrystalline grains were subdivided into:

- **Angular particles**: defined by prominent concavities formed by vesicles with peaks formed by brittlely-fractured vesicle walls

- **Curvi-Planar particles**: defined by planar and curvi-planar fracture surfaces that cross cut vesicles showing little to no deformation and intercept to form sharp edges

- **Fluidal particles**: defined by exterior features indicating surface tension or hydro/aerodynamic reshaping of the grains while molten

- **Sheared-Vesicle particles**: defined by the presence of tube vesicles and are typically bounded by curvi-planar fractures.

Fluidal and Sheared-vesicle particles are not observed in Microcrystalline grains. Elongate Tube-Vesicle particles were split in to a different set of second order componentry groups:

- **Elongate Tube-Angular particles**: defined by elongated forms, with concave surfaces, defined by brittle-fractured bubble walls

- **Elongate Tube-Ribbed particles**: defined by surface ribs the run parallel to the vesicle and clast elongation direction, the surfaces of which are smoothly undulating and typically unmarked by vesicles

- **Elongate Tube-Fluidal particles**: are elongate, unmarked by vesicles, with flowing molten surfaces that form peaks or droplet like features, and show evidence of ductile necking

Optical componentry is extremely time-consuming when done across the full grain size range; it was carried out on 28 samples from around the caldera. These samples were chosen as representative of the surrounding seafloor and represent most of the thickness of the ash deposit at each sample site as inferred from video observation from Jason.
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2.5. SEM investigation

Grain morphology and microtextures were investigated using secondary electron (SE) and back-scattered electron (BSE) methods on a Zeiss Sigma VP® Field-Emission-Gun Scanning Electron Microscope at the University of Otago Centre for Electron Microscopy.

2.5.1. SEM SE based point counting componentry

For SE (morphological) imaging grains were mounted on an SEM stub using carbon tape and then were carbon coated. Imaging was undertaken using a 15 keV accelerating voltage, working distance when imaging varied from 7.1 to 9.5 mm. SEM SE imaging was undertaken in two phases, initially general morphological imaging of particles was conducted allowing their visual description. The second phase involved more systematic creation of montaged SEM SE image maps of samples from which SEM componentry could be conducted. The initial morphological imaging was conducted on a range of samples on grains sizes smaller than and equal to 3 φ (125 µm).

Systematic creation of montaged maps was undertaken on 20 samples from around the caldera, chosen to be representative of a range of locations, units, and depositional environments. For each sample SEM SE image maps were collected of grain fractions 3 φ (125 µm), 4 φ (63 µm), and smaller than 4 φ (63 µm) in size. The magnification was varied sample by sample to balance image detail vs map size (acquisition time). Point counting was then undertaken on the SEM SE montaged image maps, using a step size approximately 1.5 times the average grain size. At each point the grain was grouped by its morphology into one of six secondary morphological subgroups; Angular, Curvi-planar, Fluidal, Sheared Vesicle, Elongate Tube-Angular, and Elongate Tube-Fluidal. Point counting was undertaken until at least 400 points had been grouped, for each size fraction, or the grains had run out.

Visual componentry via point counting was chosen over 2D shape parameter analysis as it was found to better reflect the morphological variation indicative of fragmentation in the samples at Havre.
2.5.2. **SEM BSE Microtextural observations**

SEM BSE imaging was undertaken on grains mounted in carbon coated polished briquettes for qualitative description of groundmass and vesicle microtextures. Particles from a range of samples and grain sizes were imaged to compare inter-sample and -grain size variation in the groundmass and vesicle population of the ash produced at Havre. In addition, particles from the three first order, and seven second order componentry classes were imaged to compare and qualitatively quantify variation between them.

Initial work was undertaken to quantify particle based on image process analysis for comparison with other studies (e.g. Dellino and La Volpe 1996; Dellino and Liotino 2002; Riley 2003; Liu et al. 2015b, a). Initial analysis was undertaken on SEM BSE images of grain size fractions at 3 φ (125 µm), 4 φ (63 µm), and smaller than 4 φ (63 µm) using the ImageJ macro from (Liu et al. 2015b) based on methods outlined in (Liu et al. 2015a). The results from this analysis applied to the Havre sample set defined a widely spread data set from which no conclusions on fragmentation mechanism could be drawn (beyond fragmentation been driven by a range of mechanisms). Additionally, when trying to match the grain shape data back to particles that could be visually identified as been ‘curvi-planar’, ‘angular’, ‘fluidal’, ‘elongate tube-vesicle’, etc. there was some disconnect. At the time the influence of vesicles on grain shape, and the range of vesicle populations observed in the 2012 Havre ash were postulated to account for this discrepancy. It was therefore decided to conduct SEM SE based point counting on samples since particle morphological signatures show strong implications on fragmentation mechanism. This is an area of ongoing work.

2.6. **Synchrotron microtomography**

X-ray tomography is a well-established method for the non-destructive 3D visualisation and analysis of solids. The method works based on differences in the X-ray attenuation values of material been
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examined. A sample is exposed to a high-powered X-ray source and rotated stepwise over 180° recording the shadow projections of the sample at each step. 2D horizontal slices that stack vertically through the sample are then constructed using the shadow projections with a back-projection procedure. This procedure produces a series of 2D slices through the object that was imaged that stack vertically. The 2D slices can then be examined individually or reconstructed to form a 3D rendering of object in question. The resolution of tomography scans is measured by voxel size, equivalent to a 3D pixel.

Tomography was undertaken at the Australian Synchrotron in Hutch 2B using the ruby detector, with an accelerating voltage of 30 keV on grains -1 to 1 φ (2 to 0.5 mm) in diameter. Prior to analysis samples were clean in an ultrasonic bath for ~ 2 minutes and subsequently dried overnight. Grains where then either glued in layers into a syringe or placed in a syringe separated from one another by layers of foam. For each run 1810 scans were taken over a 181° rotation with a sample source distance of 10 cm. Each analysis scanned a volume of approximately 2 cm height which was centred on layers of ash grains. Back-projection was undertaken on software accessed through MASSIVE. The reconstruction gave an image resolution of 6 μm per voxel side.

Analysis was undertaken using Avizo® Fire 8.1. The image stack was thresholded into particle and void space (air), to allow faster processing and analyses of data. By cropping an internal volume of the particle, with no clast outer surfaces, percentage porosity could be calculated by measuring the amount of grain and void space. I attempted several methods of analysing particle morphology; however, none were satisfactory. When measuring particle surface area in Avizo® Fire 8.1 the software algorithms consider all connected surfaces, and thus due to the open interconnected porosity internal vesicle surface area was also measured. Although tools do exist as part of Avizo® Fire 8.1 that can attempt to “close” open porosity, then mechanism by which the algorithms undertake this lead to the loss of significant particle shape information. As of yet a solution to this problem has not been found.
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2.7. Major element glass chemistry

Major-element chemistry of the glass was determined by electron probe microanalysis (EPMA) (WDS) using a JEOL JXA-8230 SuperProbe® Electron Probe Microanalyser at Victoria University, Wellington. Quantitative measurements were calibrated using international standards of similar composition (VG-568 and ATHOG). Measurements were undertaken using a 10 µm beam. Na⁺ and K⁺ were analysed first to reduce alkali loss.

Energy-dispersive X-ray spectroscopy (EDS) was also conducted using a Zeiss Sigma VP® Field-Emission-Gun Scanning Electron Microscope at the University of Otago. An accelerating voltage of 20 keV was used, with an aperture diameter of 120 mm, and a working distance of 8.5 mm. Factory standards were found to give the most accurate results for groundmass glass, plotting exactly within the glass range of the Havre 2012 eruption as determined using EPMA. AZtec EDS software, from Oxford Instruments, was used to analyse the spectra, and calculate the wt% of major elements.

2.8. Synchrotron radiation Fourier-Transform Infrared Spectroscopy (SR-FTIR)

Synchrotron radiation Fourier transform infrared spectrometer (SR-FTIR) analysis was conducted at the Australian Synchrotron using the Infrared Microspectroscopy (IR) beamline. The IR beamline uses a Bruker V80v Fourier-transform infrared spectrometer and a Hyperion 2000 IR microscope attached to a synchrotron source. Measurements were conducted on 15-40 µm thick doubly polished wafers, created from 2 and 3 φ ash grains. Wafer thickness was determined using an Eee® digital indicator with a resolution of 1 µm. Wafers were created by mounting an ash grain to a glass slide using Crystalbond® cement. The ash grain was then polished to a smooth flat surface, finishing on a 1 µm diamond grit-impregnated paper. The crystal bond was then dissolved using acetone and the grain flipped and mounted polished face flush to the glass slide in crystal bond. The grain was then polished two produce a doubly polished wafer.
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Absorbance spectra were collected from point scans of the groundmass glass. Point scans were undertaken in transmission mode over a wavelength range of $\nu = 6000$–700 cm$^{-1}$ with a spatial resolution of 4 µm. Absorbance spectra were visual examined using OPUS Spectroscopy Software® and samples in which the 3550 cm$^{-1}$ H$_2$O$_{total}$, and 1630 cm$^{-1}$ H$_2$O$_{mol}$ peaks were overly noisy were discarded. In ‘good’ samples the height of the 3550 cm$^{-1}$ and 1630 cm$^{-1}$ peaks were measured from a baseline chosen by eye. Concentration of the total dissolved water and the dissolved molecular water were determined through application of the Beer-Lambert law for the 3550 and 1630 cm$^{-1}$ wavelength peaks, respectively (Wysoczanski and Tani 2006). The species dependence of the 3500 cm$^{-1}$ was also accounted for (McIntosh et al. 2017).

The SR-FTIR data has also been examined to assess for sample hydration. Molecular water is by far the more mobile of the water species at low temperature (Zhang et al. 1991; Anovitz et al. 2006, 2008; Zhang and Ni 2010; McIntosh et al. 2014). Assuming equilibrium degassing, H$_2$O$_{mol}$, measured using the 1630 cm$^{-1}$ peak, was used to calculate the temperature of apparent equilibrium ($T_{ae}$) (Zhang et al. 1997; Ihinger et al. 1999; McIntosh et al. 2014). $T_{ae}$ is approximately equal to the glass transition temperature ($T_g$) (Zhang et al. 1997; Ihinger et al. 1999; McIntosh et al. 2014). The $T_{ae}$ value calculated was then compared with the calculated magma temperature. Although cooling rate impacts on the $T_g$, its value will only vary by 10’s of degrees (Gottsmann and Dingwell 2001, 2002). If the $T_{ae}$ calculated from the H$_2$O$_{mol}$ varies greatly from the eruption temperature additional molecular water has likely been added to the glass’s structure following quenching. The glass can therefore said to be hydrated, and the total water content will be inflated from its equilibrium at quenching (Zhang et al. 1991; Anovitz et al. 2006, 2008; Yokoyama et al. 2008; Bindeman and Lowenstern 2016). For a hydrated glass it is possible to reconstruct the equilibrium speciation of the magma using the OH wt%, assuming OH was not altered by hydration below the $T_g$ (Zhang et al. 1991; McIntosh et al. 2014).
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Chapter 3

Characteristics and stratigraphy of submarine-erupted silicic volcaniclastic deposits, Havre volcano, Kermadec Arc, New Zealand.

This chapter presents work prepared for publication as a manuscript.

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Author contributions:

RJC organised and led the cruise. All authors were present on the 2015 cruise and collected the samples. JDLW established the scope of the project. All sample analysis was conducted by APM. APM wrote the manuscript and produced the figures. All authors read through draft and assisted in editing of the manuscript.
Chapter 3

3.1. Abstract

In 2012 the largest deep subaqueous silicic eruption of the last century occurred 700-1220 meters below sea level at Havre volcano, Kermadec Arc New Zealand. Pre- and post-eruption bathymetry surveys, along with sampling by a remote-operated vehicle showed 14 seafloor lavas and three major seafloor clastic deposits, making the 2012 Havre eruption ideal for the study of processes and products of subaqueous volcanism. This paper presents results and inferences on stratigraphy, fragmentation, and dispersal from a stratified seafloor ashy deposit, the Ash and Lapilli (AL) Unit.

Seafloor images show the Ash and Lapilli unit consists of multiple subunits, all ash-dominant. Sampling destroyed stratigraphy in all but two samples, yielding mixed samples of all subunits present at the location. By combing seafloor imagery with granulometry and componentry from both the stratified samples and the remaining mixed samples, the spatial distribution and relative stratigraphy of the subunits throughout the study area was established. We distinguish, from base to top, Subunits 1 (S1), S2, S3, S4a, and S4b.

The relationship of the four subunits to other seafloor products of the 2012 Havre eruption allow us to build a relative stratigraphic framework showing a temporal development in the eruption mechanisms. We show that explosive fragmentation of a glassy vesicular magma generated a buoyant thermal plume and dilute density currents from which Subunit 1 and 2 were deposited respectively. Following a time break (days/weeks?) effusion of lava on the southern caldera rim lead to ash generation initially through syn-extrusive ash venting, quenching, brecciation, and comminution (S3 and S4b), and following this through gravitational collapse (S4a). The results presented here provide a detailed insight into the development of eruption, fragmentation and depositional processes through the largest deep subaqueous eruption of the last century.
3.2. Introduction

Volcanic eruptions into a deep subaqueous environment are complex. Submarine eruptions are significantly modulated by the physical properties of water both indirectly (hydrostatic pressure, increased viscosity of water relative to air), and directly (rapid heat transfer, rapid volume expansion of water) (Head and Wilson 2003; White et al. 2003; Cas and Giordano 2014; White et al. 2015). Hydrostatic pressure will suppress the rates of volatile exsolution and expansion, and possibly explosive expansion and related fragmentation (Fisher 1984; Staudigel and Schmincke, 1984). Rapid heat transfer on direct contact between magma and water however, can induce both explosive (Zimanowski et al. 1997; Austin-Erickson et al. 2008) and passive fragmentation (van Otterloo et al. 2015).

The cost and difficulty of collecting well constrained samples linked directly to subaqueous eruptive dynamics, and the complexity of modelling these processes, (both physically and computationally) has led to much debate regarding eruptive processes in the subaqueous environment (Head and Wilson 2003; Allen and McPhie 2009; Schipper et al. 2010; Rotella et al. 2013; White and Valentine 2016). Due to the complexity and cost, much of our understanding comes from studies of uplifted subaqueous volcanic successions in Japan (Kano et al. 1996; Cas et al. 2003; Jutzeler et al. 2015); the Mediterranean (Allen and McPhie 2000; Allen and Stewart 2003; Stewart and McPhie 2004; Allen and McPhie 2009); and Australia (Cas 1978; Simpson and McPhie 2001); among others.

In this paper we present data on proximal ash deposits sampled three years after the deep submarine eruption of Havre volcano (Carey et al. 2014; Carey et al. 2018). Sampling by remote-operated vehicle (ROV) Jason, guided by high-resolution bathymetry from the autonomous underwater vehicle (AUV) Sentry (Carey et al. 2018), shows that different layers of ash formed during the eruption. The differing distributions of the ash layers, together with textural differences among deposits and their ash particles, allow us to interpret their origins during the eruption sequence. This raises questions regarding the mechanisms operating during the 2012 Havre eruption, and provides a
detailed stratigraphic framework of the eruption allowing us to the temporal evolution in processes of the largest silicic subaqueous eruption of the last century.

3.3. Geological Setting

Havre is a fully submerged volcano located at 31° 05'S 179 °5'W (-31.10, -179.03) along the Kermadec arc (Fig. 3.1.) (Wright et al. 2006). The edifice rises from a 1500-2000 meters below sea level (mbsl) sea floor to a peak that is truncated by a caldera at 600 mbsl (Wright et al. 2006). The caldera is four km long and three km wide, elongate northwest-southeast, with an average depth below caldera rim of 900 m (Wright et al. 2006). The caldera floor is at 1500 mbsl (Fig. 3.1.) (Wright et al. 2006).

In 2012 an eruption occurred at Havre volcano (Carey et al. 2014; Jutzeler, Marsh, et al. 2014; Carey et al. 2018). At 1050 18 July 2012 UTC (universal standard time) an eruption plume and hot spot were observed in satellite images emanating from a point source above Havre. An image taken at 2151 on the same date showed an extensive pumice raft, 400 km² in area. Satellite imagery indicates that eruptive activity affecting the sea-surface, including origination of an atmospheric plume, a pumice raft and a plume of ash stained discoloured water, extended over 21.5 hours. An image taken at 0209 20th July 2012 (CST) shows the pumice raft had detached from its source and the atmospheric plume had ended, indicating the eruption had either ended or was no longer powerful enough to produce effects at the sea surface. From 17th to 21st July frequent earthquakes of magnitude three to five were also recorded from Havre. After 21st July 2012 there is no further evidence of volcanism at Havre in satellite imagery.

A comparison of bathymetry surveys conducted in 2002 (Wright et al. 2006) and after the eruption, on 26th October 2012, shows that there had been large scale topographic changes on the volcano summit (Carey et al. 2014; Carey et al. 2018). Several dome/cone shaped features had been produced on the southern caldera rim along with a large bulge on the southwest caldera wall. The
Fig. 3.1. The location of Havre volcano along the Kermadec Arc with an insert showing a MODIS (Aqua) image taken at 0126 on 19th July of the 2012 Havre eruption. A 1 m scale resolution bathymetry map of the Havre caldera and summit (Carey et al. 2018) overlain on a lower resolution (35 m) bathymetry map of the whole of Havre volcano (Carey et al. 2014). Overlain is shown the bathymetry differences between the 2002 (Wright et al. 2006) and Oct 2012 surveys (red = material added, purple = material removed). The locations of all clastic samples taken at Havre are shown along with the sampling method used. Samples in which detailed componentry was undertaken are labelled. The outlines of the ABL Unit and GP Unit outlines are marked in white.

Seafloor products of the Havre eruption were subsequently mapped at high resolution, imaged and sampled in a 2015 cruise using ROV Jason and AUV Sentry (Fig. 3.1.). Fourteen lavas were erupted (A to P) along the southern and southwestern caldera rim, following two apparent structural lineaments (Fig. 3.1.). Three clastic units were mapped on the seafloor. The first is a widespread deposit composed of giant pumice clasts greater than 1 m in diameter (Giant Pumice Unit (GP Unit)) extending and coarsening to the NW and inferred to enclose Dome OP in the southwest (Fig. 3.1.). The contact of the GP Unit with the pre-2012 substrate however was not observed in detail, or sampled, during the cruise. Surrounding Dome OP is a locally dispersed lobate unit, the Ash, Lapilli and Block Unit (ALB Unit) (Fig. 3.1.). The ALB Unit overlies the GP Unit, with several lobes extending from Dome OP on the edifice and into the caldera. The Ash and Lapilli Unit (AL Unit), which over most of its extent consists almost entirely of ash, is the most widespread unit, found in every collected sample. The AL Unit overlies the GP Unit and is composed of four subunits, the upper three of which also overlie the ALB Unit; the relationship of the basal subunit of the AL Unit to the ALB Unit however is unclear (Fig. 3.1.). Since the basal contact of the GP Unit was not observed in detail it is not known whether the AL Unit represent the earliest ash deposit from the eruption, or whether further ash deposits underlie the GP Unit. The AL Unit is focus of the present study (Fig. 3.1.).
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3.4. Results

3.4.1. Establishing strata of the AL Unit

Seafloor observations did not reveal natural vertical exposures through the AL Unit (Fig. 3.2a-d.). Layering of the AL Unit however was observed after sampling at several sites (HVR042, HVR132, HVR163, HVR196, HVR229, HVR232 and HVR272) that produced small vertical incisions. In addition, two push cores (HVR134 and HVR159) preserved distinct layers. By combining observations of the seafloor and preserved samples four distinct layers have been identified in the AL Unit.

Sampling from the tops of giant pumice clasts at sampling sites HVR132 and HVR163 exposed the most complex stratigraphy, showing four layers in the AL Unit with similar features and thicknesses at both locations (Fig. 3.2e-f.). At these locations, 2.5 km apart on the caldera floor, the basal layer directly on top of the giant pumice is about 4 cm thick, relatively coarse-grained (coarse ash) and light in colour. Overlying this layer is a 2 cm thick highly cohesive layer with approximately 5 vol.% of coarser dark particles. Above this layer is an approximately 2 cm thick, coarser-grained coarse ash layer, comprised of dominantly dark coloured grains with rare light-coloured grains (Fig. 3.2e, 2f). The top layer is approximately 4 cm thick. Similarity to the second layer the top layer is highly cohesive and fine-grained. In both locations the upper surface of the top layer appears light brown, while lower parts appear light grey.

Sampling of HVR196, proximal to Dome OP to the southeast, from on top of a GP clast exposed two layers in the AL Unit (Fig. 3.2c.). The basal layer directly overlying the GP is an approximately 2 cm thick layer of white fine lapilli and coarse ash in a grey matrix of finer ash. Overlying this is a 1 to 2 cm thick grey fine grained cohesive layer, the upper surface of which appears light brown. The two layers observed here have similarities with the coarse basal and fine grained top layers observed in HVR132 and HVR163, however importantly only a single cohesive fine-grained layer is observed.

Sampling from on top of GP clasts at HVR229, HVR232, and HVR272 all on the northwest caldera rim exposed only a single layer (Fig. 3.2d.). At HVR229 the layer is 1.5 cm thick, while at HVR232 and
Fig. 3.2. Images of the seafloor taken on front facing cameras showing images of AL Unit stratigraphy exposed during sampling from on top of GPs at locations HVR132 (a), HVR163 (b), HVR196 (c), and HVR272 (d). At both HVR132 (a) and HVR163 (b) a similar stratigraphy can be observed showing four layers with comparable deposit thicknesses, apparent grain size and colour. At HVR196 (c) only two layers can be seen, however their characteristics appear similar to the bottom and tops layers in HVR132 and HVR163. At HVR272 (c) only a single layer can be seen. (e and f) show a clastic deposit consisting of lapilli and ash with dominantly elongate tube morphologies at HVR070 (e) and overlying the carapace of Lava G (f). In (g-k) the variation in the AL Unit coverage overlying lavas around the caldera is shown. Thick deposits of the AL Unit can be seem overlying an apparently older part of Lava N (g) and overlying a lava produced prior to the 2012 Havre eruption (h). Over the more recent part of Dome N (i), along with Domes M (j), and I (k) however the AL Unit is thinner and patchier. Dome morphologies were chosen to by similar so as to provide a consistent context for observing variations in the AL Unit.

HVR272 the layer is approximately 3 cm thick. The layer is a cohesive light grey fine ash deposit, the upper surface of which appears light brown. The single layer observed at HVR229, HVR232, and HVR272 has similar characteristics to the top layer observed in HVR132 and HVR163.

In addition to the four layers present in HVR132 and HVR163 a further deposit was observed on the seafloor overlying Lava G and at location HVR070 approximately 150 m to the northwest of Lava G (Fig. 3.2e-f.), containing highly elongate tube-pumice ash and lapilli. At HVR070 this deposit was approximately 0.5 m thick. At both locations the deposit of highly elongate tube-pumice ash and lapilli was overlain by a cohesive fine-grained layer, however no other layers were observed.

Images of the AL Unit overlying lavas produced in the 2012 Havre eruption show variation in deposit coverage and thickness over similar lava morphologies compared with lava produced prior to the 2012 eruption. The AL Unit overlying the southern lobe of Lava N and a pre-2012 lava on the southern caldera rim shows thick and consistent accumulations (Fig. 3.2g-h.). Comparatively the AL Unit overlying Lava H, I, K, M, the central part of N, and OP (Fig. 3.2i-k.) show thin patchy deposits.
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Pushcore HVR159 from the southwest caldera rim, was taken through a thick deposit of the AL Unit, outside the GP Unit boundary and away from any significant slopes. At site HVR159, the push core sampled and preserved at least four layers, the upper three of which are visible in Fig. 3 (the basal part of the sample mostly collapsed when removed from the push core). Observations of the mostly collapsed basal part of pushcore sample HVR159 show that it is a light-coloured layer between 4 and 5 cm thick composed of fine to coarse ash. Stratigraphically overlying this layer is a 1 cm thick layer that is rich in elongated coarse ash (Fig. 3.3.). A 2.5 cm thick dark coloured medium to coarse ash layer with a sharp basal contact then overlies this layer (Fig. 3.3.). The uppermost layer is then overlain by a cohesive 2 cm thick layer at the top of HVR159 (Fig. 3.3.).

Pushcore HVR134 was taken on the caldera floor, northeast of Lava C through a thick deposit of the AL Unit at the base of a slight slope. The seafloor sample HVR134 is taken from an area that shows significant hummocks and undulations ahead of Lava C. The push core shows only two layers, a basal grey cohesive fine-grained layer approximately 20 cm thick. This is overlain by an 8 cm thick dark grey layer.

In pushcore HVR159 similar layer stratigraphy, thicknesses, and characteristics to those in vertical exposures through the AL Unit at HVR132 and HVR163, are observed (Fig. 3.2a-b and 3.3.). The basal fine to coarse ash layer, the 2.5 cm thick dark layer, and the cohesive upper layer are all consistent with seafloor observations. In pushcore HVR134 the layers have similar characteristics, however their stratigraphic relationship and thicknesses contrast with those observed in push core HVR159 and in seafloor images of the AL Unit.

3.4.2. Granulometry and componentry of HVR134 and HVR159

A critically useful feature of the two samples that did preserve layering (HVR134 and HVR159), is that the layers have distinctive characteristics. Both HVR134 and HVR159 pushcores were subsampled from their base and top. The material that remained between the subsamples formed the bulk
sample. Granulometry and componentry was conducted on both the bulk and subsamples of HVR134 and HVR159 from which common layer characteristics could be established.
Fig. 3.3. Images taken of pushcore sample HVR159 displaying the stratigraphy. (a) shows a broad view of the upper part of HVR159 in which three layers can be seen. These layers match in colour, apparent grain size and thickness what was observed on the seafloor at locations HVR132 and HVR163. (b) A zoom in on the layer rich in elongate tube particles. This is reflected in the componentry data where both HVR134 and HVR159 show a concentration of Elongate Tube particles in the bulk sample.

Granulometry results of bulk samples from both HVR134 and HVR159 show two main grain size peaks, determined using GRADISTATv8 (Blott and Pye 2001), at 6 to 5 φ (16 to 32 µm), dominate in HVR134, and at 2 to 1 φ (250 to 500 µm), dominate in HVR159. A minor peak also occurs between 0 to -1.5 φ (1 to 2.8 mm) (Fig. 3.4.). In both samples the basal subsample displays a bimodal grain size distribution with modes at 2 to 1 φ and 0 to -1.5 φ (Fig. 3.4.). The top subsample of both HVR134 and HVR159 displays a unimodal distribution with the mode at 6 to 5 φ. The granulometry modes can be used to identify specific layers in mixed samples taken from around the caldera.

Componentry was conducted in 1 φ steps from -2 to 2 φ (4 mm to 250 µm) for 27 samples. Grains were counted into three first-order groups: Glassy Vesicular, Microcrystalline, and Elongate Tube-Vesicle particles (Fig. 3.5.). These first-order groups are subdivided into secondary classes based on particle vesicularity and morphology (Fig. 3.5.). No lithic component was recognised in any sample examined.

Glassy Vesicular grains are white to creamy grey glass of moderate to high vesicularity (Fig. 3.5.).

Microcrystalline grains are black to dark grey, microcrystalline, and weakly to non-vesicular (Fig. 3.5.).

Elongate Tube-Vesicle clasts are white to cream-coloured glass with elongate shapes defined by tube vesicles; they have a woody/fibrous appearance (Fig. 3.5.).

Componentry of individual layers preserved in the HVR159 and HVR134 can be examined and quantified similarly to granulometry (Fig. 3.4.). Componentry results in HVR159 show an increase in the proportion of microcrystalline clasts from 11% in the base subsample, to 23% in the bulk sample (middle), to 62% in the top. A simultaneous decrease in glassy vesicular clasts is observed over the
Fig. 3.4. The grain size distribution and componentry data for sample HVR159 and HVR134 and their subsamples. In both samples similar patterns can be observed; an upwards fining in grain size, a
decrease in Glassy Vesicular particles with a concurrent increase in microcrystalline particles from bottom to top and a concentration of Elongate Tube particles in the bulk sample. The grain size range over which componentry was undertaken is displayed for both samples.

same trend from 83% at the base, 66% in the middle, and 25% at the top. The broad trend of increasing microcrystalline clasts and decreasing glassy vesicle clasts from the base to the top of samples is repeated in sample HVR134. For Microcrystalline clasts increase from 42% (base) to 45% (middle), to 54% (top). Componentry results from HVR159 bulk sample (middle) also shows a higher percentage of Elongate Tube-Vesicle clast, 10% compared to 5% and 4% in the base and top respectively (Fig. 3.4). The enrichment in Elongate Tube-Vesicle particles in the bulk sample of HVR134 however, is not well resolved.

3.4.3. Nomenclature and stratigraphy of layers within the AL Unit

By combining seafloor observations with qualitative descriptions, quantitative grainsize and componentry data from preserved stratigraphy in push cores, the characteristics and stratigraphic relationship of four subunits within the AL Unit have been identified. The fact that granulometry and componentry results in pushcore HVR134 and HVR159 equate well with visually identified layering in the AL Unit, indicates the pushcore layering is primary, resulting from deposition and not formed during sampling (Jutzeler, et al. 2014). The results of the identification and characterisation of four AL Unit subunits are outlined below.

Subunit 1 (S1) – The basal layer in seafloor images of the AL Unit and in sample HVR159 is a at least 6 cm thick light-cream coloured deposit of coarse ash (Fig. 3.2a-c and 3.3). The bimodal grainsize distribution has two modes at 2 to 1 φ and 0 to -1.5 φ (the large mode is subdued) (Fig. 3.4). Overall S1 is dominated by glassy vesicular clast types. Subunit 1 is also shown in seafloor images HVR132 and HVR163 to directly overlie the GP Unit.
Fig. 3.5. Optical images showing the componentry classes and their morphological subclasses in the -1 φ to 0 φ (2 to 1 mm) range.
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Subunit 2 (S2) – Subunit 2 overlies S1 across a gradational contact, observed on the seafloor as a cohesive fine-grained deposit (Fig. 3.2a-c and 3.3.). Observations from HVR132 and HVR163 show two cohesive fine-grained layers, the first overlying S1 (S2a), and the second at the top of stratigraphy (S2b), separated by subunit 3 and 4 (Fig. 3.2a-c and 3.3.). Subunit 2 as a single layer will be presented in more detail below however, beyond the dispersal of S3 and S4, S2 occurs as a single deposit with the upper and lower layers indistinguishable from one another. This can be seen at HVR196 where only a single cohesive fine-grained layer is observed (Fig. 3.2c.). Subunit 2a is approximately 2 cm thick, while S2b ranges between 2 to 10 cm thick. Direct sampling of S2b shows a unimodal grain size mode of 6 to 5 φ. The fineness of S2 means it was not possible to conduct granulometry on it.

Subunit 3 (S3) – In pushcore HVR159 S3 is an approximately 3 cm thick layer rich in elongate clasts in the middle of stratigraphy (Fig. 3.3.). The granulometry of S3 will be presented in detail below as it shows a strong spatial fining trend. Subunit 3 is composed of Elongate Tube-Vesicle clasts. Subunit 3 occurs overlying S2a, although the contact has not been directly observed.

Subunit 4 (S4) – A 2 cm thick layer composed of medium/coarse dark coloured ash observed at seafloor locations HVR132 and HVR163 (Fig. 3.2a-b.), as well as in pushcore HVR159 (Fig. 3.3.). Subunit 4 directly overlies S3 across a sharp boundary. The granulometry of S4 is hard to resolve since it has not been directly sampled. Subunit 4 is composed of microcrystalline particles.

3.4.4. Defining subunits in ‘mixed’ samples

We use the observed seafloor stratigraphy at HVR132, HVR163, along with the combination of grainsize and componentry characteristics from pushcore HVR159 to establish the presence or absence of subunits in most other samples where stratigraphy was not preserved (mixed samples). For example, the presence of the 6 to 5 φ grain size mode indicates the presence of S2 subunit (Fig. 3.6a.). After the establishment of subunits in mixed samples, we trace the distribution of these
subunit-diagnostic features across the study area (Fig. 3.6b.). An important consideration of this method is that at each location the proportional depth of sampling within the overall deposit is unknown (Fig. 3.6a.), and deeper layers may not have been sampled; however broad trends can be established.

Fig. 3.6. A schematic diagram of an idealized method for describing several subunits in mixed samples. (a) In a mixed sample the bulk grainsize characteristics will be reflective of the sum of each individual layers sampled. If layer unique granulometry features are known the bulk grainsize distribution can be examined to assess which are present in that sample. (b) By plotting the samples and which layers are present spatially the distribution of layers around the study area can be established. Patterns in the spatial distribution of the granulometry may reflect the presence or lack of a subunit from a certain location allowing mapping of deposit dispersal even in mixed samples.
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Granulometry results from 81 mixed samples of the AL Unit show that they are composed of more than 90% ash with complex multi-modal grain size distributions. The grainsize distributions are unimodal and bimodal, with common modes identified at 6 to 5 φ (16 to 32 µm) associated with Subunit 2, and between 3.5 and 0.25 φ (88 to 840 µm) associated with Subunit 1 (Fig. 3.7.). Seven ash-dominated samples also show grain size modes in the 0 to -2 φ (1 to 4 mm) range. The presence of subunits as defined by granulometry, were confirmed with the presence of the various component types which are also distinct to subunits (see section 3.4.3, Fig. 3.8.).

Subunit 1

Seafloor images show S1 ranges in thickness from 2 to 6 cm (Fig. 3.2a-c.). Common seafloor ripples and strong seafloor currents encountered during ROV Jason dives suggest deposit reworking and may indicate that observed layers may not preserve their full, original thickness. The thickness of HVR159 (6 cm) is therefore taken as the most representative since it is the only one not taken away from GP clasts. The glassy vesicular clasts that characterise subunit 1 have been identified in all clastic samples, indicating it was dispersed across all the study area and extends beyond it (Fig. 3.8.). The grain size mode of S1 in pushcore HVR159 was identified as between 2 to 1 φ. Examination of the mixed samples show a consistent slight variation of this mode between 0.25 φ and 3.5 φ documenting a fining trend towards the northwest (Fig. 3.7.).

Subunit 2

Subunit 2 is divided into lower (a) and upper (b) when separated by subunits 3 and 4 that form more localised deposits. Where S3 and S4 are not present, S2a and S2b cannot be distinguished from one another, and S2 occurs as a single layer. Subunit 2a was observed at sites HVR132 and HVR163 (Fig. 3.2e-f.). At both locations the subunit thickness is approximately 2 cm (Fig. 3.2e-f.). Subunit 2b is 4-10 cm thick on the caldera floor and 2 to 3 cm thick on the south, east and west caldera rims. Subunit 2 particles are characterised by a consistent grain size mode at 6 to 5 φ. Subunit 2 can be identified in all clastic samples south of a boundary that roughly follows the east-west trend of the northern caldera wall (Fig. 3.7.). To the south, east and west, Subunit 2 is present to the edge of the
Fig. 3.7. Showing the decrease in the grain size of the modal peak between 0.25 φ and 3.5 φ across the Havre caldera. Representative grain size distributions are shown with the 0.25 φ to 3.5 φ and 5 to 6 φ modes indicated. The lack of the 5 to 6 φ mode on the northern caldera rim can be seen in samples HVR229 and HVR272. The red line in the grain size distribution denotes the cross over from Mastersizer ® to sieving data.

investigated area with no notable change in thickness or grain size. It is inferred to extend well beyond the area, similarly to S1.

Subunit 3

Subunit 3 can be observed in-situ in HVR159 where it has a thickness of approximately 3 cm. In mixed samples, componentry shows that Elongate Tube-Vesicle particles are a minor fraction (6 to 15%) of samples taken from the southwest caldera rim, caldera floor, and a single sample taken on the northeast caldera rim (Fig. 3.8.). The highest concentration of Elongate Tube-vesicle clasts occurs at HVR070 (50%), where the deposit is comprised of ash to coarse lapilli approximately 0.5 m thick (Fig. 3.2e and 3.8.). A similar deposit is observed overlying Lava G (Fig. 3.2f.); however, this was not sampled. Three sample taken south of Dome OP in the southeast of the study area also show very minor fractions (4 to 6%) (Fig. 3.8.). Combining granulometry and componentry results of mixed samples a fining trend in Elongate Tube-Vesicle clasts away from HVR070 is revealed.

Subunit 4

Subunit 4 is an approximately 2 cm thick layer observed on the seafloor at locations HVR132, and HVR163 along within pushcore HVR159. The componentry shows that the microcrystalline grains that are used to define Subunit 4 occur in two disconnected areas (Fig. 3.8.). The first area trends northeast across the caldera floor from the southwest caldera rim and is called subunit 4a (Fig. 3.8.). The second area is observed around Dome OP and is called subunit 4b (Fig. 3.8.). All direct in-situ observations of Subunit 4 were recorded within the subunit 4a area (Fig. 3.8.). Componentry result
however show no distinguishable difference between the clast characteristics of S4a and S4b (Fig. 3.8.).

Subunit characteristics are summarised in table 3.1.

Fig. 3.8. Spatial distribution of componentry data plotted by sample location. Concentrations in Elongate Tube-Vesicle particles (green) can be observed around Lava G. Microcrystalline particles (red) concentrations are present around Dome OP and the lava flows A-E on the southwest caldera wall, and caldera floor. Glassy Vesicular particles (green) are found in every sample and are the dominate component of the overall deposit.

3.4.5. Grain morphology

Glassy vesicular, elongate tube-vesicle, and microcrystalline particles are split into subclasses based on morphology or vesicle form. Glassy vesicular grains show four subclasses; curvi-planar, angular, and fluidal particles (Fig. 3.5 and 3.9.). Curvi-planar clasts are defined by planar and curvi-planar surfaces that intersect to form sharp edges (Fig. 3.5a. and 3.9a-c.) and include both platy and sub-
<table>
<thead>
<tr>
<th>Subunit</th>
<th>Dispersal</th>
<th>Stratigraphic relationships</th>
<th>Depositional characteristics</th>
<th>Grainsize/Componentry characteristics</th>
</tr>
</thead>
<tbody>
<tr>
<td>S1</td>
<td>Entire study area</td>
<td>Basal contact not seen. Overlies the GP Unit. Relationship the ALB Unit unknown</td>
<td>At least 5 cm thick at all sites observed. No evidence observed of thinning trends. Drapes topography. No internal sedimentary features observed.</td>
<td>Grain size of 0.5 to 3.5 φ. Composed of glassy vesicular ash. No lithic clasts.</td>
</tr>
<tr>
<td>S2</td>
<td>Sharp boundary at the northern caldera wall. South of the boundary deposit extends to the edges of the study area.</td>
<td>S2 is split in to lower (S2a) and upper (S2b) sections. Forms current seafloor. Overlying S1 and ALB Unit. S2a has diffusive contact with S1. S2b overlies domes.</td>
<td>S2a thickness of ~ 2 cm on the caldera floor. Thickness poorly resolved on rim. S2b thickened on the caldera floor 3 – 14 cm thick vs ~2 cm on rim. Thickest on Lava C (10-14 cm). Ripples occasionally observed on upper surface (seafloor). Appears to drape GP clasts to some degree. No internal sedimentary features observed.</td>
<td>Characteristic grain size of 6 φ. Composed of glassy vesicular ash. No lithic clasts.</td>
</tr>
<tr>
<td>S3</td>
<td>Deposit extends NE-SW across the caldera with boundaries approximately at either caldera wall. In proximal locations overlying S2a. Distally a diffusive layer at the top of S2a.</td>
<td>Thickens towards the area of Lava G. &gt;0.5 m thick at HVR070. Diffusive layer of unknown thickness at HVR283 (most distal). S3 deposited on topography ~50 m higher than Lava G. Topography has little influence on grain size or thickness. S3 drape topography. No internal sedimentary features observed.</td>
<td>Maximum grain size drops from &gt;70 mm on Lava G to &lt;250 µm at HVR283 (most distal). Characterized by elongate tube-vesicle particles. No lithic clasts.</td>
<td></td>
</tr>
<tr>
<td>S4a</td>
<td>Deposit extends NE-SW across the caldera. Thins towards the NE Overlying S3. Overlain by S2b, Sharp boundary with S3 and S2b.</td>
<td>Subunit ~2 cm thick on the caldera floor (HVR132 and 163) and on the SW rim (HVR159). Topography does not appear to affect thickness. S4 drapes topography.</td>
<td>Characterized by microcrystalline ash. No lithic clasts.</td>
<td></td>
</tr>
<tr>
<td>S4b</td>
<td>Surrounding Dome OP and extending towards the north down slope. Overlying the ALB Unit.</td>
<td>Visually the deposit appears to thin away from Dome OP. Deposit elongated downslope to the north. Deposit poorly observed.</td>
<td>Metre scale blocks proximal to Dome OP. Maximum grain size reduces away from Dome OP Characterized by microcrystalline ash No lithic clasts.</td>
<td></td>
</tr>
</tbody>
</table>
equant blocky clasts. Vesicles in curvi-planar clasts are cross-cut by fracture surfaces, which show no
def ormation around the bubble (Fig. 3.9a.). Angular clasts have prominent concavities defined by
brittle-fractured vesicle walls (Fig. 3.5b. and 3.9d-f.). Fluidal clasts have exterior features indicating
surface tension or hydro/aerodynamic reshaping of the grains while molten. Fluidal clasts include
both those with a wholly fluidal form, and those that preserve a single fluidal surface (Fig. 3.5c and
3.9g-h.). Fluidal particles are commonly cross cut by undeformed curvi-planar fracture surfaces (Fig.
3.9g-h.).

Microcrystalline particles, by contrast, show only two subclasses; curvi-planar and angular grains (Fig.
3.5e-f.). Curvi-planar clasts are typically weakly- to non-vesicular, defined by planar and curvi-planar
surfaces that intersect to form sharp edges and include both platy and sub-equant blocky clasts (Fig.
3.5e.). Angular clasts generally show moderate-vesicularity, which defines complex particle shapes
the result from vesicle walls (Fig. 3.5f.).

Elongate-tube particles are categorised into three different subclasses; elongate tube-angular,
elongate tube-ribbed and elongate tube-fluidal (Fig. 3.9j-l.). Elongate-tube angular particles are
elongate, with concave surfaces defined by brittle-fractured bubble walls (Fig. 3.9j.). Elongate-tube
ribbed grains show surface ribs the run parallel to the vesicle and clast elongation direction (inferred
to be outer tube-vesicle walls) (Fig. 3.9k.). The surface ribs have smoothly undulating surfaces and
are typically unmarked by vesicles. Elongate-tube fluidal particles are elongate, unmarked by
vesicles, with flowing molten surfaces that form peaks or droplet-like features, and show evidence of
ductile necking (Fig. 3.9l.).

Point counting by particle morphology was conducted on scanning electron microscope (SEM)
secondary electron (SE) maps for grains sizes of 3 φ (125 µm), 4 φ (63 µm), and smaller than 4 φ (63
µm) from samples that show negligible microcrystalline clasts. Removing elongate tube-vesicle
(previous page) Fig. 3.9. Scanning electron microscope SE images of the different componentry classes; Curvi-planar (a-c), Angular (d-f), Fluidal (g-i) and Elongate Tube (j-l). Curvi-planar particles can be both moderately vesicular (a) and dense (b and c), fracture surfaces cross cut vesicles. Angular particles are bound by fractured vesicle walls with their morphology dominantly controlled by vesicle texture (d-f). Fluidal particles are defined based on features that are indicative of molten behaviour syn/post-fragmentation. This includes ductile reshaping (g), particle welding (h), and post fragmentation vesicle inflation (i). (j-l) A range in the clast surface textures of Elongate Tube particles can be seen. Note the different scale across images.

particle show that S1 and S2 are composed dominantly of Curvi-planar particles in the less than 3 φ size range (Fig. 3.10.). SEM componentry shows that Curvi-planar particles make up between 50 and 86% of the total sample, with a relatively consistent over each grain size (Fig. 3.10.). Over the same grain size range angular clasts in S1 and S2 compose between 2 and 45% showing an increase in percentage with decreasing grain size going from an average of 12% at 3 φ to 22% at smaller than 4 φ (Fig. 3.10.). Fluidal clasts compose between 3 and 35% of clasts in S1 and S2, showing a decrease with particle size going from an average of 19% at 3 φ to 7% at smaller than 4 φ (Fig. 3.10.).

3.4.6. Qualitative microtextural descriptions

Microtextural analysis has been conducted on ash of the AL Unit for a range of grain sizes (-1 φ (2 mm) to particles smaller than 4 φ (63 μm)). Each grain componentry class shows distinct vesicle and microlite textures described below. In all clast types phenocrysts compose less than 5% area and generally comprise clusters of euhedral plagioclase and pyroxene 70-300 μm in size.

The groundmass of glassy vesicular clasts is composed of more than 90% glass, with a microlite population of acicular plagioclase and pyroxene (Fig. 3.11a-b.). Glassy vesicular clasts are typically moderately- to highly-vesicular and show a wide range of vesicle populations, textures, and degree of vesicle deformation. Vesicles are typically sub-round to round in 2D and range in cross-sectional diameter from less than 6 μm up to approximately 500 μm. Vesicles smaller than 20 μm in diameter
Scanning electron microscope point counting componentry data for fluidal, angular, and curvi-planar particles from samples composed dominantly of glassy vesicular clasts (S1 and S2), shown by sample.

are typically isolated, while larger vesicles display more-complex shapes resulting from coalescence and bubble interaction. Vesicles in Fluidal Glassy Vesicular clasts exhibit a range of features indicating ductile behaviour of the melt during and after fragmentation, such as inflated bubble walls which deformed outer clast surfaces, and dense rims around highly vesicular clast cores (Fig. 3.11g-h.). Some fluidal grains also display several domains in single clasts defined by vesicular cores surrounded by a convex dense fluidal rim (Fig. 3.11h.).

The groundmass of microcrystalline grains is formed of 8-35% acicular plagioclase, pyroxene and Fe-Ti oxide microlites (Fig. 3.11c-d.). Plagioclase microlites display swallowtail and hopper forms (Fig. 3.11c-d.). The characteristics and textures of microcrystalline particles varies greatly between grains (Fig. 3.11c-d.). A single grain also shows apparent mingling of two melts of differing microlite populations (Fig. 3.11d.) Both vesicle and groundmass hosted cristobalite can be observed in approximately 20% of observed microcrystalline clasts. Vesicles in microcrystalline clasts are generally isolated from one another and have ragged forms that result from the interaction of bubble walls with the microlite population (Fig. 3.11d.).

Elongate Tube-Vesicle clasts have greater than 95% groundmass glass with dominantly acicular pyroxene microlites and minor plagioclase (Fig. 3.11e-f.). Elongate Tube-Vesicle clasts show generally weak- to moderate-vesicularities. Vesicles are generally highly elongate showing tube to pipe like morphologies in 3D, with lengths of between approximately 10 µm to traversing the whole length of clasts. The microlites in Elongate Tube-Vesicle clasts are aligned parallel with the vesicle- and clast-elongation direction (Fig. 3.11e-f.). The smallest vesicles in some particles (<20%) (<10 µm) have circular cross-sectional forms and appear undeformed. Asymmetrical strain shadows can be
observed around phenocrysts with vesicles and microlites wrapping around in distinctly flow-like patterns. In the strain shadows vesicles display rounded to sub-rounded forms.

Fig. 3.11. Scanning electron microscope BSE images illustrating representative microtextures of glassy vesicular (a-b), microcrystalline (c-d), elongate tube-vesicle (e-f) components. In addition, an image of a fluidal grain (g) and a particle from sample HVR054 (h) is shown. Glassy vesicular particles
show rounded vesicles and dominantly glassy groundmasses. Microcrystalline particles show a range
of both groundmass crystallinity and vesicle textures (c-d), in (c) three microcrystalline grains can be
observed each showing differing crystallinities show plagioclase (Plg) and pyroxene (Pyx) microlites.
Groundmass hosted cristobalite (C) is commonly observed (c), textures of mingling between melt of
different crystallinities are rarely observed (d). Elongate tube-vesicle grains are generally glassy with
the alignment of the sheared vesicles and microlites (e-f). Around phenocrysts strain shadows (SS)
can occasionally be observed (f). In (g) the fluidal particle shows a Pele’s Tear like structure with a
highly vesicular core and a dense glassy rim.

3.5. Interpretations

3.5.1. Timing, eruption, and pyroclast transport processes

Subunit 1

Subunit 1 drapes topography, which suggests deposition from suspension in the water column. The
wide distribution of this subunit (Fig. 3.12) requires that the height from which the grains settling
must have been shallower than 700 mbsl, the highest point on the caldera rim. Thinning of Subunit 1
is not observed, however fining of this subunit away from Dome OP is indicative of eruption form a
source vent now covered by Dome OP (Fig. 3.7.). The lack of any apparent internal stratification
indicates that the deposition of S1 occurred as a relatively continuous phase (Fig. 3.2a-d.).

The dominance of glassy vesicular ash in S1 indicates fragmentation of a relatively homogeneous
source (Fig. 3.4 and 3.8.). The range in vesicle population and textural characteristics in the S1 ash
does suggest variability in vesicle nucleation and growth conditions during ascent (Fig. 3.11a-b, g-h.).
The extremely low microlite content indicates a high degree of magma undercooling (Fig. 11a-b, g-
h.). The modal grain size of S1 between 0.5 and 3.5 φ suggests energetic fragmentation of the
magma (Fig. 3.4 and 3.6.). The dominance of curvi-planar ash morphologies indicates that
fragmentation was driven by direct magma water interaction (Fig. 3.10.). Fluidal rhyolitic ash grains
observed in S1 however indicate a range of fragmentation mechanisms were in operation (Fig. 3.10.).
These grains also suggest unusual magma rheology during fragmentation in the Havre eruption.
Wholly fluidal clasts cannot have been produced by abrasion from larger particles indicating a primary volcanic fragmentation mechanism.

Subunit 1, the raft pumice, the GP Unit, and the ALB Unit all show broad microtextural similarity been composed of dominantly glassy material which displays variable vesicle populations and textures (Rotella et al. 2015) (Fig. 3.11a-b, g-h.). These deposits additionally contrast with S3 and S4, both of which show distinctive componentry signatures. Stratigraphically S1 was deposited directly on top of the GP Unit. Observations of S1 however were all taken form on top of GP clasts. There is therefore some uncertainty as to whether this contact represents the base of S1, or whether S1 and the GP Unit occur as a single deposit. The relationship of S1 to the ALB Unit is unknown. The much thinner and patchier AL Unit deposit overlying the caldera rim lava domes (Fig. 3.2g-k.), along with the dearth of S1 signature in sample HVR255 (Fig. 3.8.), taken from on top of Dome OP, suggest that S1 was deposited prior to the effusion of the caldera rim lavas. Subunit 1 is therefore inferred to be associated with the eruption phase during which the raft pumice and the GP Unit were produced. The presence of a discoloured water plume associated with the Havre pumice raft in MODIS images indicates significant quantities of ash in the water column during the eruption of the raft pumice and the GP Unit (Fig. 3.1.). The discoloured water plume however could be the result ash generated by abrasion of the raft pumice.

Subunit 1 is inferred to be a settling out deposit following particle dispersal in a buoyant plume emanating from an eruption at the vent below Dome OP. Modal S1 grains, 500 to 125 µm, would have settled through 700 to 1500 m of seawater over 3 to 52 hrs after release from the top of the water column (Ferguson and Church 2004). To produce the observed stratigraphic relationship of S1 overlying GP clasts we suggest that S1 and the GP Unit were generated during distinct events, with intervening time sufficient to allow the settling of the GP Unit prior to the deposition of S1. Alternatively, to produce both S1 and the GP Unit in a single event would require rapid emplacement of the GP, thereby implying rapid water logging of a significant volume of vesicular pumice which seems unlikely. Fragmentation is inferred to have been driven by a range of processes, dominantly
resulting from direct magma water interaction, and not abrasion of the pumice raft. The presence of fluidal ash in S1 further suggests that magmatic fragmentation must have also occurred isolated from direct contact with water.

Fig. 3.12. Top Bathymetric map showing Havre caldera along with the distributions/outlines of the clastic deposits produced during the 2012 Havre eruption over the study area. Subunit 1 has been found over the entire study area. Subunit 2 northern boundary white wide dashed, S3 boundary yellow, S4a and b boundary in blue, GP Unit white narrow dashed, ALB Unit boundary solid white.
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The inferred locations of the sources for each subunit are denoted by stars. Note that every clastic unit identified extends beyond the study area. The trend of the idealized stratigraphic cross section (below) is shown. Below An idealized stratigraphic cross section through the Havre eruption, showing the temporal and spatial evolution in eruptions and deposition from various vents, along with changes in eruption style. Three eruption phases have been identified in the Havre eruption reflecting changes in style and location from dispersed effusive to fragmental focused on a single vent and back to dispersed effusive.

Subunit 2 (a and b)

Subunit 2 has a sharp boundary along northern caldera wall and is thickened on the caldera floor compared to the caldera rim (Fig. 3.12.), indicating a strong topographic control on its deposition. Subunit 2 shows no internal stratification or grading indicating continuous deposition (Fig. 3.2a-c.). The lack of S2 deposits on the northern caldera rim suggest that it was erupted from a vent on the southern caldera rim. The extremely fine modal grain size of S2 would result in particle settling over approximately 1 to 3 months in still water (Ferguson and Church 2004) from a height of 500 m above the depositional surface (the height of the caldera walls). The formation of vertical density currents (e.g. Fiske et al. 1998; Manville and Wilson 2004) or particle aggregation (Wiesner et al. 1995) would speed up the rate of deposition. Strong currents encountered during ROV Jason dives however suggest particle settling could have taken several months following the initial eruption. The presence of S3 and S4 as discrete layers within S2 is consistent with an inference of deposition over an extended period.

Subunit 2 is composed of glassy vesicular ash, indicating fragmentation of a highly undercooled broadly homogeneous source. The modal grain size of 6 to 5 φ indicates highly energetic fragmentation. At this size particles are typically smaller than vesicles, as such grain morphology is not necessarily diagnostic of fragmentation mechanism.

Similarly to S1, S2 shows a broad microtextural similarity to the raft pumice (Rotella et al. 2015), the GP Unit, and the ALB Unit all show broad microtextural similarity been composed of dominantly
glassy material which displays variable vesicle populations and textures. Subunit 2 also overlies S1 across a gradational contact (Fig. 3.2a-c.) suggestive of continuous deposition and indicating that S1 and S2 were likely produced in the same event. Subunit 2 is inferred to have been deposited from continuous/repeated dilute water supported turbulent suspension flows coming off the eruption column that generated S1. Flows spread radially from the vent below Dome OP. Flows entering that caldera were reflected from the steep northern caldera wall as a migrating bore, ponding in the topographic low of the caldera (Pickering and Hiscott 1985; Pickering et al. 1992; Edwards et al. 1994; Mulder et al. 2009; Talling et al. 2012). Subunit 2 is inferred to be the distal and dilute equivalent deposits of the ALB Unit, which shows a dispersal outline distinctive of having been deposited from flows (Fig. 3.13.). Slowing of a flow and condensation of any volatile components would result in rapid deposition of coarse suspended material producing the ALB Unit. The remaining dilute water supported turbulent flow would then spread from which S2 was deposited. The deposition of S2 is inferred to be slow, accumulation over months, however the initial generation of particles occurred rapidly. While S2 was been deposited subsequent eruptive activity lead to the rapid emplacement of S3 and S4, while the deposition of S2 continued afterwards. Subunit 2b is expected to eventually grade up in to a normal pelagic seafloor sediment as the volcanic sediment sources slow down post eruption.

Subunit 3

Subunit 3 drapes topography indicating deposition by particle settling out of the water column. The thinning and fining relationships within S3 indicate an eruptive source proximal to Lava G (Fig. 3.12.). No vent structure can be seen in bathymetry around Lava G and the number of sample locations in the AL Unit is too small to confidently assign a precise source location (Fig. 3.13b.). Rapid thinning and fining away from Lava G suggest a relatively passive eruptive mechanism.
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Subunit 3 is composed of elongate tube-vesicle clasts that are defined by tube/pipe vesicles (Fig. 3.9j-l and 3.11e-f.) indicating fragmentation of an extensively sheared source. The lack of microlites also indicates the pre-eruption magma source was highly undercooled (Fig. 3.11e-f.). Particle morphologies in S3 indicate both brittle and viscous fragmentation (Fig. 3.9j-l.).

Stratigraphically S3 is deposited overlying S2a indicating that the eruption and deposition of S3 occurred following the pyroclast forming eruptive phase during which the raft pumice, the GP Unit, the ALB Unit, S1, and S2 were formed. Inferring Lava G to be the source of S3 indicates that eruptive activity was occurring at Lava G post pyroclast forming eruptive phase of the Havre eruption.

Subunit 3 is inferred to a fallout deposit associated with explosive-effusive ash venting during the effusion of Lava G. Proximal deposits of S3 are inferred to have also been formed by quench fragmentation and comminution of the lava’s pumiceous carapace followed by dispersal in a thermally driven buoyant plume. The eruptive mechanisms of S3 however are a topic of an ongoing study. The inference of ash venting as an eruption mechanism for S3 implies that Lava G was still being extruded at the time of deposition (Schipper et al. 2013; Cole et al. 2014; Black et al. 2016). The sharp upper contact of S3 with S4a indicates a sudden termination of ash venting associated with the onset of S4a deposition.

Subunit 4

Subunit 4a drapes topography both on the caldera rim and caldera floor suggesting a fallout style of deposition. The microcrystalline clasts componentry signature of S4a appears concentrated around Lava C on the caldera floor, suggesting a proximity to the subunits source (Fig. 3.8.). Stratigraphically the timing of S4a is well constrained overlying S3 across a sharp boundary and overlying 2 cm of S2a at both HVR132 and HVR163 (Fig. 3.2a-b.) and in pushcore HVR159 (Fig. 3.3.). The sharp basal contact of S4a suggests a rapid depositional onset at the same time as the termination of the production of S3. The consistent 2 cm thickness and inferred slow accumulation rate of S2a indicates
deposition of S4a began a long time (weeks?) after the termination of the pyroclast forming phase of the Havre eruption.

The deposition of S3 indicates that at least Lava G had already been erupted prior to the eruption of S4a. The microcrystalline clasts that characterise both S4a and b indicate a relatively dense crystalline source (Fig. 3.11c-d.), strongly suggestive of fragmentation of the dense crystalline lava cores. Both S4a and b therefore give a stratigraphic indication that lavas were on the seafloor, however are not necessarily indicative of the onset of lava effusion.

From the rapid onset of S4a deposition associated with the termination of S3, its concentration at the base of the southwest caldera wall around Lava C, and the fragmentation of dense crystalline lava cores we suggest S4a was formed by a collapse of the of the caldera wall below Lavas G, H and I. Deposition of S4a on the northeast caldera rim (HVR283) on the opposite side to the lavas indicates however that S4a is not simply the product of mass wasting. In the case of Lava G this collapse included its source vent thereby sharply shutting off the production of S3 and sharply truncating the northern edge of the lava flow (Fig. 3.13b.). The northern edges of Lava H and I also slow truncation along a scallop shaped scarp (Carey et al. 2018) (Fig. 3.13b.). The collapse fed a debris avalanche, the deposits of which can be seen in bathymetry of the caldera floor (Carey et al. 2018) (Fig. 3.13b.). Water interaction with the hot exposed core of Lava G would have led to molten fuel coolant interaction (MFCI) and quench fragmentation. Subunit 4a was then deposited by fallout.

Subunit 4b

Subunit 4b is deposited relatively proximal surrounding Dome OP, with the deposit elongated downslope to the north (Fig. 3.12 and 3.13c.). This indicates a subunit source of Dome OP with transport been driven dominantly by gravity. Stratigraphically S4b overlies S1 and S2a however its relationship to S3 and S4a remains unclear.
Fig. 3.13. Detailed bathymetry features. (b) shows the location around Lavas G to I. A scarp shown in dashed white sharply cuts Lava G, while Lavas hand I appear truncated a sharp scarp face is not observed. At the base of the gully below these lavas large blocks are observed inferred to have been generated by a debris avalanche generated by collapse of the caldera wall around the source vent of Lava G. Exposure of the insulated hot core of Lava G resulted in magma water interaction driven fragmentation generating S4a. (c) shows a detailed view of Dome OP the central part of which shows a rugged uneven surface formed by exposed lava. The slopes of the dome however appear smooth.
indicating they are composed of clasts smaller than 1 m in diameter. These slopes are formed by a lava breccia, the fine component of is inferred to be S4b.

From the gravity driven dispersal and surrounding of Dome OP we infer S4b to be the fine-grained component of a lava breccia formed during the extrusion of Dome OP. Fragmentation is therefore inferred to have occurred by quenching, brecciation and comminution of the erupting lava. Slopes of lava breccia composed of meter and decimetre scale blocks can be observed in bathymetry proximal to and on the slopes of Dome OP. Particles formed are inferred to have been gravitationally transported down slope, been transported into the caldera to the north (Fig. 3.12 and 3.13c.).

3.5.2. Eruption & timing of Havre depositional events: constraints from the AL Unit

The initial phase of the Havre eruption produced Lavas A-E on the caldera floor (Fig. 3.1.) at some time after the 2002 bathymetric survey of Wright et al. (2006) and prior to the activity observed at the sea surface 18th – 19th July 2012 (Carey et al. 2018).

The second eruptive phase was from the vent now filled by Dome OP and produced several seafloor clastic deposits; the GP Unit, the ALB Unit, S1 and S2 (Fig. 3.12.). The pumice raft is also inferred to have been generated during this phase (Carey et al. 2018; Manga et al. 2018). Seafloor stratigraphy indicates that the GP Unit was deposited first, directly overlying it is S1 and the ALB Unit which are both in turn overlain by S2. The relationship of the ALB Unit to S1 remains unclear. Strong constraints can be placed on the eruption timing of the raft pumice, however its relationship to seafloor deposits remains unclear and an area of ongoing investigation. Based on the similarity in erupted material from a single vent and the indication of an energetic eruption in all the deposits from this phase we suggest that the units of the second eruptive phase were produced within the time of the pumice raft generation, ± days depending on the exact eruption mechanics.
Deposition of the GP Unit, the ALB Unit, and S1 occurred rapidly (hours), while the deposition period of S2 is unusual due to its extremely small modal grain size. Slow settling rates and sea floor currents mean that the observed thicknesses of S2 may have accumulated on the timescale of months. Over time S2 will likely grade up into a seafloor normal pelagic sediment for the Kermadec Arc as the high sediment input rates of the 2012 Havre eruption begin to diminish.

Following the end of the second eruption phase from the vent now filled by Dome OP, ash generation in the Havre eruption switched to a dominantly effusive source (Fig. 3.12.). Timing between the end of ash generation the pyroclast forming phase and the start in the second effusive phase is unclear. The 2 cm accumulated thickness of S2a undying S4a however suggests potentially weeks between the generation of each deposit. During the 2009 Chaitén and the 2011-12 Cordón Caulle rhyolitic eruptions (both in Chile) a transitional phase of explosive-effusive activity occurred after the main plinian phase and prior to the onset of pure lava effusion (Castro et al. 2012; Schipper et al. 2013). This phase is defined by simultaneous lava effusion and vigorous outgassing generating ash venting and vulcanian explosions (Schipper et al. 2013). The generation of S3 by ash venting suggests that Lava G was being activity extruded days/weeks after the end of the second eruptive phase (Schipper et al. 2013; Cole et al. 2014; Black et al. 2016). The deposition of S3 may then also point to the onset of effusive activity following the second eruptive phase at Havre.

The deposition S4 gives a timing for gravitational collapse of the southwest caldera wall. Subunit 4 therefore cannot be used as a wider tie line to pinpoint the onset of lava effusion. Lavas H and I however are not sharply cut by the collapse scarp indicating post collapse effusive activity at these domes (Fig. 3.13c.). In contrast Lava G is sharply cut by the scarp, indicating eruptive activity ceased prior to or at the collapse event. This points to some stratification in the timing of the end of eruptive activity of the southern rim caldera lava flows. The timing of effusive activity relative to the Oct 2012 bathymetric survey however is unknown.

The deposition period of S2 is unusual due to its extremely small modal grain size. Slow settling rates and sea floor currents mean that the observed thicknesses of S2 may have accumulated on the
timescale of months. Over time S2 will likely grade up into a seafloor normal pelagic sediment for the Kermadec Arc as the high sediment input rates of the 2012 Havre eruption begin to diminish.

3.5.3. Gas rich vesicular magma eruption phase

The second phase of the Havre eruption is inferred to have been the most intense. Sourced from the vent now covered by Dome OP during this phase eruption of a glassy gas rich vesicular magma is inferred to have produced the GP Unit, the ALB Unit, S1 and S2, along with the pumice raft (Carey et al. 2018) (Fig. 3.12.). A transitional effusive eruption style unique to the subaqueous environment has been previously inferred for the eruption of the pumice raft and the seafloor GP Unit (Manga et al. 2018). The results and implications regarding S1 and S2 presented here however have large implications on fragmentation and eruption style during this phase of the Havre eruption.

Ash grain size and shape data from S1 and S2 point to energetic fragmentation dominantly driven by direct interaction between magma and water (Fig. 3.4 and 3.7.). The presence of ash with wholly fluidal morphologies is further suggestive of primary fragmentation of magma as opposed to quenching or abrasion of ash from a larger body (i.e. giant/raft pumi- ce). These inferences point towards a Neptunian like eruption style with explosive fragmentation driving an overlying thermal plume from which water supported density currents are generated (Allen and McPhie 2009).

Inference presented here are in disagreement to the work of Manga et al. (2018). However, the exact association of the ALB Unit, S1 and S2 with the raft pumice and the GP Unit is unknown, and there are several processes that could account for this discrepancy. For example, both eruptive styles could have occurred separately, eruption of the GP Unit first with the ALB Unit, S1 and S2 eruption occurring after the GP had settled. It is conceivable in such this case that the raft pumice be associated with either eruption style. Future inferences on the eruption mechanisms during the Havre eruption must therefore account for energetic primary of ash as part of a potentially complex intense eruptive phase.
3.5.4. *Ash and Lapilli Unit volume estimates*

The stratigraphy of the AL Unit presented in Fig. 3.12. represents the intra-caldera and near-caldera deposits. All the subunits described here however extend beyond the study area in one direction or another. There is no evidence of thinning in S1, S2 or S4, suggesting that these deposits may extend well beyond the study area, hindering any firm volume estimate. Recent work on the widespread seafloor ash component of the 1650 CE Kolumbo eruption did not add significantly to the overall eruption volume (Fuller et al. 2018). The 2012 Havre eruption products however, were entirely contained within the water column in contrast to the Kolumbo eruption which shalled during its eruption generating a significant subaerial component (Cantner et al. 2014; Fuller et al. 2018). The fine component of the 2012 Havre eruption may therefore represent a much more significant component of the overall eruptive volume. This is hinted at by the plume of discoloured water observed in MODIS imagery on 18-19th July 2012 also showing the pumice raft and atmospheric steam plume (Carey et al. 2014; Jutzeler, Marsh, et al. 2014; Carey et al. 2018). This points to a significant population of fines that were carried away from the volcano. The low settling velocities of 5-6 φ particles in seawater, and lateral distribution by density currents, favour long-distance transport of extremely fine ash in the water column. A volume of 0.063 km$^3$ was presented by Carey *et al.* (2018) for the bulk AL Unit within the study area. We suggest this value represents a lower end current best estimate for the AL Unit volume within the 35 km$^2$ study area. Future studies in the medium and distal environments would be required for more accurate subunit specific volumes estimates and of bulk AL Unit.

3.7. *Conclusions*

The results and inferences of the AL Unit presented here provide stratigraphic and eruption mechanism constraints on the 2012 Havre eruption. The ash-dominated AL Unit produced during the
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2012 Havre eruption is composed of four distinct subunits. The diversity in deposit thicknesses and extent, grain morphologies, and microtextural features indicates that the Havre eruption was a multiphase event. The subunits were generated by four different processes from three locations with eruption style developing through time. Initial deposits, S1 and S2, were generated in a phase of energetic, explosive fragmentation of a highly undercooled vesiculating magma. Following this the Havre eruption switch to an effusive style with ash generation first by syn-extrusive ash vent (S3), and then by gravitational collapse of the caldera wall and interaction between the exposed hot lava and seawater (S4a). Additionally, extrusion of Dome OP generated a proximal breccia through quenching, brecciation, and communition (S4b).

The stratigraphic relationship of the four subunits to the other seafloor deposit provides a detailed relative temporal framework from which the full eruption time line can be established. Subunit characteristics allow some estimation of time differences between eruption events.

Results from S1 and S2 point towards explosive fragmentation occurring during the 2012 Havre eruption during the same phase the GP Unit and raft pumice were produced, in contradiction to previously published results (Manga et al. 2018). Detailed information on the exact association between the GP Unit and raft pumice (effusive?) (Manga et al. 2018), and S1, and S2 (explosive?) however is yet unclear and the focus of ongoing investigation.
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Chapter 4

Weakly pyroclastic and passive ash generation during the effusion of a subaqueous silicic lava flow, Havre Volcano, Kermadec Arc.

This chapter presents work prepared for publication as a manuscript.

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Author contributions:

RJC organized and led the cruise. APM, JDLW, RJC, and FI were present on the 2015 cruise and collected the samples. JDLW established the scope of the project. Clastic sampling analysis (granulometry, microtextural, morphological, etc.) was conducted by APM. Results on Lava G and its pumiceous carapace were collected by RM and FI. Simple modelling was undertaken by APM. APM wrote the manuscript and produced the figures. In its current state JDLW and RJC have read through draft and assisted in editing of the manuscript.

4.1. Abstract
The 2012 deep subaqueous silicic eruption of Havre Volcano was the largest of the last century, producing significant seafloor clastic and effusive products, along with a 400 km² sea-surface pumice raft. Detailed post-eruption AUV mapping along with ROV observations and sampling make Havre ideal for the study of processes and products of interaction between silicic magma and water. Here we examine a wide spread seafloor ash and lapilli deposit that is closely associated with a lava flow, both produced during the 2012 Havre eruption.

Subunit 3 drapes topography across the southern and northern caldera rim, as well as on the caldera floor. It shows a strong thinning and fining trend away from Lava G on the southern caldera rim. Subunit 3 is composed of ash grains that are morphological and microtexturally similar to the pumiceous carapace of Lava G, showing woody fibrous forms defined by pipe vesicles. Microtextures preserve evidence of shearing with alignment of microlites and tube/pipe vesicles.

Subunit 3 is inferred to have been generated by both quenching and communition on the brittle lava flow crust and by subaqueous explosive-effusive activity during the eruption of Lava G. Dispersal occurring due to thermal plumes generated by the heat from the underlying lava flow. Simple 1D modelling of thermal plume allows us to show that the higher heat flux generated by fragmental volcanism is required to produce the dispersal and grain size features of the seafloor Subunit 3.

4.2. Introduction

Effusive silicic volcanism is a comparatively passive process in which dense lava is extruded either flowing away from the vent or piling up over it. Ash generation is typically associated with energetic fragmentation occurring in explosive volcanism, however its contemporaneous formation during effusive activity via non-explosive brecciation and communition of a brittle crust has long been noted (e.g. Fink 1983; Manley 1996). More recent work has highlighted a protracted explosive-effusive transitional state (Castro et al. 2012; Schipper et al. 2013), also referred to as ash venting (Norton et al. 2002; Cole et al.)
During this phase weakly pyroclastic activity occurs simultaneously with lava effusive as a result of intensive outgassing through permeable channels formed due to shear localization enhancing of vesicle coalescence (Castro et al. 2012; Schipper et al. 2013; Cole et al. 2014; Kendrick et al. 2016).

Deep subaqueous eruptions are significantly modulated by the physical properties of water both indirectly (hydrostatic pressure, increased viscosity of water relative to air), and directly (rapid heat transfer) (Wohletz 1983; Head and Wilson 2003; White et al. 2003; Cas and Giordano 2014; White et al. 2015), in addition to the influences of magma composition and rheology (Walker and Croasdale 1971; Dingwell and Webb 1990), and volatile content of the magma and magma flux (Gonnermann and Manga 2003; Namiki and Manga 2008). Rapid heat transfer on contact between magma and water can lead to both passive and explosive fragmentation of magma through quenching and fuel coolant interaction with no involvement of magmatic volatiles (Wohletz 1983). In the subaqueous environment the proximity of lava and water during effusive volcanism is typically inferred to result in extensive quench fragmentation (e.g. van Otterloo et al. 2015). It has also been postulated that thermal plumes resulting from the heat of the underlying lava flow may disperse ash formed non-explosively to produce extensive ‘settling out’ clastic deposits.

Here we examine in detail a lava flow and associated clastic deposit produced during the 2012 deep subaqueous silicic eruption at Havre volcano. By examining the macroscopic deposit features, microtextural characteristics, and ash morphology we inferred the eruption and dispersal mechanisms of the ash unit associated with the lava flow produced during the 2012 Havre eruption. Inferences on particle dispersal are then tested through comparison with a simple 1D model of a thermal plume produced over subaqueous volcanism. The Havre data set is detailed and well constrained (Carey et al. 2018), providing the best-case study for examining ash generation from a range of mechanisms during deep subaqueous silicic volcanism.
### 4.3. The 2012 Havre eruption and its seafloor products

Havre is a fully submerged volcano located along the Kermadec Arc, west of the Kermadec Ridge (I. C. Wright et al. 2006). The volcano forms a 1 km high edifice that is truncated by a summit caldera 3 by 4 km in diameter (Fig. 4.1.). The caldera floor is relatively flat at approximately 1500 msbl with the walls rising at least 500 m on all sides (Fig. 4.1.).

In 2012 Havre produced the largest deep-water (>500 mbsl) silicic eruption of the last century (Carey et al. 2014; Jutzeler et al. 2014; Carey et al. 2018). The 2012 eruption produced a 400 km$^2$ pumice raft, along with and ocean hot spot, discoloured water plume, and atmospheric steam plume over about 21.5 hours (Carey et al. 2014; Jutzeler et al. 2014; Carey et al. 2018). Comparison between pre-eruption bathymetry from 2002 (I. C. Wright et al. 2006) and post-eruption survives in October 2012 (Carey et al. 2014) and March 2015 (Carey et al. 2018) revealed large scale changes to the seafloor.

Seafloor surveys conducted with autonomous underwater vehicle (AUV) Sentry and remotely operated vehicle (ROV) Jason in 2015 reviled that the large-scale seafloor changes were the result of the emplacement of 14 lavas from vents at depths between 900 to 1200 meters below sea level (mbsl) (Carey et al. 2018). Three major seafloor clastic units were also identified; ‘Giant Pumice (GP) Unit’, ‘Ash Lapilli Block (ALB) Unit’, and the ‘Ash and Lapilli (AL) Unit’, (Carey et al. 2018). The ash deposit we examine here is a subunit of the AL Unit described below.

At Havre a range of physical and chemical data sets, as well as AUV and ROV observations on a major silicic deep subaqueous eruption have been collected (I. C. Wright et al. 2006; Carey et al. 2014; Jutzeler et al. 2014; Rotella et al. 2015; Carey et al. 2018). The Havre data set is unparalleled its detail and is unique among studies on deep silicic eruptions. It therefore offers the best case study for examining subaqueous silicic eruption processes.
4.3.1. The Ash and Lapilli unit

The AL Unit is a wide spread deposit, produced during the 2012 Havre eruption, and composed of four distinct subunits (Carey et al., 2018). The subunits that composed the AL Unit show dominantly ash sized grains with a minor amount of lapilli. The four AL subunits were defined from mixed samples of the whole deposit using geographical variations in grain size and componentry (Chpt. 3.).

From the lowest subunit to the top, Subunit 1 (S1) is the basal layer of the exposed AL Unit, apparently directly overlying the GP Unit in stratigraphy. Subunit 1 has been dispersed over the entire study area to a thickness of at least 5 cm and certainly beyond. The formation mechanism of S1 is somewhat uncertain, however its is inferred to be the result of some type of pyroclastic volcanism from the vent below Dome OP with dispersal occurring in an overlying thermal plume (Chpt. 3.). The timing of this is inferred to be simultaneous with the production of the observed sea surface pumice raft (Chpt. 3.).

Subunit 2 directly overlies S1 and has a deposit boarder that follows the northern caldera wall, to the west, south, and east however it is dispersed beyond the study area. Subunit 2 thickens on the caldera floor compared to the caldera rim. Subunit 2 is locally divided into lower (a) and upper (b) parts because Subunits 3 (S3) and 4 (S4) have local dispersal. Subunit 2 has a grain size mode of 32 – 16 µm and is thus inferred to have been deposited over several months following production. Subunit 2 is inferred to have been deposited by a dilute density current formed in the same event that S1 was produced.

Subunit 3 is the focus of this study. It is characterised by distinctive elongate particles defined by tube vesicles. Subunit 3 has been locally dispersed on the southern caldera rim, caldera floor, and northern
caldera rim however, it appears to be related to Lava G located on the southern caldera rim. The aim here is to determine its formation and dispersal mechanisms, this will be discussed below.

Subunit 4 directly overlies S3 and has been deposited in two distinct areas; the first from the southwest caldera rim trending over the caldera floor to the northeast, while the other forms a roughly circular shape around Dome OP on the southeast caldera rim. The particles that composed S4 are inferred to have been fragmented from the crystalline core of the 2012 Havre lava flows. The deposit surrounding Dome OP is inferred to be the ash component of the lava’s talus slope. The caldera floor component of the deposit is inferred to be the result of a mass wasting deposit resulting from a collapse southwest of the caldera wall.

4.4. Methods

This chapter draws on microtextural observations using SEM, and major-element chemistry, to support its interpretations. The methods by which these results were collected are outlined in Chapter 2. Methods.

4.5. Results

4.5.1. Macroscopic deposit characteristics of Subunit 3

Subunit 3 was deposited over the entire caldera floor with a significant deposit thickness proximal to Lava G on the southwest rim of the caldera. Minor deposits of S3 also extend onto the northern and eastern caldera rims (Fig. 4.1.). Samples taken northwest and southeast of the caldera do not contain evidence of S3 (Fig. 4.1.). Two parallel subunit boundaries running northeast-southwest tangential to the caldera are thereby defined (Fig. 4.1.). To the northeast and southwest S3 extends beyond the study area (Fig. 4.1.). Subunit 3 is thickest near Lava G and thins away from it. The full subunit thickness on top of
lava G was not measured, however ROV images show that the subunit thickly drapes Lava G’s pumiceous carapace beneath it (Fig. 4.2a.). At sample locations HVR061 and HVR070, approximately 150 m away and level with Lava G, S3 is 0.5 m thick (Fig. 4.2b.). In sample HVR159, taken approximately 700 m from Lava G, S3 forms a diffuse layer within stratigraphy 2 cm thick. At more distal locations from Lava G S3 does not form a discrete layer and is only recognisable, from componentry, as a diffuse layer interbedded in the upper part of S2a (Chpt. 3.) (Fig. 4.2b.).

With increasing distance from Lava G the maximum grain size of S3 decreases rapidly, from blocks 70 mm in diameter at sample location HVR070, to 4 mm particles at sample location HVR159 (Fig. 4.3.). At sample locations HVR134 and HVR163 on the caldera floor the maximum grain size is 1 mm and 1.5 mm respectively. Distally, at sample location HVR283, approximately 4 km from Lava G, the maximum grain size is 500 µm (Fig. 4.3.).

Subunit 3 drapes topography and shows no observable sedimentary structures (Fig. 4.2.). At location HVR031 a deposit of S3 is found approximately 47 m higher and 1 km laterally from Lava G (Fig. 4.2b.). At location HVR283 S3 is found level with Lava G, approximately 4 km away, at the top of the northeast caldera wall (Fig. 4.1.).

4.5.2. Grain morphologies

Subunit 3 is composed entirely of Elongate Tube-Vesicle clasts. These particles are characterised by elongate clast morphologies showing both simple straight and complex twisted forms, with tube vesicles running parallel to the long axis (Fig. 4.4a, b.). Particles show and brittle fracture surfaces that cut perpendicular to the elongation direction of the tube vesicles exposing generally roughly oval to circular clast cross sections (Fig. 4.4.). Elongate Tube-Vesicle clasts are split into three groups based on variations in their surface morphologies (Fig. 4.5.). Elongate Tube-Angular particles: defined by elongated forms, with concave surfaces, defined by brittle-fractured bubble walls (Fig. 4.5a-e.). Elongate Tube-Ribbed
Fig. 4.2. (a) Subunit 3 overlies the Lava G carapace and is overlain by S2b deposits. Large blocks, which may have formed at the same time as S3, can be seen apparently fragmented off the Lava G carapace. Distance between the two red dots is 10 cm. (b) Seafloor images of S3 from location HVR070. Subunit 3 is exposed with no overlying deposits and underlain by an older deposit formed prior to the 2012 eruption. (c) Sample location HVR033 on top of the same dome sample HVR031 was taken. Elongate tube-vesicle particles can be seen in this small talus slope, demonstrating the presence of S3 on top of the pre-2012 lava dome.

Fig. 4.3. Granulometry of samples in which S3 is present. The weight percentage of Elongate tube-vesicle grains in sieve fractions from -1 to 2 φ (2 mm to 250 µm) in 1 φ steps is shown in yellow. The value below the sample number shows the percentage of the whole sample that Elongate tube-vesicle make.
particles: defined by surface ribs the run parallel to the vesicle and clast elongation direction, the surfaces of which are smoothly undulating and typically unmarked by vesicles (Fig. 4.5k-o.). Elongate Tube-Fluidal particles: are elongate, unmarked by vesicles, with flowing molten surfaces that form peaks or droplet like features, and show evidence of ductile necking (Fig. 4.5f-j.). The peaks or droplets may indicate ‘pull back’ features where the molten connection to another fibre was broken and reformed by surface tension. Curvi-planar brittle fracture surfaces often cut across perpendicular to the elongation direction of both Elongate Tube-Ribbed and -Fluidal clasts (Fig. 4.5f-o.).

Fig. 4.4. Optical microscope images of Elongate Tube-Vesicle particles from sample HVR159. Particles in (a-c) show typically straight forms, while those in (d) shows more twisted forms.
Fig. 4.5. SEM SE images showing Elongate Tube-Angular (a-c), -Ribbed (d-f) and -Fluidal particles (g-i) from a range of samples; HVR070 – a, d, g, i; HVR031 – b; HVR159 – c; HVR283 – f, h; HVR163 – e. Grains from each class show a range of features associated with that group. Elongate Tube-Angular particles show brittlely fractured bubble walls cross cut by curvi-planar fracture surfaces (a-c). (c) also shows a twisted shape. Elongate Tube-Ribbed particles show a large range of variability with ribs that may not extend the full particle length (e) and variable spacing (f). Fluidal grains show a wide range of features including elongated fibres welded together (g), drips and pull back features (h and i), and evidence of ductile necking (i).
4.5.3. Microtextural descriptions

Phenocrysts in Elongate Tube-Vesicle clasts are rare and generally comprise clusters of euhedral plagioclase and pyroxenes 120-70 µm in size. The groundmass of Elongate Tube-Vesicle particles is glassy composed of more than 95% glass with approximately 5% microlites (Fig. 4.6.). Microlites are acicular plagioclase and pyroxenes with variable relative proportions. The plagioclase preserve swallowtail and hopper ends. Microlites show a strong alinement with vesicle and clast elongation orientation. Vesicles in Elongate Tube-Vesicle particles show dominantly tube/pipe like morphologies, with oblong through to needle like cross-sectional forms. Vesicle cross-sectional length varies from approximately 10 µm up to pipes that traverse the whole length of approximately 4 mm long particles. Vesicles with cross-sectional lengths more than approximately 80 µm show evidence of coalescence, with complex sub-angular to sub-rounded shapes and features indicative of bubble wall retraction (Fig. 4.6a-c.). Bubbles smaller than 80 µm long have simple oblong sub-rounded shapes (Fig. 4.6b.). The smallest vesicles in some particles (<10 µm) have circular cross-sectional forms and appear undeformed (Fig. 4.6b.).

Asymmetrical strain shadows can be observed around micro-phenocrysts with vesicles and microlites wrapping around in distinctly flow-like patterns (Fig. 4.6b.). Vesicles in strain shadows have rounded to sub-rounded forms and show little to no evidence of deformation.

Elongate tube-vesicle grains show weak through to moderate vesicularities (Fig. 4.6a and d.). The lower vesicularity particles on average appear to have more elongate vesicles with more needle like morphologies (Fig. 4.6d-f.). Additionally, they do not show the fine, undeformed population of bubbles, instead vesicles of all sizes show tube/pipe forms (Fig. 4.6e and f.).
Fig. 4.6. Microtextural characteristics Elongate Tube-Vesicle grains over a range of SEM magnifications. SEM images taken approximately along the vesicle elongation axis. (a) Shows an elongate tube-vesicle particle with an asymmetrical strain shadow around a euhedral plagioclase micro-phenocryst. (b) Shows half a strain shadow around a micro-phenocryst cluster. Highly elongate needle like vesicle forms can be seen in (c). In (d) vesicle deformation has occurred over all vesicle sizes. In all images there is a strong alignment between vesicles and microlites.

4.5.4. Characteristics of the pumiceous carapace and core of Lava G

Lava G is one of 14 lavas produced during the 2012 Havre eruption (Carey et al. 2018). Located on the southwest rim of the caldera (Fig. 4.1.), Lava G is characterized by large pressure ridges and comprises a dense coherent core, and pumiceous carapace, forming a total volume of approximately $4 \times 10^5$ m$^3$. The flow is sharply truncated at its northern edge along a larger scallop shaped scarp (Fig. 4.7.). Two samples
were taken from the pumiceous carapace of Lava G and three from the flow interior, on the truncated northern margin (Fig. 4.7.).

**Fig. 4.7.** (a) Detail bathymetry map of the southwest caldera rim showing the location of clastic, and pumiceous carapace samples. The locations of the ROV photos in Fig. 2. are denoted by blue stars. The yellow line shows the path of the profile through Lava G shown in (b) with 0 vertical exaggeration. The 2002 (orange) and 2015 (blue) bathymetry profiles are shown, along with the rough locations of samples from Lava G.

Lava G’s pumiceous carapace is between 4 m and 5 m thick overlying the 20 m thick dense coherent core. The carapace shows a woody texture that appears fibrous, defined by elongate vesicles. The groundmass has a porphyritic texture, with approximately 15% phenocrysts. In order of abundance the phenocrysts comprise euhedral plagioclase and alkali feldspar, irregular magnetite, anhedral quartz, and minor amounts of tabular pyroxene. Phenocrysts are 0.25-1 mm in size and are commonly found in
clusters (Fig. 4.8.). The groundmass is dominantly glassy and microlites form approximately 4% of the total groundmass area. Microlites are dominantly plagioclase with minor amounts of clinopyroxene. Microlites have acicular forms, with plagioclase microlites showing swallowtail and hopper ends (Fig. 4.8.). Only slight variations in relative proportions of components are observed between the two carapace samples. Texturally however, the carapace appears quite heterogeneous on the small scale, with rapid spatial variations in the degrees of elongation and vesicularity.

**Fig. 4.8.** SEM BSE images showing the microtextures of the Lava G pumiceous carapace from samples HVR086 (a and b) and HVR286 (c and d) at different magnifications. (a) Shows large variation in shear direction/amount with variation in vesicle deformation across the picture, at the centre on the image a micro-phenocryst cluster can be seen disrupting the shear orientation. In (c) fluidal vesicle internal walls can be observed, however they rarely define isolated hairs. In all images the smallest vesicle fraction appears undeformed and rounded.
The carapace has a vesicularity of 60-65% (Fig. 4.8). Vesicles generally show evidence of shearing and range in size from long axis of between 35 µm to 8 mm. Larger vesicles typically have subangular-to-subround forms while at the smaller scale vesicles appear subspherical to cylindrical (Fig. 4.8.). Strain shadows are observed around microphenocrysts.

The core of Lava G was sampled at depths of approximately 6 m, 11 m, and 19 m below the lava flow surface (Fig. 4.7.). These samples show little difference in the phenocryst content with those described form the pumiceous carapace. Comparison of the three samples taken from the lava flow core show an increasing proportion of microlites with depth, reaching an observed maximum of 70% total groundmass area. Over this depth the microlite components and textures remain largely the same, with dominate acicular plagioclase showing swallow tail and hopper ends with minor amounts of acicular clinopyroxene. Vesicularity also changes with depth from 40% at 6 m below the lava flow surface, 35% at 11 m, and 30% at 19 m. At sample HVR287, 6 m below the surface, small bubbles, smaller than 50 µm, show spherical to rounded forms, while bubbles 0.2-3.5 mm in size show irregular shapes. In HVR288 and HVR289 bubbles typically appear irregular and rounded. These samples also show an unusual texture not observed in the other samples, where most large bubbles are surrounded by a rim of void space with interconnected microlites. Vesicle and groundmass hosted cristobalite is also present in both HVR288 and HVR289.

4.6. Discussion

4.6.1. Identifying the source of Subunit 3

Subunit 3 shows a strong coarsening and thickening trend towards Lava G, suggesting that its source was located on or proximal to the lava flow (Fig. 4.3.). Limited sampling around Lava G, however, prevents pinpointing of a precise source location. Visual comparison of the Lava G pumiceous carapace
and breccia, with the Elongate Tube-Vesicle grains that compose S3, show striking similarities (Fig. 4.2a and b.). Microtextures described in grains from S3 and those in the pumiceous carapace of Lava G show a strong correlation (Fig. 4.6 and 4.11.). In addition, similarities in particle morphology, microlite population and textures, along with parallels in vesicle form suggests S3 is related to the pumiceous carapace of Lava G (Fig. 4.6 and 4.11.). Similar woody and fibrous textures have not been observed from any other products of the 2012 Havre eruption (Carey et al. 2018). The production of S3 is therefore inferred to be related to the eruption of Lava G, and specifically to the formation of the pumiceous carapace.

Subunit 3 is the third of four subunits in the stratigraphy of AL Unit, deposited following the pyroclastic phase of the Havre eruption during which the GP Unit, S1, the ALB Unit, and S2 were deposited (Fig. 4.1.) (Chpt. 3.). Subunit 3 occurs as a diffusive layer within the upper part of S2a, indicating that some time-period elapsed after the pyroclastic phase prior to the onset of the eruption of S3.

Based on flow morphology Lava G was truncated by a collapse of the caldera wall, during which the source vent, inferred to be to the northeast of the flow, was destroyed (Fig. 4.7.). There was no bathymetric change in this area between October 2012 and March 2015 (Carey et al. 2018), suggesting that the collapse took place during, or immediately following Lava G’s emplacement, prior to the 2012 bathymetry survey. Subunit 4 overlies S3 across a sharp boundary and is inferred to have been produced during the caldera wall collapse. The sharp boundary between S3 and S4 suggests the processes generating S3 were terminated by the collapse.

Subunit 3 overlies Lava G indicating that the subunits’ deposition followed development of a solid surface on the lava (Fig. 4.2a.). Subunit 3’s stratigraphic position overlying the inferred seafloor deposits of the July eruption seen at the sea surface (Carey et al. 2018; Chpt. 3.) brackets the formation time of Lava G, and therefore S3, to between 20th July 2012 (pumice raft) (Jutzeler et al. 2014) and October 2012 (2012 bathymetry survey) (Carey et al. 2014).
4.6.2. Groundmass and microtextures

Microlite and vesicle populations and textures can be used to infer magma undercooling (Swanson 1977; Hammer et al. 2000), overpressure, and deformation conditions during ascent. The paucity of microlites and the acicular, hopper and swallow tail form of those present in a glassy groundmass indicate a high degree of undercooling, in both Elongate tube-vesicle clasts and the pumiceous carapace of Lava G (Swanson 1977; Hammer et al. 2000) (Fig. 4.6.). This contrasts with the highly crystalline microtextures observed in the core of Lava G which show extensive evidence of groundmass crystallization, suggestive of a lower degree of undercooling (Swanson 1977; Hammer et al. 2000) and may reflect crystallization of the insulated flow core.

Vesicles in both Elongate Tube-Vesicles clasts and the Lava G carapace show evidence for extensive shearing across almost all bubble sizes (Fig. 4.6 and 4.11.). The differences in vesicle form from needle like to oblong shapes reflects variations in the degree of shearing. Needle-like vesicles appear to be restricted to Elongate Tube-Vesicles clasts, suggesting these particles underwent increased degrees of shearing compared with the carapace (Fig. 4.6.). Fine undeformed spherical vesicles, observed in some Elongate Tube-Vesicles clasts, and in the pumiceous carapace, indicates the end of shearing preceded the end of vesiculation in these particles. This implies a variable point of shear termination during the ascent of magma that produced S3 and the pumiceous carapace of Lava G. The tube-like forms of vesicles suggest that the source of the Elongate Tube-Vesicles clasts was highly permeable, in the direction of the vesicle long axis (H. M. N. Wright et al. 2006; Wright et al. 2009).

Particles from S3 are ubiquitously display evidence for extensive shearing during ascent. Asymmetrical strain shadows, observed in both Elongate Tube-Vesicles particles and the pumiceous carapace around phenocrystals, indicate deformation occurred as simple shear (Fig. 4.6.) (Durney and Ramsay, 1973). Simple shear implies that shearing occurred ‘against’ something because of variable magma velocity,
commonly inferred to be the influence of the conduit margin (Mastin 2005; Okumura et al. 2006; Wright and Weinberg 2009; Cole et al. 2014; Elizabeth Gaunt et al. 2014; B. Dingwell et al. 2015).

4.6.3. Grain morphology

The external morphology of particles is reflective of processes operating during and after fragmentation (Heiken 1972; Wohletz 1983; Wohletz and McQueen 1984). Elongate Tube-Angular particles are bound by fracture surfaces that cut tube vesicle walls, and are indicative of brittle fracturing (Heiken 1972). This is typical for silicic magmas, whose high viscosity and long Maxwell times means that at the rate of fracture propagation the magma will behave elastically (Dingwell and Webb 1990). Ribbed and fluidal particles in contrast show evidence of viscous shaping of particles during or after fragmentation (Walker and Croasdale, 1971).

On ribbed particles, the ribs appear to parallel the tube vesicles, yet lack jagged partial walls of tube vesicles once connected to those of the parent magma that would indicate they were cracked loose from that magma in a brittle state. The magma is inferred to have been sufficiently ductile at the time of fragmentation that such partial walls were eliminated post-fragmentation by retraction and reshaping by pull-back processes, leaving the preserved ribs (Walker and Croasdale 1971; Gonnermann 2015). Fluidal particles, although lacking the surface ribs, show ‘pull-back’ features indicative of viscous retraction of melt bridges, suggestive of a ductile magma at the time of fragmentation (Walker and Croasdale 1971; Gonnermann 2015)

The fact that ribbed and fluidal particles show these surfaces over their whole circumference indicates they were not brittlely fragmented off a larger body that underwent a previous phase of ductile deformation. Both ribbed and fluidal particles preserve clear evidence of viscous shaping during fragmentation. The preservation of tube vesicles and elongate forms, in both ribbed and fluidal particles
however, indicates limited surface tension reshaping (Wohletz 1983). This suggests ductile shearing played a large role during fragmentation (Marti et al. 1999). This is consistent with morphological features on several grains that appear to be implicit of ductile necking during fragmentation (Fig. 4.5o.).

4.6.4. Eruption mechanism

The microtextural similarity between Elongate tube-vesicle particles and the Lava G carapace suggests that S3 was formed by fragmentation of the pumiceous carapace. The thick S3 deposit, and its large grain size overlying the pumiceous carapace appears to form a surface breccia on Lava G (Fig. 4.2a.). The formation of a surface breccia is consistent with observations of subaerial silicic lavas where brittle fragmentation occurs because of comminution the lava flow surface (Fink 1983). Although in the subaqueous realm water would cushion comminution processes (White et al. 2003), the addition of quench fragmentation of the lava is to result in the production of a surface breccia. The lack of explosive craters would seem to rule out explosive activity from under the pumiceous carapace as the source of S3 (Fink and Manley 1989).

The presence of ribbed and fluidal particles in S3 however, are not consistent with a brittle fragmentation mechanism. Morphologically these particles are indicative of viscous shaping during or post fragmentation (Walker and Croasdale 1971). In the subaqueous realm the upper layer of Lava G would have been rapidly quenched following extrusion, forming an elastic layer. Communion and quenching both involve the brittle fragmentation of an elastic material due to impacts (Manley 1996) or thermally induced stress (van Otterloo et al. 2015). For ribbed and fluidal particles to have been fragmented out brittlely, they would have had to have been present in the carapace as distinct volumes prior to quenching (Fig. 4.8c.). Detailed images of the pumiceous carapace of Lava G however do not appear to show fluidal fibre like structures that could be brittlely fragmented out (Fig. 4.8c.).
Alternatively, ribbed and fluidal clasts must have been produced by a different mechanism than comminution and quenching of the elastic pumiceous carapace.

The production of ash synchronously with effusive volcanism has been noted from several lava flow and dome forming eruptions e.g. Mount St. Helens (Kennedy and Russell 2012), Soufrière Hills Volcano, Montserrat (Bonadonna et al. 2002; Cole et al. 2014), Volcán de Colima (Kendrick et al. 2016), Santiaguito (Holland et al. 2011) volcán Chaitén (Castro et al. 2012), and Cordón Caulle (Schipper et al. 2013). Termed both ‘hybrid explosive-effusive activity’ and ‘ash venting’ this weakly pyroclastic style of volcanism is inferred to occur by strong outgassing through permeable tuffisite channels resulting from shear localization (Schipper et al. 2013; Kendrick et al. 2016; Saubin et al. 2016). Hybrid explosive-effusive activity or ash venting is an important process in the transition between explosive and effusive activity, modulating system permeability and therefore outgassing of the ascending magma column, increasing or decreasing the likelihood of explosive vs effusive eruptions (Castro et al. 2012; Farquharson et al. 2016; Kendrick et al. 2016; Saubin et al. 2016). Such activity has not previously been noted in subaqueous silicic volcanism, however, the theoretical impact of the environment on the processes at play and magma conditions are not such that this should be prevented.

Comparison between ash from S3, with that produced by hybrid explosive-effusive activity in the rhyolitic eruption of Cordón Caulle (Chile, 2011-2012), shows strong visual similarities in particle microtextures and morphology (Fig. 4.4a and b.) and (Fig. 6a and b. from (Schipper et al., 2013)). At Cordón Caulle the sampled particles occurred over a range of grains sizes (~2.5 mm - <<0.1 mm), showing elongate forms defined by tube vesicles that run parallel to the clast long axis, were mostly aphyric, and showed high along axis permeabilities (Schipper et al., 2013). Along with the strong visual and textural similarities the grains described by (Schipper et al. 2013) were also produced synchronously with the effusion of a highly silicic lava flow, similar to Lava G. Based on these
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morphological and textural similarities, along with the comparable volcanological settings it is inferred that S3 was at least in part generated by subaqueous ash venting during the effusion of Lava G.

The grains that make up S3 are inferred to have a bimodal source with both passive comminution and quenching of the pumiceous carapace along with weakly pyroclastic ash venting during the extrusion of Lava G.

4.6.5. Fragmentation processes of fluidal particles

The Elongate Tube-Ribbed and -Fluidal grains are both inferred to have been generated during ash venting. These grains show evidence of ductile shearing and reshaping post fragmentation indicating that prior to reshaping the particle must not have come in to contact with water else they would have been rapidly quenched. Ash venting is weakly pyroclastic, and thus would provide a gas filled region from which water was excluded, where viscous shaping of grains could occur. The viscous behaviour of ash at Havre is unexpected and is at least in part related to the decreased magma viscosity resulting from eruption under high hydrostatic pressure (9 MPa) that reduces exsolution at similar depths below the vent compared to subaerial eruptions. The magma erupted during the 2012 Havre eruption would have an equilibrium solubility of approximately 1 wt% (Newman and Lowenstern 2002), giving a vent depth viscosity of $10^{6.7}$ Pa s (Giordano et al. 2008). The decreased viscosity associated with eruption at depth however, does not appear to be sufficient to account for the formation of the highly viscous Elongate Tube-Ribbed and -Fluidal grains during traditional fragmentation mechanisms (i.e. stress induced (Alidibirov 1994; Zhang 1999; Spieler et al. 2004; Mueller et al. 2008) and strain rate induced (Papale 1999)).

To account for the highly viscous forms of the Elongate Tube-Ribbed and -Fluidal grains we define a fragmentation mechanism unique to ash venting, that allows slow deformation of magma in an
isothermal environment, protected from the ambient water. These particles are suggested to have formed in the vent region above an underlying highly permeable zone of shear localization. The rush of the gas through outgassing passageways weakly but persistently deforms the passageway walls, resulting in stretching and viscous reshaping of parts of the passageway walls. Viscous behaviour is allowed due to the continuous gas flow maintaining high temperatures in the passageways and excludes the ambient water, along with the slow rate of deformation. Fragmentation may occur viscosely due to attenuation or brittlely due to the passage of pressure waves or quenching resulting from continued extrusion.

Particle shaping in a gas stream is consistent with the lack of large microtextural differences between the effusive (pumiceous carapace) and pyroclastic (ribbed and fluidal particles) eruption components. Fragmentation from a shear zone also explains the ubiquitously sheared nature of particles in S3. This process would also allow particle formation in the subaqueous environment despite rapid quenching post fragmentation due to contact with the ambient water. Strong outgassing would prevent the invasion of water into the permeable channel whilst also providing a mechanism for viscous shaping of clasts into elongate forms. Brittle fractures that cross cut ribbed, and fluidal particles may result from late stage brittle breakup of the shear zone close to the vent, quenching on contact with the ambient water or from clasts impacts while in the gas jet.

4.6.6. Depositional processes of Subunit 3

In the subaqueous environment dispersal and deposition of volcanioclastics by a plume is not necessarily indicative of any style of volcanism (White et al. 2003). Although both air and water can convect when heated producing plumes, waters higher density and viscosity means it can more readily entrain and transport particles (Kaminski et al. 2005). As such heating of the water from an effusive source can drive a convective plume that could entrain and disperse clasts.
Subunit 3 drapes topography suggesting that it was deposited via particle settling through the water column. The S3 deposit is also slightly elongated (Fig. 4.1.), implicit of a plume ascending into a current that trended northeast. The deposit shape of S3 contrasts with that observed in the GP Unit, and the pumice raft which show a northwest trending elongation, indicating a switch in current direction before the eruption of S3.

Subunit 3’s presence in sample HVR031, taken 47 m higher than its source at Lava G, and in HVR283, found roughly level with Lava G on the opposite caldera rim 4.7 km away, are consistent with the interpretation of deposition from a plume, and suggest the fallout height was at least 47 m above Lava G (Fig. 4.1. and 4.6.). The topographic relationship of these samples to Lava G, along with particle settling velocities can be used to estimate the height of the plume from which S3 was deposited.

The method of determining required plume height is outlined in Fig. 4.9. The spatial relationship of samples HVR031, HVR105, HVR159, and HVR283 relative to their source at Lava G are used as input distances (Fig. 4.1.). Using the average size of Elongate tube-vesicle clasts at each location (HVR031 – 500 µm, HVR105 – 500 µm, HVR159- 250 µm, and HVR283- 250 µm) terminal grain settling velocities (w) were calculated using a universal equation (Eq. (4) -Ferguson and Church, 2004).

\[ W = \frac{R g D^2}{C_1 v + (0.75C_2 R g D^3)^{0.5}} \]

Where the specific gravity of a particle is R, g is the acceleration due to gravity, D is particle diameter, v is the kinematic viscosity of the fluid, and C_1 and C_2 are particle shape factors (Ferguson and Church 2004). For smooth spheres C_1 and C_2 are set as 18 and 0.4 respectively, extreme angular particles have values of 24 and 1.2, and values of 18 and 1.0 are used for intermediate grains of varied shape (Ferguson and Church 2004). To reflect the variability and extreme eccentricity of particle shapes of Elongate tube-vesicle particles values for C_1 and C_2 of both the extreme angular and intermediate particles (24 and 1.2, and 18 and 1.0 respectively) are used. The average grain size of Elongate tube-
vesicle particles at each site was used giving values of 500 µm at HVR031, and 250 µm at HVR283 (Fig. 4.3.). Settling velocities for a range of vesicularities are calculated, in each case all vesicles assumed to be water-filled during transport (Allen et al. 2008; Fauria et al. 2017), this gave a particle densities of 2600 kg m$^{-3}$ for 0% vesicularity, 2126 kg m$^{-3}$ for 30%, and 1652 kg m$^{-3}$ for 60%.

Grain settling velocities of 0.068 – 0.030 ms$^{-1}$ were calculated for particle diameters of 500 µm, and 0.022 – 0.008 ms$^{-1}$ for particle diameters of 250 µm (Ferguson and Church 2004). A lateral current velocity of 0.2 m/s is used. During the 2015 Havre cruise currents of approximately 0.1-0.3 m/s were noted at several locations. A particle release height of ~410-470 m for HVR031, ~570-670 m for HVR105, ~140-180 m for HVR159, and ~720-880 m for HVR283 above vent is required for deposition at these locations (Fig. 4.9.). From this we infer a required plume height of up to 880 m above Lava G (Fig. 4.9.).
Fig. 4.9. Graphic outline of the process by which the release height range for a grain was calculated for deposition at HVR031, HVR105, HVR159, and HVR283. Given a lateral current velocity and a distance the transport time can be calculated. Then by multiplying this with the calculated range of settling velocities of the maximum grain size, the total depth settled during transport was determined. By adding this value to the height of the deposit location above source level, the required release height of a grain to produce the deposit observed was ascertained. This gives a required plume height to produce S3, which is equal to 885 m above Lava G.

4.6.7. Modelling of a thermal plume during the production of Subunit 3

Based on grain morphology a bimodal source for S3 is inferred, involving both comminution and quenching of the pumiceous carapace, along with ash venting. To determine the contribution of each process to the overall deposit of S3 a basic 1D model of the physical characteristics of a plume generated by both heating from a lava flow surface and by ash venting is produced. By comparing the axial velocity of each plume, determined using the methods of (Morton et al. 1956), with the calculated particle settling velocities rough fallout heights are calculated. Particle fallout heights are then compared with the inferred plume heights calculated based on sample locations HVR031 and HVR283 to assess how the plume dispersing S3 was formed.

Thermal plumes are caused by heating of the surrounding fluid by a persistent source lowering its density that results in vertical motion of the buoyant fluid (Morton et al. 1956). Entrainment of the cold ambient fluid during ascent leads to a decrease in buoyancy of the ascending mixture causing it to eventually reach a point of neutral buoyancy, at which point the mixture will spread laterally (Morton et al. 1956). The height the plume reaches, and its axial velocity are a function of both the source heat flux (H) and the density gradient (N) of the ambient fluid (Morton et al. 1956). A solution to for the model of (Morton et al. 1956) is shown in Fig. 4.10, where $H = 1 \text{ MWm}^{-2}$ and $N^2 = 2 \times 10^{-6} \text{ s}^{-2}$. The value for $N$ come from a water profile taken by King et al., (2012) north of Tonga and is consistent throughout all
calculations undertaken here. It should also be noted however, that N varies greatly with location and can change with time (King et al., 2012). The value used here represents the closest location to Havre volcano that could be found.

**Fig. 4.10.** Showing the solution to the model of Morton *et al.* (1956) for an underwater thermal plume driven by source heat flux of 1 MWm\(^2\). The schematic diagram shows the physical processes occurring at different points relative to the graph.

For calculating the plume characteristics due to heating from a lava flow surface a 1 m\(^2\) lava flow surface is considered as a source. Convective heat transfer is by far the most effect mechanism of heat transfer in the subaqueous environment, and as such convective heat flux values for various temperatures of submarine lava flows are used from Griffiths and Fink (1992). Measurement of subaerial silicic lava flow surfaces typically show low temperature, with the core is insulated by a cooled carapace (Bernstein *et al.* 2013). Brecciation of this carapace during effusion, however, will locally expose the hotter interior of the flow to the ambient atmosphere (Bernstein *et al.* 2013). Thermal imaging of the rhyolitic lava dome produced during the 2008-09 eruption at Chaitén volcano, Chile,
reveal local temperature highs of up to 400 °C (Bernstein et al. 2013). Along with this, permeability and fluidal flow within a pumiceous and brecciated carapace will increase the heat flux generally from that expected based on the surface temperature. The heat flux structure on a lava flow surface is therefore likely to have complex structure with spatial variations and temporal evolution, due to the dynamic stresses applied by the motion of the lava flow.

Convective heat flux values of 1 MW m$^{-2}$ to less than 1 kW m$^{-2}$ are used to calculate axial velocities by height, corresponding to thermal plumes generated by lava flow surface temperatures of 850 °C to less than 300 °C (Griffiths and Fink 1992). The comparison of the plume axial velocity profiles with settling velocities of saturated Elongate tube-vesicle clasts of diameter 500 µm to 250 µm give fallout height results which are shown in Fig. 4.11a. Using a maximum exposed lava temperature of 400 °C (Bernstein et al. 2013), fallout heights of between 10 – 30 m and 40 – 50 m for particles 500 µm to 250 µm in diameter are calculated respectively (Fig. 4.11b.). These are by far insufficient to deposit ash at HVR031 and HVR283 (Fig. 4.9.).

Magma is a relatively poor conductor of heat, as such heat flux is largely dependent on surface area. The eruption of a fragmented magma during ash venting would therefore result in a much higher heat flux. Ash of 1 - 0.1 mm has a surface area $10^3$ to $10^4$ times that of an unfragmented block of lava 1 m$^3$ in volume. For Havre, with an inferred temperature of 850 °C (Carey et al. 2018) this would result in an increase in convective heat flux to 1-10 GW m$^{-2}$. Heat contained within volatiles, assuming 30% vesicularity full of gas, is approximately 0.1 MW, and is comparatively negligible. Source heat fluxes of 1-10 GW m$^{-2}$ produce thermal plumes capable of transporting 500 µm to 250 µm grains approximately 380 m to 680 m above source (Fig. 4.11a.). The resulting transport heights are consistent with the range of required plumes heights calculated based on the observed spatial of distribution of S3 around Havre caldera. In the subaqueous environment rapid quenching of pyroclasts and condensation of volatiles (Deardorff et al. 2011) means that all heat is transferred to the ambient environment at the source. The
modelling of a subaqueous thermal plume resulting from weakly pyroclastic activity, such as ash venting, therefore does not need to account syn-ascent heating and the model presented by Morton et al. (1956) is applicable.

Fig. 4.11. Showing the outputs of the plume analysis conducted using the methods of (Morton et al. 1956). (a) The relative fallout heights of 30% vesicularity water logged particles of a range of size vs source heat flux for plumes generated by radiation from a lava flow surface. (b) A graph showing the max plume height vs the source heat flux, the heat flux range of a lava flow and ash venting source are shown. The inferred required particle fallout heights to produce S3 from Fig. 9. are shown.

The plume inferred here is inferred to be similar to that observed during the weakly explosive eruption at NW-Rota 1 (Deardorff et al. 2011). Here rapid condensation of the volatile phase and the high pressure and density of the water lead to stalling of the initial expansion driven phase of the jet within a meter of the vent (Deardorff et al. 2011). A second phase of stable buoyant ascent of the plume was
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(a) High plume

- High local heat flux due to ash and gas
  - Ambient water stalls initial momentum of plume
  - Fragmentation of permeable channel

(b) Lava G

- Increasing degree of shear
  - Permeable channel
  - Sheared magma
  - Central magma conduit

(c) Ash and lapilli dispersed in thermal plumes

- High local heat flux from ash venting forms high plume

- Low heat flux from lava flow surface, forming a low convective region

- Ash formed by quenching and disintegration of the carapace

Shear localization on conduit margin leads to increased bubble coalescence forming a permeable channel that allows outgassing of the ascending silicic magma.
Fig. 4.12. A schematic model of the eruption processes of SEAV. (a) Shear localisation during magma ascent leads to fracturing and vesicle coalescence forming a permeable channel. Outgassing occurs along the permeable channel. Near the vent, shear-induced fragmentation breaks apart walls of the permeable channel, fragments of which are entrained into the gas stream. (b) On eruption of the gas stream the gas condenses, and ash is rapidly quenched due to the high thermal conductivity of water. The transferred heat drives a convective plume up to 680 m high. (c) Fragmentation of the pumiceous carapace occurs due to quenching and dynamic stressing of the brittle crust by movement of the lava flows molten interior. Heating of the overlying water drives a convective cell above the lava flow. The insulating effect of the pumiceous carapace; however, results in a low heat flux that drives a cell tens of metres high. Fine particles generated on the pumiceous carapace are entrained in and dispersed by this cell. This generates two overlapping deposits composed of related particles producing by different processes.

then observed resulting from heating of the water (Deardorff et al. 2011). The plumes from this activity reached less than 100 m into the water column above the vent (Deardorff et al. 2011).

Comparison of the inferred required dispersal height to produce S3 with the modelled plumes from both a lava flow surface and ash venting strongly implies that ash venting was required to disperse grains sufficiently to produce S3 (Fig. 4.12.). The low thermal conductivity of magma severely limits its heat transfer rate, resulting in a strong dependency of heat flux on surface area. Evidence from particle morphology and dispersal strongly imply that S3 is dominantly the result of ash venting. Shearing of the magma during ascent in conduit margin shear zones lead to the development of a permeable outgassing channel (Fig. 4.12.). Outgassing lead to low strain rate isothermal deformation of the magma leading to the production of viscous clasts (Fig. 4.12.). Ash venting occurring because of outgassing generated a high thermal plume that was able to disperse particle over several kilometres (Fig. 4.12.). A lower thermal plume was also generated over the cooling lava flow that resulted in more localized dispersal of ash (Fig. 4.12.). The loss of the lava flow source during a caldera wall collapse event resulted in the abrupt termination of ash venting.
4.7. Conclusions

Subunit 3 was generated during the effusion of Lava G. The linking of this single seafloor ash unit back to an identified lava has allowed the study of synchronous explosive effusive activity occurring in the subaqueous environment. The Elongate Tube-Vesicle (ETV) grains which make up S3 are inferred to have been formed by both brittle brecciation of the pumiceous carapace of Lava G, and by ash venting occurring synchronously with lava effusion. Fragmentation during ash venting resulted in both brittle fragmentation and isothermal viscous shaping of particles within tuffisite veins due to shear, induced by the continuous flow of outgassing volatiles.

Dispersal of S3 occurred in thermal plumes driven by the underlying heat of volcanism producing a topography draping fallout deposit. By simple modelling of thermal plumes produced above the ash venting and above the brecciated surface of a lava flow and comparison of the results with the seafloor distribution of S3 it was shown that to form the observed S3 deposit required a plume driven by ash venting. The strong dependency on thermal plume high on driving heat flux and the dependency of heat flux from magma on surface area means that fragmental activity is required to produce any significant dispersal of even fine ash. Plumes driven by the surface heat of a lava flow alone might only result in tens of meters of particle dispersal.
Chapter 5

Unusual fluid behaviour of a silicic magma during fragmentation in a deep subaqueous eruption, Havre 2012

This chapter presents work prepared for publication as a manuscript.

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Author contributions:

RJC organized and led the cruise. All authors were present on the 2015 cruise and collected the samples.

JDLW established the scope of the PhD project. All sample analysis was conducted by APM. APM wrote the manuscript and produced the figures. All authors read through drafts and assisted in editing of the manuscript.
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5.1. Abstract

Magma can respond to applied stresses in either a viscous or elastic manner, dependent on timescale over which strain is accommodated through diffusive rearrangement of its molecular structure, this is a function of the system chemistry, temperature, and deformation rate. The higher silica content and lower temperature of silicic magmas produces longer accommodation timescales. As such the high strain rates produced in pyroclastic volcanism typically results in brittle fragmentation.

The 2012 Havre eruption was the largest deep subaqueous eruption of the last century, occurring from a depth of 700-900 m depth. A population of silicic ash that displays features indicative of syn-/post-fragmentation viscous deformation has been observed in two subunits (Subunit 1 and 2) of the seafloor Ash Lapilli Unit produced during the 2012 eruption. The viscous ash composes approximately 20-30% of S1 and S2, the remaining ash shows brittle fragmentation. The grains are glassy with less than .5% crystal content and shows a range of vesicle textures. Viscous features include evidence of surface tension rounding, particle to particle welding, and post fragmentation vesicle inflation.

The silicic nature of the Havre magma and the high strain rates generated during fine fragmentation are at odds with the viscous behaviour of the melt, implying unusual eruption conditions. The occurrence of viscous and brittle ash together indicates a duality in fragmentation mechanisms and suggests local variations eruption conditions. Here we show that while the mechanism allowing viscous deformation remains unknown the presence of these grains likely implies an explosive eruption mechanism driving a gas supported jet.

5.2. Introduction

Magma is traditionally modelled as a viscoelastic fluid, whose rheological response to stress is a function of its chemistry, temperature and the rate of deformation (strain rate) (Dingwell and Webb 1990;
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Giordano et al. 2008; Gonnermann 2015). Chemistry and temperature define the magma’s Maxwell time, the period over which strain is accommodated following deformation; through diffusive rearrangement of the molecular structure. Macroscopically this is seen as viscous behaviour (Dingwell and Webb 1990; Gonnermann 2015). Deformation applied faster than the Maxwell time leads to elastic behaviour of the magma, followed by brittle failure past a critical stress (Dingwell and Webb 1990; Gonnermann 2015). The deformation rate required to exceed the Maxwell time is referred to as the critical strain rate (Dingwell and Webb 1990; Gonnermann 2015).

The ability of magma at a given temperature and pressure to behave in both elastic and viscous manners has important implications for fragmentation processes (Dingwell and Webb 1990; Namiki and Manga 2008). Low silica and high temperatures typical of mafic magma result in relatively short Maxwell times, meaning high strain rates can be accommodated by molecular rearrangement. Brittle fragmentation can therefore only be induced by the extremely high strain rates produced during direct magma water interaction (Zimanowski et al. 1997; Dürig et al. 2012) or highly energetic eruptions (Namiki and Manga 2008). Fragmentation of mafic melts often occurs due to fluid instabilities (Namiki and Manga 2008) following which surface tension, aerodynamic, and vesiculation processes can lead to reshaping of clasts (Porritt et al. 2012). In contrast, the lower temperatures and high silica of silicic magmas mean the critical strain rate required to produce brittle fragmentation is extremely low. During silicic explosive volcanism the induced strain rates are high and as such fragmentation is typically brittle (Zimanowski et al. 2003).

A significant population of ash particles defined by features indicative of post-fragmentation viscous reshaping of silicic ash while molten, have been identified in two subunits of the seafloor Ash and Lapilli (AL) Unit generated during the 2012 subaqueous eruption of Havre volcano (Carey et al. 2018). Reduced volatile exsolution because of eruption under high hydrostatic pressure will lower the magma viscosity. The small size of the fluidal ash at Havre implies energetic fragmentation, and thus high strain rates
(Zimanowski et al. 2003; Kueppers et al. 2006; Liu et al. 2015), contrary to the expected behaviour of silicic magmas, which fragment brittlely when rapidly strained (Webb and Dingwell 1990; Gonnermann 2015). This is compounded by the abundant ambient water which direct interaction of the magma with will drive high strain rates through energetic fracturing (Zimanowski et al. 1991; Zimanowski et al. 1997; White et al. 2003; Mastin et al. 2004; Schmid et al. 2010). The presence of the fluidal ash grains is therefore implicit of unusual eruption conditions allowing low strain rate fragmentation, or potentially resulting from variations in magma chemistry, or temperature.

To our knowledge, silicic ash deposits with a significant population of fluidal fine-ash grains have not yet been recognised elsewhere. Their presence is highly unusual and has important implications for the rheology of the Havre magma and for the fragmentation and eruption mechanisms. Here the fluidal particles are examined in detail to infer the mechanism(s) by which these unusual grains form, and what this implies for phase 2b of the 2012 Havre eruption.

5.3. Geological Setting

Havre is a fully submerged volcano located along the Kermadec Arc with a peak at approximately 650 metres below sea level (mbsl) (Wright et al. 2006). The volcano is truncated at approximately 900 mbsl by a caldera 3 by 4 km in diameter that has a relatively flat floor at 1500 msbl (Fig. 5.1.).

In 2012 the largest subaqueous silicic eruption of the last century took place at Havre volcano on the Kermadec Arc, from an inferred vent at a depth of approximately 900 mbsl (Carey et al. 2014; Jutzeler et al. 2014; Carey et al. 2018) (Fig. 5.1.). The eruption produced a sea-surface hot spot, pumice raft, and plume of discoloured water along with an atmospheric vapour plume from a point source over 21.5 hrs. (Carey et al. 2014; Jutzeler et al. 2014; Carey et al. 2018). A scientific cruise conducted in 2015 undertook detailed mapping, imaging, and sampling of the seafloor products form the eruption (Carey et al. 2018).
Fig. 5.1. The location of Havre volcano along the Kermadec Arc with an insert showing a MODIS (Aqua) image taken at 0126 on 19th July of the 2012 Havre eruption. A 1 m scale resolution bathymetry map of the Havre caldera and summit overlain on a lower resolution (35 m) bathymetry map of the whole of Havre volcano. Overlain is shown the bathymetry differences between the 2002 and Oct 2012 surveys (red = material added, purple = material removed). The locations of all clastic samples taken at Havre are shown along with the sampling method used. Samples in which detailed componentry was undertaken are labelled. The outlines of the ABL Unit and GP Unit outlines are marked in white.

The deposit freshness and completeness of the data set acquired during the Havre cruise, combined with pre-existing bathymetry (from a survey conducted in 2002 (Wright et al. 2006)) allow for a well-constrained and detailed interpretation of deep subaqueous silicic volcanism.

The seafloor deposits of the 2012 Havre eruption include 14 lavas, and three clastic units; the Giant Pumice (GP) Unit, the Ash Lapilli Block (BLA) Unit, and the Ash Lapilli (AL) Unit (Carey et al. 2018) (Fig. 5.1.). The AL Unit is composed of four subunits; from base to top these are Subunit 1 (S1), S2, S3, and S4. Subunit 1 and 2 are composed of glassy vesicular coarse (0.5 to 2 mm) and fine (16 to 32 µm) ash respectively. While S1 is a settling out deposit draping topography S2 is topographically constrained inferred to have formed from a dilute density current(s). Subunit 1 and 2 are inferred to have formed during a single continuous event along with the ALB Unit, where an explosive eruption drove dispersal of S1 in a thermal plume overlying a gas jet. The ALB Unit and S2 were then deposited from density currents formed by partial condensation of the gas jet from which S1 was dispersed. The fluidal ash observed at Havre is contained in S1 and S2.

Formation of S3 and S4 is inferred to have occurred following phase 2b during a period of effusive activity. Subunit 3 is inferred to have been generated from a lava on the southwest caldera rim by simultaneous ash venting and brecciation of its pumiceous carapace with dispersal in thermal plumes.
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Subunit 4 is composed of particles inferred to have been fragmented directly from lava flows following a collapse of the southwest caldera wall.

5.4. Methods

Ash samples were collected from deposits on the seafloor in 2015 using remotely operated vehicle (ROV) Jason. The sampling methods employed generally destroyed layering in the AL Unit. As a result, all but two samples were ‘mixed’ (i.e. they contain grains from two or more subunits mixed together as one mass). Samples were dried, and hand sieved from -4 φ to 4 φ in 0.5 φ steps. Subunit boundaries were determined by combining granulometry and componentry results and tracing signature features as outlined in Chpt. 3.

For this study we were focused on ash from S1 and S2, scanning electron microscopy (SEM) componentry and imaging was thus conducted on samples that were dominantly (>85%) composed of S1, S2, and S3. The elongate tube-vesicle grains that compose S3 were easily recognised and removed so that point counting data reflected dominantly grains form S1 and S2. SEM componentry results are therefore focused on the morphology of glassy vesicular particles.

Splits from the 3 φ, 4 φ and smaller than 4 φ size fractions, from 20 samples, were mounted in polish epoxy briquettes and on SEM stubs and carbon coated. Back Scatter Electron (BSE) and Secondary Electron (SE) montaged image maps were collected from both sample types with a Zeiss Sigma VP FEG SEM at the University of Otago using an accelerating voltage of 15keV and a working distance of 8.5 mm. Energy-dispersive X-ray spectroscopy (EDS) maps of some ash grains were also collected to assess the relative content of major elements across the particle.

SEM componentry was conducted by point counting on SE montaged images, using a step size approximately 1.5 times the average grain size. At each point the grain was grouped in to one of four
componentry classes (fluidal, tube, curvi-planar and angular), defined by particle morphology and surface texture (Chpt. 3.). Point counting was undertaken until at least 400 points had been grouped, for each size fraction.

Glass major-element chemistry was measured by electron probe microanalysis (EPMA) using a JEOL JXA-8230 SuperProbe Electron Probe Microanalyser at Victoria University, Wellington.

5.5. Results

Curvi-planar and angular grains dominate S1 and S2, however there exists a non-trivial component of fluidal grains identified in every sample thus far examined. Fluidal grains are defined by least one outer surface that exhibits features indicating surface tension or viscous reshaping of the grains while molten (Fig. 5.2.). In S1 and S2 the fluidal grains display features such as smooth unbroken rounded surfaces, inflated vesicle walls bulging beyond general clast surfaces, grains welded to each other, and drawn out melt fibres (Fig. 5.2. and 5.3.). These features are common in the fluidal ash in S1 and S2, along with a much rarer feature indicating viscous processes; partial deflation of inflated bubbles. In deflation features a concave bowl is seen in the middle of a bulged-out vesicle wall; the depression is inferred to result from partial collapse of an inflated vesicle as the internal pressure was removed. Many fluidal grains are truncated by secondary brittle fractures (Fig. 5.2.).

The results of SEM componentry show that for grains sizes of 3 φ (125 µm), 4 φ (63 µm), and smaller than 4 φ (63 µm), S1 and S2 are composed dominantly of curvi-planar particles (Fig. 5.4.). At 3 φ (125 µm), SEM componentry shows that curvi-planar particles make up between 52 and 81% of the total sample, with an average of 69%. At 4 φ (63 µm) curvi-planar particles compose between 59 and 86% of the total sample, with an average of 71%. For particles smaller than 4 φ (63 µm) curvi-planar particles
(previous page) Fig. 5.2. Secondary electron images of Fluidal grains from Havre. (a) a section through a grain with a fluidal smooth outer surface and displaying a vesicular core and a dense rim, (b) a grain showing a smooth outer surface, vesicle bulges can also be observed. The lines points to a bulge in which a concavity can be observed, a deflation feature. (c) a grain showing a fluidal surface and a vesicular surface, inferred to be the edge of a dense rim vesicular core grain that has disintegrated (d) two particles with a single smooth undulating fluidal surface each, with brittle fracture surfaces exposing cross section through inflated vesicles on the particle edges, (e) two Fluidal particles welded together (f) two grains that have highly fluidal exteriors.

compose between 50 and 82% of the total sample, with an average of 70%. The variance of both the total range and average percentage of curvi-planar particles in S1 and S2 is remarkably consistent.

On average the percentage of angular particles has an inverse relationship with the size of grains analysed (Fig. 5.4.). At 3 φ (125 µm) SEM componentry shows angular particles compose between 2 and 24% of the total sample, with an average percentage of 12%. At 4 φ (63 µm) angular particles compose between 6 and 28% of the total sample, with an average of 14%. For particles smaller than 4 φ (63 µm) the percentage of angular particles shows a significant increase, composing between 11 and 45% of the total sample, with an average of 22%.

The percentage of fluidal grains in S1 and S2 decreases with increasing grain size (Fig. 5.4.). At 3 φ (125 µm) SEM componentry shows fluidal particles compose between 11 and 35% of the total sample, with an average of 19%. At 4 φ (63 µm) fluidal particles compose between 6 and 28% of the total sample, with an average of 16%. For particles smaller than 4 φ (63 µm) fluidal particles compose between 3 and 16% of the total sample, with an average of 7%.

When examining the percentage change of curvi-planar, angular, and fluidal particles in individual samples there is some variance from the averaged trend. The observed variance however is typically
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*Fig. 5.3.* (a) Smoothly undulating outer surface defined by inflated edge vesicles, (b) an example of a dense rim vesicular core particle, the core shows extensive evidence of vesicle coalescence. (c) complex dense rim vesicular core particle displaying several inflated domains. (d) two particles showing inflated vesicle walls, (e) particle showing a highly fluidal edge (left) and cross cut by a brittle fracture (right), vesicles in this particle appear slightly deformed. (f) a particle showing a fluidal bulge along with a brittle fracture surface. (g) EDS maps showing relative variation of several major elements across the particle shown in (f).

restricted to single points, and no sample thus far examined displays opposite or strongly different trends (Fig. 5.4.).

To assess whether fluidal particles could have come from disintegration of Giant Pumice clasts also produced during the 2012 Havre eruption, we crushed part of a Giant Pumice clast from the GP Unit and conducted SEM componentry using the same method employed for the ash. In contrast to the sample of S1 and S2, SEM componentry results for the crushed GP showed that GP granulate was dominantly composed of angular particles (Fig. 5.4.). The percentage of angular particles decreases with grain size showing 82% at 3 φ (125 µm), 74% at 4 φ (63 µm), and 50% for all particles smaller than 4 φ (63 µm). The opposite trend was observed in curvi-planar particles whose percentage increased from 12% at 3 φ (125 µm), to 22% at 4 φ (63 µm), and to 48% for all particles smaller than 4 φ (63 µm). Fluidal particles in the GP granulate show the same negative trend with grain size, however composed only a trace amount of the overall samples 5% at 3 φ (125 µm), 4% at 4 φ (63 µm), and 2% for all particles smaller than 4 φ (63 µm). Result are summarised in table 5.1.

The fluidal particles in S1/S2 do not contain any phenocrysts. They are very glassy, with less than 5% area of acicular plagioclase and pyroxene microlites which are typically less than 10 µm in length although occasionally up to 40 µm (Fig. 5.3.). Vesicularity varies within and between fluidal particles, with some grains showing a vesicular core and dense glassy rim (Fig. 5.3.). Vesicles in fluidal particles are often
Fig. 5.4. SEM componentry data for fluidal, angular, and curvi-planar particles from samples of S1 and S2 (circles) and the Giant pumice granulate (triangles).
rounded displaying quite simple forms. In fluidal particles that contain a vesicular core, however, vesicle shape can be more complex, preserving evidence of bubble coalescence.

<table>
<thead>
<tr>
<th>Sample</th>
<th>Points counted (no.)</th>
<th>Fluidal (no. %)</th>
<th>Angular (no. %)</th>
<th>Curvi-planar (no. %)</th>
</tr>
</thead>
<tbody>
<tr>
<td></td>
<td>3 φ 4 φ 5 φ</td>
<td>3 φ 4 φ 5 φ</td>
<td>3 φ 4 φ 5 φ</td>
<td>3 φ 4 φ 5 φ</td>
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<tr>
<td>HVR159 Base</td>
<td>396 543 652</td>
<td>12.9 21.4 11.8</td>
<td>11.4 13.0 18.2</td>
<td>75.6 65.6 70.0</td>
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<td>12.3 16.0 20.0</td>
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<td>6.7 16.6 17.8</td>
<td>81.3 69.0 77.8</td>
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<td>1.9 7.8 13.4</td>
<td>80.1 86.0 82.5</td>
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<td>24.4 27.7 29.7</td>
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<td>9.5 9.6 21.2</td>
<td>75.4 73.6 73.2</td>
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<td>517 591 539</td>
<td>18.6 19.4 4.9</td>
<td>14.1 17.4 44.8</td>
<td>67.3 63.2 50.4</td>
</tr>
<tr>
<td>HVR122</td>
<td>401 474 485</td>
<td>35.3 10.1 9.1</td>
<td>12.4 15.4 26.9</td>
<td>52.3 74.5 63.9</td>
</tr>
<tr>
<td>HVR124</td>
<td>504 520 651</td>
<td>13.3 7.4 3.8</td>
<td>13.5 15.9 22.5</td>
<td>73.3 76.7 73.7</td>
</tr>
<tr>
<td>HVR191</td>
<td>698 655 711</td>
<td>25.4 27.9 10.1</td>
<td>16.4 10.2 26.1</td>
<td>58.2 61.9 63.8</td>
</tr>
<tr>
<td>HVR229</td>
<td>581 582 572</td>
<td>19.2 23.4 15.7</td>
<td>10.6 5.8 10.7</td>
<td>70.2 70.9 73.6</td>
</tr>
<tr>
<td>HVR283</td>
<td>607 638 553</td>
<td>26.4 20.4 8.5</td>
<td>10.8 8.4 18.1</td>
<td>62.9 71.2 73.5</td>
</tr>
<tr>
<td>Average</td>
<td></td>
<td>18.9 15.8 7.4</td>
<td>12.0 13.6 22.5</td>
<td>69.1 70.6 70.2</td>
</tr>
<tr>
<td>Max</td>
<td></td>
<td>35.3 27.9 15.7</td>
<td>24.5 27.7 44.8</td>
<td>81.3 86.0 82.5</td>
</tr>
<tr>
<td>Min</td>
<td></td>
<td>10.7 6.2 3.4</td>
<td>1.8 5.8 10.7</td>
<td>52.3 59.5 50.4</td>
</tr>
<tr>
<td>GP granulate</td>
<td>348 701 709</td>
<td>5.5 3.6 1.8</td>
<td>82.1 74.0 49.7</td>
<td>12.4 22.4 48.5</td>
</tr>
</tbody>
</table>

Table 5.1. – Results of SEM SE morphological point counting

Glass chemistry of ash from S1/S2 shows consistent major element compositions, with silica contents of 72-75 wt%. Energy-dispersive X-ray spectroscopy (EDS) chemical mapping shows no variation in the groundmass glass major element composition between ash formed brittlely or viscously. No spatial variation can be seen either in ash which shows both brittle and viscous features. An eruption temperature of 850 °C has been inferred for the Havre magma by Manga et al. (2018), based on two-pyroxene Fe-Mg exchange in compositions measured for clinopyroxene and orthopyroxene and assuming equilibrium conditions (Putirka 2008). The eruption that generated S1 and S2 is inferred to have taken place at a vent depth of approximately 900 mbsl, equivalent to 9 MPa of applied hydrostatic
pressure. Equilibrium glass saturation for these conditions would give a volatile water content of approximately 1 wt% (Newman and Lowenstern 2002). Using the Giordano et al., (2008) method, the calculated viscosity for 850 °C Havre magma with this water content and at this vent depth is $10^{6.7}$ Pa s (Table. 5.2.).

<table>
<thead>
<tr>
<th>Sample</th>
<th>Giant pumice</th>
</tr>
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<tbody>
<tr>
<td>SiO$_2$</td>
<td>71.92</td>
</tr>
<tr>
<td>TiO$_2$</td>
<td>0.47</td>
</tr>
<tr>
<td>Al$_2$O$_3$</td>
<td>14.01</td>
</tr>
<tr>
<td>Fe$_2$O$_3$</td>
<td>3.38</td>
</tr>
<tr>
<td>MnO</td>
<td>0.12</td>
</tr>
<tr>
<td>MgO</td>
<td>0.67</td>
</tr>
<tr>
<td>CaO</td>
<td>2.58</td>
</tr>
<tr>
<td>Na$_2$O</td>
<td>5.14</td>
</tr>
<tr>
<td>K$_2$O</td>
<td>1.62</td>
</tr>
<tr>
<td>P$_2$O$_5$</td>
<td>0.08</td>
</tr>
<tr>
<td>LOI</td>
<td>1.27</td>
</tr>
<tr>
<td>Total</td>
<td>99.71</td>
</tr>
<tr>
<td>H$_2$O total (wt%)</td>
<td>1</td>
</tr>
</tbody>
</table>

<table>
<thead>
<tr>
<th>Temperature (°C)</th>
<th>Log Viscosity (Pa s)</th>
</tr>
</thead>
<tbody>
<tr>
<td>750</td>
<td>$10^{8.1}$</td>
</tr>
<tr>
<td>850</td>
<td>$10^{6.7}$</td>
</tr>
<tr>
<td>950</td>
<td>$10^{5.6}$</td>
</tr>
<tr>
<td>1050</td>
<td>$10^{4.7}$</td>
</tr>
<tr>
<td>1150</td>
<td>$10^{3.9}$</td>
</tr>
</tbody>
</table>

**Table. 5.2.** Whole rock major element chemistry data for the giant pumice from (Carey et al. 2018), along with the viscosity range calculated for a range of temperatures. The approximate saturation water content is used for a vent of 900 mbsl. Viscosity calculations undertaking using the methods of (Giordano et al. 2008).
5.6. Discussion

5.6.1. Implications of fluidal features for fragmentation processes

The ash that composes S1 and S2 shows evidence of both brittle fragmentation (curvi-planar and angular grains) and viscous shaping (fluidal grains). The dominance of curvi-planar and angular grains suggests that fragmentation during the Havre eruption was dominantly brittle. The lack of evidence for viscous retraction of melt bridges in fluidal grains however, and their low presence in the GP granulate indicates that viscous shaping was a primary process, occurring syn- and/or post-fragmentation. Here we explore the implications of this bimodal fragmentation process during phase 2b of the 2012 Havre eruption (Chpt. 3.).

The dominance of curvi-planar (blocky and platy) morphologies and the common occurrence if stepped or river line fractures in ash is typically associated with magma water interaction (MWI), and specifically molten fuel coolant interaction (MFCI) (Wohletz 1986; Dellino and La Volpe 1996; Austin-Erickson et al. 2008). Both features are common to ash from S1 and S2 suggestive of extensive MWI fragmentation during the eruption of these units at Havre. However, some uncertainty as to the timing of the MWI exists. We have previously inferred that S1 and S2 were produced during an explosive phase of the Havre eruption, where a gas supported jet was generated (Chpt. 1.). The MWI signature may therefore reflect either a primary fragmentation (Zimanowski et al. 1991; Zimanowski et al. 1997; Austin-Erickson et al. 2008) or a secondary quench signature (Liu et al. 2015). Extensive MWI during fragmentation is consistent with the environment in which S1 and S2 were erupted and with inferences of other studies of silicic ash erupted in a subaqueous environment (Kano et al. 1996).

In contrast the fluidal grains in S1 and S2 indicate viscous melt behaviour, something that the extremely highly strain rate generated during MWI would preclude. The microtextures and shapes of fluidal particles in S1/S2 indicate a range of viscous behaviours post- and maybe syn-fragmentation (Walker and Croasdale 1971; Porritt et al. 2012). Features such as inflated outer bubble walls suggest that vesicle
growth continued after fragmentation, while the melt was behaving viscously, deforming the particle (Fig. 5.2. and 5.3.). Rounded and smoothly undulating surfaces unbroken by vesicles are indicative of surface-tension driven particle reshaping (Fig. 5.2. and 5.3.). Particles with a rounded, smoothly undulating dense unbroken rim and highly vesicular core suggest particle shaping due to both surface tension and post-fragmentation vesiculation. Vesiculation of the particle core would have led to stretching of the outer surface, which was rounded continually by surface tension. The features described above are common in samples of S1 and S2 and observed in grains 3 φ (125 µm), 4 φ (63 µm), and less than 4 φ (63 µm) in size.

Viscous vesicle expansion appears to have operated in concert with surface tension shaping in the formation of the fluidal ash at Havre. Modelling conducted by Wadsworth et al. (2017) defined a particle size limit to surface tension driving rounding following heating by a lightning strike. Rounding limits of ~30 to 10 µm radius were defined for ash generated during the 2009 eruption at Redoubt, Alaska, and the 2010 eruption of Eyjafjallajökull, Iceland, with heating of 4000 K. The fluidal ash at Havre however extends to particles sizes well beyond this range (at least 125 µm), and thus cannot be solely explained because of surface tension shaping. The dominance of one process over the other could then lead to the range of fluidal ash shapes observed i.e. rounded undulating walls forming from surface tension vs inflated vesicle walls from viscous vesicle expansion.

The duration of viscous behaviour is limited by heat loss (Dingwell and Webb 1990; Gonnermann 2015). The high surface-to-volume ratio of vesicular ash means these particles would have rapidly quenched (Wohletz 1983; Schmid et al. 2010), following which viscous deformation would be prohibited. Viscous shaping of particles is inferred to be a syn- or post-fragmentation feature therefore indicating the presence of a window following fragmentation and prior to quenching during which viscous shaping occurred (Fig. 5.5.).
The majority of fluidal grains observed in S1 and S2 are truncated by undeformed brittle fractures, demonstrating that they formed after the glass had solidified (Fig. 5.5.). These later brittle fractures are
inferred to result from quenching when hot viscous particles came into direct contact with liquid water (Carlisle 1963; van Otterloo et al. 2015).

Fluidal particles in S1 and S2 demonstrate that viscous reshaping of ash occurred syn- and/or post-fragmentation during their eruption. Surface-tension rounding and viscous vesicle expansion in a liquid-water-free environment appears to have driven the fluidal behaviour. Despite the presence of these fluidal grains in S1 and S2, fragmentation was dominantly brittle in nature with extensive MWI. The formation of both brittle and fluidal silicic ash in the same eruption implies variation in fragmentation mechanisms or conditions either temporally or spatially.

5.6.2. Implications for magma viscosity

Bimodal rheological behaviour during inferred pyroclastic activity with fluidal shaping of particles syn- and/or post-fragmentation in a magma the composition erupted at Havre is unusual. It implies spatial or temporal variations in magma chemistry, temperature and/or strain rate to allow viscous behaviour of a silicic magma on very short timescales (e.g. Dingwell and Webb, 1990; Giordano et al., 2008; Gonnermann, 2015). This is further compounded by the eruption occurring subaqueously where abundant ambient water would rapidly quench magma on contact, preventing further viscous shaping (van Otterloo et al. 2015). Reduced volatile exsolution resulting from eruption under hydrostatic pressure will decrease magma viscosity (Simpson and McPhie 2001; Cas et al. 2003; Busby 2005) (Table. 5.2.). Alone however, this does not account for the significant amount fluidal grains and rheological bimodality observed in the ash of S1 and S2.

Fluidal behaviour of silicic magma during pyroclastic eruptions has been observed in several subaerial deposits (e.g. Furukawa and Kamata 2004), typically inferred to result from high eruption temperatures (Branney et al. 2008; Self et al. 2008), or unusual chemical compositions (Dingwell et al. 1985; Stevenson et al. 1993; Dingwell et al. 1998).
The products of the 2012 Havre eruption do not show peralkaline compositions, and broad scale major element chemical variation is negligible (Carey et al. 2018). Additionally, no evidence of small scale chemical variation in the glass of fluidal particles from S1 and S2 has been observed. There is also no evidence to suggest an excess of volatile components that could account for the fluidal grains in S1 and S2 (e.g. H$_2$O, Cl, F, and S) (Carey et al. 2018). Differences in chemical composition can therefore be ruled out as a mechanism by which the fluidal particles in S1 and S2 were formed.

The eruption temperature at Havre of 850 °C has been petrologically calculated (Manga et al. 2018) on the assumption of equilibrium (Putirka 2008). Volcanic eruptions however are inherently non-equilibrium events, as such this temperature may record a deeper state of the magma system (Putirka 2008). At the inferred eruption temperature, the magma’s calculated viscosity seems to preclude viscous shaping over short timescales. A temperature increase during ascent on the order of several hundred degrees, required to produce viscous behaviour of the Havre magma during fragmentation however seems highly unlikely. During magma ascent and eruption there are several processes that can induce localized heating of the magma that could account for the fluidal ash, such as; crystallisation, shear zone development, and lightning. Local fluctuations in temperature induced in the conduit acting in concert with the overall reduced viscosity of magma erupted under hydrostatic pressure could potentially account for the fluidal ash population.

In Table. 5.3. several geological processes by which viscous shaping of fine ash could be induced are assessed, examining the evidence for and against each process in the context of the 2012 Havre eruption. The processes considered here are; viscous dissipation from a localised shear zone, volcanic lightning, and low strain rate in a tuffisite vein.

All the processes considered in Table. 5.3. fail to satisfactorily explain the fluidal particles observed in S1 and S2. Both viscous dissipation and volcanic lighting require a gas jet, consistent with an inferred pyroclastic formation mechanism for S1 and S2 (Chpt. 1.). In the case of viscous dissipation however
### Chapter 5

<table>
<thead>
<tr>
<th>Process</th>
<th>Process and products</th>
<th>Implications</th>
<th>For</th>
<th>Against</th>
<th>References</th>
</tr>
</thead>
<tbody>
<tr>
<td><strong>Viscous Dissipation</strong></td>
<td>- Heat generated by shearing&lt;br&gt;- In localized shear zones, such as those often inferred to form on conduit margins, heating of up to 250 °C could be caused&lt;br&gt;- Textural heterogeneities in tephra of explosive silicic eruption are sometimes attributed to the localized heating in conduit margin shear zones</td>
<td>- Formation of shearing zones in Havre magma&lt;br&gt;- Fluidal particles erupted from hotter veins in the magma column&lt;br&gt;- Would require explosive volcanism to exclude ambient water from fragmentation front and initial dispersal pathway&lt;br&gt;- Ash would still cool rapidly in gaseous environment therefore likely require hydrodynamic fragmentation</td>
<td>- Plenty of evidence for shearing in deposits of the Havre eruption&lt;br&gt;- A localized process therefore explains why we see fluidal and viscous behaviour</td>
<td>- No positive microtextural evidence in any of the 2012 Havre products for extensive syn-ascent heating&lt;br&gt;- No evidence for fluidal ash in silicic eruptions where viscous dissipation is inferred&lt;br&gt;- Hydrodynamic fragmentation unlikely in magma with the viscosity of that erupted at Havre</td>
<td>(Mastin, 2005; Hess et al., 2008; Polacci et al., 2001; Lavallée et al., 2015; Rosi et al., 2004; Wright and Weinberg, 2009)</td>
</tr>
<tr>
<td><strong>Volcanic lighting</strong></td>
<td>- Heating of ash by radiation from lighting in a jet/plume to 3000 to 5000 K or higher&lt;br&gt;- Leads to surface tension shaping of particles into glass spherules&lt;br&gt;- Heating/cooling vs rounding timescales limit rounding to particles &lt;~80 µm in diameter</td>
<td>- Lightning would require a gaseous environment to occur therefore requiring a gas jet and explosive volcanism&lt;br&gt;- Primary fragmentation brittle, lighting heating leads to viscous shaping of already brittle particles</td>
<td>- Explosive volcanism consistent with inferred eruption mechanism for S1 and S2&lt;br&gt;- A localized process therefore explains why we see fluidal and viscous behaviour&lt;br&gt;- Viscous shaping occurring during</td>
<td>- Particles in S1 and S2 extend to sizes beyond inferred rounding limit for lightning heating&lt;br&gt;- Fluidal grains compose much larger proportion of deposit that than seen in other cases where lightning is inferred</td>
<td>(Genareau et al., 2015; Wadsworth et al., 2017)</td>
</tr>
</tbody>
</table>
Chapter 5

Table 5.3. Outlining the processes by which local viscous behaviour over short timescales in a silicic melt may be induced. Here the implications of these processes and general arguments for and against each process in the context of the Havre eruption are also presented.

| Low strain rate – tuffisite veins | - At sufficiently low strain rates any magma can behave viscously  
- During pyroclastic activity slow deformation could perhaps occur in ash trapped in tuffisite veins then later released  
- Evidence for vesiculation driven viscous rounding of silicic ash can be seen in tuffisite veins in other settings  
- Slow particle deformation in Tuffisite veins  
- Require an unknown period post energetic brittle fragmentation in which to deform slowly  
- Unlikely to have occurred during pyroclastic volcanism  
- Does not require heating for which no microtextural evidence can be found  
- Water would be excluded from tuffisite veins  
- Tuffisite veins requires effusive/transitional volcanism, not consistent with implications on formation mechanism of S1 and S2  
- Viscous rounding in tuffisite veins inferred to be occurring during sintering, not conducive to subsequent dispersal  
- Seems like a very high percentage of Fluidal ash to be only generated by slow deformation while trapped in a tuffisite | rapid ascent, consistent with inference of vesiculation and surface tension rounding driving fluidal behaviour | - No evidence of glass volatile element migration  
- S1 and S2 Fluidal particle morphologies diverse, something not seen in the subaerial lightning spherules |

In natural deposits glass spherules compose <5% of the ash grains

[Tuffen et al., 2003; Saubin et al., 2016; Kendrick et al., 2016; Castro et al., 2012; Schipper et al., 2013; Mcgowan, 2016]
there is no positive evidence to suggest its occurrence during the Havre eruption, while the fluidal grains in S1 and S2 at Havre do not match with size and amount of lightning formed glass spherules in subaerial eruptions. Low strain rate deformation in tuffisite veins seems somewhat unrealistic and does not fit with inferences on the eruptive processes for S1 and S2. The exact processes by which the S1 and S2 fluidal grains form therefore remain uncertain.

5.6.3. Implications of fluidal ash-grain features for eruption processes at Havre

Although the process(es) by which fluidal particles formed at Havre remains uncertain, their presence has important implications for eruption processes. In the ash-size range, water-contact quenching is extremely rapid due to the high surface area to volume ratio of vesicular ash, and the rapidity of heat transfer into the water (Potuzak et al. 2008; van Otterloo et al. 2015). Viscous shaping had to have occurred prior to quenching. The fragmentation front and initial dispersal pathway must have been veiled from the ambient liquid water to allow viscous shaping prior to quenching.

Veiling of the fragmentation front and initial dispersal pathway of a deep subaqueous eruption requires a high heat and/or mass flux to vaporise and/or physically exclude the ambient water. Two potential processes by which veiling might occur are considered here; the first is a gas jet driven by intense pyroclastic volcanism (Fig. 5.6.), the second is within a tuffisite vein (Fig. 5.7.).

Subunit 1 and 2 are inferred to have been generated by a pyroclastic eruption in which ash was dispersed both in a thermal plume above a gas-thrust (S1) and from dilute density currents formed by condensation and collapse of the gas supported jet (S2). The gas jet model is implicit of explosive volcanism with a high heat and mass flux excluding the ambient water and instantly vaporising any that becomes entrained (Fig. 5.6.). In doing so a core gas supported region is sustained, the height of which is a function of the heat flux and the vent width (Head and Wilson 2003; Allen and McPhie 2009). Primary
fragmentation in this model is both hydromagmatic and magmatic, with the magmatic component driving a gas-supported jet (Allen and McPhie 2009). Viscous deformation of ash occurs post-fragmentation in the gas jet (Fig. 5.6.). Decompression as the particle ascends though the jet leads to viscous vesiculation acting with surface tension that generates a range of viscous features. As the viscous particles are ejected from the jet quenching of some ash generates brittle fractures cross cutting previous viscously formed features (van Otterloo et al. 2015) (Fig. 5.6.).

![Fig. 5.6. A schematic diagram outlining the inferred geometry by which Fluidal particles can deform post fragmentation in a gas supported explosive jet. A heating process is not in inferred on this diagram. Those particle that are at an elevated temperature are shielded from the ambient environment in the gas supported jet allowing them to deform above the glass transition. Particles that are not at an](image-url)
elevated temperature within the jet will not be fluidly deformed but will receive a quench over pressure on contact with the water.

A second model suggests low-strain-rate vesiculation and surface tension reshaping of particles trapped in tuffisite veins (Fig. 5.7.). Tuffisite veins form due to shear failure resulting from stiffening of the magma along the conduit edge (Castro et al. 2012; Kendrick et al. 2016; Saubin et al. 2016). When open, the veins are filled with a mix of ash and volatiles, which when connected to the surface can drive ash venting (Schipper et al. 2013). Over time, deposition of molten particles on the wall edges of a vein/crack in lava leads to compaction and sintering (Mcgowan 2016). Viscous pyroclasts have been observed in sintered tuffisite veins, inferred to have formed by in situ vesiculation (Mcgowan 2016). Low strain rate deformation is envisaged to occur in particles that are ‘stuck’ in the vein, and after some period are released and erupted (Fig. 5.7.). On eruption particles are rapidly quenched (Fig. 5.7.). Low strain deformation in a tuffisite vein however is not consistent with inferences on the formation mechanism of S1 and S2.

**Fig. 5.7.** A schematic diagram for the formation of Fluidal particles in a tuffisite vein (Kendrick et al. 2016). (a) Fragmentation and exsolution will occur as a tuffisite vein advances due to shear fracturing of the magma. (b) Once a tuffisite system becomes linked to the surface vent, outgassing of the volatile ash mixture occurs leading to ash venting at the surface. Hot molten particles may stick together in places and on the crack edge leading to sintering. The isothermal environment and low strain rate of particle
sintered grains leads to continued exsolution of some particles resulting in rounding. (c) Continued outgassing causes erosion plucking some Fluidal particles from the wall resulting in their dispersal. The lower overall viscosity of the Havre magma due to the hydrostatic pressure may result in a much higher percentage of these particle been formed than when compared to subaerial systems.

During the eruption of S1 and S2 fragmentation was dominantly brittle with magma water interaction processes playing a significant role (Fig. 5.5.). The exact timing of the magma water interaction fragmentation however is uncertain. At least 20% of the particles produced did not encounter water during fragmentation or their initial dispersal and were able to undergo viscous deformation, likely post-fragmentation (Fig. 5.5.). Eruption models of a gas supported explosive jet and eruption from a tuffisite vein are both considered to enable at least partial veiling of the fragmentation front and initial dispersal pathway. The gas jet model, implicit of explosive volcanism, is favoured here due to its consistency with previous inferences on the eruption mechanisms of S1 and S2 (Fig. 5.6.). In addition the gas jet model seems more realistic, having been described in the literature on subaqueous volcanism (Allen and McPhie 2009). The physical conditions and processes that were at play to allow viscous deformation of a silicic magma on short time scales are unknown. They are inferred to be the result of locally increased eruption temperatures, potentially resulting from fragmentation of high temperature sheared zones, or due to lightning strikes occurring within the gas jet.

5.7. Conclusions

Ash from S1 and S2, seafloor deposits produced in the 2012 Havre eruption, both show dominantly brittlely formed morphologies, however they also preserve extensive evidence of viscous reshaping syn- or post-fragmentation, and prior to quenching. The fluidal ash extends to fine grains sizes which are typically indicative of energetic fragmentation, something not conducive to viscous deformation. Viscous
reshaping appears to have been driven by both viscous vesicle expansion and surface tension rounding which may suggest a post fragmentation origin. The decreased viscosity resulting from eruption under high hydrostatic pressure does not account for the fluidal behaviour of the ash at high strain rates alone. Several potential mechanisms to either increase the magma temperature locally (lighting or viscous dissipation) or allow low strain rates (deformation in a tuffisite vein) are considered to explain the presence of the fluidal grains. Neither of the heating methods explain all the observed data, while the tuffisite model fails to explain the stratigraphic relationship of S1 and S2 to the GP Unit. The dependence of viscosity on temperature means that viscous behaviour is limited by heat loss. The high quench rate of fine vesicular ash when in contact with water means that particles must have been veiled from the ambient environment during fragmentation and their initial dispersal. It is suggested that the fluidal grains where veiled from contact with the ambient water by a gas supported jet, driven by explosive volcanism. The mechanism leading to locally increased magma temperature to cause the fluidal behaviour is still unknown. Identification of the potential heating mechanisms that lead to viscous behaviour are the focus of further work.
Chapter 5
This chapter presents initial results on a population of foreign grains in the AL Unit at Havre

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Author contributions:

RJC organised and led the cruise. All authors were present on the 2015 cruise and collected the samples. JDLW established the scope of the project. All sample analysis was conducted by APM.

APM wrote the manuscript and produced the figures. All authors read through draft and assisted in editing of the manuscript.
Chapter 6

6.1 Abstract

Particle dispersal in the subaqueous environment lends itself to long distance transport of fine ash because of the comparatively low submerged specific gravity, and water's high viscosity relative to air. A population of foreign grains has been identified in a range of seafloor samples from the 2012 Havre ash samples. The foreign grains are typically fresh, with only a few displaying evidence of glass hydration rims around the particle edge and along open cracks and vesicles. The groundmass glass composition ranges from basalt through to dacite (51 to 70 wt% SiO₂). Microlites present are generally acicular plagioclase and pyroxene, with their proportions differing between particles. Groundmass microtextures also vary, displaying differing degrees of microcrystallinity, textures, and microlite shapes. Comparison of the groundmass-glass major elements with bulk rock samples taken from various volcanic centres along the Kermadec Arc shows a similar fractionation trend. In several cases, the chemistry of grains overlaps with that of published analyses from specific centres e.g. Raoul, Hungaroa, Healy, Macauley, etc. However, the overlaps are not consistent across the full range of major elements, complicating direct correlations.

The observed chemical and microtextural variation of the foreign grains, together with the lack of compositional banding or foreign inclusions in the products of the 2012 Havre eruption suggest the foreign grains come from sources other than the 2012 eruption. The freshness of the particles also indicates these grains are not an excavated country-rock lithic component from the older Havre volcano. Even less likely as a source is break-up of the caldera wall rocks, which host hydrothermal systems. Two other sources are considered as the potential origin of the foreign grains; the first is scouring and re-sedimentation of particles initially deposited on the various pre-2012 surfaces of Havre Volcano. The second source considered is recent volcanism and/or scouring of the flanks of other volcanic centres along the Kermadec Arc, followed by transport to Havre. Neither of these two possibilities can be proven based on current data. Future investigation of trace element composition of these particles would potentially allow positive identification of eruptive sources.
Chapter 6

6.2. Introduction

Particle dispersal and transport processes are the lens through which eruption dynamics are inferred from their deposits. The processes by which fine volcanic particles are dispersed in the subaerial environment contrasts with those that occur subaqueously due to the differences in the physical properties of the ambient medium (Cashman and Fiske 1991; Fiske et al. 1998; Allen et al. 2008; Deardorff et al. 2011). However, when considering the dispersal of fine particles steam charging and resulting buoyant ascent (Allen et al. 2008; Fauria et al. 2017) can be disregarded, as the high surface to volume ratios and the low heat mass of the ash means quenching is rapid.

Water is 1000 times denser than air and 100 times more viscous (White et al. 2003; Cas and Giordano 2014; White et al. 2015). These physical realities have important implications on particles transport processes. One of the more significant impacts on this is the large reduction in particle settling velocities (Ferguson and Church 2004). For individual particles this can results in several orders of magnitude difference in settling times compared to subaerially, with the finest grains potentially settling out over months or years in still water. Although the formation of vertical density currents can greatly increase settling rate (Fiske et al. 1998; Manville and Wilson 2004), turbulence in the water column could act to keep fine particles suspended. Particles settling over potentially months could be transport extremely long distances in local currents producing a potentially quite unusual dispersal pattern.

Particle entrainment is partly a function of specific gravity and the viscosity of the ambient fluid (Valyrakis et al. 2013). Particle entrainment will thus be more effective in the subaqueous environment compared to the subaerial. The formation of asymmetrical ripples, scour and moating around blocks, and crag and tail structures (among other features) to water depths of at least 1500 mbsl demonstrate the presence of currents with sufficient velocity to entrain particles deep in the water column (Wright 2001).

These impacts on particle dispersal processes have important implications on how we interpret deep sea ash deposits compared to subaerially. Potentially distal dispersal from re-entrainment or effusive
eruptions could lead to wide spread deposits with local eccentricates resulting from local currents. This must be considered when examining subaqueous ash deposits on the modern seafloor or in ancient uplift successions.

Here are examined a population of fresh foreign ash particles recovered from the seafloor deposits of the Ash and Lapilli Unit (AL Unit) of the 2012 Havre eruption, but not chemically or microtexturally associated with the eruption. The foreign grains are microtexturally described and the glass major element assessed in the context of Kermadec Arc volcanism. An attempt is made to infer the most probable sources of the particles.

6.3. Methods

This chapter draws on microtextural observations made from SEM images, and major-element chemistry, to support its interpretations. The methods by which these results were collected are outlined in Chapter 2. Methods.

6.4. Results

6.4.1. Foreign grains

During microtextural and glass major element chemistry investigation of samples of the AL Unit, a population of foreign grains was identified. The foreign grains have mostly been imaged in the 3-4 φ (125 – 63 μm) size range, with the largest particle thus far observed in the 1 φ (500 μm) size range. Foreign grains are present in a range of samples from around the Havre study area (Fig. 6.1.). Foreign particles have also been identified in the base subsample of HVR159 (representative of Subunit 1 (S1)) and, Sample HVR070 (representative of Subunit 3 (S3)) demonstrating their presence throughout stratigraphy. The lack of foreign grains from samples is likely more a function not having specifically looked for them there as opposed to an actual lack of them.
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The foreign grains imaged so far typically show fresh looking unaltered groundmass glass. Some particles show evidence of slight glass alteration, with different coloured rims around the outer particle surface, in open pores/vesicles, and along cracks (Fig. 6.1.). The microlite populations of the foreign grains vary greatly. Typically, they comprise plagioclase and pyroxene, but relative proportions differ grain by grain, as do textural and morphological features of the population of microlites (Fig. 6.1.). Plagioclases larger than ~ 400 µm$^2$ show tabular forms, while those smaller than this are typically acicular (Fig. 6.1.). The plagioclase generally appear more anorthite-rich (brighter in BSE) than those in the Havre eruption, and in some particles, show simple rims (Fig. 6.1.). Pyroxene microlites have different shapes in different clasts, with some clasts showing highly acicular microlites while others have notably more-tabular forms (Fig. 6.1.). In some grains, euhedral to subhedral oxide/sulphide microlites are also observed. Groundmass microcrystallinity varies from 5% to upwards of 50% (Fig. 6.1.).

Particle vesicularity is generally relatively low, but the small size of the foreign grains means that their vesicularity may not represent that of the magma fragmented to produce them. In low-microcrystallinity particles, vesicles are round to oval in 2D. In particles with abundant microlites, vesicle shape is constrained by microlites, and they have ragged shapes.

Major-element analysis of the groundmass glass has been conducted on the foreign grains at Havre using both EMPA (WDS) and EDS methods. The glass chemistry shows that foreign grains span a range of compositions from basalt through to dacite/rhyolite (Fig. 6.2.). The major-element glass chemistry of the Havre foreign grains shows similar trends to those defined for the whole of the Kermadec Arc (Fig. 6.2.). The results plotted in Fig. 6.2. are from whole rock analyses of samples taken from various volcanic centres along the Kermadec Arc (Wright et al. 1996; Wright and Gamble 1999; Wright et al. 2006; Barker et al. 2013), whereas here the groundmass glass of the foreign grains have been analysed. Comparison of groundmass glass vs whole rock is not like for like. The inclusion of crystals in the whole rock will lead to a more mafic composition compared to the glass. Due to variation in the composition of this crystal content however that change is not predictable.
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(Fig. 6.1). SEM BSE images showing microtextural variation in the foreign grains. All scale bars are 10 µm, unless otherwise denoted. A range of groundmass microcrystallinities can be observed from low (f) to high (k, l). The range in microtextural features point to contrasting decompression histories. Grains typically appear quite fresh, with (g) been the only example of alteration of the grain edge.

This can be seen where the glass chemistry of foreign grains appears to cluster with analyses of material from eruptive centres on the Kermadec Arc. The clustering however, does not appear to be consistent between plots of different elements (Fig. 6.2.). Typically, the more silicic points in Fig. 6.2. come from aphyric pumice (Wright et al. 2006; Barker et al. 2013), which will more closely reflect glass composition. The comparison of major element chemistries however is not used to infer specific source, only to demonstrate the similarities in composition of the whole Kermadec Arc and the foreign grains identified in the AL Unit at Havre.

![Fig. 6.2. A total Alkali vs Silica (TAS) diagram showing the groundmass glass composition from the products of the 2012 Havre eruption (using EPMA), and the foreign grains. Foreign grain major element data was collected using both EPMA and EDS shown in open red and green triangles respectively. The results from the Havre deposits are then compared with whole rock data from various eruptive centres in the southern part of the Kermadec Arc, data from Wright et al. (1996); Wright and Gamble (1999); Wright et al. (2006); and Barker et al. (2013).]
Sample HVR054 was taken from the base of a large gully in the southwest caldera wall below lavas G-I (Fig. 6.3.). The sample is composed of fine particles that were collected via vacuum sampling from an extremely cohesive deposit, not consistent with any of those identified in the AL Unit, on top of a dark dense block. Images of the seafloor around the location of sample HVR054 show dominantly dark dense blocks, with occasional large pumice blocks, overlain by the cohesive fine deposit. Downslope from the mouth of the gully a large bolder field can be seen in the Sentry map (Fig. 6.3.).

Examination of sample HVR054 using an optical microscope showed pyrite makes up ~5-10% of the total deposit (Fig. 6.4.). Many of the ‘grains’ were clumps of extreme fine particles. Few primary ash particles from the 2012 Havre eruption were observed, often those that were observed were coated by fines (Fig. 6.4.). SEM BSE analysis showed that grains in HVR054 are generally composed of clumps of smaller particles aggregated together in an extremely fine-grained matrix (Fig. 6.4.). The clumps do not show any internal structure and are inferred to be secondary features formed during sample drying as opposed to primary volcanic aggregates. These clumps often incorporate fresh glassy vesicular and microcrystalline particles from S1 and S4, along with pyrite crystals, and rare examples around foreign grains (Fig. 6.4.). The presence of pyrite was confirmed used EDS.

Discussion

The 2012 Havre eruption produced a chemically consistent magma, showing only a few percent variation in whole rock silica values, across both lavas and clastic material (Carey et al. 2018). Banding has been observed in lapilli and blocks from the pumice raft, Giant Pumice Unit, and the Ash, Lapilli, Block Unit. However, bands are defined by variations in groundmass microcrystallinity and vesicularity with no change in composition (Carey et al. 2018). The chemical homogeneity of the
magma produced in the 2012 Havre eruption compared with the diversity seen in the compositions of the foreign grains suggests they were not produced in the Havre eruption (Fig. 6.2. and 6.5.). The

Fig. 6.3. A detailed bathymetric model of the southwest caldera wall in the vicinity of Lavas G to I, showing the caldera wall collapse structure, and the location of sample HVR054 (Carey et al. 2018).
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The inferred scarp line is marked by a white dashed line. The scarp has been covered by the talus slope Lava H indicating continued extrusion following the collapse. The down slope deposit can be seen on the floor of the gully, including a concentration of blocks at the base of a reverse slope.

fact that the foreign grains are observed dispersed through the stratigraphy of the AL Unit, their apparent freshness, and that only fine particles have been observed indicates they are not a lithic component.

Fig. 6.4. SEM BSE images showing particle from sample HVR054. (a) and (d) show primary 2012 Microcrystalline and Glassy Vesicular grains coated in a mix of extremely fine-grained mud, larger altered grains, and pyrite crystals. (b) a close-up view of the fine-grained mud matrix of the agglomerate particles from HVR054. (c) showing a foreign grain and an altered particle with a thin mud coating.

The apparent freshness of the foreign grains indicates a relatively recent formation (Fisher and Schmincke 1984; Jacobsson and Moore 1986). The lack of alteration also indicates they cannot have
spent any considerable amount of time proximal to a consistent source of heat due to the rate of palagonitisation at high temperature (Fisher and Schmincke 1984; Jacobsson and Moore 1986). Over their lifetime volcanoes are fluxed by high temperature fluids resulting in alteration, especially of the clastic component due to its typically high surface area and permeability (Fisher and Schmincke 1984; Jacobsson and Moore 1986). At Havre, hydrothermal venting appeared to have been focused on the caldera ring fault. As such, the caldera walls would likely be highly altered, as indicated by the presence of abundant pyrite in sample HVR054, formed by collapse of the caldera wall (Fig. 6.4.). Fragmentation and dispersal from the Havre caldera wall (or similar processes at other volcanoes) can therefore be ruled out as the source of the foreign grains. Two other potential sources for the foreign grains are considered here. The first is that particles have been scoured from pre-2012 deposits on the outer flanks of Havre volcano and transported to the caldera in local currents. Alternatively, particles could have been sourced from other volcanic centres; either directly from volcanism, or due to scouring and entrainment, followed by long distance transport in regional currents.

Turbulence in deep ocean currents resulting from their interaction with the Havre edifice may lead to scouring of the outer flanks (Boehlert 1988). The high viscosity of water and the lower specific gravity of submerged grains means they are more easily entrained in flows compared to the subaerial environment (Valyrakis et al. 2013). During the Havre cruise, ripples were observed on the seafloor sediment, and strong currents were occasionally encountered when maneuvering ROV Jason; such features suggest a relatively dynamic seafloor environment. Strong seafloor currents could entrain loose ash on the seafloor. Once the ash is entrained, the turbulence of bottom currents would keep particle suspended for some distance. Particles could then be dispersed continuously over the edifice from which they were sourced as well as for some distance around it. The continuous re-sedimentation rate would result in synchronous deposition with the seafloor deposits of the 2012 Havre eruption.
The Havre edifice is constructed of volcanics that display a range of compositions from basalt through to rhyolite (Wright et al. 2006). The outer flanks of Havre are described as been composed of plagioclase-bearing to aphyric lavas of basalt to dacitic composition (Wright et al. 2006). Any fine grained deposits on the flanks of the Havre edifice resulting either from pyroclastic activity synchronous with the lava effusive, or from fragmentation of the lavas themselves (Resing et al. 2011; Portner et al. 2014; Portner et al. 2015) are likely to have similar compositions. Rapid quenching and transport away from the heat source of the eruption will slow the alteration process significantly (Fisher and Schmincke 1984; Jacobsson and Moore 1986). Wright et al., (2006) however also described the outer flanks of Havre volcano as having been mantled by weathered pumice. The degree of weathering is not quantified; however, alteration has clearly begun. There is no evidence as to the timing of volcanism prior to the 2012 eruption at Havre (Wright et al. 2006). It is therefore unknown the length of time that material may have been sitting on the Havre edifice prior to entrainment and dispersal. It seems likely however that weathered pumice would indicate that fines exposed on the Havre flanks would also have begun to weather. However, the foreign grains examined here generally show little to no evidence of alteration/weathering (Fig. 6.1.).

An alternative explanation is that particles are not sourced from Havre, but instead have come from various volcanic centres along the Kermadec Arc (Fig. 6.5. and 6.6.). Particle could have been entertained by scouring, as described above, and then transported long distance in arc parallel currents, such as has been described in the Kermadec Islands region (Sutton et al. 2012). The extreme groundmass freshness suggests that the foreign grains may have been sourced from recent volcanism, and potentially transported directly from such activity. The Kermadec arc is an area of active volcanism from several centres (Worthington et al. 1999; Chadwick et al. 2008; Watts et al. 2012; Wormald et al. 2012; Metz et al. 2016). At Monowai, subaqueous activity includes production of discoloured water plumes, which clearly show that fine particles are been dispersed, potentially long distances (Fiske et al. 1998; Chadwick et al. 2008; Watts et al. 2012; Wormald et al. 2012; Metz et al. 2016). Dispersal and transport from several volcanic centres would explain the microtextural
and chemical diversity, whilst also explaining the particle freshness. Chemically the foreign grains glass major elements are consistent with the Kermadec arc’s fractionation trend (Fig. 6.2. and 6.5.). True fingerprinting by source would require trace element analyses. In the case of the multiple eruptive centres source however, we are talking about the dispersal of fine particles over hundreds
Fig. 6.5. A Harker diagram showing variation in several major elements vs silica for groundmass glass ash produced in the 2012 Havre eruption (grey circles), the foreign grains (orange circles), and data from a range of volcanic centres along the Kermadec Arc (also shown in Fig. 6.2.) (Previous Havre data blue circles; other volcanic centres yellow circles) (data from (Wright et al. 1996; Wright and Gamble 1999; Wright et al. 2006; Barker et al. 2013)). The foreign grains typically plot on a similar trend to that defined by the broad arc excepting Na$_2$O and FeO (systematic error?).

of kilometers from potentially low intensity eruptions or simply scouring of the seafloor. Although particle settling rate is much reduced in the subaqueous environment, the distances involved in such transport are beyond what would be expected. There therefore may be additional processes at work here that have yet to be identified.
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Previous page Fig. 6.6. A regional bathymetric map of the southern Kermadec Arc showing the location of the volcanic centres in Fig. 6.2 and 6.5. Subaqueous volcanic centres are named and shown as blue triangles, green triangle denotes the location of subaerial volcanic centres.

The presence of foreign grains within the AL Unit not scoured from the Havre eruption has important implications for other studies examining subaqueously produced ash deposits. The fact that the foreign grains are observed throughout the AL Unit indicates a relatively steady rate of deposition over the whole study area during the eruption. When examining seafloor deposits, especially in a low depositional rate environment, the fact that particles may have been transported from potentially extremely distal sources must be accounted for. I speculate that in the distal environment of the Havre eruption the lower rate of deposition could lead to foreign grains composing a far greater percentage of the overall deposit than that seen here. In distal deposits of subaqueous eruptions such a sediment source could account for a non-trivial amount of the overall deposit.

6.6. Conclusion

The presence of foreign grains in the Havre deposit may have important implications on long distance dispersal of fine particles in a subaqueous volcanic arc environment. There is currently insufficient information to explain the presence of these grains in the 2012 Havre deposits. Fingerprinting of the source of these grains would require trace element analyses of the particles, and comparison with various eruptive centres on the Kermadec Arc. Fingerprinting of the foreign source would likely provide the best constraints to any further interpretation. The particle freshness however indicates that they were generated recently and may hint at dispersal from an eruption(s) that occur prior to the 2012 Havre eruption somewhere within the Kermadec Arc.
Chapter 7

SR-FTIR results

This chapter presents preliminary work.

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Author contributions:

RJC organised and led the cruise. APM, JDLW, and RJC were present on the 2015 cruise and collected the samples. JDLW established the scope of the PhD project. ARLN and IMM assisted APM greatly in the processing and interpretation of the results. APM wrote the chapter and produced the figures. All authors read through and assisted in editing of this chapter.
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7.1. Abstract

The applied hydrostatic pressure during subaqueous volcanism reduces volatile exsolution, impacting greatly on magma ascent conditions and eruption processes. The volatiles preserved in subaqueous eruption products inform us about exsolution, fragmentation and quenching processes; these in turn can provide strong controls when examining eruption processes. In 2012 a subaqueous eruption from Havre volcano occurred from a depth of 900-1200 meters below sea level (mbsl), producing a range of seafloor and sea surface products. Pre-eruption bathymetry, along with a detailed follow up cruise mean that Havre is the ideal eruption the examine a range of subaqueous volcanic processes. This chapter focuses on ash sampled from Subunits 1 (S1), 2 (S2) and 3 (S3) of the seafloor Ash and Lapilli (AL) Unit produced during the 2012 Havre eruption. Using Synchrotron radiation-Fourier transform infrared microspectroscopy (SR-FTIR) on Havre ash from which we can infer quench depths, we aim to provide constraints on eruption dynamics, as well as to gain insights on fragmentation and dispersal processes.

SR-FTIR data from grains of S1 and S2, produced during phase 2b, show strong differences to those analyzed from S3, produced by syn-extrusive ash venting at Lava G. The S3 grains are generally quite highly hydrated and show well defined $H_2O_{\text{total}}$ range. In contrast, a blocky grain from S1 or S2 shows no hydration; while a fluidal grain from the same unit is highly hydrated. Using OH content, reconstructed total water contents show a range of 0.54-0.78 wt% for S3, values of 0.89 to 1 wt% and 0.77 wt% were recorded for the blocky and fluidal particles from S1 or S2 respectively. Using an eruption temperature of 850 °C these reconstructed $H_2O_{\text{total}}$ values suggest quench depths of 250 to 500 mbsl for S3, for S1 or S2 the blocky grains indicate a quench depth of 6.8 to 850 mbsl and the fluidal grain 510 mbsl.

The disparity between apparent an apparently shallower quench depth compared to vent depth implies that fine particles were dispersed vertically up to several hundred meters prior to contact with the water. Veiling of fine ash for several hundred meters through the water column is inferred to be the
result of dispersal in a gas supported jet. Such a jet would be implicit of explosive volcanism and is consistent with previous inferences regarding the deposition of S1 and S2, along with the presence of a population of fluidal particles. A jet between 400 m and 650 m high however does not fit with inferences regarding the eruption mechanisms of S3. We also therefore note that volatile solubility is a function of both pressure and temperature. An increase in magma temperature could lead to increased degassing, resulting in a falsely shallow quenching signal. In the eruptions of S1 and S2 along with S3 there are inferred to be processes operating that could lead to several hundred degrees of syn-eruptive heating.

7.2. Introduction

Magmatic volatiles play a significant role in driving volcanism, as well as modulating the eruption style through impacts on magma rheology and the storage or build-up of over-pressure. The solubility of volatiles in a silica melt is controlled by pressure, temperature, melt composition and volatile species (Newman and Lowenstern 2002; Wallace et al. 2015). For silicic magmas water is the main volatile species composing approximately 95% of all dissolved gases (Burgisser and Degruyter 2015).

The pressure dependency of volatiles in magma can be exploited to calculate the pressure at which exsolution was stopped, using experimentally determined solubility models (Zhang, Belcher, et al. 1997; Zhang 1999; Newman and Lowenstern 2002). Equilibration of magma \( \text{H}_2\text{O} \) content to ambient pressure conditions occurs in less than a second (Zhang and Ni 2010), until cooled below the glass transition, at which point the chemical system becomes viscously ‘locked in’. By measuring \( \text{H}_2\text{O} \) content in the ash and comparing results with experimentally determined solubility models, we can estimate the depth at which the fragmented magma first quenched to ash. For subaqueous and subglacial volcanic eruptions, quench pressure can be used to estimate the depth of overlying water or ice, and to constrain interpretations of eruption processes (e.g. Tuffen et al. 2010; Nichols et al. 2014).
The small volume high surface area of ash means that on contact with water it will quench rapidly and will thus in theory preserve a record of the quench depth during the 2012 Havre eruption. Quench depths of the ash can be used to constrain and inform discussions regarding the dynamics of the 2012 Havre eruption.

Initial results of SR-FTIR from ash from the 2012 Havre eruption are presented here. Interpretations and the constraints these would place on the inferred eruption mechanisms for S1 and S2 are also shown.

**7.2.1. The 2012 Havre eruption and its seafloor products**

Havre is a fully submerged volcano located along the Kermadec Arc, west of the Kermadec Ridge (Wright et al. 2006). The volcano forms a one-kilometre high edifice that is truncated by a summit caldera four and a half kilometres in diameter (Fig. 4.1.). The caldera floor is relatively flat at approximately 1500 msbl with the walls rising at least 500 m on all sides (Fig. 4.1.).

In 2012 the largest deep-water (>500 mbsl) silicic eruption ever recorded occurred at Havre volcano (Carey et al. 2014; Jutzeler et al. 2014). The eruption was imaged by Moderate-Resolution Imaging Spectroradiometer (MODIS) satellite images taken on 18th and 19th July 2012, coordinated universal time (UTC) (Carey et al. 2014; Jutzeler et al. 2014). Over the period of ~21.5 hours, from 18th to 19th July 2012, a pumice raft ~400 km² in size was produced along with a plume of discoloured water (Carey et al. 2014; Jutzeler et al. 2014). Accompanying the production of the pumice raft over the same period, an atmospheric plume emanating from a point source was observed above Havre (Carey et al. 2014; Jutzeler et al. 2014). A MODIS image taken at 1050 18th July showed a thermal hot spot above Havre, approximately 4-6 °C hotter than the surrounding water (Bernard 2012). From 17th to 21st July frequent earthquakes of magnitude three to five were also recorded from Havre (Carey et al. 2014; Jutzeler et al. 2014). Following the appearance of the pumice raft, vapour plume and hot spot imaged on 18th – 19th
July, no further sea-surface activity was imaged above Havre. Comparison between pre-eruption bathymetry from 2002 (Wright et al. 2006) and a survey undertaken post-eruption in October 2012 revealed large scale changes to the seafloor (Carey et al. 2014; Jutzeler et al. 2014). A repeat survey undertaken in 2015 showed no further bathymetric changes, confirming that eruptive activity had ceased prior to the October 2012 survey (Carey et al. 2018).

The 2012 Havre eruption occurred from a series of vents along the southern caldera wall and rim in three structural lineaments (Carey et al. 2018). Vent depth ranges from 900 to 1200 mbsl (Carey et al. 2018). ROV and AUV surveys conducted in 2015 showed that the large-scale seafloor changes were the result of the emplacement of 14 lavas (Carey et al. 2018). Along with the lavas three major clastic units were identified; ‘Giant Pumice Unit (GP)’, ‘Ash Lapilli Block Unit (ALB)’, and the ‘Ash and Lapilli Unit (AL)’, (Carey et al. 2018).

7.2.2. The Ash and Lapilli unit

The Ash and Lapilli (AL) Unit is a widespread deposit, produced during the 2012 Havre eruption, and composed of four distinct subunits (Carey et al., 2018; Murch et al in prep). The subunits that composed the AL Unit show dominantly ash sized grains with a minor number of lapilli (the content of the following section is drawn from Chpt. 3). The four AL subunits were defined from mixed samples of the whole deposit using geographical variations in grain size and componentry. From the spatial distribution of specific, and recognisable, grain size nodes and componentry classes we could extract virtual subunit outlines. Subsampling of two samples that preserved original layering allowed us to link the virtual subunits with the seafloor stratigraphy. By comparison with seafloor images of the ash stratigraphy, taken using ROV Jason, we were able to confirm the validity of this method and provide several points of known relative stratigraphy. The results of the above analysis showed four distinct subunits can be
identified, with varying formation mechanisms, sources, and dispersal patterns. The methods by which subunits in the AL Unit were established are outlined in detail in Chpt. 3.

From the lowest subunit to the top, Subunit 1 (S1) is the basal layer of the exposed AL Unit, apparently directly overlying the GP Unit in stratigraphy. It has been identified in every seafloor clastic sample taken at Havre and must have been dispersed beyond the study area in every direction. Seafloor images show it has a thickness of at least 5 cm. Subunit 1 is defined by glassy vesicular (GV) grains and has a grain size mode at -1 to 1 φ (2 mm to 500 μm). Particle shapes are dominantly curvi-planar with minor amounts of fluidal, and angular particles. The formation mechanism of S1 is currently under debate, however its close microtextural similarity to the raft and giant pumice suggests it is related to their production. The presence of fluidal particles in S1 has shown that it is at least partly a primary fragmentation deposit, and not the result of comminution.

Subunit 2 (S2) directly overlies S1 and has a deposit border that follows the northern caldera wall. It has a characteristic grain size mode of 5 to 6 φ (32 to 16 μm). Scanning electron microscope images show that the grains in this size range are typically GV grains, with rare microcrystalline particles. Below approximately 10 μm the particles are typically dense, with curvi-planar and shard morphologies.

Subunit 2 is locally divided into lower (a) and upper (b) parts because Subunits 3 (S3) and 4 (S4) have local dispersal. Where S3 and S4 are not present however S2a and S2b cannot be distinguished from one another. Subunit 2 thickens on the caldera floor at 4-10 cm thick compared to approximately 3 cm on the caldera rim. Subunit 2 is inferred to be the fine-grained deposit associated with a density current, coming from the vent below Dome OP.

Subunit 3 is characterized by the distinctive grains called Elongate Tube-Vesicle (ETV). These particles have been deposited over the southern caldera rim, the caldera floor, and in a single sample on the western caldera rim. The deposit fines and thins away from its inferred source at Lava G on the
southwest caldera rim, going from a mode of -1 φ (2 mm) proximal to Lava G to a mode smaller than 2 φ (250 µm) 4 km away (Chpt. 4.). Subunit 3 is inferred to have been formed by ash venting that occurred synchronously with the effusion of Lava G, with particle dispersal occurring in a convective plume driven by the heat of the underlying volcanism.

Subunit 4 (S4) directly overlies S3 and consists of microcrystalline grains. The grain size mode of this subunit is poorly constrained but is likely in the range of approximate -1 to 2 φ (2 mm to 250 µm). It has been deposited in two distinct areas; the first from the southwest caldera rim trending over the caldera floor to the northeast, while the other forms a roughly circular shape around Dome OP on the southeast caldera rim. The deposit surrounding Dome OP is inferred to be the ash component of the lava’s talus slope, resulting from brecciation and quench fragmentation of the lava dome (Chpt. 3.). The deposit that crosses the caldera floor is inferred to be the result of either brecciation and quench fragmentation of the domes followed by dispersal in a plume or a mass wasting deposit resulting from collapse of the caldera wall.

7.3. Methods

This chapter draws on synchrotron radiation-Fourier transform infrared microspectroscopy (SR-FTIR) results in support of its interpretations. The methods by which these results were collected are outlined in Chapter 2. Methods.

7.4. Results

Synchrotron radiation-Fourier transform infrared microspectroscopy (SR-FTIR) was carried out on wafers created from a range of 2 φ (250 µm) ash particles from S1, S2, and S3 of the AL Unit, formed during the
2012 Havre eruption. Doubly polished wafers 15-40 µm thick were prepared from glassy particles. Wafers of GV curvi-planar and fluidal, along with ETV particles were created from representative samples taken from around the caldera (Table 7.1). Few of the tiny wafers made and analysed to date have yielded reliable results. The high vesicularity of many of the examined grains meant that finding an area of vesicle and microlite free glass from which to take a measurement was challenging. In many cases likely contained/intercepted vesicle walls or microlites which scattered the beam altering results. Results presented here rest on high-quality spectra, which were obtained for only four grains of the 19 analysed, two from subunits 1 or 2 (grains 13, and 26) and three from subunit 3 (grain 15, 16(a), and 16(b) (Fig. 7.1.).

The SR-FTIR analysis of glassy particles yields $H_2O_{\text{total}}$ groundmass contents of 0.64 wt% to 1.05 wt% (Table 7.1) (Fig. 7.2.). Glassy Vesicular particles (grains 13 and 26) have $H_2O_{\text{total}}$ contents of ~1.03 and 0.95 to 1.05 wt% respectively (Fig. 7.2.). In contrast the ETV particles (grains 16(a), 16(b), and 15) show lower $H_2O_{\text{total}}$ content of 0.64 to 0.84 wt% (Table. 7.1.) (Fig. 7.2.). Dissolved CO$_2$ was not detected and is inferred to be in concentrations below the detection limit for the wafer thicknesses, which are approximately 40-60 ppm. Water speciation results were compared with the experimentally determined temperature-dependent water speciation model for a haplogranite composition from Nowak and Behrens, (2001). This shows what concentrations of OH and $H_2O_{\text{molecular}}$ that are expected for a given $H_2O_{\text{total}}$ concentration at a given temperature (Fig. 7.2.). The Nowak and Behrens, (2001) speciation model predicts that the ratio of OH/$H_2O_{\text{molecular}}$ is dependent on temperature and the $H_2O_{\text{total}}$. At higher temperatures and lower $H_2O_{\text{total}}$ the OH/ $H_2O_{\text{molecular}}$ ratio will increase, while decrease in temperature along with increasing $H_2O_{\text{total}}$ will result in lower and even decimal ratios. Using this speciation model, a temperature of apparent equilibrium ($T_{ae}$) for the measured water speciation can be determined, which is approximately equal to the glass transition/quenching temperature ($T_g$) of that particle (Zhang, Jenkins, et al. 1997; Zhang and Ni 2010). The $T_{ae}$ values vary, however, ETV particles mostly fall between
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(previous page) Fig. 7.1. Two representative SR-FTIR spectra (from grain 26 (left) and grain 16(a) (right)) on which the 3550 cm$^{-1}$ H$_2$O$_{total}$ and 1630 cm$^{-1}$ H$_2$O$_{mol}$ absorbance peaks are marked. Base lines for the peaks are also marked.

540 and 660 °C (Table. 7.1.). $T_{ae}$ values for grain 26 (GV blocky) are the highest of all particles examined, between 710 and 760 °C, while grain 13 (GV fluidal) returned the lowest measured $T_{ae}$ at 500 °C (Table. 7.1.).

The glass transition is the point at which the system becomes viscously ‘locked in’, following which the rate of volatile diffusion slows by several orders of magnitude. This is typically inferred to occur at a viscosity of between $10^{12}$ - $10^{13}$ Pa s (Gottsmann and Dingwell 2001; Gottsmann and Dingwell 2002; Giordano et al. 2008; van Otterloo et al. 2015). It is taken as the point at which degassing ceases. Rather than being a defined point however, the glass transition occurs over a temperature range of more than 100 °C (Gottsmann and Dingwell 2001; Gottsmann and Dingwell 2002). $T_g$ is associated with sharp changes in the relationship of the specific volume and coefficient of expansion to temperature (Gottsmann and Dingwell 2001; Gottsmann and Dingwell 2002; Giordano et al. 2008; van Otterloo et al. 2015). The temperature at which the glass transition is crossed is a function of composition (including volatile content), and cooling rate. Cooling rate variation in melt of similar composition has shown a range in $T_g$ of approximately 80 K, for cooling rates of between 0.000017-0.105 (K/s) (Gottsmann and Dingwell 2001; Gottsmann and Dingwell 2002), where high cooling rates lead to high $T_g$. In both cases eruption temperatures were hotter than the $T_g$.

The inferred Havre eruption temperature is ~850 °C (±20 °C) (Manga et al submitted 2018), measured using clinopyroxene orthopyroxene Fe-Mg exchange in measured clinopyroxene and orthopyroxene compositions assuming equilibrium conditions (Putirka 2008).
<table>
<thead>
<tr>
<th>Sample</th>
<th>Grain number</th>
<th>Grain Morphology</th>
<th>Thickness (µm)</th>
<th>1630 cm^{-1}</th>
<th>3500 cm^{-1}</th>
<th>Measured H₂O₆ wt% (1630)</th>
<th>Measured OH wt%</th>
<th>Measured H₂O₆ (3500) wt%</th>
<th>Tₑₑ in °C</th>
<th>reconstructed unhydrated H₂O₆ wt% (T=850 °C)</th>
<th>Pressure from unhydrated H₂O₆ (MPa) (T=850 °C)</th>
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Table 7.1. Showing the result from the FTIR analyses. A density of 2350 g/L was used when calculating the water species content. When calculating $T_{ae} W = 32.47$. Quench pressure was then determined using Volatilecalc (Newman and Lowenstern 2002).

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The derived $T_{ae}$ for the Havre ash (500-760 °C) differs significantly from the petrologically inferred Havre eruption temperature (850 °C). The $T_{ae}$ values for grain 26, from S1/S2 fall within the error and potential cooling rate range for the $T_g$ (up to 100 °C, based on the error (±20 °C) and the $T_g$ range observed by Gottsmann and Dingwell (2001; 2002) (80 °C)) of the eruption temperature.

Experiments have shown however that at eruption temperatures the interconversion of $H_2O$ species happens almost instantly. In Zhang et al. (2000) a rhyolitic melt with a $H_2O_{total}$ of 2.5 wt% at 1000 K was cooled at 70 K/s, however the produced $T_{ae}$ was approximately 800 K. During cooling the decrease in
temperature prior to the $T_g$ drove the $\text{OH}/\text{H}_2\text{O}_{\text{molecular}}$ ratio down due to the temperature dependency of this ratio (Nowak and Behrens 2001), and the extremely fast interconversion rate at magmatic temperatures. For the $T_{ae}$ to equal eruption temperature requires an extreme cooling rate as even a small amount of time above the $T_g$ is enough for interconversion of OH to $\text{H}_2\text{O}_{\text{molecular}}$ to produce a lower than expected $T_{ae}$, this is referred to as the quench effect (Zhang and Ni 2010).

At extremely high quenching rates the ‘quench effect’ might be negated as there is no time for the interconversion of water species. Water has a higher specific heat capacity, thermal conductivity, and thermal diffusivity than air, as such cooling rates in subaqueous eruption are expected to be much higher. Calorimetric analysis of *limu o Pele* grains from Loihi Seamount, Hawaii and Axial Seamount, Juan de Fuca Ridge have shown quenching rates of $10^{5.31}$ K s$^{-1}$ and $10^{4.3}$ to $10^{6}$ K s$^{-1}$ (1mm grains) and $10^{3.9}$ to $10^{5.1}$ K s$^{-1}$ respectively (Potuzak et al. 2008; Helo et al. 2013). These results represent different end member compositions and eruption styles but point toward extremely high quench rates for pyroclastic subaqueous activity.

Due to the high rate of cooling expected for ash emplaced directly into seawater (Potuzak et al. 2008; Helo et al. 2013), if the Havre ash erupted directly into the seawater its measured $T_{ae}$ should be close to the eruption temperature (Gottsmann and Dingwell 2001; Gottsmann and Dingwell 2002). However, in the samples successfully analysed here, there is therefore significant discrepancy between the $T_g$ (=Tae) of the glass and the petrologically inferred eruption temperature. The implication of the quench effect however is that the divergence of $T_{ae}$ from the eruption temperature at Havre could be explained by almost instantaneous interconversion of species at magmatic $T$ above the $T_g$. This is especially true for grain 26 showing $T_{ae}$ of between 711 and 755 °C. Grain 26 is therefore considered to be the result of the quench effect, where quenching was not rapid enough to prevent interconversion of species prior to $T_g$ between approximately 711 and 755 °C. The remaining particles however show much lower $T_{ae}$ values and thus would require significant temperature variations for their result to be consistent with primary
conditions. An alternative explanation is that the result from the clinopyroxene orthopyroxene Fe-Mg exchange barometer may reflect deeper parts of the magmatic system as opposed to the temperature of the magma at the vent owing to the rates of diffusion in the pyroxenes, which mean that the system does not instantaneously equilibrate as it ascends (Putirka 2008). To produce the observed $T_{ae}$ results however would require hundreds of degrees of variation in magma temperature, which seems extremely unlikely.

The discrepancy between $T_{ae}$ and the eruption temperature is instead inferred to be mostly a signature of glass hydration, apart from grain 26 where it is accounted for by the quench effect (Anovitz et al. 2008; Yokoyama et al. 2008; Zhang and Ni 2010; McIntosh et al. 2014). Hydration describes the process by which molecular water is added to a glass’s structure at temperatures below the $T_g$ (Zhang et al. 1991; Anovitz et al. 2006; Anovitz et al. 2008; Yokoyama et al. 2008; Zhang and Ni 2010; McIntosh et al. 2014). The addition of molecular water increases the $H_2O_{total}$ value but does not affect the $OH$ content (McIntosh et al. 2017).

Assuming equilibrium saturation and a temperature the measured $OH$ content can be used to estimate the original $H_2O_{total}$ content from a speciation model of the assumed value of $T_g$ (McIntosh et al. 2014, McIntosh et al. 2017). For instantaneous quench $T_g$ is equal to the eruption temperature of 850 °C, and the measured $OH$ content would be equivalent to original pre-hydration $H_2O_{total}$ groundmass content between 0.54 wt% and 1.00 wt% (Table. 7.1.). If $T_g$ were lower, for example due to slow cooling, then the same measured $OH$ content would correspond to higher original $H_2O_{total}$ content due to the dependency of the $OH/H_2O_{molecular}$ ratio on temperature and the extremely high rate of exchange at eruption temperatures (Zhang and Ni 2010). For example, a $T_g$ of 600 °C would give $H_2O_{total}$ of 0.65 and 1.05 wt%.
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VolatileCalc (Newman and Lowenstern 2002) was then used to determine the quench pressure of the Havre glass from these reconstructed \( \text{H}_2\text{O}_{\text{total}} \) values, using a temperature of 850 °C. Calculated quench pressures are between 2.5 and 8.5 MPa, equivalent to water depths of 250 and 850 m. For a lower \( T_g \) of 600 °C calculated quench pressures are between 2.1 and 5.3 MPa, equivalent to water depths of 210 and 530 m (Fig. 7.3.).

Elongate Tube-Vesicle particles from S3 form a distinct cluster, with reconstructed water contents of between 0.54 and 0.78 wt%, implying quench pressures of 2.5 to 5.2 MPa assuming \( T_g = 850 \) °C, equivalent to 250 to 520 mbsl (Table. 7.1.). Where \( T_g = 600 \) °C water contents of between 0.58 to 0.78 wt%, implying quench pressures of 1.6 to 3.4 MPa, equivalent to 160 to 340 mbsl (Table. 7.1.). The two particles from S1 and S2 differ from the S3 grains (Table. 7.1.). The blocky particle, grain 26, has a reconstructed water content of 0.89 to 1 wt% where \( T_g = 850 \) °C, corresponding to quench pressures between 6.8 and 8.5 MPa (680-850 mbsl; Table. 7.1.). Using a quench effect implied \( T_g \) of 700 °C grain 26 shows quench pressures of 5.7 to 6.9 MPa equivalent to 570 to 690 mbsl. The fluidal grain from S1 or S2, 13, however shows a reconstructed water content of 0.77 wt% where \( T_g = 850 \) °C, implying a quench depth of 510 mbsl (5.1 MPa; Table. 7.1.). Using a \( T_g \) of 600 °C grain 13 has a reconstructed water content of 0.89 wt%, with a quench pressure of 3.8 MPa equivalent to 380 mbsl.

7.5. Discussion

The small size and high surface to volume ratios of ash particles means that on contact with cold seawater they would have quenched more or less instantaneously (Wohletz 1986; Potuzak et al. 2008; van Otterloo et al. 2015). The calculated quench depths are therefore taken to reflect the point at which the particles first encountered liquid water. The initial SR-FTIR analysis undertaken produced useable results for only four particles, so the discussion presented here is tentative; work with the aim of
Fig. 7.3. Schematic diagram showing the physical constraints placed on the eruption model inferred from the presence of fluidal ash by the FTIR derived quench depths.
obtaining more information about quench depths for Havre ash particles is ongoing. The implications of those FTIR results however, provide crucial constraints on eruption processes. The inferred quench depths are used here to establish the nature of volcanism that formed particles in S1, S2 and S3.

7.5.1. *Inferences on the explosive pyroclast-forming phase of the Havre eruption*

Below I examine inferences from SR-FTIR results and place them in the context of what has been concluded in previous chapters about the 'explosive' pyroclast-forming phase.

Subunits 1 and 2 were produced during what I infer was the most energetic period of the 2012 Havre eruption, referred to here as the phase 2b (Chpt. 3). The depth of the vent for the pyroclast-forming phase of the eruption is inferred to be ~900 mbsl (based on 2002 bathymetry (Wright et al. 2006)), equivalent to hydrostatic pressure of 9 MPa (Wright et al. 2006; Carey et al. 2014; Carey et al. 2018). The measured saturation pressure for grains from S1 and S2 suggest quenching at depths of between 850 and 510 mbsl (Fig. 7.3.). This implies that grains quenched in the water column at heights of 50 to 390 m above the vent site (Fig. 7.3.). For the particles to reach these heights before quenching, they would need to be isolated from enclosing seawater, because they would quench rapidly upon contact (Fig. 7.3.). Those values at the deeper end of the quench range however, are within error of the eruptive vent depth, and thus may have quenched on exiting the vent (Fig. 7.3.).

"Veiling" of large pyroclasts by steam has been described for emerging buoyant pumice, and inferred to take place during subaqueous volcanism (Reynolds and Best 1957; Allen et al. 2008; Fauria et al. 2017). Veiling could allow large clasts to be transported higher into the water column prior to being further fragmented and the new clasts quenched. Fluidal ash at Havre is, however, inferred to be primary (Chpt. 5) (Fig. 7.3.). Formation of fluidal ash by cracking or abrasion from a larger steam-charged clast therefore cannot explain the quench depth of the analysed fluidal particle (Grain 13). A small amount
might travel upward virtually in contact with large clasts, veiled by their steam during initial upward transport, but this seems an insufficient explanation for the proportion of fluidal particles identified (Chpt. 5).

An alternative mechanism to isolate ash from the enclosing water column in a sustained subaqueous gas-supported eruption jet (Head and Wilson 2003; Allen and McPhie 2009). A subaqueous eruption jet requires both a large heat source to vaporise incoming ambient seawater, and a high mass flux gas-driven jet to physically exclude water; both imply explosive volcanism (Head and Wilson 2003; Allen and McPhie 2009) (Fig. 7.3). Mixing on a jets margins produces a complex zone of exchange that leads to a decline in the jets heat energy with height, causing it to rapidly pinch out (Head and Wilson 2003; Allen and McPhie 2009). The height to which the jet penetrates ambient seawater water is a function of the jets mass and heat flux, along with the width of the vent (Head and Wilson 2003). Particles in the central part of the jet will be shielded from contact with the water until they are pulled out of the jet by mixing at the edges, or when the jet thins to zero (Fig. 7.3). Based on the saturation pressures of grain 13, a jet up to 390 m high is suggested to have formed during phase 2b of the 2012 Havre eruption (Fig. 7.3).

The lack of a subaerial pyroclastic plume shows the eruption did not breach the sea surface. The inference of a gas supported eruption jet is consistent with conclusions based on the presence of fluidal grains in S1 and S2 (Chpt. 5), which indicate that ash grains deformed viscously following fragmentation, and prior to quenching. A subaqueous eruption jet would drive an overlying convective plume that could transport fine particles high into the water column, which would then settle out. Such a mechanism is consistent with the inferred fallout origin of S1. Condensation of the jet due to fluctuations in eruptive power could lead to density current formation, the inferred source of S2. Ash reaching surface waters would provide a compelling explanation for the plume of discoloured water seen at the sea surface in MODIS imagery (Jutzeler et al. 2014).
Production of such an intense gas-supported eruption jet driving the ash-forming 'explosive' pyroclast-forming phase(s) of the Havre eruption is surprising. Work by Manga et al., (2018) suggests that the mass flux during the production of the GP Unit, and the pumice raft was insufficient to induce fragmentation of the magma by vesicle over pressure or critical strain rate. More information is needed to reconcile the implications of ash grains versus those of the large pumice clasts. A key uncertainty, resulting from the absence of raft pumice on the seafloor (Carey et al. 2018), is the relative timing of when S1/S2, the pumice raft, and the seafloor giant pumice deposit were formed.

There is much uncertainty when attempting to compare the ash from S1/S2 with lapilli and block sized equivalents due to large differences in particle settling rates. The water content range in S1/S2 does not overlap with any other product from the Havre eruption. Pumice from the GP Unit, and the ALB Unit have low OH concentrations equivalent to hydrostatic quench depths of 0 to 100 mbsl (Mitchell et al. 2018). Pumice from the AL Unit displays quench depths of 0 – 400 mbsl, the pumice raft has shown quench pressures of between 100 – 350 mbsl (Mitchell pers. comm.). Fourier Transform InfraRed spectroscopy results from lapilli and blocks from the GP Unit, pumice raft, and the ALB Unit all preserve evidence of hydration (Mitchell et al. 2018). The differences in the measured dissolved volatile contents of the S1/S2 ash vs lapilli and blocks may reflect differences in particle-cooling history controlled by particle size. The lower pre-hydration water contents and shallower quench depths of the pumice and lapilli could be related to the larger initial internal heat of the larger clasts, which allowed them to remain above the $T_g$ and continue degassing during transport high into the water column (Allen et al. 2008; Fauria et al. 2017). By contrast primary ash is quenched instantly on contact with water, recording the initial depth of direct water contact whether in the conduit, at the vent, or as particles exit a gas-thrust jet.

With samples having been collected only a few years after the eruption the degree of hydration may also reflect differences in particle cooling histories (Anovitz et al. 2006; Anovitz et al. 2008; Yokoyama et
Chapter 7

al. 2008; McIntosh et al. 2014; Bindeman and Lowenstern 2016). The diffusion of water into a glass is heavily temperature dependent, with higher temperatures leading to more rapid water diffusion (Zhang et al. 1991; Bindeman and Lowenstern 2016). We can assume S1/S2, GP Unit, and the ALB Unit were quenched around the same time and have since been in the ambient environment for approximately the same amount of time (±days/weeks) (Carey et al. 2018). The wide variation in amount of hydration thus cannot reflect large differences in glass age. Instead, the sensitivity of hydration rate to temperature (Anovitz et al. 2006; Anovitz et al. 2008; Yokoyama et al. 2008; Zhang and Ni 2010; McIntosh et al. 2014; Bindeman and Lowenstern 2016) suggests that the main variation in hydration may have occurred rapidly as part of eruption processes and during initial cooling of particles (Mitchell pers. comm.). In the case of the blocks and lapilli hydration may have occurred as water infiltrated into the pumice and was heated and even vaporised, allowing rapid high-temperature diffusion of water. Using equations laid out in Carslaw and Jaeger (1959), Kano et al. (1996) calculated that cooling of coarse lapilli (diameter of 64mm) to ambient temperature from 850 °C would take approximately 66 minutes. For a Giant Pumice clast greater than 1 m in diameter we can assume cooling may be on the time scale of hours. At eruption temperature (>800 °C) volatile equilibration can occur in less than a second (Zhang and Ni 2010). For a water content of 1 wt% in the 300-400 °C range hydration rims of approximately 1 µm can form over approximately 3-4 minutes (Zhang and Ni 2010; McIntosh et al. 2014). Clast cooling over hours therefore allows significant time for hydration of the groundmass glass operating over seconds at high temperatures to occur. Once the particle has cooled to ambient temperature however the hydration rate slows significantly, and hydrated rims of ~1 µm might take 30 years to form (Anovitz et al. 2006; McIntosh et al. 2014).

Grain 13 and 26 are both from S1/S2 in the AL unit, however they show strongly contrasting hydration signals. Grain 13 shows a quench depth of 510 mbsl, and a high degree of hydration, while grain 26 displays a low degree of hydration and deep quenching 850-680 mbsl. The deep quenching of grain 26
suggests it may have encountered water at the vent depth, where it was rapidly cooled through the glass transition. The fluidal morphology of grain 13 suggests it is the result of primary fragmentation and underwent initial dispersal in a water free environment (Chpt. 5). The shallow quenching of grain 13 has been inferred to be the result of transport in a gas supported jet. Transport in a steam rich hot environment may have led to brief but intense hydration.

In closing this section, it is worth stating there remains considerable uncertainty about the origin of fluidal grains in S1 and S2. One line of interpretation points towards large local increases in the magma temperature, possibly due to shear heating (Rosi et al. 2004; Mastin 2005; Hess et al. 2008) or lightning (Genareau et al. 2015; Wadsworth et al. 2017). Volatile solubility is partially controlled by system by temperature with higher temperatures reducing solubility (Lavallée et al. 2015). If the magma temperature of fluidal grains was substantially higher the thermally-induced increase in degassing could reduce the volatile content for magma at the same temperature producing an anomalously shallow quench depth. More data are needed, and this is an area of ongoing investigation.

7.5.2. Constraints on inferred links between the eruption of Lava G and Subunit 3 ash

Lava G, the inferred source of S3 ash, is located at approximately 950 mbsl. Fourier Transform InfraRed spectroscopy results on particles from S3 indicate they quenched at depths between 520 and 250 mbsl. The groundmass volatile contents of ash grains from S3 can be compared with lapilli sized equivalents from the same deposit (Mitchell et al. 2018). The OH content of the lapilli and the reconstructed ash results from grains 16 (a), 16 (b), and 15 form S3 overlap, appearing to confirm that these are the products of the same deposit.

All the analysed products of S3 display shallow quench depths, however its inferred eruption mechanism is effusive, with syn-extrusion ash venting (Schipper et al. 2013; Cole et al. 2014; Black et al. 2016) and
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brecciation of the Lava G carapace producing particles that were then dispersed in thermal plumes of hot water (Chpt. 4). Although the thermal plume from SEAV has been shown to reach a height of up to 300 mbsl (Fig. 4. 12.), particles would be transported in liquid water and thus have quenched before entering the plume. The SR-FTIR results are therefore difficult to reconcile with the inferred eruption mechanisms of Lava G. I speculate that the increased degree of degassing may have resulted from increased temperatures of magma that produced the S3 ash grains. Volatile solubility is inversely proportional to magma temperature, so a temperature increase will lead to a lower solubility and thus a lower equilibrium water content (Ghiorso and Sack 1995; Lavallée et al. 2015). When considering the ribbed and fluidal particles of S3, there was initial consideration that high temperatures were involved in their formation, resulting from strong heating in localised shears associated with the ash-venting process (Rosi et al. 2004; Mastin 2005; Hess et al. 2008). Presently, there is no positive microtextural evidence to support such shear localisation, as such further work is required to assess the viability of this hypothesis.

7.6. Conclusion

The SR-FTIR analysis of ash from S1, S2, and S3 produced during the 2012 Havre eruption provide an initial dataset, with further work ongoing. The interpretations presented here are thus tentative, and hint at more-complex scenarios. Some of the basic interpretations of this initial dataset however, have important implications for inferring eruption dynamics. The equilibrium quench pressures for particles generated during primary fragmentation show strong evidence for hundreds of meters of upward transport of grains in a water-free environment during the pyroclastic phase. I suggest that quenching of particles high in the water column is the result of explosive activity that generated an intense gas-thrust jet during the Havre eruption. At the height of the eruption the gas-thrust jet, with a peak height of at
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least 390 m above the vent at 510 mbsl, is inferred to have resulted in effects at the sea surface. These surface effects link S1 and S2 with the phase of eruption that produced the pumice raft. An alternative mechanism is also considered, however, by which apparent shallow-quenching signatures could be the result of heating-induced volatile exsolution. Such thermally induced exsolution is also postulated to account for the shallow quench signatures of S3, which are otherwise inconsistent with its inferred origin during an effusive eruptive phase.
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Discussion and conclusions
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8.1. Reprise and discussion-point summary

The 2012 eruption of Havre Volcano was the largest deep subaqueous silicic eruption ever recorded (Carey et al. 2014; Carey et al. 2018), and the first. Pre-eruption bathymetric mapping, tow-camera imaging, and dredge sampling of the Havre edifice (Wright et al. 2006), along with the detailed follow up study of the seafloor eruptive products (Carey et al. 2018) make this eruption a natural laboratory for the examination of how magma interacts with water during eruptions in deep subaqueous settings. A 2015 cruise used autonomous underwater vehicle (AUV) Sentry and remotely operated vehicle (ROV) Jason in tandem to map, image, and sample the seafloor deposits of the 2012 Havre eruption (Carey et al. 2018). The data set collected during the 2015 cruise has spawned a range of projects that aim to examine in detail various parts of the eruptive deposits and processes, e.g. lavas, pumice, edifice chemistry, ash, ingestion of water into pumice, modeling of subaqueous eruption plumes, and more (Carey et al. 2018).

The results presented in this thesis focus on analysis of the fine lapilli and ash component (-3 \( \phi \) and smaller particles) of the 2012 Havre eruption. The discussion here integrates results from ash analysis with results presented to date (January 2018) by other groups undertaking analyses of different components. By doing so I aim to address the following questions outlined initially in section 1.4.:

- What does detailed stratigraphy of the ash deposits comprising the Ash and Lapilli Unit (AL Unit) reveal about relative timing of deposition (and production) of particles from the 2012 Havre eruption?

- What were the eruption, fragmentation, and depositional mechanisms by each ash subunit in the Ash and Lapilli (AL) Unit was produced?

- What mechanisms formed the population of fluidal grains observed in two subunits at Havre?

- What is the form of an eruption model that explains the eruptive and depositional processes active during the 2012 Havre eruption?
- What was(were) the dispersal mechanism(s) that added a population of foreign grains to Havre ash deposits? Where were their eruptive sources?

8.2. Stratigraphy and relative timing

The 2012 eruption of Havre volcano was a complex multiphase eruptive episode (Carey et al. 2018). Through detailed analysis of the Ash and Lapilli (AL) Unit I identified four distinct subunits of ash that have been dispersed widely over the study area. Changes in the depositional characteristics of each subunit reflect both differences in depositional processes, and changes in the source and location of activity through the eruption. Additionally, changes in particle microtextures and morphologies show variation in the eruptive processes during the 2012 Havre eruption. By examining the subunit stratigraphy along with the subunits' relationship to other products at Havre I was able to infer the relative timing of deposition and the different types of eruptive activity (Fig. 8.1.). This information supports an assessment of changes in the eruptive style and locations of active vents through most, if not all, of the 2012 eruptive episode at Havre volcano. Three main phases can be identified from observations and sampling of Havre eruption products, defined broadly as either effusive or explosive (pyroclast-forming) (Fig. 8.1.). Before going further, it should also be noted that key stratigraphic uncertainties remain. First, the basal contact of the eruption's deposits with pre-2012 eruption stratigraphy was neither observed nor sampled. It is not known if any ash deposits were formed prior to the second phase, described below. Second, the temporal relationship of the pumice raft with the seafloor products of the Havre eruption can only be inferred. Within standard deviation, no clear distinctions have been identified in chemical composition or microtextural features that could link the raft pumice with, or distinguish it from, any of the seafloor pumice deposits. Importantly however, most ash is texturally distinct from all the pumice, and hence is not an abrasion product of the raft.

Phase 1: An initial effusive phase began at some point between 2002 and July 2012 (Wright et al. 2006; Carey et al. 2014; Carey et al. 2018). During phase 1 five lava flows were emplaced on the
Fig. 8.1. The upper panel shows the outline of the lavas (green) produced during the 2012 eruption at Havre plotted on the 1 m resolution AUV Sentry map of the caldera. In white are shown the outlines of the ALB (dashed), and GP (solid) Units. In addition, stars denote the inferred eruption locations for each of the four subunits that make up the AL Unit; S1 (red) and S2 (brown) from the vent now filled by dome OP, S3 (purple) from Lava G, and S4 (yellow) from the caldera wall below Lavas G to I. The lower panel shows the detailed inferred stratigraphy of the 2012 Havre eruption. Stars denote the timing and cross-sectional location of the events that produced each of the four subunits that make up the AL Unit, using the same colours as the upper panel. Phase 1 has no observed stratigraphic
deposits associated with it. The seafloor component of phase 2 began with the deposition of the GP Unit sub-phase 2. Subunit 1, the ALB Unit, and S2 were then deposited after a time gap in a single event, sub-phase 2. After a time gap following the deposition of S2 the phase 3 of the Havre eruption began with the onset of effusion at Lava G, and likely the other southern caldera rim lavas and domes. During the effusion of Lava G S3 was generated. The formation of S3 and Lava G was terminated by a collapse in the caldera wall bellows Lava G to I, resulting in the formation of S4. In the west S4 formed as a talus slope produced during the effusive of Dome OP. The Dome OP component of S4 has no stratigraphic overlap with S3 or S4, so the relative timing of its formation is unknown. The deposition of S2 continued through phase 3. In addition, the steady input of foreign grains to the Havre caldera has resulted in them been dispersed through the 2012 seafloor clastic deposits.

southwest caldera wall, and one lava on the southern caldera rim (Carey et al. 2018). Little is known regarding the relative timings of this initial stage (Fig. 8.1.). No clastic deposits have been identified associated with this stage, nor were the bases of the lavas observed or sampled.

Phase 2 (with sub-phases): The next phase of eruption is inferred to have been multiphasal with evidence of explosive volcanism, producing seafloor deposits comprising the Giant Pumice (GP) Unit, S1, the Ash, Lapilli, and Block (ALB) Unit, and S2 (Fig. 8.1.). The microtextural similarity of these seafloor deposits to the sea-surface pumice raft produced on 18th – 19th July 2012 suggest the raft was also generated during phase 2 (Carey et al. 2018; Manga et al. 2018). The production of the pumice raft during phase 2 provides the only approximate indication of the timing and duration of explosive seafloor activity in the Havre eruption.

The GP Unit is the lowest seafloor deposit observed from phase 2 (Fig. 8.1.). The GP Unit’s basal contact could not be sampled, and was not observed, so it is not known whether the GP Unit represents the start of phase 2, or whether it is underlain by an earlier pyroclastic deposit. The GP Unit is directly overlain by S1. Near Dome OP the ALB Unit then overlies S1. Subunit 2 caps the seafloor phase 2 clastic succession, overlying the ALB Unit close to Dome OP, and overlying S1 where the ALB Unit is not present (Fig. 8.1.). The pumice raft has no direct stratigraphic link to any of the
The seafloor clastic units produced during phase 2 of the 2012 Havre eruption are inferred to have been deposited by a range of mechanisms (Carey et al., 2018; Chapter 3, 5). Both S1 and the GP Unit are inferred to comprise clasts deposited from suspension. Ash from S2 and the ALB Unit are both inferred to have been emplaced by density currents. From deposit characteristics and the stratigraphic relationships between them I have defined two distinct depositional events during phase 2 of the Havre eruption. The ash from S1 directly overlies the GP Unit. The clasts that comprise these two deposits show contrasting settling velocities (and for the giant pumice, time-varying ones (Allen et al. 2008; White et al. 2009; Fauria et al. 2017)). Assuming there is no unseen S1-type ash lying beneath the GP Unit, the contrast in settling velocities mean that the GP Unit and S1 must have been generated during distinct events, with an intervening period sufficient to allow the settling of large pumice clasts to form the GP Unit prior to the deposition of S1. In contrast the contact of S1 with the ALB Unit appears more complex and is not well documented. The contact between S1 and S2 appears gradational, suggesting the two formed during a single ongoing period of deposition as opposed to two events. The gradational contact between S1 and S2, and the inferred genetic link between S2 and ALB Unit suggest that S1, S2, and the ALB Unit were deposited continuously during a single event. In contrast the GP Unit is inferred to have been erupted in a different, distinct event, prior to the deposition of S1. The inference that the ALB Unit is unrelated to the GP Unit is supported by the dispersal mechanism of large gas charged pumice blocks. The large size of the GP means their transport and dispersal was largely controlled by their own buoyancy resulting from internal gaseous volatiles (Allen et al. 2008; Fauria et al. 2017). Buoyancy-driven dispersal (followed by water-saturation-controlled settling) of large particles would not result in the formation of density currents, the inferred depositional mechanism of the ALB Unit. The apparent depositional break between the seafloor deposits of phase 2 indicates two sub-phases to the eruption with the GP Unit deposited during sub-phase 2a and S1, S2, and the ALB Unit deposited during sub-phase 2b.
In sub-phase 2a of the 2012 Havre eruption the GP Unit was deposited by particle settling out of suspension. After a time break of at least a couple of hours sub-phase 2a began with the deposition of S1 by particle settling, followed by the deposition of the ALB Unit and S2 from density currents with no apparent time break.

The temporal relationship of the pumice raft to the seafloor stratigraphy cannot be directly deduced from deposits characteristics or observations. Sub-phase 2a, during which S1, S2, and the ALB Unit were generated, was highly energetic capable of vaporising or excluding ambient water for at least 390 meters above the erupting vent. An eruption generating a jet of this magnitude would have produced an overlying convective cell that could have transported ash to the sea surface. This would have resulted in the generation of a plume of discolored water, commonly observed above active subaqueous volcanoes and inferred to result from suspended fine particles (Fiske et al. 1998; Kano 2003; Chadwick et al. 2008; Watts et al. 2012). A large area of discolored water was associated with the Havre pumice raft (M. Jutzeler et al. 2014; Carey et al. 2018). All pumice clasts generated in the eruption would have been initially gas-filled and buoyant in the water column and ascended toward the sea surface. I infer that the high jet helped most of pumice clasts to reach the sea surface where entrainment of air allowed them to remain buoyant and form the pumice raft (Allen et al. 2008; Fauria et al. 2017). Pumice clasts that became water-logged before reaching the sea surface would have settled rapidly and become entrained in ash rich density currents coming off the main column and emplaced as part of the ALB Unit, on top of both the GP Unit and S1.

**Phase 3 (with sub-phases):** Following phase 2, the 2012 Havre eruption entered another effusive phase in which nine lavas were produced, dominantly on the southern caldera rim, along with the deposition of two clastic subunits; subunit 3 (S3) and subunit 4 (S4) (Fig. 8.1.). Lavas produced during phase 3 were fully emplaced by October 2012 (Carey et al. 2014; Carey et al. 2018). The two ash subunits are linked with different sources. Ash of S3 appears to have been deposited during extrusion of Lava G. Uncertainty remains about whether S4 formed during later lava extrusion, or due to collapse of the caldera wall below Lava G to I. The fact that S3 does not directly overlie S1,
however, and the slow settling rate of S2 particles, indicate a time gap between the end of phase 2 and the onset of fragmentation during Lava G extrusion (Fig. 8.1.). The morphology of the scarp that cuts Lava G vs Lavas H and I suggest that H and I were been actively erupted after the eruption of Lava G ended. This suggests some temporal stratification in the effusive eruptions of phase 3.

8.3. Summary: Formation of subunits of the Ash and Lapilli Unit

8.3.1. Subunit 1

Subunit 1 (S1) is widely distributed across all seafloor sites sampled on Havre, with no detected thinning or fining trends. It is composed of glassy vesicular grains, dominantly blocky but also with angular and fluidal grains. Individual sample grain-size modes for particles of this subunit range from -1 to 1 φ (2 mm to 500 µm). The dominantly blocky morphology of the S1 particles is suggestive of formation by magma-water interaction (MWI), and the particle population shows some similarities with that of the subaqueous Shinjima Pumice (Kano et al. 1996).

The presence of fluidal grains and their lack in the GP or raft pumice suggests that S1 comprises primary ash produced by strong fragmentation during the Havre eruption (Fig. 8.2.). Synchrotron radiation-Fourier transform infrared microspectroscopy (SR-FTIR) results imply that some ash grains did not quench until they were approximately 390 m above the vent, while the presence of fluidal grains also suggest particle were veiled from the ambient water for some period after fragmentation.

I have therefore inferred that a gas supported submarine eruption jet was sustained during S1 production (Fig. 8.2.). Ash of S1 was deposited by settling from high in the water column following upward gas-jet transport (Fig. 8.2.).
Fig. 8.2. Summary figure showing schematic models of the inferred eruption(s) for each of the four subunits that make up the AL Unit. The star relates each subunit back to Fig. 8.1, where it can be temporally and spatially located. Subunit 1 and 2 were both formed as part of the same eruption during phase 2b. Explosive volcanism produced a gas jet at least 390 m high. Fragmentation was driven both by magmatic and hydromagmatic processes. The gas jet allowed some particles that...
were at elevated temperatures to be viscously reshaped as they were veiled from direct contact with the ambient water and quenching. Pumice blocks generated during this eruption were initially dispersed in the gas jet following which they ascended under their own buoyancy to form the pumice raft. The gas jet drove a convecting plume above it which was able to transport ash high into the water column forming a plume of discoloured water to the sea surface, this was observed along with the pumice raft (Carey et al. 2014; M. Jutzeler et al. 2014; Carey et al. 2018). Later in the eruption variations in mass and/or heat flux lead to condensation of parts of the jet. The rapid densification caused by volatile quenching formed density currents. Any water-logged pumice blocks that were caught up in the jet ‘collapse’ were rapidly deposited as the flow slowed and spread forming the ALB Unit. Fine particles from these density currents were lofted into the water column forming a ‘cloud’ from settling particle settling occurred over weeks to months. Subunit 3 was generated during the effusion of Lava G. Shear localisation on the conduit edge lead to the development of permeable outgassing channels. Isothermal viscous shaping of melt fibres occurred in the outgassing channels as a result of the continuous flow of gas. These particles were then fragmented and dispersed initially in a gas and ash plume that rapidly transformed in to a high water supported thermal plume. Particle settling out occurred from high in the water column forming a fallout deposit. Lower plumes were generated above the brecciated carapace of Lava G dispersing fines tens of meters. Subunit 4 was generated following collapse of the caldera wall below Lavas G to I. Fragmentation occurred through a range of mechanisms with abrasion, decompression, and quenching of lava blocks caught up in the collapse. Exposure of hot magma/lava following the collapse would also have led to water magma interaction resulting in fragmentation. Dispersal occurred both as density currents and in plume rising from the exposed lava/magma.

8.3.2. Subunit 2

Subunit 2 (S2) directly overlies S1 across a locally gradational contact. It is extremely fine-grained, with modal grain size of 6 φ (approximately 16 µm) dominated by dense blocky and shard-like grains. This extremely fine grained deposit suggests highly energetic fragmentation (Zimanowski et al. 2003), and it thus represents a significant release of energy during the Havre eruption.

The S2 deposit is localised to sites south of a line that parallels the northern rim of the caldera, but extends in the south, east and west to the edge of the investigated area with no notable change in grain size. Subunit 2 is substantially thicker inside the caldera compared to on the rim. Locally, S2 is
divided into two parts, separated by the localised Subunits 3 and 4. Termination of this subunit along the northern caldera wall suggests that deposition was topographically constrained. We infer that S2 is related to settling of the finest-grained particles that were lofted into the water column by dilute density current(s) during sub-phase 2b (Fig. 8.2.). Topographic control of its northern extent suggest that S2 was deposited by a density current from a source on the southern caldera rim.

The presence of S3 and S4 as discrete layers within S2 indicates that it was deposited over a significant period of time. The very fine ash grainsize of S2 would result in extremely low (0.000132 – 0.000513 m/s in motionless seawater) rates of particle settling such that particles settling from just below sea surface would require five to 75 weeks to reach the seafloor 500 m below.

8.3.3. Subunit 3

Subunit 3 has been dispersed over the southern caldera rim, on the caldera floor, and in a single sample on the northern caldera rim. The deposit shows strong thinning and fining trends pointing to a source at Lava G on the southern caldera rim. The particles that compose S3 show distinctive elongate morphologies, with tube vesicles running parallel to the clast elongation direction. These particles bear morphological and microtextural resemblances to the pumiceous carapace of Lava G.

Subunit 3 contains particles from both syn-extrusive ash venting (SEAV) and carapace brecciation from Lava G (Fig. 8.2.). Dispersal was by thermal plumes rising from the hot extruding dome lava, and the subunit exemplifies a fine-grained suspension deposit sourced from a subaqueous effusive lava flow (Fig. 8.2.).

A population of grains in S3 display evidence of fluidal behavior during particle shaping. This implies a novel fragmentation mechanism to account for viscous behavior despite the high viscosity of this magma and enclosing seawater. I infer that this mechanism involves, relatively slow, isothermal viscous shearing of melt fibers or particles on the wall of outgassing pathways by the continuous flow
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of hot volatiles (Fig. 8.2.). This process is aided by the decreased magma viscosity due to the reduced exsolution as a result of eruption under high hydrostatic pressure.

8.3.4. Subunit 4

Subunit 4 is dispersed in two groups the first around Dome OP, and the second over the southern caldera rim, caldera floor, and the northern caldera rim. There is no stratigraphic link between the two parts of S4. No thinning or fining trend has been identified. Ash from S4 is composed of microcrystalline grains inferred to have been fragmented from the dense crystalline core of the lava produced during the 2012 Havre eruption (Fig. 8.2.). The local dispersal of S4 means it occurs stratigraphically within S2.

The two parts of S4 were produced by two different processes. Around Dome OP it was formed by fine-scale brecciation and quench fragmentation, followed by particle dispersal down the dome slopes. Elsewhere, on the caldera floor, S4 is inferred to have been formed by collapse of the caldera wall near Lava G to generate a density current that dispersed particles over the caldera floor (Fig. 8.2.).

8.4. Formation of silicic fluidal ash

A significant proportion of primary ash grains identified in S1 and S2 at Havre show evidence of fluidal behavior during or quickly following fine-scale fragmentation; their origin remains obscure. Fluidal behavior of silicic magma has, rarely, been inferred elsewhere (Mueller and White 1992; Simpson and McPhie 2001; Furukawa and Kamata 2004; Busby 2005; Self et al. 2008). In those cases the rhyolite appears to have been erupted under unusual circumstances: high temperature (Branney et al. 2008; Self et al. 2008), under high pressure (thus with increased volatile content) (Furukawa and Kamata 2004; Busby 2005), high content of network modifying volatiles (e.g. fluorine (Congdon and Nash 1991)), or a combination (Mueller and White 1992; Simpson and McPhie 2001).
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The fluidal behavior is inferred from thin and expansive lava flows (Busby 2005) and formation of spatter/cow-pat bombs (Mueller and White 1992; Simpson and McPhie 2001; Furukawa and Kamata 2004; Self et al. 2008). I know of no other descriptions of fluidal rhyolitic ash similar to that at Havre.

Fluidal ash in S1 and S2 shows no apparent chemical or volatile divergence from the other Havre eruption products. They also show no obvious microtextural differences, apart from the fluidal features, to brittlely produced particles. The fine grain size of many fluidal particle is suggestive of energetic fragmentation rather than during low strain rate fragmentation (Zimanowski et al. 2003).

The fluidal grains are best explained by large local increases in temperature, such as produced by lightning strikes on ash (Genareau et al. 2015; Wadsworth et al. 2017), or by shear heating in volcanic conduits (Rosi et al. 2004; Mastin 2005; Hess et al. 2008). Those mechanisms, however, produce particles with different sizes and microtextures from the Havre ones, so the fluidal grains in S1 and S2 remain an enigma.

8.5. An ash-informed unified eruption model of the Havre eruption

An explosive, pyroclast-forming mechanism is inferred for phase 2b (+phase 2a?) of the 2012 Havre eruption producing a complex multi-unit seafloor succession comprise of S1, S2, and the ALB Unit (+GP Unit?) (Fig. 8.2.). Current discussions among Havre cruise scientists have coalesced around two end-member models. Here I present my current interpretations of the eruption dynamics of this phase based on the work presented in this thesis. I also compare Havre with other deposits and models for deep silicic submarine volcanism. Finally, I attempt to reconcile the current (as of May 2018) end member models for the 2012 Havre eruption.

Havre is the most intensely studied deep subaqueous silicic eruption to date. Unlike other products of subaqueous volcanism studied, the pyroclastic deposits of the Havre eruption were emplaced over a firmly delimited period of 10 years. Within that time, it is all but certain that all the products from phases 3 were formed between 19th July and 26th October of 2012 (Carey et al. 2018). There is a high
probability that the phase 2b (+2a) deposits examined here (S1, S2, and the ALB Unit (+ GP Unit)) resulted largely from activity limited to a single day during the production of the pumice raft, on 18th – 19th July 2012. This level of temporal control is common, even a bit deficient, for modern subaerial eruptions, but orders of magnitude tighter than available for other submarine-emplaced deposits (e.g. Kano et al. 1996; Allen and McPhie 2000; Allen and McPhie 2009; Schipper et al. 2010; Rotella et al. 2013; Rotella et al. 2015). There is, in addition, exceptionally resolved spatial information from the AUV mapping and sampling program; the former matches or exceeds information available for many modern subaerial eruptions, the latter is unprecedented for a modern submarine deposit and unachievable in ancient successions.

Numerous subaqueous eruption processes have been proposed based on studies of ancient uplifted successions (Kano et al. 1996; Allen and McPhie 2000; Bear and Cas 2007; Allen and McPhie 2009; Jutzeler et al. 2014), and of deposits on the modern seafloor (Fiske et al. 2001; Wright et al. 2006; Allen et al. 2010; Schipper et al. 2010; Rotella et al. 2013; Rotella et al. 2015). The main models considered here are these: Neptunian (Kano 2003; Allen and McPhie 2009; Rotella et al. 2015; White et al. 2015), Tangaroan (Rotella et al. 2013), explosive lava dome destruction (Kano 1996; Allen and McPhie 2000; Kano 2003; Allen et al. 2010; White et al. 2015), and Poseidic (Schipper et al. 2010; White et al. 2015) eruptions. Although Poseidic eruptions were defined in a basaltic system they represent an important subaqueous end member model for entirely hydromagmatic fragmentation (Schipper et al. 2010; White et al. 2015).

Some key features of various subaqueous eruption models are summarised in Table. 8.1, along with equivalent features for the Havre phase 2b model presented here, and an alternative effusive phase 2a model (Manga et al. 2018). The Havre eruption phase 2b model presented here appears to most closely resemble a Neptunian eruption, but contrasts in several important features such as the occurrence of GP at the base of the sequence, and the lack of a Neptunian-style pyroclastic flow sequence. Several of the eruption models lack information in several categories, precluding direct comparison.
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In the Tangaroan model, ash is not explicitly accounted for, because fragmentation is driven by viscous detachment of buoyant magma rising in the vent. Because fragmentation and dispersal are buoyancy-driven, the Tangaroan model does not predict formation of density currents or any gas-thrust column.

Results presented in this thesis favors a dominantly explosive mechanism to form the pyroclastic deposits, which drove a gas thrust that penetrated the water column for at least 390 m above the vent (Fig. 8.2.). Ash from S1, and the pumice raft are inferred to have been produced at the same time, after deposition of the GP Unit, via dispersal in the gas thrust (Fig. 8.2.). Particles of S1 were entrained in an overlying convection cell and transported vertically upward from the point at which the gas-thrust decayed, with particle fallout then occurring from high in the water column. At the sea surface S1 ash formed the light-scattering plume of discolored water (i.e. Chadwick et al., 2008; Fiske et al., 1998; Kano, 2003; Watts et al., 2012) associated with the pumice raft (M. Jutzeler et al. 2014). Pumice clasts meanwhile ascended under their own buoyancy to the sea surface to form the raft (Allen et al. 2008; Jutzeler et al. 2014; Carey et al. 2018). Any pumice clasts that became water logged during ascent and began to sink would have become entrained in density currents produced off the gas jet because of fluctuations in mass/heat flux leading to condensation of the volatiles (Fig. 8.2.). Ash of S2, and the BLA Unit were deposited from these density currents resulting from column collapse. The GP Unit is suggested to have formed prior to the main eruptive episode (or possibly at the same time if there is S1 ash, unobserved and unsampled, beneath the GP Unit). This largely ash-based model, if GP were simultaneously erupted, resembles the model for a Neptunian eruption (Allen and McPhie 2009). However, further seafloor work at Havre seamount however would be required to differentiate these hypotheses.
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<th>Grain shape characteristics</th>
<th>Microtextural features</th>
<th>Eruption inferences</th>
<th>References</th>
</tr>
</thead>
</table>
| Neptunian      | Model derived from examination of ancient uplifted seafloor deposits | Deposits show similar stratigraphic sequences  
*Base* - Coarse lithic breccia facies  
± erosional base  
Up to several m’s thick  
Massive/normally graded  
*Middle* - Pumice lapilli-rich facies  
Up to 10’s m thick  
Laterally extensive  
Massive beds  
*Top* - Ash and GP facies  
Up to 10’s m thick  
Laterally extensive  
Laminated ash  
Deposit volumes between 1 – >200 km² | Giant pumice clasts to ash  
Lithics concentrated at the base of the sequence  
Grades up into juvenile material  
*Middle* – dominantly pumaceous lapilli  
*Base* – dominantly nonjuvenile clasts  
± dense juvenile material/deep water sediment particles  
*Top* – Composed of juvenile ash/ash and GP | All juvenile material microtexturally and compositionally similar  
High vesicularities in the pumaceous material (60-85 vol%) | The Neptunian model is inferred to be the subaqueous equivalent of a plinian eruption  
A rapidly erupting vesicular magma drives a gas jet above which a buoyant plume is driven  
Condensation of volatiles and water logging of pumice causes collapse of the plume driving density currents  
The *base* and *middle* facies are inferred to be the lithic rich base and juvenile rich core of a water supported density current  
Ash and large non-permeable pumice clasts are buoyantly transported vertically settling out over time - the *top* facies | (Kano et al. 1996; Allen and McPhie 2009; White et al. 2015) |
| Tangaroan      | Model derived from dredged pumice blocks and lapilli | Not observed | Lapilli to blocks recovered however dredging may have altered | Lower density clasts are generally rounded  
Higher density clasts appear | Glassy pumice  
Distinct lower and higher density pumices  
Some clasts contained both | Vesiculating magma is erupted to fast to produce a dome  
Instead blebs buoyantly detach  
Bleb margins quench while the Centre continues to vesiculate | (Rotella et al. 2013) |
<table>
<thead>
<tr>
<th>Chapter 8</th>
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</thead>
<tbody>
<tr>
<td><strong>original grain size</strong></td>
</tr>
<tr>
<td>Finer grain sizes not recovered</td>
</tr>
<tr>
<td><strong>blocky to subangular and show prismatic jointing</strong></td>
</tr>
<tr>
<td><strong>modes with a gradient between them Homogeneous composition</strong></td>
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<tr>
<td><strong>as it rises through the water column Bleb fragments dispersing fragments over the seafloor</strong></td>
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<tr>
<td><strong>Lava dome</strong></td>
</tr>
<tr>
<td>Model derived from examination of ancient uplifted seafloor deposits and from lava dome samples collected in situ from the seafloor</td>
</tr>
<tr>
<td>3m to 3cm thick moderately to well-sorted beds</td>
</tr>
<tr>
<td>Large grain size beds show tabular forms</td>
</tr>
<tr>
<td>Smaller grain size beds show wedge shapes</td>
</tr>
<tr>
<td>Ash through to 3m pumice blocks</td>
</tr>
<tr>
<td>Dominantly composed of pumice pebble to bolder size &lt;10 wt.% ash</td>
</tr>
<tr>
<td>Low lithic content (&lt;1 wt.%)</td>
</tr>
<tr>
<td>Cobble to bolder sized blocks show internal polyhedral joints with quenched margins</td>
</tr>
<tr>
<td>Pebble to cobble pumice angular to subrounded</td>
</tr>
<tr>
<td>Pumice vesicularity 36 to 86 vol.%</td>
</tr>
<tr>
<td>Large clasts formed by buoyant dispersal following brecciation of the pumaceous carapace which then form settling deposits</td>
</tr>
<tr>
<td>Explosive destruction of the lava dome by magma water interaction or explosive decompression generates smaller pumice clasts forming wedge shaped density current deposits, in addition to resedimentation downslope</td>
</tr>
<tr>
<td>(Kano 1996; Allen and McPhie 2000; Kano 2003; Allen et al. 2010; Jutzeler et al. 2015; White et al. 2015)</td>
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<tr>
<td><strong>Poseidic</strong></td>
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<tr>
<td>Model derived from samples collected in situ from the seafloor</td>
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<tr>
<td>Sample taken from the flanks of a cone near the summit plateau of Loihi</td>
</tr>
<tr>
<td>40 cm bombs to fine ash</td>
</tr>
<tr>
<td>Dominantly lapilli sized</td>
</tr>
<tr>
<td>Fine ash component shows a population of active particles</td>
</tr>
<tr>
<td>Low lapilli vesicularities</td>
</tr>
<tr>
<td>No textural indicators of magmatic fragmentation</td>
</tr>
<tr>
<td>Closed system volatile coupled decompression pathway</td>
</tr>
<tr>
<td>Based on basaltic magma Coupled volatile magma decompression lead to rapid ascent and vesiculation of the magma resulting in weakening Thermohydraulic explosions below the vent level drove primary fragmentation of the magma with little to no magmatic component Dispersal occurred in jets resulting from thermohydraulic explosions</td>
</tr>
<tr>
<td>(Schipper et al. 2010; White et al. 2015)</td>
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<tr>
<td><strong>Effusive eruption</strong></td>
</tr>
<tr>
<td>Model derived from Partitioning of erupted material</td>
</tr>
<tr>
<td>Pumaceous lapilli to giant</td>
</tr>
<tr>
<td>Pumice clasts show surface</td>
</tr>
<tr>
<td>Raft pumice and GP modal</td>
</tr>
<tr>
<td>Model suggest extrusion of eruption magma occurred at a</td>
</tr>
<tr>
<td>(Manga et al. 2018)</td>
</tr>
<tr>
<td>Model for phase 2a of the 2012 Havre eruption</td>
</tr>
<tr>
<td>---------------------------------------------</td>
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<tr>
<td>Explosive eruption model for phase 2b of the 2012 Havre eruption</td>
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</table>

Table 8.1. Comparison between depositional, granulometry, grain shape, and microtextural characteristics of various pyroclast forming subaqueous eruption styles inferred from uplifted deposits and seafloor sampling
8.6. In defense of an explosive mechanism for phase 2b of the 2012 Havre eruption

The detailed study of the 2012 Havre eruption has sparked many questions regarding the mechanisms of large deep subaqueous silicic eruptions (Carey et al. 2018). At the time of writing one area this discussion has focused in on is the exact eruption dynamics of what has been defined here as phase 2 of the Havre eruption. That is the phase during which most of clastic material was produced i.e. the GP Unit, the ALB Unit, S1, S2, and the pumice raft. The work presented here points towards energetic fragmentation of the magma and the presence of a gas jet, strongly implicit of explosive volcanism. However, Manga et al. (2018) have suggested an alternative entirely effusive eruption mechanism. Here I attempt to reconcile the two models.

The model presented by Manga et al. (2018) is based on modeling of conduit conditions and pumice water logging processes informed by studies on pumice composition and vesicularity characteristics, along with inferred mass eruption rates and environmental conditions at the vent. The conduit model suggests that the inferred mass eruption rate was not high enough for magmatic fragmentation to occur at 900 mbsl for a magma of the composition erupted in the 2012 Havre eruption. Instead the vesiculating magma is inferred to erupt onto the seafloor as a spine where coarse fragmentation into metre scale blocks occurs by quenching. The gas charged pumice is initially buoyant in the water column ascending vertically. Pumice water logging occurs on variable timescales dependent on the interconnectedness of porosity and its size (Allen et al. 2008; Fauria et al. 2017). The clasts of the GP Unit had in general higher interconnectivity resulting in water logging and sinking (Manga et al. 2018). In contrast the raft pumice had lower interconnectivity meaning clasts where able to stay buoyant to the sea surface and entrain air, potentially remaining floating indefinitely.

The effusive model presented by Manga et al. (2018) however does not account for many of the results of analysis of the ash presented here. Granulometry and particle shape are strongly implicit of energetic fragmentation, while fluidal silicic ash and a shallow quenching signature imply the presence of a gas jet, neither of which is consistent with an entirely effusive eruption model. It also
seems unlikely that density current deposits, S2 and the ALB Unit, would form in a system in which dispersal is driven entirely by the individual buoyancy of clasts. In addition, the eruption volume estimates currently represent ‘at least values’ since every seafloor deposit extends off the study area.

Both models appear to present strong evidence for explosive or effusive volcanism during phase 2 of the 2012 Havre eruption, however there appear to be two options for reconciliation. The first, generally assumed here, is to split sub-phase 2a (the GP Unit) and sub-phase 2b (S1, S2, and the ALB Unit) into two distinct events, based on the stratigraphic relationship and differences in settling velocity of GP and ash. The pumice raft in this case could be related with either event/sub-phase. Alternatively, if S1 and the GP Unit occur as a single mixed deposit then phase 2 may occur as a single eruptive event. In this case the two eruption models could represent end members along a spectrum.

Based on current data I favor an interpretation of sub-phase 2a and 2b occurring as two distinct eruptions. An initial eruption dispersed the GP Unit over the Havre caldera. Following settling of the GP Unit a more intensive eruption occurred generating explosive volcanism with an overlying gas jet from which density currents were generated.

8.7. Foreign grains: origin and distal dispersal

A small number of foreign grains have been identified scattered in trace quantities through the 2012 Havre ash deposit. They are not inferred to have been produced in the eruption, and I considered two possible sources for these grains. The first possibility is that they were scoured from young foreign lava flows and/or volcaniclastic deposits exposed on the outer flanks of Havre (Wright et al. 2006). The second option is that they were transported long distances from various volcanoes around the Kermadec Arc.

The apparent freshness of the foreign grains seems to preclude a third possibility, that are lithic fragments entrained from ‘country rock’ of the caldera wall strata. The concentration of
hydrothermal systems along the caldera ring fault, and evidence of hydrothermal alteration in a caldera collapse deposit, means that fresh volcanic glass is likely to have been rapidly altered. For grains to remain relatively unaltered in the ocean they must have been relatively rapidly transported away from continuous heat sources i.e. vents, and hydrothermal systems. These considerations make the vent walls and volcano substructure an unlikely source for fresh glassy particles.

If the foreign particles are from scouring of relatively fresh lava/volcaniclastics on the volcano's outer flanks, then fine particles were entrained and carried upslope to be incorporated into the sampled ash. Transport of these particles would be in turbulent currents, potentially present but not documented, around the Havre edifice.

Glass geochemistry, however, reveals greater diversity than known from Havre itself, and suggests that these grains are derived from multiple Kermadec volcanic sources. Although the specific sources of these particles remain unknown, the implied long-distance transport from the nearest neighboring volcanoes along the arc has implications for the study of small, fresh, submarine deposited volcanic particles. The rate of deposition in the proximal environment during the Havre eruption would have been very high. I speculate that in the distal environment the lower rate and volume of sediment input mean that foreign grains may compose a higher percentage of the overall deposit. Studies of ash from distal locations may therefore have to account for the steady input of ash from a range of sources due to scouring and re-sedimentation.

8.8. Limitations/weaknesses

The study of the 2012 Havre eruption is the most comprehensive yet of a deep silicic submarine eruption or its deposits (Carey et al. 2018). There nevertheless remain some limitations to the overall study. The 2015 cruise focused dives on an area approximately 4 km² around the summit caldera of Havre (Carey et al. 2018) and extracted an excellent dataset on the lava and proximal clastic deposits of the eruption (Carey et al. 2018). A consequence of the necessary focus on new volcanic features
and the caldera is that the medial deposits remain unstudied; every clastic deposit so far identified at Havre extends beyond the 2015 study area. The medial and distal stratigraphy of the Havre deposits and their implications remain masked. Deposits S1, S2, S4, and the GP Unit show little evidence of thinning or fining and thus may extend far beyond the area examined. If the volume and extent of material, particularly ash, dispersed to seafloor deposits beyond the study area is large, documenting it could substantially change volume estimates and in turn inferences of eruption dynamics. From another cruise (R/V Sonne cruise SO255) undertaken in the Kermadec arc since the 2012 eruption there are hints of extremely wide dispersal of at least Havre pumice (pers comm; Jutzeler 2017). Whether ash is also present at these distal pumice sites, and if so what its thickness and grainsize might be, remains unknown and would require additional ROV dives or core/grab sampling of the uppermost seafloor deposits.

8.9. Future work

Despite the regional and local contextual information available for the 2012 eruption of Havre (Wright et al. 2006; Barker et al. 2013; Carey et al. 2014; M. Jutzeler et al. 2014; Rotella et al. 2015), and with the unprecedented detail in which this deep subaqueous eruption’s deposits have been studied, there remain many unanswered questions and areas for future work (Carey et al. 2018). Fragmentation is partly assessed here, but is the focus of planned work that will build on these results. In the last decade much progress has been made in the use of microtomography as a tool to understand geological processes (Ketcham and Carlson 2001; Jerram and Higgins 2007). In volcanology microtomography is increasingly often used to analyse vesicle populations (Berg et al. 2016). With the use of synchrotron sources, in situ temporal analysis of high temperature experiments has been undertaken. A few recent studies have also analysed grain shape using tomodirgy (Rausch et al. 2015; Vonlanthen et al. 2015; Dioguardi et al. 2017), and suggest another approach to characterising grain shape and interpreting fragmentation mechanisms at Havre. This
would require the development of a new set of grain morphology standards specific to the analysis of microtomography.

There remains significant divergence in views of the role and exact mechanisms of MWI, and in how they can be diagnosed (Zimanowski et al. 2003; Austin-Erickson et al. 2008; Liu et al. 2015; White and Valentine 2016; Liu et al. 2017). The Havre samples would provide an excellent test bed with which to examine MWI during a silicic eruption.

The 2015 Havre cruise has given us a good understand of the proximal clastic deposits and lavas produced in the eruption. Any future seafloor studies at Havre should therefore include observation and sampling of the medial and distal eruption deposits.

More-detailed FTIR examination of the Havre ash could reveal answers about the formation mechanism, specifically of S1 and S2. The rapidity with which ash quenches can be used to infer the depth at which is first came in to contact with liquid water. By examining a large range of grains from various deposits formed at Havre a quench depth range could be established which would provide extensive insight into ash formation and eruption processes. The variation in grain size and thus quenching time means that glass hydration time scales can be investigated in detail. Hydration studies could also provide more insight to particle cooling histories and thus eruption conditions and processes.
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