Melting of the glacier base during a small-volume subglacial rhyolite eruption: evidence from Bláhnúkur, Iceland.

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Abstract

Although observations of recent volcanic eruptions beneath Vatnajökull, Iceland have improved understanding of ice deformation and meltwater drainage, little is known about the processes that occur at the glacier base. We present observations of the products of a small-volume, effusive subglacial rhyolite eruption at Bláhnúkur, Torfajökull, Iceland. Lava bodies, typically 7 m long, have unusual conical morphologies and columnar joint orientations that suggest emplacement within cavities melted into the base of a glacier. Cavities appear to have been steep-walled and randomly distributed. These features can be explained by a simple model of conductive heat loss during the ascent of a lava body to the glacier base. The heat released melts a cavity in the overlying ice. The development of vapour-escape pipes in the waterlogged, permeable breccias surrounding the lava allows rapid heat transfer between lava and ice. The meltwater formed percolates into the breccias, recharging the cooling system and leaving a steam-filled cavity.

The slow ascent rates of intrusive rhyolitic magma bodies provides ample time for a cavity to be melted in the ice above, even during the final 10 m of ascent to the glacier base. An equilibrium cavity size is calculated, at which melting is balanced by creep closure. This is dependent upon the heat input and the difference between glaciostatic and cavity pressure. The cavity sizes inferred from Bláhnúkur are consistent with a pressure differential of 2-4 MPa, suggesting that the ice was at least 200 m thick. This is consistent with the volcanic stratigraphy, which indicates that the ice exceeded 350 m in thickness.

Although this is the first time that a subglacial cavity system of this type has been reconstructed from an ancient volcanic sequence, it shares many characteristics with the modern firn cave system formed by fumarolic melting within the summit crater of Mount Rainier, Washington. At both localities it appears that localised heating at the glacier base has resulted in heterogeneous melting patterns. Despite the different rheological properties of ice and firn, similar patterns of cavity roof deformation are inferred. The development of low-pressure subglacial cavities in regions of high heat flux may influence the trajectory of rising magma, with manifold implications for eruptive mechanisms and resultant subglacial volcanic landforms.

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Introduction

Considerable progress has been made in recent years in understanding subglacial volcanic phenomena. Observations of the 1996 Gjálp eruption beneath Vatnajökull, Iceland have shed light on ice deformation, melting rates and meltwater drainage patterns that accompany small-volume subglacial basaltic eruptions (Guðmundsson et al., 1997; Alsdorf & Smith 1999). Meanwhile, detailed lithofacies analysis of ancient subglacial volcanic sequences has helped to reconstruct the mechanisms of much larger-volume basaltic eruptions (e.g. Smellie & Skilling, 1994; Werner et al., 1996; Smellie, 1999). The principal hazards during subglacial volcanic eruptions are caused by the accumulation of meltwater, which can interact explosively with rising magma, and which may drain suddenly, causing catastrophic floods (Major & Newhall, 1989; Chapman et al., 2000). It is thus of paramount importance to understand the mechanisms and rates of melting that occur during different types of subglacial eruption. A vexed question is the mechanism of energy exchange between magma and ice. To date, little is known about processes that occur at the ice-magma interface. This paper presents observations from Bláhnúkur, a subglacial rhyolite volcano at Torfajökull, Iceland. Lava bodies at Bláhnúkur may provide a unique record of melting patterns at the glacier base during a small-volume rhyolite eruption.

Melting of ice during subglacial eruptions

Subglacial volcanic eruptions are dynamic multi-component systems, in which ice, water, steam and intact and fragmented magma can interact. Various mechanisms of energy exchange between magma and ice may occur, depending upon the magma properties, eruption rate, and environmental factors (e.g. presence of meltwater, effective pressure, architecture of subglacial cavities). Since these environmental factors are themselves largely determined by the local melting rate (Hooke, 1984; Hooke et al, 1990; Hoskuldsson & Sparks, 1997; Fountain & Walder, 1998), there is coupling between eruptive and melting mechanisms (Tuffen et al., 2001). Consequently, the style of magma-ice interaction may be both spatially and temporally heterogeneous, being sensitive to fluctuations in eruption rate and cavity position (Tuffen et al., 2001; Tuffen, 2001).

To date, there have been few detailed studies of energy exchange mechanisms in a subglacial setting (Allen 1980; Hoskuldsson & Sparks, 1997; Lescinsky & Fink, 2000). Hoskuldsson & Sparks develop a conceptual model, in which energy is transferred from effusive lava 'pillows' to ice via convecting meltwater. It leads to low estimates of ice melting rates: 10^{-5} m s^{-1} in basaltic and 10^{-6} m s^{-1} in rhyolitic eruptions. Considerably higher melting rates of 10^{-3} m s^{-1} occurred during the 1996 Gjálp eruption in Iceland (Guðmundsson et al., 1997). This is though to be due to rapid heat exchange between fragmenting magma and meltwater, and rapid heat exchange between the convecting, heated meltwater and the surrounding ice. To improve the understanding of energy exchange during a specific phase of a ancient subglacial eruption, it is necessary to construct models that account for all features of the lithofacies produced. Recent fieldwork on subglacial lithofacies in Iceland suggests that convecting meltwater models (Hoskuldsson & Sparks, 1997; Lescinsky & Fink, 2000) are inappropriate for effusive rhyolitic eruptions, since there is evidence of meltwater drainage, rather than accumulation (Tuffen et al., 2001). This comes from 1) the presence of juvenile sediments at the base of the subglacial rhyolite sequence thought to be have been deposited by flowing water, and 2) the massive, poorly-sorted nature of the bulk of the fragmental deposits, which indicates that reworking within standing water did not occur (c.f. Skilling, 1994). However, little is known about the thermodynamics of alternative melting mechanisms (e.g. convecting steam or abrasion of ice by ash particles during subglacial explosions).

Evidence from ice caves in volcanic areas

Observations of subglacial volcanic eruptions can help to reconstruct ice-melting and drainage patterns (Guðmundsson et al., 1997). However, the glacier base is generally obscured during an eruption, and extreme hazards prevent researchers from entering any subglacial cavities. Thankfully, useful insights into melting patterns and cavity morphologies can be gained from glacier-bearing volcanoes currently in a quiescent phase. Fumarole-melted ice and firn caves are known to exist at Mount Rainier, Washington, Mount Baker, Washington, Mount Wrangell, Alaska and Mount Erebus, Antarctica (Kiver & Steele, 1975). Of these, the summit firn caves of Mount Rainier are by far the best documented (e.g. Kiver & Mumma, 1971; Kiver & Steele, 1975; Zimbelman et al., 2000). A cave system totalling 1.9 km in length has been mapped to date within the summit craters (Fig. 1), beneath snow and firn up to 120 m thick. Although the mechanical properties of firn differ from those of ice (e.g. Paterson, 1994), the patterns of basal melting and deformation are likely to be similar. Kiver & Steele (1975) made the following observations:

1. Meltwater drips almost continuously from firn walls and ceilings, but percolation into the permeable substrate appears to prevent the accumulation of standing water. Drainage occurs because the water table is significantly below the base of the firn.
2. Rates of cavity closure by firn deformation and enlargement by fumarolic melting were in approximate equilibrium during the period 1970-1974. Deformation and melting rates were estimated at 2-3 m per year ($\sim 10^{-7}$ m s^{-1}). Deformation was entirely ductile, despite a pressure difference of up to 1 MPa between the 'glaciostatic' pressure of the firn roof and the cavity at atmospheric pressure.
3. The position of cavities is strongly influenced by the spatial distribution of basal heating, with individual cavities related to single fumaroles. However, some sections of cavities appear to be maintained by motion of warm air.
4. Cavities occur as slope-normal 'perimeter passages' and slope-parallel cavities (Fig. 1). Both types are roughly semi-circular in cross section and bounded by steep, scalloped firn walls. There is well-established airflow from one cavity to the next.
5. Talus on cavity floors is close to the angle of repose (30-40°), and highly unstable.
6. Smooth-walled conical 'steam cups' develop above the most powerful fumaroles.

Experimental and numerical simulations of cavity evolution

Hoskuldsson & Sparks (1997) used blocks of ice and PEG wax and a basal heat source to simulate the formation of subglacial cavities during volcanic eruptions. Cavities entirely filled with convecting meltwater had conical morphologies, whereas those in which an air gap developed became ‘pancake-shaped’. This reflects slower melting rates of ice in contact with air than ice in contact with convecting meltwater. In a study of melting during 'normal' conditions (i.e. no geothermal heat), finite element analysis was used to investigate the evolution of a subglacial conduit carrying a variable meltwater flux (Cutler, 1998). During periods of high meltwater discharge, conduits may become only partially filled with meltwater. This acts to focus melting low on the walls, forming broad, low conduits (c.f. Hooke, 1984). Pressure in conduits may become atmospheric, if a hydrological connection is established with the glacier snout, or to the glacier surface via a system of fractures. Thermal energy released in a subglacial eruption may cause heating or even boiling of meltwater, depending upon the relative rates of magma-water and meltwater-ice energy exchange (Hoskuldsson & Sparks, 1997) and the cavity pressure. Melting by heated meltwater may be an order of magnitude more rapid than melting caused purely by the mechanical energy of flowing meltwater at the pressure melting point (Clarke, 1982). Little is currently understood about subglacial melting mechanisms during eruptions. The shape of evolving cavities may be controlled by the patterns of convection within meltwater or steam in the cavity (Hoskuldsson & Sparks, 1997).

Cavity pressure and hydrology

Hoskuldsson & Sparks (1997) predicted that eruption of rhyolite within a water-filled subglacial cavity would cause positive pressure changes. This is because the volume reduction upon melting of ice is insufficient to accommodate the volume increase caused by injection of magma into the cavity. However, overpressure could not be sustained if the subglacial substrate is permeable, as meltwater would be driven into the substrate, with a flux proportional to the local pressure gradient (Dullien, 1992). Poorly-consolidated fine-grained ash, which forms the substrate at Bláhnúkur, has a relatively high permeability (Ascolese et al., 1993), which would favour re-equilibration of subglacial pressure by porous flow.

The position of the water table at the glacier base controls patterns of water flow (Björnsson 1988), including flow between subglacial cavities and the glacier substrate. If the water table is below the subglacial volcanic edifice, low pressure air-filled cavities may develop, and the edifice will tend to be rather dry, with only minor magma-water interaction. A water table at or above the edifice will favour the development of water-filled cavities at higher pressures. The latter scenario is the more likely for eruptions beneath temperate glaciers and ice sheets such as those in Iceland (e.g. Björnsson, 1988).

Low-pressure conditions are known to develop in subglacial drainage conduits close to the glacier snout during periods of high water flux (Hooke, 1984; Fountain & Walder, 1998). This is due to the melting rate exceeding the deformation rate, leading to conduit enlargement and the presence of an air gap, which will be at atmospheric pressure if there is a hydraulic connection with the glacier snout (Hooke, 1984). Rapid melting by geothermally-heated meltwater will strongly favour the establishment of low-pressure conditions (Clarke, 1982). A vexed question is whether low-pressure conditions may develop in the vent area if a subglacial eruption occurs far from the glacier snout, since there is the problem of backfilling from regions of high subglacial water pressure (Hooke, 1984). If so, pressure in cavities adjacent to such low-pressure conduits may be significantly less than glaciostatic.

It is not currently possible to estimate the magnitude of pressure gradients within the permeable substrate, due to uncertainties in the flow regime, hence no temporal or spatial variations in cavity pressure will be considered. Furthermore, the position of the water table during the eruption is unclear. Hence, for the purposes of the model, we assume that cavity pressure may vary between glaciostatic (c. 2-4 MPa) and atmospheric (0.1 MPa).

Deformation of ice above subglacial cavities

If the pressure in a subglacial cavity is less than the glaciostatic pressure, the ice will deform (e.g. Paterson, 1994). Ice deformation may be brittle or ductile, depending upon the strain rate, confining pressure and the ice structure (e.g. Jones, 1982; Mizuno, 1998). Ductile deformation is best approximated by Glen's flow law (Glen, 1955), in which

$$\varepsilon'_{xy} = A\tau_{xy}^{n} \tag{1}$$

where ε' is the shear strain rate, τ is the shear stress, n is a constant (~3) and A is determined by the properties of the ice. Nye (1953) showed that the roof of a hemispherical cavity of radius r beneath a glacier of thickness h will deform in a ductile manner at a rate r'_c given by the relation

$$r'_c = -r\left[\frac{\Delta P}{nB}\right]^n \tag{2}$$

where effective pressure ΔP = glaciostatic pressure (P_i) – cavity pressure (P_c), n is Glen's flow law constant, ~3, and B is an ice viscosity parameter. However, this relation only applies when $r \ll h$. When the cavity size approaches that of the glacier thickness, high strain rates trigger brittle failure of the ice, as observed during the 1996 Gjálp eruption (Guðmundson et al., 1997; Alsdorf & Smith, 1999). Since this paper considers cavities of radius <10 m beneath ice 100-1000 m in thickness, it is assumed that $r \ll h$, and ice deformation will thus be ductile and approximated by equation (2). Some justification for the assumption of ductile ice deformation comes from observations of the evolution of ice cauldrons above regions of geothermal melting (MT Guðmundsson, personal communication). The ice surface is often smooth and unfractured during the early stages of cauldron formation, indicating that ductile deformation of the surface is occurring. The glacier base is likely to also deform in a ductile manner at this stage, as ductile behaviour is favoured by a high confining pressure (Jones, 1982). Cavity sizes during the initial stage of ice cauldron development may be similar to those considered in this paper.

Implications for cavity evolution during subglacial eruptions

An important conclusion gained from Mount Rainier is that subglacial melting may be strongly heterogeneous, and controlled by the distribution of heat at the glacier base. Furthermore, meltwater accumulation within subglacial cavities may be prevented if the substrate is sufficiently permeable and the water table is below the glacier (firn) base. Evolving subglacial cavities may attain an equilibrium size and shape, determined by the balance between inward creep deformation and melting. The rate of melting is influenced by the mechanism of energy transfer from magma to ice (e.g. water/steam/air/ash slurry), which is in turn controlled by the subglacial hydrological system and eruption mechanisms. In order to reconstruct the melting processes that occurred during the subglacial volcanic eruption at Bláhnúkur, it is thus necessary to evaluate the field evidence for:

1. Morphology and distribution of subglacial cavities
2. Accumulation or drainage of meltwater
3. Mode of heat loss from magma (e.g. fragmentation, cooling and crystallisation)
4. Presence of a vapour phase.

Field observations

Geological setting of Bláhnúkur

Torfajökull central volcano, the largest silicic centre in Iceland, is located in south-central Iceland at the southern terminus of the eastern rift zone (Fig. 2a). Torfajökull has erupted >250 km^3 of mildly alkaline to peralkaline rhyolites over the last ~1 Ma (McGarvie, 1984). Eruptions during glacial and interglacial periods have produced a variety of volcanic landforms (McGarvie, 1985). The youngest subglacial eruption at Torfajökull is thought to have occurred at Bláhnúkur, close to Landmannalaugar in the north of the complex (Fig. 2a, 2b).

Bláhnúkur is a 350 m-high conical edifice consisting of a veneer of subglacial rhyolite >50 m thick that overlies older, altered rhyolite and till. Ubiquitous evidence for magma-water and magma-ice interaction suggests that the eruption was entirely subglacial (Tuffen et al., 2001). Detailed descriptions of the lithofacies and eruption mechanisms are given elsewhere (Tuffen et al., 2001; Tuffen, 2001). We focus on the lava lobe lithofacies, which preserves strong evidence for unusual patterns of lava-ice interaction.

Lava lobe lithofacies

Much of the subglacial sequence at Bláhnúkur consists of rhyolitic lava lobes set in poorly sorted breccias (Furnes et al., 1980, Tuffen et al., 2001), shown in Fig. 3a. Lava lobes are conical to irregular in shape and vary between 5 and 20 m in length (Fig. 3a). They can be subdivided into flow lobes and feeder lobes (Tuffen et al., 2001). Flow lobes are commonly cylindrical to prismatic in shape, with the long axis aligned parallel to the modern day slope (Fig. 3b). They consist of a black obsidian carapace 0.05-0.5 m thick that envelops a pale grey microcrystalline rhyolite interior. A banded zone up to 0.2 m thick at the obsidian-microcrystalline rhyolite contact consists of sheared blebs of pale crystalline lava 0.01-0.1 m long enclosed within dark crystal-poor obsidian.

The obsidian carapace of the upper part of flow lobes is frequently cut by polygonal columnar joints 0.07-0.1 m apart, and normal to a steeply inclined near-planar surface which dips approximately down the maximum modern day slope (Fig. 3b). The orientation of fifty-eight of these near-planar surfaces were measured and are plotted in Fig. 4. The vast majority of columnar jointed surfaces dip at between 50 and 80° from the horizontal. Joints penetrate into the microcrystalline rhyolite core of most flow lobes. The bases of flow lobes dip more gently downslope (typically 20-30°). Intact black obsidian at the bases of these flow lobes grades downward into grey, highly fractured, perlitised obsidian, which becomes increasingly fragmented and grades into massive, poorly sorted breccia. Columnar joints are absent.

The breccia consists of 0.5-30 cm wide angular clasts of pale grey perlitised obsidian in an ash matrix (Tuffen et al., 2001). Ash shards are typically 10-100 μm in diameter, glassy, and have blocky-to-cuspate morphologies (Fig. 3c). The breccia is cut by an anastamosing network of veins 5-30 mm wide and up to 10 m in length (Fig. 3d). These are entirely filled by ash, which contains spherical to elongate vesicles up to 20 mm in length (Tuffen et al., 2001). Veins terminate at the intact obsidian of the lower margin of flow lobes. Vesicularity varies from <5% in the microcrystalline rhyolite core to 15-40% in the perlitised lower carapace and breccias.

Feeder lobes are irregular to sheet-like, oriented roughly perpendicular to flow lobes, and typically 5-20 m across. Where exposed, feeder lobes are seen to be linked to one or more flow lobes (Fig. 3e). Feeder lobes have hackly jointed, microcrystalline cores and 0.1-0.5 m thick obsidian margins, which grade outwards into the massive breccias described above. The banded obsidian-microcrystalline transition zone common in flow lobes is absent. Vesicularity is generally less than 5%.

Interpretation

The orientation of columnar joints in lavas indicates the direction of heat loss during cooling (DeGraff et al., 1989). The upper carapaces of flow lobes thus appear to have chilled against steeply inclined, near-planar surfaces. Our interpretation is that the flow lobes were emplaced at the base of a glacier, where they flowed and chilled against the walls of cavities melted into the basal ice (c.f. Furnes et al., 1980). Similar rhyolitic lava lobes within a submarine volcanic succession lack columnar joints (Yamagishi & Dimroth, 1985). There is strong corroborative evidence for a subglacial eruptive setting, such as the presence of facetted clasts within diamicton in the volcanic sequence at Bláhnúkur (Tuffen et al., 2001). Similar columnar-jointed lava flows elsewhere are interpreted as an ice-contact feature (Lescinsky & Sisson, 1998). The position of flow lobes suggests that the subglacial cavities were randomly distributed. The consistently steep orientation of a number of inferred cavity walls suggests a common melting mechanism.

Meanwhile, flow lobe bases share many features with shallow-level peperitic intrusions (e.g. Hunns & McPhie, 1999). Perlitisation, blocky ash shards and matrix vesicles are all evidence for magma/water interaction (Lorenz, 1974; Heiken & Wohletz, 1985; Hunns & McPhie, 1999). We therefore infer that lobes rose towards the glacier base within poorly-consolidated, water-bearing (and possibly water-saturated) breccias. Flow lobes reached the glacier base and show ice-contact features, whereas feeder lobes solidified within the breccia. The fate of meltwater released during the melting of subglacial cavities is likely to have been controlled by the local pressure patterns. The cavity may have filled up with meltwater until the cavity pressure exceeded the pore pressure in the underlying breccia, whereupon meltwater was expelled by porous flow (Dullien, 1992). A steam-filled cavity may develop if the cavity pressure is low and the meltwater can be heated sufficiently rapidly. Steam-filled cavities would be most likely to develop under thin glaciers (≤ 100 m), but the presence of steam in cavities beneath much thicker ice cannot be ruled out until the thermodynamics of energy exchange are better constrained.

The poorly sorted breccias lack evidence for aqueous reworking, supporting the view that standing bodies of meltwater did not accumulate in cavities. Instead, cavities were probably filled with either convecting meltwater or steam. This may explain the steepness of the inferred ice walls, since a cavity only partially filled with meltwater is expected to develop a broad, low shape (Hoskuldsson & Sparks, 1997; Cutler, 1998). Cavity morphology may also be influenced by the patterns of convection within the meltwater or steam. Steam-filled cavities (steam cups) on Mount Rainier have smooth, steeply inclined walls (Kiver & Steele, 1975). The orientation of ice walls may have been controlled by the angle at which meltwater ceased to run down the wall and began to drip off.

Ash-filled veins are interpreted as vapour-escape pipes, which formed during brittle failure of poorly consolidated breccia. Veins were filled by the ash matrix of the breccia, which was remobilized by rising steam. Vapour escape was probably accompanied by the development of numerous fumaroles, as observed in ignimbrites (Sheridan, 1970). The position of fumaroles would have been highly localised, being concentrated close to the heat source (an individual magma body). We propose that cavities may have melted in the ice above a rising magma body during the last ~10 metres of its ascent to the glacier base. Rapid, focussed transfer of heat from magma to ice by convecting steam within vapour-escape pipes facilitated melting ahead of the advancing magma (Fig. 5). Cavities acted as moulds for the advancing magma, which flowed and chilled against the ice walls. The ability of ice to deflect moving lava flows has been observed during a basaltic eruption in Alaska (Vinogradov & Murav'ev, 1988).

Model for melting of basal ice during lava lobe emplacement

In order to test the viability of our interpretations, we present a simple heat transfer model, which predicts the evolution of subglacial cavities during the emplacement of lava lobes. The model is constructed to be consistent with all the field evidence gathered from the lava lobe lithofacies. The symbols and constants used are listed in Table 1.

Assumptions

1. Ice is incompressible and deforms in accordance with Glen's law
2. Ice is at 0 °C
3. Rising magma loses heat through conduction, aided by the formation of joints
4. Convecting steam carries this heat to the glacier base in vapour-escape pipes
5. Vapour-escape pipes develop during the final 10 m of magma ascent to the glacier base, and localised melting of ice occurs, forming subglacial cavities
6. Cavities have attained approximate size equilibrium when they are entered by lava bodies
7. Meltwater may flow into the permeable substrate - with flow governed by local pressure patterns, which are determined by the position of the groundwater table and the proximity of low-pressure meltwater conduits.
8. Cavity pressure may be less than or equal to glaciostatic pressure, and possibly as low as atmospheric pressure
9. All of the thermal energy of heated meltwater is transferred to the ice (i.e. meltwater leaves the system at 0 °C)
10. The ice roof does not thin significantly during basal melting, hence glaciostatic pressure is constant.

Formulation of the model

1. Heat flux from cooling body of magma

In order to simulate the rise of lava lobes to the glacier base, we consider the rise of a cylindrical magma body of radius R with a hemispherical upper surface towards isothermal ice at 0 °C (Fig. 5). In order to simplify calculations, the magma body is assumed to be isothermal (T_m = 850 °C) and loses heat by conductive cooling from the hemispherical upper portion (Fig. 5). An isothermal magma body is deemed appropriate because it is likely to have lost only a fraction of its heat on the timescales considered (see later in text). Furthermore, the magma temperature is likely to have

been buffered by latent heat released during microlite crystallisation, which can be shown texturally to have occurred during lobe emplacement (Tuffen, 2001).
The conductive heat flux Q_h from an isothermal hemispherical magma body is given by

$$Q_h = Sk_l \Delta T, \tag{3}$$

where S is a shape term, k_l is the thermal conductivity of magma, and ΔT is the temperature difference between the magma and surrounding breccia (Holman, 1997). Estimating the effective ΔT is not straightforward. We assume that the breccia was maintained close to 100 °C, the temperature of steam at atmospheric pressure, because of the field evidence for steam fluxing through the breccia. Field observations suggest that the magma body developed a fractured, chilled rind during its ascent and emplacement. The effect of this rind on the conductive heat flux is twofold:- its lower temperature will reduce the temperature contrast between the magma and the breccia (ΔT), but fracturing will increase the surface area available for energy exchange, hence increasing S. The relative magnitudes of these two effects can be estimated:-
a) Chilled rind
We assume that the fractures in the obsidian rind propagated inwards as far as magma at the glass transition temperature T_g (~400 °C, Stevenson et al., 1995), and the outermost portion of the rind was in thermal equilibrium with the surrounding breccia (T = 100 °C, Fig. 6). Assuming a linear temperature gradient within the obsidian, the average temperature of the fractured obsidian surfaces is therefore roughly (400 °C + 100 °C) / 2 = 250 °C. Hence ΔT, the temperature difference between the obsidian rind and the surrounding breccia, is 150 °C. Without a chilled rind, ΔT is 750 °C. Since the conductive heat output Q_h is proportional to ΔT, the presence of a chilled rind reduces Q_h by a factor of about 5.
b) Fractured rind
In accordance with field observations, we assume that fractures propagated a distance z_j = 0.1 m into the obsidian rind, and were spaced a distance w_j = 0.1 m apart (Fig. 6). The number of regular hexagonal columns of side length w_j on a hemispherical surface, N, is given by hemisphere surface area / column surface area

$$N = \frac{2\pi R^2}{3\sqrt{3}w_j^2/2} \tag{4}$$

where R is the hemisphere radius. The total surface area of each column is thus

$$A_{column} = 6z_j w_j + 3\sqrt{3}w_j^2/2 \tag{5}$$

and the total surface area of the fractured hemisphere

$$A_f = \frac{2\pi R^2}{3\sqrt{3}w_j^2/2}\left[6z_j w_j + 3\sqrt{3}w_j^2/2\right] = 2\pi R^2\left[1 + \frac{4z_j}{\sqrt{3}w_j}\right]. \tag{6}$$

Therefore, as

$$\frac{A_f}{A_0} = 1 + \frac{4z_j}{\sqrt{3}w_j} = 3.3 \tag{7}$$

where $z_j = w_j = 0.1$ m, fracturing of the obsidian rind is expected to increase the conductive heat flux by a factor of 3.3.

Combining the two effects, rind cooling and fracturing will reduce the heat flux by a factor of 5/3.3 = 1.5. Many parameters in the heat flux model have been only roughly estimated, hence we assume that the effects approximately cancel out, and ignore the chilled rind in our heat flux calculations. Instead, we consider the conductive cooling of an isothermal magma body at 850 °C, with $\Delta T_{eff} = T_{magma} - T_{steam} \approx 750$ °C. The conductive heat flux Q_h from a hemisphere with radius R = 2.5 m is thus ~ 0.15 MW.

2. Energy transfer from lava to ice

The presence of vapour-escape pipes within the lava lobe lithofacies at Bláhnúkur suggests active and locally focussed steam flux through the breccia during lava emplacement (Tuffen et al., 2001). We interpret that conductive heat released from magma generated a steam envelope in the adjacent breccias, and that the steam formed travelled within vapour-escape pipes to the base of the glacier. From the length of vapour-escape pipes observed at Bláhnúkur, we expect that this process began to operate efficiently once the magma bodies had risen to within 10 metres of the glacier base, and that heat was then transferred almost instantaneously from the magma to the overlying ice. The fraction of released heat that causes melting of ice, E_h, is given by $E_h = Q_m / Q_h$. E_h is thus assume to be 0 at z > 10m (no heat reaches the glacier base) to 1 at z ≤ 10 m (all the heat reaches the glacier base).

3. Magma ascent rate

It is necessary to estimate the ascent rate of magma bodies. Since the magma is denser than the surrounding breccia, its rise must be triggered by a connection to an underlying over-pressured magma chamber. A lower limit on the magma rise rate can be obtained by considering the dimensions of lava lobes. Lobes are typically 100 m^3 in volume. The length of feeder lobes suggests that magma bodies have risen ~10 m through breccia to the glacier base. The minimum ascent rate $V_{l\,min}$ is thus the distance travelled/cooling time. Taking 10^5-10^6 s as a reasonable cooling time (c.f. Hoskuldsson & Sparks, 1997), we acquire $V_{l\,min} = 10^{-4}$-10^{-5} m s^{-1}.

4. Rate and distribution of ice melting

As discussed above, we anticipate that heat from the rising lava lobes will be carried by convecting steam to the glacier base. If all the thermal energy of the magma is used to melt ice, the volume of ice melted per second *a* is given by

$$a = \frac{Q_m}{L_i \rho_i}, \tag{8}$$

where L_i is the heat of fusion of ice. The mean meltback rate of the ice walls

$$r'_m = a / A \tag{9}$$

where A is the ice wall area. For simplicity, we assume that the ice cavity is hemispherical, thus for a cavity of radius r,

$$A = 2\pi r^2. \tag{10}$$

Hence the melting rate is proportional to the inverse square of the cavity radius

$$r'_m = \frac{a}{2\pi r^2} = \frac{Q_m}{2\pi r^2 L_i \rho_i}. \tag{11}$$

5. Cavity closure by ice deformation
If glaciostatic pressure exceeds cavity pressure, the cavity walls are likely to close by visco-plastic deformation (Nye 1953), as discussed earlier. The deformation rate r'_c is given by the relation

$$r'_c = -r\left[\frac{\Delta P}{nB}\right]^n \tag{12}$$

(Nye, 1953) where effective pressure ΔP = glaciostatic pressure (P_i) – cavity pressure (P_c), n is Glen's flow law constant, ~3 (Glen, 1955) and B is an ice viscosity parameter.
An equilibrium radius can be thus be calculated, at which the melting and closure rates are equal and opposite. This is obtained by combining equations (11) and (12):

$$r_{eq} = \left[\frac{Q_m (nB)^n}{\Delta P^3 2\pi L_i \rho_i}\right]^{1/3} \tag{13}$$

which reduces to

$$r_{eq} = \left[\frac{Q_m C}{\Delta P^3}\right]^{1/3} \tag{14}$$

where C is a constant, of value $(nB)^n / 2\pi L_i \rho_i = 2.1\times10^{15}$ J kg^{-3} s^5, where n = 3. The equilibrium cavity radius is thus weakly dependent on the basal heat input and inversely proportional to the effective pressure. The summit firn caves of Mount Rainier appear to be close to equilibrium, and r'_c is estimated at 10^{-7} m s^{-1} from measurements of the ablation rate (Kiver & Steele, 1975). Deformation rates of glacier ice at Bláhnúkur are likely to have been orders of magnitude higher, since the effective pressures were probably greater and ice deforms more readily than firn (Paterson, 1994).

6. Equilibrium cavity radii

Fig. 7a shows equilibrium cavity radii as a function of effective pressure and the radius of spherical lava bodies at ΔT.= 750 °C. Equilibrium radii are weakly proportional to lava body size, but strongly influenced by the effective pressure. For a range of magma sphere sizes consistent with the volume of Bláhnúkur feeder lobes (1.5-4 m), equilibrium cavity radii of 2-4 m, inferred from flow lobes, require effective pressures of 2-4 MPa. The equilibrium cavity radius is only weakly proportional to the lava-breccia temperature contrast ΔT (Fig. 7b). Cavity sizes applicable to Bláhnúkur develop at effective pressures of 2-4 MPa for all reasonable values of ΔT.

7. Timescales of cavity enlargement and lava rise

The rate of change of the cavity radius is obtained by combining equations (11) and (12):

$$\frac{dR}{dt} = r'_m - r'_c = \frac{Q_m}{2\pi R^2 L_i \rho_i} - R\left[\frac{\Delta P}{nB}\right]^n, \tag{15}$$

which is a second order differential equation of the form

$$\frac{dR}{dt} - \frac{M}{R^2} + NR = 0 \tag{16}$$

where M is a melting term of value $Q_m/2\pi L_i \rho_i$ and N is a deformation term of value $[\Delta P/nB]^n$. Thus the cavity radius at time t, R_t is given by

$$R_t = R_0 + \int_0^t \frac{dR}{dt} dt \tag{17}$$

and R_0 is assumed to be zero. Analytical solutions of (17) indicate that approximate equilibrium is attained within 5×10^5 s (Fig. 7c). This results is in accordance with the response time of subglacial cavities to changes in meltwater flux obtained by finite element modelling (Cutler, 1998). Assuming that magma bodies rise at a velocity of 10^{-4} m s^{-1}, melting will occur during the 10^5 s that a magma body takes to rise the final 10 metres to the glacier base, and the lava lobe will then enter the cavity. At 10^5 s, the rate of wall recession (dR/dt) has fallen to $<10^{-5}$ m s^{-1} (Fig. 7c), which is an order of magnitude slower than the lava lobe advance rate. Ice walls with thus be effectively stationary when flow lobes enter cavities, and advancing lavas will be moulded and chilled against the ice. It is assumed that the melting rate will not greatly increase once the lava has enter the cavity, and that the lava will solidify against the near-stationary ice wall (Vinogradov & Murav'ev, 1988; Lescinsky & Sisson, 1998).

8. Application to Bláhnúkur lava lobe lithofacies

The formation of steep-walled steam- or water-filled subglacial cavities is consistent with many features of the lava lobe lithofacies, including lobe morphologies, joint patterns, magma-water interaction at lobe bases, and the presence of vapour-escape pipes. We suggest that lobes were emplaced within cavities with an effective pressure in the range 2-4 MPa. This may correspond to a cavity at atmospheric pressure beneath ice 200-400 m thick. If the cavity pressure were much greater than atmospheric, this may imply that the ice was considerably thicker. Thus the model predicts that the cavity pressure was significantly less than glaciostatic, possibly caused by the high permeability of the substrate, and the steep slope of the glacier base, which may favour the development of low-pressure meltwater conduits (Hooke, 1984). The palaeo-ice thickness has been independently estimated at >350 m from the volcanic stratigraphy (Tuffen et al., 2001). An obvious problem is that there is no clear relationship between the stratigraphic position (elevation) of lava lobes at Bláhnúkur and their size, which the model predicts. More detailed measurements of ice-contact lobes at Bláhnúkur are necessary to test this relationship. Analyses of the volatile contents and vesicle populations of Bláhnúkur lobes would provide information on the degassing patterns within lobes, and may help to constrain the pressure of emplacement, and possibly the position of the groundwater table during the eruption.

A cavity system of this nature is likely to be peculiar to effusive subglacial eruptions with a low magma flux, which favours the slow rise of intrusive magma bodies within permeable 'hyaloclastite' breccias. Drained cavities could only develop during a basaltic eruption if meltwater were heated significantly above 0 °C (Björnsson, 1988; Hoskuldsson & Sparks, 1997). This is due to ice deformation above the region of melting, which forms a 'potential trap' and prevents the escape of meltwater. Such hindered drainage is not thought to apply to subglacial rhyolite eruptions, because of positive pressure changes (Hoskuldsson & Sparks, 1997). This may explain the lack of evidence for meltwater accumulation within subglacial rhyolite sequences (Tuffen, 2001). Similar melting patterns may occur during quiescent degassing at other volcanoes, as observed at Mount Rainier, and may precede many subglacial eruptions. Extremely rapid melting may occur at high eruption rates, as heat is rapidly transferred from lava to ice via mechanisms such as explosive magma-water interaction (Guðmundsson et al., 1997). The isolated cavity system may be quickly overwhelmed, and much larger ice vaults formed (Tuffen, 2001).

Implications of localised melting during subglacial eruptions

Localised melting during a subglacial eruption may create heterogeneous pressures at the glacier base, varying from atmospheric pressure within cavities to glaciostatic pressure (~MPa) elsewhere. Magma may rise preferentially towards low-pressure domains, and therefore become focussed in regions of high heat output, such as a caldera wall or crater rim. A prolonged period of high heat flux would favour the development of low-pressure domains. This is a possible explanation for the evolution of subglacial 'constructional caldera' complexes such as Askja (Brown et al., 1990) and Sollipulli (Gilbert et al., 1996), at which caldera walls are built up by a ring of subglacial domes.

Magma may be extremely rapidly decompressed if it enters a low-pressure cavity, with a pressure drop of several MPa possible over a timescale of seconds. This may lead to non-equilibrium degassing and unusual fragmentation patterns. Cavity pressures may increase suddenly if they become filled with tephra, with the potential for volatile resorption and rewelding of pyroclasts. To date, one locality has been discovered at Torfajökull at which it appears that an explosive eruption has entered a low-pressure subglacial cavity (Tuffen 2001).

Conclusions

Rhyolitic lava lobe lithofacies at Bláhnúkur display evidence for lava/ice interaction. Lobes are inferred to have been emplaced within steep-sided, steam-filled cavities melted into the basal ice. A simplified model shows that a cavity may be melted in the overlying ice during the ascent of a magma body to the glacier base. The model suggests that the cavities at Bláhnúkur formed beneath ice 200-400 thick, in accordance with field observations. Heterogeneous melting patterns during a subglacial eruption may influence the trajectory of rising magma and the resultant volcanic landforms.

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