Melting of Ice by Magma-Ice-Water Interactions During Subglacial Eruptions as an Indicator of Heat Transfer in Subaqueous Eruptions

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Eruptions within glaciers are characterized by fast cooling of volcanic deposits, rapid melting of ice and heating of meltwater. Heat transfer rates in subglacial eruptions may be monitored through melting rates of ice and simple calorimetric calculations used to infer heat fluxes and estimate the efficiency of heat transfer from magma. Cooling models of effusive basaltic eruptions forming pillow lava indicate that thermal efficiency of such eruptions is 10–45%, and highest when the eruption rates are low and pillows are exposed to surrounding meltwater for a comparatively long time. When magma fragmentation occurs by non-explosive granulation or explosive activity the glass particles formed have diffusion times mainly in the range 10-3 s to 102 s depending on grain size, the mean being of the order of 1 s. Limited observational data on ice-melting rates and models of cooling times suggest that the efficiency of heat transfer from fragments may commonly be 70–80%. Correspondingly, total heat transfer rates associated with fragmentation are several times higher than for pillow lava at the same eruption rate. The contrasts in efficiency imply that variation in heat transfer rates during fragmentation may closely correlate with variations in magma eruption rate, whereas for pillow lava eruptions changes in heat transfer lag well behind changes in eruption rate. Though pillows may still have molten cores when buried in a growing volcanic pile, the temperature of volcanic glass created during subaqueous fragmentation should be no greater than 250–300°C at the time of deposition.

INTRODUCTION

Interaction of magma and water in subaqueous eruptions can cause rapid quenching of magma into glass, pillow lava formation with partial to almost complete crystallization, yet in some cases sheet lavas still form [e.g. *Moore*, 1975; *Wohletz*, 1983; *Griffiths* and *Fink*, 1992]. The rapid cooling reflected in the formation of fine-grained glass particles, quenched, glassy, pillow rinds and, in some cases almost fully quenched and glassy sheet lavas, implies fast heat transfer from magma to the surroundings, i.e. ocean, lake or groundwater. The quantification of these processes (heat transfer rates etc.) is difficult, especially so in the ocean due to a paucity of well-instrumented observations of submarine eruptions.

In most cases eruptions under glaciers are essentially subaqueous, with the same cooling processes occurring as in eruptions under open water. Volcanic areas characterized by hyaloclastite formations and pillow lava were formed during the Pleistocene in Canada, Iceland and Antarctica [*Hickson*, 2000; *Kjartansson*, 1959; *Smellie*, 1999]. High stratovolcanoes throughout the world have partly ice-covered slopes, and ice-(water-)magma interaction influences their eruption behavior [*Major* and *Newhall*, 1989]. In Iceland, eruptions within glaciers are common and hyaloclastite mountains are still being formed under ice caps [*Gudmundsson et al.* 1997, 2002; *Larsen*, 2002]. For eruptions of stratovolcanoes where magma-ice interactions take place, the ice is often thin, and most of the energy is usually lost to the atmosphere through the eruption plume; only a minor part is dissipated through melting of the shallow ice covering the slopes. In contrast, when eruptions occur within large glaciers, ice caps or large ice-filled calderas a large part of the magmatic heat is used to melt ice (Figure 1) and calculations of melting rates and heat transfer can be made through measurements of the volume of ice melted. This provides important insight into rates of cooling processes in subaqueous eruptions in lakes and in the ocean.

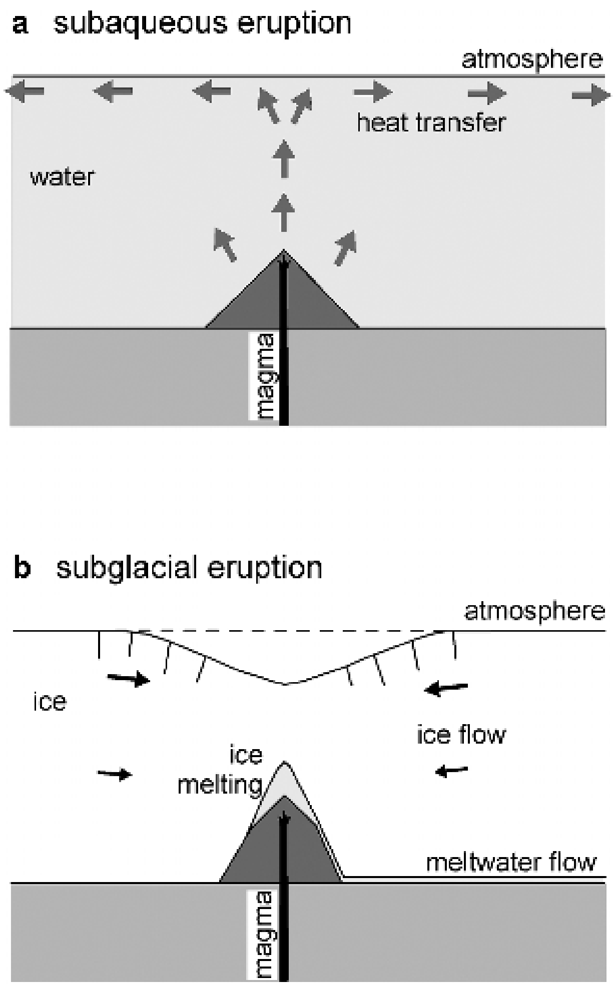

Figure 1. Schematic settings for, a) a subaqueous eruption, and, b) a fully subglacial eruption. In both cases the volcano and the water reservoir form a closed system. However, in a) heat transfer rates are very difficult to measure, whereas in the subglacial case fairly accurate estimates can be done by estimating meltwater production and/or ice loss, based on a combination of visual observations at the eruption site and possibly stream gauging downstream.

SUBGLACIAL ERUPTIONS

Most eruptions under glaciers lead to jökulhlaups or lahars when meltwater is released. Such flooding is one of the main hazards associated with subglacial eruptions [*e.g. Björnsson*, 1975]. Several observations of ice-magma interaction have been reported from stratovolcanoes [*e.g. Major and Newhall,* 1989]. Direct observations of eruptions in which ice melting takes up a large part of the thermal energy of the eruption are not as widely reported. Recognition of exclusively subglacial eruptions is rare. Most observed eruptions within glaciers in Iceland have an early phase during which the activity is confined under the ice (Figures 2 and 3). Usually, the eruption melts its way through the ice cover and then becomes explosive, typically producing surtseyan, phreatomagmatic, activity (Figure 4), and sustaining eruption plumes that carry tephra into the atmosphere [*Larsen et al.* 1998, *Larsen* 2002]. Confirmed exclusively subglacial eruptions include the eruption in Gjálp, Vatnajökull in 1938 [*Björnsson* 1988; *Gudmundsson and Björnsson*, 1991]. The subaerial phase of the more recent Gjálp eruption, in 1996, constituted only a minor part of the eruption in terms of magma and energy transport [*Gudmundsson et al.* 1997, 2003]. In both eruptions >95% of the total available thermal energy was used for ice melting, with the great majority of heat for melting transferred through water. Three eruptions have been directly observed at Grímsvötn, Iceland, in 1934, 1983 and 1998 [*Áskelsson*, 1936; *Tryggvason*, 1960; *Grönvold* and *Jóhannesson*, 1984; *Sigmundsson et al.,* 1999]. These eruptions rapidly opened a pathway for magma through 100–150 m of ice and were subsequently characterized by surtseyan activity (Figure 4). Some melting of ice occurred in these eruptions. However, the volume of tephra deposited from the eruption plumes was such a large fraction of the total magma erupted that the heat dissipated to the atmosphere from the eruption plumes was probably greater than that used to melt ice.

The best-documented example of an eruption within a thick ice cap is the Gjálp eruption in 1996 (Figure 3). It lasted from September 30 until October 13. Ice thickness at the site was 550–750 m before the eruption, during which a 6 km-long and up to 450 m-high hyaloclastite ridge was formed by 0.8 km^3 of mostly volcanic glass, equivalent to 0.45 km^3 of magma [*Gudmundsson et al.* 2002] of basaltic icelandite composition [*Steinthorsson et al.* 2000]. During the eruption about 3 km^3 of ice were melted. About 2 km^3 of ice was melted in the first four days, indicating a heat

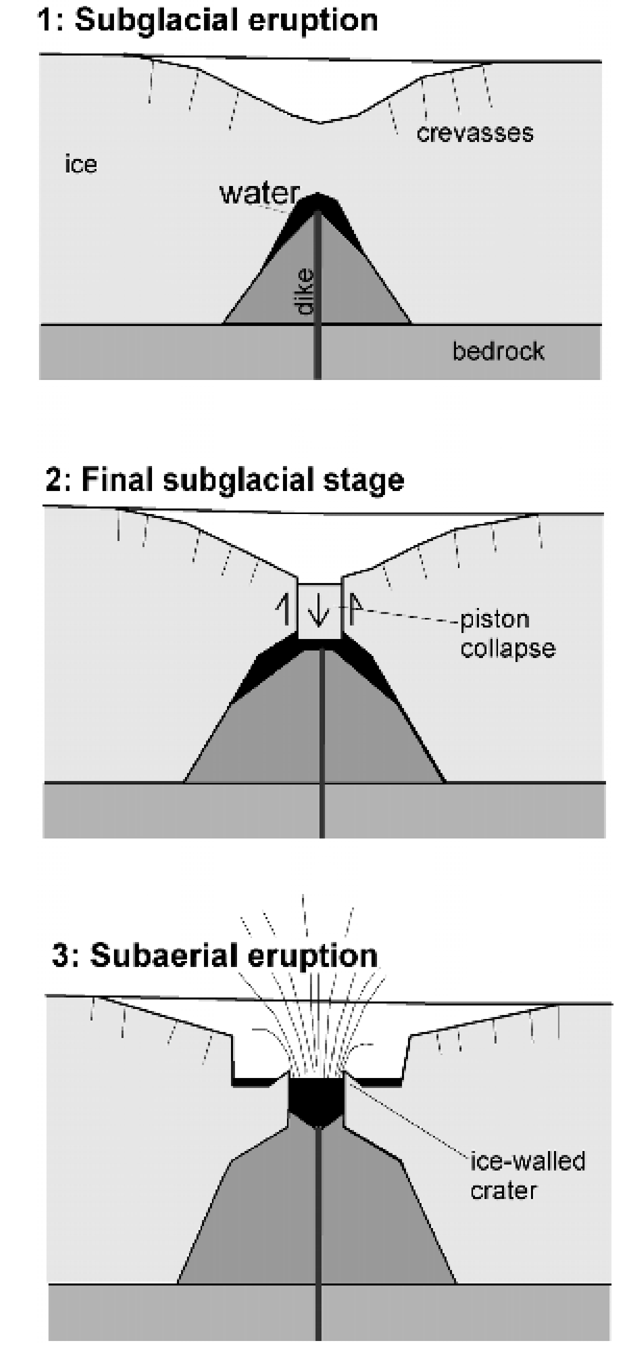

Figure 2. Schematic sections showing how volcanic eruptions may penetrate thick ice, based on observations from the Gjálp eruption in Iceland in 1996. In (1), heat from the eruption rapidly melted the overlying ice, and as water drained away subglacially, depressions formed in the ice surface. (2) As the eruption advanced brittle failure of the ice overlying the vent caused collapse of a piston of ice, which was rapidly melted. In (3) a subaerial vent has opened through the hole left by the melted ice piston. [Modified from *Gudmundsson et al.,* 1997].

Figure 3. The surface expression of a fully subglacial eruption in Gjálp on October 1 1996, about 15 hours after its start. Depressions were forming in the ice surface over subglacial vents.

transfer rate of about 2×10^{12} W [*Gudmundsson et al.* 1997]. The corresponding heat flux, found by dividing the heat transfer rate by the area of the volcanic edifice, was 5–6×10^5 W m^{-2}; about an order of magnitude too high to be explained by cooling and solidification of pillow lavas, and hence implying turbulent mixing of quenched ash fragments and meltwater [*Gudmundsson et al.* 1997; 2003].

The morphology of erupted products from past subglacial eruptions provides indications of heat transfer rates during the formation of these deposits, which are found in several volcanic regions around the world [e.g. *Mathews*, 1947, *Bemmelen* and *Rutten*, 1955; *Kjartansson*, 1959; *Jones*, 1969; *Furnes et al.*, 1980; *Smellie*, 1999; *Hickson*, 2000;

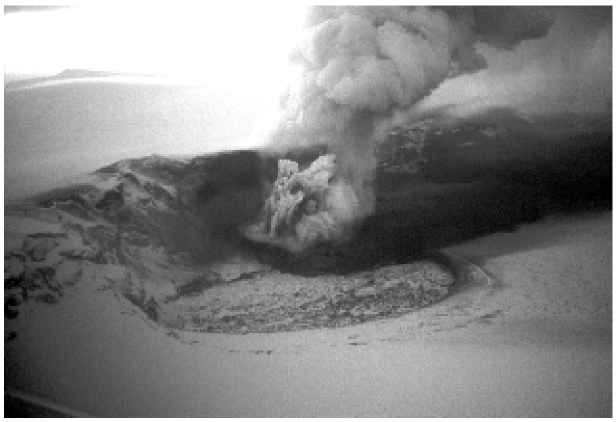

Figure 4. The Grímsvötn eruption in December 1998 was englacial, but the vents ejected material subaerially from the start because ice thickness at the eruption site was only 50–150 m. The photo shows the eruption in its fifth day. A tuff cone was forming from surtseyan explosive activity. A large part of the erupted tephra banked up against ice walls, demonstrating that it had lost most of its heat as it was deposited. The cliff is about 300 m high.

Tuffen et al. 2001]. For basalts, the complete stratigraphic sequence formed in such eruptions can be divided into four main units:

Unit 1: Pillow lavas usually make up the basal part of the sequence. The pillow lavas are erupted in the initial effusive phase when water pressure in the subglacial vault is high. The pillows are commonly 0.5–1.0 m in diameter [e.g. *Jones*, 1969] and have a glassy rind and crystalline core.

Unit 2: On top of the basal pillow lavas, fragmented volcanic glass piles up when granulation and explosive activity become the dominant styles of activity as the volcanic edifice rises and confining pressure is reduced. In hyaloclastite mountains formed during glacial periods the volcanic glass has to a large degree altered to palagonite [e.g. *Jones*, 1969; *Jakobsson*, 1979; *Smellie* and *Skilling*, 1994]. The term hyaloclastite is used here for this type of fragmented volcanic glass; it is commonly used for all clastic volcanic material produced both in explosive and non-explosive magma-water interaction [e.g. *Fisher and Schmincke*, 1984; *Werner and Schmincke*, 1999]. As the hyaloclastite pile grows the eruption may melt its way through the ice, leading to surtseyan explosive activity within an englacial lake.

Unit 3: After the crater of the growing hyaloclastite edifice has emerged from the water, the vent may eventually isolate itself from the surrounding englacial lake. The style of the eruption will then change, and effusive activity lead to subaerial flow of lava.

Unit 4: As the lavas of unit 3 enter the water a delta forms, which is composed of dipping layers of pillows, pillow fragments and hyaloclastite. *Jones* [1969] and *Jones and Nelson* [1970] introduced the term flow-foot breccia for these deltas.

The term tuya is used for mountains having lava caps and formed by eruptions within glaciers. The similar mountain with no lava cap commonly indicates that the eruption did not last long enough to become effusive during its subaerial stage, or that a relatively thin cap was removed by erosion. This applies to most of the hyaloclastite ridges formed in subglacial fissure eruptions in Iceland [*Kjartansson*, 1959; *Jones*, 1969; *Allen*, 1980], since usually only units (1) and (2) are found in the ridges. Some ridges and tuyas may lack the basal pillows, especially ridges of modest height. Both pillow lavas and hyaloclastites are diagnostic of rapid heat transfer from magma to the surroundings.

In the following discussion on heat transfer in eruption of basaltic to mildly intermediate magma, only the formation of the two units that occur in truly subglacial eruptions (units 1 and 2) is considered, when magma, water and ice form a closed system. The heat transfer data from Gjálp are an important constraint but the Gjálp eruption was not long-lived enough to create units 3 and 4 [*Gudmundsson et al.*, 1997; 2003]. Although models may be constructed of heat transfer from magma to water during formation of flow-foot breccias, this is not attempted here.

CALORIMETRY

The processes that occur during rapid cooling and solidification of magma, especially fragmentation due to magma-water interaction, are very complex. In addition to high rates of heat transfer from magma to meltwater and ice, vaporization and condensation of water will occur. Expansion of steam accelerates and imparts momentum to the magma-steam-water mix, the magma is fractured, seismic waves are generated, and energy is expended in the creation of new particle surfaces (fragmentation energy) [*Zimanowski*, 1998; *Buettner and Zimanowski,* 1998]. In the simplified treatment of calorimetry and eruptive heat fluxes provided here, only the heat transfer is considered and most of the secondary processes are ignored. This simplification can be easily justified for the truly subglacial case, in which the volcano, ice and meltwater may form a closed system where the heat of the magma is used to melt ice (Figure 5) and no heat is lost to the atmosphere through a plume. In some cases the meltwater may have a temperature above the melting point when it drains from the glacier. However, heat transfer to the overlying ice from fast-flowing meltwater at the base of a glacier is very rapid; the 8°C meltwater

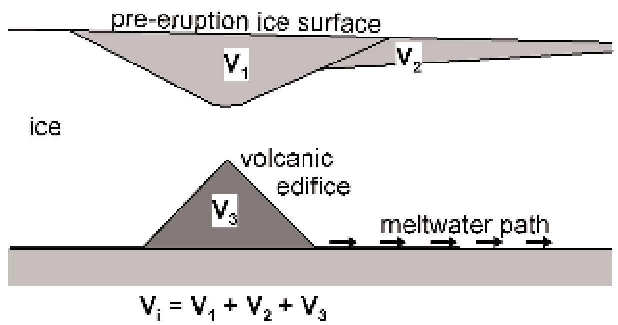

Figure 5. Schematic illustration of how large ice volumes can be melted in subglacial eruptions. V_i, can be estimated from volume of depressions in the ice surface (V_1 and V_2) and the edifice (V_3). A small but significant contribution to V_1 is the volume of void created by crevasses. The volume V_2 is created as the meltwater loses its heat during passage along the glacier bed.

released from Grímsvötn in Vatnajökull on November 5 1996 lost its excess heat over a path of only 6 km [*Björnsson et al.*, 2001; *Jóhannesson*, 2002].

In the case where magma at the liquidus crystalizes and cools down to ambient temperatures, the energy contained in a volume V_m of magma with density ρ_m is given by

$$E=\rho_m V_m \left[L_m + C_m (T_i - T_f) \right] \tag{1}$$

Here L_m is the latent heat of crystallization of the magma, C_m is the specific heat capacity of the crystalline lava, T_i is the initial magma temperature and T_f is the temperature of the surroundings. If phenocrysts make up a significant fraction of the volume of the magma, this can be taken into account by only using the latent heat of the liquid fraction. When magma is fragmented into glass and no crystallization takes place, the heat content is calculated from

$$E=\rho_m V_m C_g (T_i - T_f) \tag{2}$$

No latent heat is released, but the specific heat capacity of glass is somewhat higher than that of crystalline material of the same composition [*Bacon*, 1977]. The heat capacity decreases with decreasing temperature [*Bacon*, 1977, *Spera*, 2000] but for the large temperature changes characteristic for ice-volcano interaction, the use of a single mean value, C_g, should be sufficiently accurate in most cases.

The magmatic heat is dissipated by ice heating, ice melting, meltwater heating and vaporization of a fraction of the meltwater. While the system remains closed vaporisation can be ignored since steam will condense as it melts ice. Thus, the loss of heat during melting of a volume V_i of ice is given by

$$Q=\rho_i V_i \left[C_i (T_0 - T_1) + L_i + C_w (T_2 - T_0) \right] \tag{3}$$

where C_i and C_w are respectively the heat capacities of ice and liquid water, L_i is the latent heat of fusion for water, T_1 is the initial temperature of the ice, T_0 the melting temperature of the ice and T_2 is the final temperature of the meltwater. In most cases the meltwater will lose its heat as it flows along the glacier bed, and in the case of temperate glaciers $T_1 = T_0$. Thus, for a temperate glacier the heat dissipated may simply be $Q = \rho_i V_i L_i$.

Although heat transfer can be extremely fast, not all the magmatic heat is given off instantly. The efficiency of heat transfer [*Höskuldsson and Sparks*, 1997, *Gudmundsson et al.* 2003] can be defined as the ratio of the rate of heat transferred through the vent by magma to the rate of heat transferred to the surroundings by melting of ice and heating of the meltwater. The rate at which heat is brought through the vent, dE/dt, depends on the magma flow rate $\rho_m dV_m/dt$, so the rate of heat transferred to the surroundings is given by dQ/dt. The efficiency of heat transfer *F* is then given by

$$f=\frac{dQ/dt}{dE/dt} \tag{4}$$

This efficiency is difficult to estimate; here data on melting rates in actual eruptions are used and inferences made on the basis of models of heat flux. Melting rates were measured in the Gjálp eruption in 1996 [*Gudmundsson et al.* 1997; 2003]. Time-average values of *F* were obtained from the total mass of ice melted and total mass of erupted material [*Gudmundsson et al.* 2003]. Two different definitions of *F* were used for the Gjálp eruption. Firstly, there is the efficiency of heat transfer for ice melting, f_i. Secondly, there is the efficiency of heat transfer from magma f_m. The first value only takes into account ice melted at the eruption site itself while in the second, heating of the meltwater is also included. The values obtained for the two efficiencies were f_i = 50–60% and f_m = 63–77% [*Gudmundsson et al.* 2003]. The difference between the two reflects the residual heat of the meltwater, which had a temperature of about 20°C as it flowed from the eruption site. This heat was released along the path of the meltwater, downslope of the eruption site itself [*Gudmundsson et al*. 1997; 2003].

MECHANISMS OF HEAT TRANSFER

Effusive Eruptions

Heat transfer in effusive subglacial eruptions (those forming pillow lava) has been studied by *Allen* [1980], *Höskuldsson and Sparks* [1997], *Wilson and Head* [2002] and *Tuffen et al.* [2002]. *Allen* [1980] estimated the heat flow from pillows and concluded that it was rapid enough to melt a volume of ice larger than the volume of effused lava, thereby refuting an argument against tuya formation in subglacial eruptions put forward by *Einarsson* [1966]. *Höskuldsson and Sparks* [1997] and *Wilson and Head* [2002] used solutions to the Stefan problem [*Carslaw and Jaeger*, 1959; *Turcotte and Schubert*, 1982] to study heat fluxes from a layer of pillows. Stefan's problem deals with the heat flux from a solidifying infinitely-wide and thick layer (a half-space) where the temperature at the layer boundary remains fixed at the ambient temperature (T_f). Heat is lost by solidification of the melt and cooling of the solidified crust that separates the melt region from the surroundings. To simplify the problem it is assumed that the magma has a single well-defined melting point T_i. The depth to the solidification front y_m, is found from

$$y_m = 2\lambda\sqrt{\kappa t} \tag{5}$$

where k is the thermal diffusivity of the lava and λ is a dimensionless parameter obtained from

$$\frac{L_m\sqrt{\pi}}{C_m(T_i - T_f)} = \frac{e^{-\lambda^2}}{\lambda erf\lambda} \tag{6}$$

Previously defined parameters are as before and erf is the error function. A temperature profile can be obtained for the crust (the region $0<y<y_m$) as well as an equation for the average heat flux over time t since the emplacement of the molten layer

$$q_{av} = \frac{\rho y_m}{t}\left[L_m + C_m\left(T_i - \frac{1}{y_m}\int_0^{y_m} T(y)dy\right)\right] \tag{7}$$

The thickness y_m is obtained from (5) and (6). To a good approximation (7) can be simplified by assuming a linear temperature gradient in the crust, yielding

$$q_{av} \approx \frac{\rho y_m}{t}\left[L_m + \frac{C_m(T_i - T_f)}{2}\right] \tag{8}$$

Using a slightly different equation for q, *Höskuldsson and Sparks* [1997] calculated heat fluxes for various extrusion rates, assuming that one layer of pillows is emplaced on top of another, with the layers having a fixed thickness. The time t in (7) and (8) is the exposure time of the pillow lava layer. This simple model probably underestimates the heat flux from a layer of cooling pillows, at least the early phase of the cooling. Pillows have a partly glassy outer surface that will lose heat more rapidly than assumed in the model. Cooling cracks also form, commonly with a spacing of ~0.1 m, allowing access of water to the interior of the pillow. Figure 6 shows the heat flux from a 0.5 m thick pillow lava layer assumed to cover an area of 2 km^2. The layer is an idealization, but 0.5 m is the typical diameter of basaltic pillows [*Jones*, 1969]. The heat flux increases with increasing extrusion rate, but the heat exchange efficiency falls due to shorter exposure time. In reality pillows buried within a pile of pillow lava continue to give off heat but this must happen at a rate much reduced compared to pillows that are directly exposed to liquid water, even though a film of steam may partly insulate pillows at the seafloor from water [*Moore*, 1975]. To acknowledge these shortfalls in the model, the heat flux from the pillow layer in Figure 6 is shown as a region, with the lower bound being according to (8) and the upper bound double that value. However, for the lowest flow rates (<100 m^{3}s^{-1}) only the results for (8) are shown, because when the exposure time approaches the same order

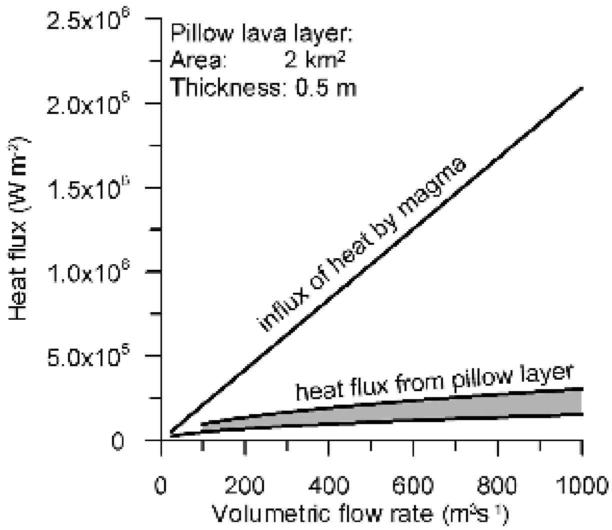

Figure 6. Estimates of heat fluxes in a subglacial basaltic effusive eruption. The influx is estimated from eq. (1) assuming magma density ρ=2600 kg m^{-3}, latent heat of fusion for magma L_m=4x10^5 J kg^{-1} [*Spera*, 2000], specific heat capacity of magma C_m=1100 J kg^{-1}°C^{-1}, T_i= 1200°C and T_f= 100°C. The heat flux from the pillow lava layer is obtained from eq. (8) assuming a thermal diffusivity of k=10^{-6} m^{2}s^{-1}.

as the time it would take to release most of the thermal energy in the layer, the average heat flux cannot be significantly higher than predicted by (8). The rather ad-hoc estimate of the upper bound in heat flux demonstrates the uncertainty in the model; the results, especially for the higher flow rates, should be regarded as order-of-magnitude estimates.

Magma Fragmentation

The abundance of hyaloclastites in subaqueously and subglacially-erupted sequences testifies to the importance of magma fragmentation in such eruptions. Clearly, heat transfer from magma/volcanic glass to the surroundings cannot be described by the solution to Stefan´s problem (equations 7 and 8). Grain size of glass particles formed in “hydrovolcanic” fragmentation is commonly in the millimeter to submillimeter range (Figure 7) [*Wohletz*, 1983, *Werner and Schmincke*, 1999]. To describe the heat flux from the magma/glass particles to the surroundings it is convenient to use the concept of diffusion time [*Colgate and Sigurgeirsson*, 1973; *Turcotte and Schubert*, 1982; *Wohletz,* 1983]. The thermal diffusion time (or thermal equilibration time) of a particle with thermal diffusivity k and diameter/thickness d is

$$\tau = \frac{d^2}{4\kappa} \tag{9}$$

It is assumed that heat flow within the particle occurs by conduction, and that convection in the surrounding fluid keeps the surface of the particle at or close to the fluid temperature. This fluid may be liquid water or steam. For a particle cooled from all sides, about 90% of its excess heat relative to the surroundings is lost during its diffusion time. In the case of subglacial eruptions the surroundings are the meltwater; for other subaqueous eruptions it is a pre-existing water body.

The intense heat transfer from magma to water should lead to steam generation but the role of steam in heat transfer in the magma-water-ice system is poorly known. It is clear from observations [e.g. *Thorarinsson*, 1967], theoretical considerations [e.g. *Wohletz*, 1983] and experiments [*Zimanowski*, 1998] that steam generation and expansion is a major factor in shallow-water explosive activity and the formation of plumes in such settings. This situation will, however, not be considered further here. The following, discussion is limitied to the closed magma-water-ice system and the effect steam may have on the rates of heat transfer from magma to the overlying ice.

Steam films may cover some and perhaps most particles [e.g. *Wohletz*, 1983] for a part of their exposure time. Steam would influence the heat transfer in two ways. Firstly, the temperature of saturated steam may be 100–200°C higher than that of the meltwater; secondly, thermal conductivity of steam is much lower than for liquid water. The first effect may provide "ambient" temperature that is much higher than that of the meltwater. The second effect reduces the rate of heat loss from the particles to the surroundings although it may be partly offset by steam convection and radiation.

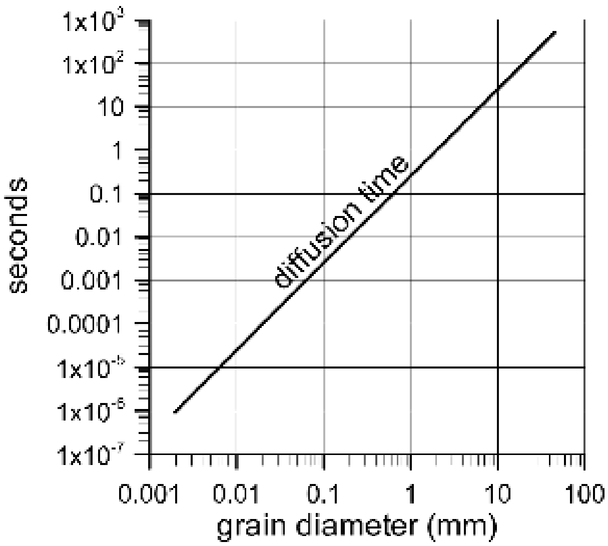

Figure 8. Dependence of diffusion time (thermal equilibrium time) on particle diameter (thermal diffusivity k $=10^{-6}$ m^2s^{-1}).

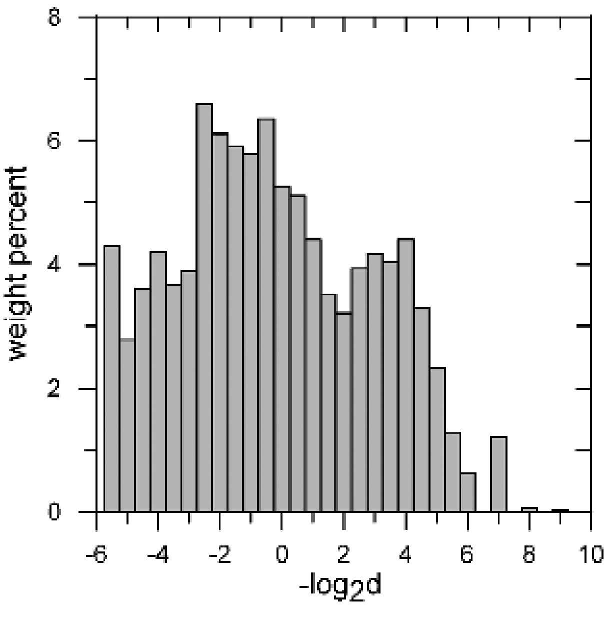

Figure 7. Grain size distribution of samples from Gjálp, collected on June 10 1997 when ice melting after the eruption had exposed the uppermost 40 m of the subglacially-formed edifice. The sites where the samples were collected correspond to the upper slopes of the edifice of stage 3 on Figure 2. The grain size d is in mm. The largest particles had diameters of 45 mm, the mean was just under 2 mm while the smallest grain size was $2x10^{-3}$ mm.

Since the heat loss is related to the square of the diameter, the heat flux is strongly dependent on the size of grains into which the magma fragments (Figure 8). Samples were taken from the upper slopes of the Gjálp edifice in June 1997 (Figure 7). It is possible that some of this material was produced in surtseyan explosive activity during the latter stages of the eruption. Thus, the top part may not be fully representative of the bulk of the edifice. However, gravity modeling of the edifice indicates that it is mainly made of hyaloclastite [*Gudmundsson et al.*, 2002] and the calculated heat fluxes [*Gudmundsson et al.* 1997; 2003] suggest that fragmentation was a major process in the eruption. The grain sizes also conform to reported analyses of subglacially-erupted hyaloclastites [*Werner and Schmincke*, 1999]. The Gjálp data are therefore used here to demonstrate cooling rates during magma fragmentation. For the Gjálp data, the mean diameter of the particles is just under 2 mm; such a particle has a heat diffusion time of 1 s. The diffusion time for the largest particles sampled for Gjalp is $>10^2$ s, but it is $<10^{-3}$ s for the finest material (Figure 9).

Heat Loss During Magma Fragmentation

The absence of good constraints on meltwater temperature, the importance of heat transfer by steam, and the exact

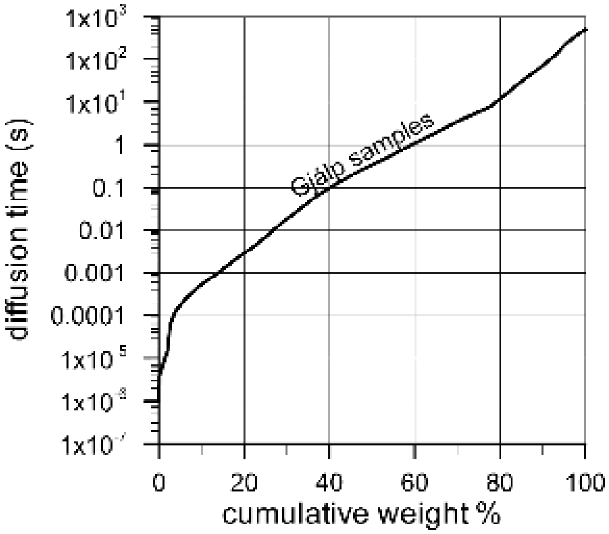

Figure 9. Diffusion time of the Gjálp samples; the curve shows the weight percent of the samples that have diffusion times equal to or smaller than the value on the vertical axis.

nature of the fragmentation process makes the following discussion somewhat speculative. However, it may be of use to consider likely processes and their possible consequences for heat transfer and temperatures in a growing subglacial volcano. In order to get some idea of the temperature of the glass particles as they settle and get buried in the volcanic pile, we need to consider the length of time that the particles are exposed to meltwater-temperatures and also the temperature of the fluid (meltwater/meltwater+steam?) in the subglacial vault above the vent. The exposure time of the particles can be divided into two stages. Firstly, there is the settling time, i.e. the time it takes a particle to rise out of the vent and then sink through the water body onto the slopes of the edifice, possibly impacting the overlying ice roof in the process. Secondly, there is the burial time, the time it takes a particle to get buried in the growing volcanic pile once it has settled. Burial will not stop heat transfer but should slow it down, provided all the heat has not been lost to the overlying fluid.

If fragmentation occurs mostly through non-explosive granulation, the rise of particles above the vent may be only a few meters; their transport may mainly be in a turbulent to laminar gravity current down the sides of the edifice onto which they settle [*Smith and Batiza*, 1989]. For high-concentration flows of hot particles, the temperature of the fluid part of the current could be close to that of saturated steam. In that case the particles may never be exposed to temperatures less than 200–250°C (boiling-point corresponding to water depths of 200–500 m, e.g. *Ingebritsen and Sanford* [1998]). Burial time (Figure 10) would be of order 10 s, suggesting that particles comprising the coarsest 20% of the total mass would retain a significant fraction of their heat as they get buried (Figure 9).

If fragmentation is explosive, the particles formed will be ejected out of the vent by rapid expansion of steam [e.g. *Zimanowski*, 1998]. The settling velocity [*Cashman and Fiske*, 1991] would be 0.1–1 m s^{-1}, with the largest grains having the highest velocity. The settling time would be 10^1–10^2 s for a water layer with a thickness of order 10 m. Only the very largest particles (~10% of mass) would have settling times short compared to their diffusion time (Figure 9); hence, only about 10% of the mass would retain a significant part of its heat by the time it is added to the top of the pile.

The above considerations indicate that probably 85–90% of the heat of the particles is lost to the fluid in the subglacial vault prior to burial in the volcanic pile. The heat released to the fluid is subsequently used to melt ice. If possible complications due to pulsating activity are ignored, some sort of equilibrium should exist between eruption rate, vault-fluid temperature and ice melting rate. Ultimately, the ambient temperature is that of the ice (~0°C) but the fluid (meltwater/meltwater+steam) in which the heat of the magma is released is hotter. Although the mean temperature of the meltwater as it flowed from the eruption site at Gjálp

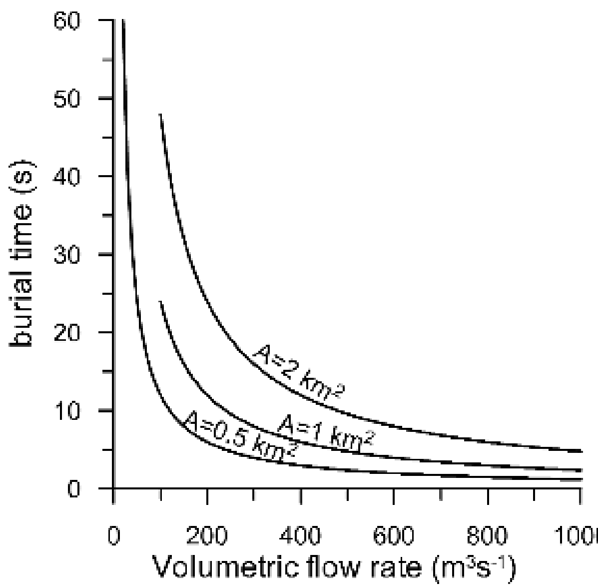

Figure 10. Time of deposition of a 4 mm thick layer as a function of volumetric magma flow rate dV_m/dt. The time is taken as equal to $Ah(1-\varphi)/(dV_m/dt)$ where A is the area over which the material is dispersed, h is layer thickness, φ is porosity of the volcanic pile (taken as 0.4).

was 20°C, its temperature was probably considerably higher during the height of the eruption [*Gudmundsson et al*., 1997; 2003]. *Höskuldsson and Sparks* [1997] calculated temperatures as high as 100°C in a convecting water body for heat fluxes that were lower than observed at Gjálp. Meltwater temperatures of 50–100°C or even higher in vigorous eruptions do, therefore, seem plausible, and gravity currents down the slope of the edifice may be considerably warmer. Thus, although a particle may have reached thermal equilibrium with the vault fluid, it may still retain some 10% of its heat relative to the ice. The heat loss from the fragmented material to the surroundings prior to burial could therefore amount to 75–80% of the available magmatic heat. Considering also that the energy used to fragment the magma, amounting to a few percent of the total thermal energy [*Zimanowski*, 1998; *Buttner and Zimanowski*, 1998], is not available for melting, 70-80% may be a realistic estimate of the instantaneous efficiency of heat transfer (equation 4) for a grain size distribution like that obtained at Gjálp (Figure 11). A similar estimate of 63–77% was obtained as an average for the Gjálp eruption by calorimetric considerations based on ice melting [*Gudmundsson et al.* 2003].

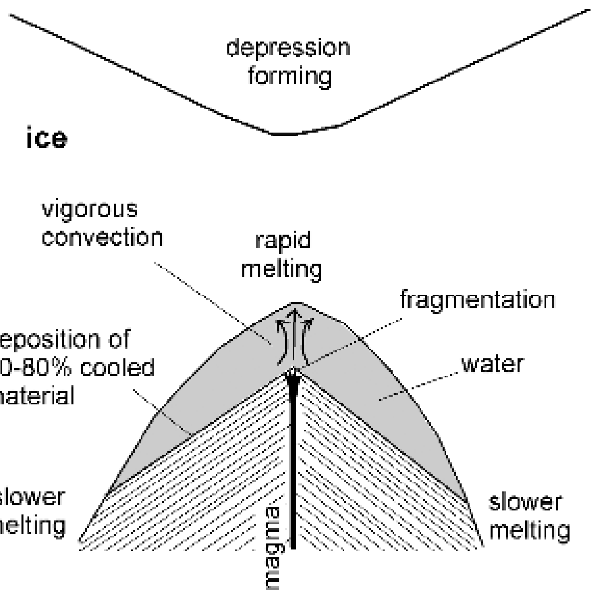

Figure 11. A schematic illustration of how heat transfer, ice melting, magma fragmentation and material deposition within a water-filled vault may occur during a subglacial volcanic eruption. The layering indicated represents material deposited at the same time; it need not reflect pulsating activity. The dip of layers is uncertain but the volcanic material may bank up against ice walls at the vault margins.

Flow-Foot Breccias

The treatment of heat transfer has here focused on the formation of pillow lava (stratigraphic unit 1) and magma fragmentation prior to emergence of a volcano in an englacial lake (unit 2). The basic mechanisms of heat transfer during formation of flow-foot breccias (unit 4) should be the same. Pillows and pillow fragments are abundant and heat fluxes to meltwater from breccias may therefore be intermediate between those suggested for pillow lavas and fragmentation.

Rhyolitic Eruptions

Studied eruptions have been basaltic, or the magma has been of mildly intermediate composition. No rhyolitic subglacial eruptions have been observed, but subglacially-erupted rhyolites are found in Iceland, notably in the Torfajökull region [*Furnes et al.*, 1980; *Tuffen et al.* 2001, 2002]. Magma composition and its effect on viscosity and magma temperature may affect fragmentation mechanisms [*Zimanowski*, 1998]. Due to lower magma temperature, and lower latent heat of fusion and heat capacity of acidic rock, the heat content of rhyolitic magma is only 60-70% of that of basaltic magma [e.g. *Höskuldsson and Sparks* 1997; *Spera*, 2000] Thus, rhyolitic magma has less ice-melting potential. Moreover, rhyolitic pillows are considerably larger than basaltic ones [*Furnes et al.*, 1980; *Tuffen et al.* 2001, 2002], and their cooling times are therefore considerably longer and heat fluxes correspondingly lower [*Höskuldsson and Sparks*, 1997]. However, melting and heat transfer rates for magma fragmentation should be somewhat reduced, but of the same order as in a basaltic eruption, given the same magma eruption rate.

MECHANISM OF MAGMA FRAGMENTATION

Experimental work on phreatomagmatic explosions has shown that fragmentation during magma-water interaction results from processes ranging from non-explosive granulation through moderately energetic steam explosions to very high-energy thermohydraulic explosions caused by molten fuel-coolant interactions (MFCI) [*Zimanowski et al.* 1997; *Zimanowski*, 1998]. The MFCI explosions result in very intense fragmentation, and the fines in the Gjálp samples (Figure 7) may have been formed in such explosions. The abundance of coarser material indicates that less-energetic fragmentation was also important. It should be kept in mind that the samples at Gjálp represent the top part of the edifice, where the overlying water column amounted to only a few tens of meters. At least for basaltic melts it is likely that at higher confining pressures, during early phases of subglacial eruptions where fragmentation takes place, that non-explosive granulation is a more important style of activity.

THERMAL EFFICIENCY AND EDIFICE TEMPERATURE

The difference in thermal efficiency as defined by (4) for the two processes of effusive activity (eruption of pillow-lava) and magma fragmentation is quite marked (Figure 12). The efficiency during pillow lava formation is 10–45%, the precise value being dependent mainly on exposure time of the pillow lavas. The efficiency of fragmentation may be more variable than shown, although the range for magma fragmentation used on Figure 12 is based on the extreme bounds obtained for the Gjálp data.

The contrasts in heat transfer rates during magma effusion and magma fragmentation lead to considerable differences in residual heat left in the pile after deposition in the growing volcanic edifice. A rapidly growing pile of pillow lava may have pillows with hot and partly molten cores. The heat in this pile will diffuse out of the pillows and transported out of the edifice by hydrothermal convection. The heat transfer rate of a subglacial or subaqueous effusive eruption will be smoothly varying, and may continue at levels comparable to those observed during an eruption, for a time considerably longer than the duration of the eruption. The peak in heat transfer from magma to water or ice may be much lower than the peak in influx of heat with the

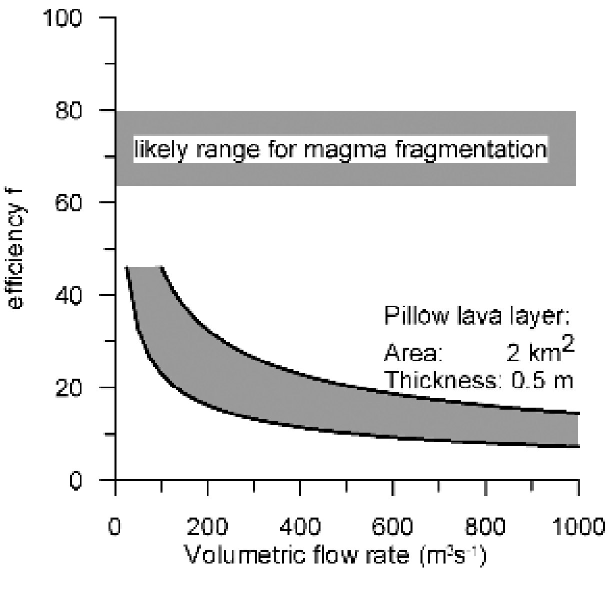

Figure 12. Bounds on efficiency of heat exchange for magma fragmentation (63–80%) and deposition of a layer of pillow lava (10–45%).

magma. In contrast, for an eruption where efficient magma fragmentation is a dominant process, heat transfer rates to water or ice at any time during the eruption will be of similar magnitude as the influx of magmatic heat. Heat flux will fall by orders of magnitude as soon as eruption stops.

An interesting consequence of the 70–80% efficiency for eruptions that produce grain-size populations similar to that of Gjálp's deposits is that temperature in the volcanic pile will initially be highly variable over distances of centimeters to decimeters. The largest grains may have temperatures of a few hundred degrees while the surrounding mass of finer material may have temperatures of 50–100°C. The efficiency suggests that a mean temperature of the pile may be 250–300°C, some 150–200°C higher than that of the surrounding meltwater. It is likely that meltwater will quickly percolate into the edifice and lead to mean temperatures of tephra and pore water of perhaps 120–180°C. If magma is intruded into the edifice during the eruption it will increase the temperature of the pile. Heating caused by intrusions during the final stages of activity in the 1963–1967 Surtsey eruption is considered to have led to the formation of a geothermal area on the island [*Axelsson et al.* 1982]. After an eruption ends, heat flux out of the pile will be relatively slow, a few orders of magnitude lower than during eruption and initial cooling. This residual heat in the pile will be eventually be removed by conduction, hydrothermal convection and groundwater advection.

In terms of heat transfer mechanisms operating, the difference between subglacial and other subaqueous eruptions should be small. As a consequence, similar heat fluxes and heat transfer rates are to be expected in both cases. Submarine eruptions will be exposed to the very large thermal reservoir of the ocean; the eruption may not alter the temperature of the surroundings to any degree [e.g. *Thorarinsson*, 1967]. Within lakes, eruptions may lead to considerable heating of the lake water. For example, in a lake with a volume of 1 km^3 the release of 50% of the thermal energy of 0.1 km^3 of basaltic magma erupted at its bottom would heat the body of water by about 40°C.

CONCLUSIONS

Heat transfer rates and heat fluxes during magma-water interaction may be studied by applying calorimetry to ice-melting rates in subglacial eruptions. For fully subaqueous/subglacial eruptions the form of eruption may be either effusive, usually leading to formation of pillow lavas, or the magma may fragment, leading to the formation of a pile of volcanic glass. Cooling models indicate heat fluxes several times lower for pillow lava formation, and heat exchange efficiencies of 10–45%. In contrast, both observations from

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the Gjálp eruption, and results from a simple cooling model indicate that fragmentation of magma into glass particles leads to a short-term efficiency of heat transfer of 70–80% and the formation of a "cold" volcanic pile that may have an average temperature of 150–300°C immediately after deposition. The remaining heat in pillow lavas (55–90% of the total magmatic heat) and a pile of fragmented glass (20–30% of the total) is lost to the surroundings by hydrothermal convection at a much slower rate than during the fast initial cooling.

Although simple heat transfer models have been applied to subglacial and subaqueous eruptions, and recent eruptions have provided important data, more work is required to gain further insight into heat transfer processes and the thermal histories of subaqueously-formed volcanic edifices. Grain size analysis and analysis of grain morphology [*Buettner et al.*, 1999] in hyaloclastites at different levels in hyaloclastite mountains may provide insight into how mechanisms of fragmentation vary with ambient pressure. Cooling models of pillow lavas need refinement, taking into account the cooling rates of the glassy outer margins and the effect of cooling cracks in pillows. In this way further understanding of ice-volcano interaction may be achieved by work on deposits of subaqueous and subglacial eruptions, field observations of eruptions, and both experimental and theoretical work.

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REFERENCES

Allen, C.C., Icelandic subglacial volcanism: thermal and physical studies, *J. Geol*., 88, 108-117, 1980.

Áskelsson, J., On the last eruptions in Vatnajökull, *Societas Scientarium Islandica*, 18, 1-68, 1936.

Axelsson, G., V. Stefánsson, G. Gudmundsson, and B. Steingrímsson, Thermal condition of Surtsey, *Surtsey Research Progress Report*, IX, 102-110, 1982.

Bacon, C.R., High temperature heat content and heat capacity of silicate glasses: Experimental determination and a model for calculation, *American Journal of Science,* 277, 109-135, 1977.

Bemmelen, R.W. van, and M.G. Rutten, *Tablemountains of northern Iceland*, E.J. Brill, Leiden, 217 pp., 1955.

Björnsson, H., Subglacial water reservoirs, jökulhlaups and volcanic eruptions, *Jökull*, 25, 1-11, 1975.

Björnsson, H., Hydrology of ice caps in volcanic regions, *Societas Scientarium Islandica*, 45, Reykjavík, pp. 1-139, 1988.

Björnsson, H., F. Pálsson, G.E. Flowers, and M.T. Gudmundsson, The extraordinary 1996 jökulhlaup from Grímsvötn, Vatnajökull, Iceland, *Eos Trans. AGU*, 82, 47, 2001.

Buttner, R., and B. Zimanowski, Physics of thermohydraulic explosions, *Phys. Rev*, E, 57, 5726-5729, 1998.

Buttner, R., P. Dellino, and B. Zimanowski, Identifying magma-water interaction from surface features of ash particles, *Nature*, 401, 688-690, 1999.

Carslaw, H.S., and J.C. Jaeger, *Conduction of heat in solids*, 2nd edition, Oxford University Press, Oxford, 510 pp., 1959.

Cashman, K.V., and R.S. Fiske, Fallout of pyroclastic debris from submarine volcanic eruptions, *Science*, 253, 275-280, 1991.

Colgate, S.A., and Th. Sigurgeirsson, Dynamic mixing of water and lava, *Nature*, 244, 552-555, 1973.

Einarsson, T., Physical aspects of sub-glacial eruptions, *Jökull*, 16, 167-174, 1966.

Fisher, R.V., and H.U. Schmincke, *Pyroclastic rocks*, Springer, Berlin. 472 pp., 1984.

Furnes, H., I.B. Fridleifsson, and F.B. Atkins, Subglacial volcanics —on the formation of acid hyaloclastites, *J. Volc. Geoth. Res*., 8, 95-110, 1980.

Griffiths, R.W., and J.H. Fink, Solidification and morphology of submarine lavas: a dependence on exstrusion rate, *J. Geophys. Res*., 97, 19729-19737, 1992.

Grönvold, K., and H. Jóhannesson, Eruption in Grímsvötn: Course of events and chemical studies of the tephra, *Jökull*, 34, 1-11, 1984.

Gudmundsson, M.T., and H. Björnsson, Eruptions in Grímsvötn 1934-1991, *Jökull*, 41, 21-46, 1991.

Gudmundsson, M.T., F. Sigmundsson, and H. Björnsson, Ice-volcano interaction of the 1996 Gjálp subglacial eruption, Vatnajökull, Iceland, *Nature*, 389, 954-957, 1997.

Gudmundsson, M.T., F. Pálsson, H. Björnsson, and Th. Högnadóttir, The hyaloclastite ridge formed in the subglacial 1996 eruption in Gjálp, Vatnajökull, Iceland: present day shape and future preservation, in *Volcano-Ice interaction on Earth and Mars, Geological Society of London Spec. Publ.*, 202 edited by J.L. Smellie and M. Chapman, pp. 319-335, 2002.

Gudmundsson, M.T., F. Sigmundsson, H. Björnsson, and Th. Högnadóttir, The 1996 eruption at Gjálp, Vatnajökull ice cap, Iceland: Efficiency of heat transfer, ice deformation and subglacial water pressure, Submitted to *Bull. Volc.*, 2003.

Hickson, C.J., Physical controls and resulting morphologic forms of Quaternary ice-contact volcanoes in western Canada, *Geomorphology*, 32, 239-261, 2000.

Höskuldsson, Á., and R.S.J. Sparks, Thermodynamics and fluid dynamics of effusive subglacial eruptions, *Bull. Volc.*, 59, 219-230, 1997.

Ingebritsen, S.E., and W.E. Sanford, *Groundwater in geologic processes,* Cambridge University Press, 341 pp., 1998.

Jakobsson, S.P., Outline of the petrology of Iceland, *Jökull*, 29, 57-73, 1979.

Jóhannesson, T., Propagation of a subglacial flood wave during the initiation of a jökulhlaup, *Hydrological Sciences*, 47, 417-434, 2002.

Jones, J.G., Intraglacial volcanoes of the Laugarvatn region, southwest Iceland - I., *Quarterly Journal of the Geol. Soc. Lond.*, 124, 197-211, 1969.

Jones, J.G., and Nelson, P.H.H., The flow of basalt lava from air into water—its structural expression and stratigraphic significance, *Geol. Mag.*, 107, 13-19, 1970.

Kjartansson, G., The moberg formation, *Geografiska Annaler*, 41, 139-143, 1959.

Larsen, G., A brief overview of eruptions from ice-covered and ice-capped volcanic systems in Iceland during the past 11 centuries: frequency, periodicity and implications, in *Volcano-Ice Interactions on Earth and Mars*, *Geological Society Of London Spec. Publ*., 202, edited by J.L. Smellie, and M.G. Chapman, pp. 81-90, 2002.

Larsen, G., M.T. Gudmundsson, and H. Björnsson, Eight centuries of periodic volcanism at the center of the Iceland Hot Spot revealed by glacier tephrostratigraphy, *Geology*, 26, 943-946, 1998.

Major, J.J., and C.H. Newhall, Snow and ice perturbation during historical volcanic eruptions and the formation of lahars and floods, *Bull. Volc*., 51, 1-27, 1989.

Mathews, W.H., “Tuyas”. Flat topped volcanoes in northern British Columbia, *American Journal of Science*, 245, 560-570, 1947.

Moore, J.G., Mechanism of formation of pillow lava, *Am. Sci*., 63, 269-277, 1975.

Sigmundsson, F., M.T. Gudmundsson, G. Sverrisdóttir, N. Óskarsson, and P. Einarsson, Style of basaltic eruptions in shallow ice/water depths: the 1998 Grímsvötn eruption, Vatnajökull ice cap, Iceland, *Eos Trans. AGU*, 46, 1084, 1999.

Smellie, J.L., Lithostratigraphy of Miocene-Recent, alkaline volcanic fields in the Antarctic Peninsula and eastern Ellsworth Land, *Antarctic Science*, 11, 347-363, 1999.

Smellie, J.L., and I. P. Skilling, Products of volcanic eruptions under different ice thicknesses: two examples from Antarctica, *Sed. Geol.*, 91, 115-129, 1994.

Smith, T.L., and R. Batiza, New field and laboratory evidence for the origin of hyaloclastite flow on seamount summits, *Bull. Volc.*, 51, 96-114, 1989.

Spera, F.J., Physical properties of magmas, in *Encyclopaedia of Volcanoes, Academic Press*, edited by H. Sigurdsson, pp. 171-190, 2000.

Steinthorsson, S., B.S. Hardarson, R.M. Ellam, and G. Larsen, Petrochemistry of the Gjálp 1996 subglacial eruption, Vatnajökull, SE Iceland, *J. Volc. Geoth. Res*., 98, 79-90, 2000.

Thorarinsson, S., *Surtsey, the new Island in the North Atlantic*, Viking Press, New York, 100 pp, 1967.

Tryggvason, E., Earthquakes, jökulhlaups and subglacial eruptions, *Jökull*, 10, 18-22, 1960.

Tuffen, H., J.S. Gilbert, and D.W. McGarvie, Products of an effusive subglacial rhyolite eruption Bláhnúkur, Torfajökull, Iceland, *Bull. Volc.*, 63, 179-190, 2001.

Tuffen, H., H. Pinkerton., D.W. McGarvie, and J.S. Gilbert, Melting at the glacier base during a small-volume subglacial rhyolite eruption: evidence from Bláhnúkur, Iceland. *Sed. Geol.*, 149, 183-198, 2002.

Turcotte, D.L., and G. Schubert, *Geodynamics*, John Wiley & Sons, New York, 450 pp., 1982.

Werner, R., and H.-U. Schmincke, Englacial vs lacustrine origin of volcanic table mountains: evidence from Iceland. *Bull. Volc.*, 60, 335-354, 1999.

Wilson, L., and J.W. Head, Heat transfer and melting in subglacial basaltic volcanic eruptions: implications for volcanic deposit morphology and meltwater volumes, in *Volcano-Ice Interactions on Earth and Mars*, *Geological Society Of London Spec. Publ*., 202, edited by J. L. Smellie, and M. G. Chapman, pp. 5-26, 2002.

Wohletz, K.H., Mechanisms of hydrovolcanic pyroclast formation: grain size, scanning electron microscopy, and experimental studies, *J. Volc. Geoth. Res*., 17, 31-64, 1983.

Zimanowski, B., Phreatomagmatic explosions, in *From magma to tephra, Elsevier, Amsterdam*, edited by A. Freundt and M. Rosi, pp. 25-54, 1998.

Zimanowski, B., R. Buttner, and V. Lorenz, Premixing of magma and water in MFCI experiments, *Bull. Volc.*, 58, 491-495, 1997.

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