Progressive cooling of the hyaloclastite ridge at Gjálp, Iceland, 1996–2005

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Abstract

In the subglacial eruption at Gjálp in October 1996 a 6 km long and 500 m high subglacial hyaloclastite ridge was formed while large volumes of ice were melted by extremely fast heat transfer from magma to ice. Repeated surveying of ice surface geometry, measurement of inflow of ice, and a full Stokes 2-D ice flow model have been combined to estimate the heat output from Gjálp for the period 1996–2005. The very high heat output of order 10^6 MW during the eruption was followed by rapid decline, dropping to ∼2500 MW by mid 1997. It remained similar until mid 1999 but declined to 700 MW in 1999–2001. Since 2001 heat output has been insignificant, probably of order 10 MW. The total heat carried with the 1.2×10^12 kg of basaltic andesite erupted (0.45 km^3 DRE) is estimated to have been 1.5×10^18 J. About two thirds of the thermal energy released from the 0.7 km^3 edifice in Gjálp occurred during the 13-day long eruption, 20% was released from end of eruption until mid 1997, a further 10% in 1997–2001, and from mid 2001 to present, only a small fraction remained. The post-eruption heat output history can be reconciled with the gradual release of 5×10^17 J thermal energy remaining in the Gjálp ridge after the eruption, assuming single phase liquid convection in the cooling edifice. The average temperature of the edifice is found to have been approximately 240 °C at the end of the eruption, dropping to ∼110 °C after 9 months and reaching ∼40 °C in 2001. Although an initial period of several months of very high permeability is possible, the most probable value of the permeability from 1997 onwards is of order 10^{-12} m^2. This is consistent with consolidated/palagonitized hyaloclastite but incompatible with unconsolidated tephra. This may indicate that palagonitization had advanced sufficiently in the first 1–2 years to form a consolidated hyaloclastite ridge, resistant to erosion. No ice flow traversing the Gjálp ridge has been observed, suggesting that it has effectively been shielded from glacial erosion in its first 10 years of existence.

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1. Introduction

Volcanic eruptions within glaciers or ice sheets create a variety of structures and morphologies, including hyaloclastite mountains and sheets (Kjartansson, 1943; Mathews, 1947; van Bemmelen and Rutten, 1955; Jones, 1969; Lescinsky and Fink, 2000; Smellie, 2000; Gudmundsson, 2005). Within large glaciers and ice sheets, high heat transfer rates from fragmented magma to ice cause melting of large volumes of ice and large outburst floods (jökulhlaups) which drain the meltwater produced (Björnsson, 1988; Major and Newhall, 1989). Subglacial volcanic activity during the Pleistocene was a major land-shaping process in Iceland, creating tuyas (table mountains) and hyaloclastite ridges that dominate large parts of the modern landscape. The present-day volcanic zones of Iceland are still partly covered with glaciers and ice caps, and subglacial volcanic activity is frequent (e.g. Larsen et al., 1998). Subglacial volcanism can be identified as an important process of land formation in other parts of the world, e.g. in western Canada (Mathews, 1947; Hickson, 2000) as well as on the Antarctic Peninsula, where large hyaloclastite regions are found (Smellie, 1999). Active volcanoes may also exist under the West Antarctic Ice Sheet (Blankenship et al., 1993).

The Gjálp eruption occurred in the autumn of 1996 within Vatnajökull (Fig. 1), the largest of Iceland’s glaciers. The
eruption began on September 30th, at about 22 h GMT, with the onset of continuous seismic tremor, and lasted until October 13th (Einarsson et al., 1997). During the eruption 3 km³ of ice melted and the Gjálp hyaloclastite ridge was formed with a volume of 0.7 km³, rising about 500 m over the pre-eruption bedrock and extending about 6 km along the bed (Gudmundsson et al., 1997, 2002a, 2004). It took 30 h for the eruption to melt its way through the 550 m thick ice at the eruption site. Gudmundsson et al. (2004) give a detailed description of the course of events during the Gjálp eruption. The main form of activity during the eruption was quenching and fragmentation of magma into volcanic glass and the observed heat transfer rate during the first ten days of the eruption reached a maximum of \( \sim 2.6 \times 10^{12} \text{ W} \) (Fig. 9 in Gudmundsson et al., 2004). Only about 2–3% of the energy released during the eruption is considered to have been lost to the atmosphere (Gudmundsson et al., 2004).

The shape of the Gjálp edifice was found to be similar to many hyaloclastite ridges formed in Iceland during the Pleistocene under an ice sheet (Gudmundsson et al., 2002a). This comparison gives rise to questions such as (Gudmundsson et al., 2002a): (1) How well can the freshly formed Gjálp ridge, an initially unconsolidated pile of volcanic glass and tephra, withstand erosion from moving ice? (2) Can palagonitization (e.g. Stroscik and Schmincke, 2002) prevent fast erosion, and what is the rate of alteration for the Gjálp ridge? (3) What role does diversion of ice flow play in the preservation of the edifice? (4) Which general conclusions about the behavior of subglacial as well as subaqueous eruptions can be made?

A pile of unconsolidated volcanic material at the base of a fast-flowing glacier, is expected to suffer heavy erosion. Large parts of a volcanic edifice may be removed over a relatively short period; it has been suggested that such edifice removal has occurred in West Antarctica (Behrendt et al., 1995) and it may have happened also in Iceland (Bourgeois et al., 1998). The Pleistocene ridges and tuyas found in Iceland are made of pillow lava, breccia and hyaloclastite, with hyaloclastite being the major component in some of the ridges (Schopka et al., 2006; Jakobsson, 1979). The vast majority is made of basalts (e.g. Johannesson and Saemundsson, 1998). Glacial, fluvi al and eolian erosion has been at work on the Pleistocene hyaloclastite mountains but systematic studies of the rates of erosion are lacking. In southwest Iceland Ar–Ar dating has revealed that partially eroded subglacially formed edifices are up to 0.7 Ma old (Jakobsson et al., 2003), showing that some hyaloclastite mountains have survived several cycles of glaciation. A common feature of these Pleistocene formations of predominantly basaltic composition is that the volcanic glass has altered into palagonite, turning the loose pile of volcanic glass into consolidated rock (Jones, 1969; Jakobsson, 1979). A similar process is apparently much slower or absent in rhyolitic hyaloclastite (Furnes, 1975; Furnes et al., 1980; Tuffen et al., 2001). This consolidation has been a key factor in preserving the edifices by making them resistant to glacier erosion. However, the rate at which this alteration occurs in the subglacial environment is unknown. A comparable case would be the evolution of the island of Surtsey that emerged in a phreatomagmatic eruption off the south coast of Iceland in 1963–64. Studies at Surtsey showed that palagonitization was strongly temperature dependent; the basaltic tephra was palagonitized to dense tuff in only 1–2 years where it was subjected to mild geothermal activity at temperatures of 80–100 °C, the alteration gradually became slower at lower temperatures (Jakobsson, 1972, 1978; Jakobsson and Moore, 1986). A strong correlation was also found between the rate of palagonitization and water temperature in experiments for basalts and basaltic andesites by Furnes (1975).

The top of the Gjálp edifice was only exposed for about a year after the eruption. It was covered by inflowing ice by the end of 1997 (Gudmundsson et al., 2002a). A deep ice cauldron remained above the submerged top (Figs. 2 and 3), gradually declining in depth. The cauldron had fresh crevasses in 2002 and was still present in 2006 and 2007. This indicates that some
heat output has occurred at this place throughout the study period. Maps have been made for every year since 1997 of the edifice form, and ice flow velocities to derive estimates of heat flow modeling and field measurements of ice depression volume.

It has been considered likely that palagonitization has to some extent taken place at Gjálp (Gudmundsson et al., 2002a) but in the absence of samples from the subglacial edifice this remains hypothetical. However, information on glacier development since the eruption and other indirect evidence on the thermal state of the edifice could constrain models for development of hyaloclastite mountains within glaciers. In this paper we apply calorimetry, ice flow modeling and field measurements of ice depression volume, edifice form, and ice flow velocities to derive estimates of heat output from the Gjálp ridge. This record is used to derive a rough 10-years temperature history of the edifice. The record provides important constraints on the post-eruption thermal evolution of subglacially formed volcanic edifices and the time scales of glacier healing after subglacial eruptions.

2. Method of heat output estimation

Our approach is to use calorimetry to derive the heat output from Gjálp. The record can be divided into two parts:

a) During the eruption. Here the heat output record is obtained from the mass of ice melted during the eruption, estimated from volume of ice depressions at the eruption site and along the flow path of the meltwater. This record already exists (Gudmundsson et al., 2002a, 2004).

b) After the eruption. Here surface mass balance and inflow of ice into the depressions need to be taken into account to derive meaningful estimates of basal melting.

After the Gjálp eruption, a separate ice drainage basin formed around the depressions created by basal melting during and immediately after the eruption (Alsdorf and Smith, 1999; Bjornsson et al., 2001; Gudmundsson et al., 2002a,b, 2004). We apply mass continuity and define the Gjálp depression (Figs. 1, 2 and 3) as our system. The rate of mass loss by melting and subglacial water drainage at the base is the parameter sought, since it is directly proportional to geothermal power. In accordance with observations of ice surface depression shape and extent, we assume that all basal melting occurs at the boundary between the Gjálp ridge and the overlying glacier. The system is bounded on the eastern, southern and western sides by ice divides (Fig. 3) while ice flows in through the northern boundary. The general mass balance of the system is given by

\[ \rho \dot{V}_{\text{ice}} = -\rho \dot{V}_{d} = \dot{m}_{\text{in}} - \dot{m}_{\text{out}} + \dot{m}_{s} - \dot{m}_{b} \]

with \( \dot{V}_{\text{ice}} \) and \( \dot{V}_{d} \) respectively the rates of change in the volume of ice and surface depression volume, \( \rho \) is the ice density, \( \dot{m}_{\text{in}} \) and \( \dot{m}_{\text{out}} \) the mass flux into and out of the system, \( \dot{m}_{s} \) the mass flux at the surface, termed surface mass balance in glaciology, and \( \dot{m}_{b} \) the mass flux at the base, i.e. mass loss caused by basal melting. It is further assumed that bedrock topography remains unchanged, in agreement with observations (see below). During the study period, \( \dot{m}_{\text{out}} = 0 \) was assumed, which is supported by the measured surface velocities (Fig. 3).

Geothermal heat flux at the base of the glacier conducts heat into the ice at the bedrock-ice interface. In case of a cold glacier or ice sheet this increases the temperature of the ice until the melting point is reached and from this point on the energy is used to melt ice. In a temperate glacier, where the ice is at the pressure melting point throughout, the energy is instantly used for melting (Gudmundsson, 2003). The Vatnajökull ice cap is temperate and thus the relation between heat output \( Q_{\text{heat}} \) [W] and ice mass melted at the base \( \dot{m}_{b} \) [kg s\(^{-1}\)] is

\[ Q_{\text{heat}} = \dot{m}_{b} L, \]

with \( L \) being the latent heat of fusion for ice \( (3.335 \times 10^5 \text{ J kg}^{-1}) \) (Petrenko and Whitworth, 1999)). The fact that \( \dot{m}_{\text{out}} = 0 \)
simplifies Eq. (1) and therefore only the terms $m_s$, $V_\dot{}$, and $m_{in}$ have to be estimated to quantify $Q_{heat}$. The mass flux terms required have been measured in the field.

3. Field methods

Extensive field work has been carried out since the eruption in 1996. Aerial observations including radar altimetry were applied in the months after the eruption (Gudmundsson et al., 2004) and from spring 1997 onward the eruption site was visited at least twice a year, in early spring and autumn. Seasonal and annual changes in glacier geometry, surface mass balance values and surface velocities were acquired during these visits.

3.1. Surface mass balance: $m_s$

Surface mass balance data for Vatnajökull has been collected continuously for the whole study period (Björnsson et al., 2002, F. Pálsson and H. Björnsson, pers. comm.) and an independent estimate was made in 2001 within the Gjálp depression by snow coring down to the tephra layers deposited in eruptions in 1996.
(Gjálp) and 1998 (Grímsvötn). The result for the Gjálp area is a mean annual balance of 1240 kg m\(^{-2}\) year\(^{-1}\) water equivalent, which leads to an annual ice layer thickness of 1.35 m using an ice density of 917 kg m\(^{-3}\).

3.2. Topography maps: V

In June each year topographic profiles have been surveyed with a GPS mounted on a snowmobile. In the field seasons from 1996 to 2003, a DGPS with sub-meter accuracy was used, but since 2004 the profiling has been done with a Trimble® Rikinematic GPS with centimeter accuracy. On the basis of the surveyed profiles, topographic maps were drawn by hand, digitized and digital elevation models (DEMs) created using the Kriging algorithm (Cressie, 1991) (Fig. 3a–c).

The 1996 pre-eruption surface topography map in combination with the annual maps allows an estimation of changes in volume (V) of the surface depression initially formed during the 1996 eruption (Fig. 3d–f).

3.3. Bedrock topography

The bedrock topography of Gjálp was measured in 1997 using radio-echo soundings, gravity and direct observations of the top part of the edifice (Gudmundsson et al., 2002a). By 2001, when the top was buried by tens of meters of ice, three drillholes made with a hot-water drill reached the top and upper eastern slope of the edifice (Fig. 4). Finally, in 2006, a radio-echo sounding profile was surveyed across the top part, along a line that coincided with the 2001 drillholes (Fig. 4). These data confirm that the top part of the edifice remained essentially unchanged after 1997. Hence, basal erosion of the edifice seems to have been absent throughout the survey period.

3.4. Surface velocities: m

Since 1997, surface ice flow velocities have been measured each summer using a network of stakes installed and positioned with GPS in spring and remeasured and retrieved in autumn. This approach allows estimation of average summer surface velocity at the location of the stakes. In June 1997 about 20 stakes were installed but due to heavy crevassing in late summer only 8 could be retrieved. In 1998, velocity at 22 stakes was obtained, 12 in 1999, 27 in 2001 and 2002, 28 in 2003 and 2004 and 2005. Between autumn 2002 and 2003, the true annual horizontal velocity could be measured by installing a 6 m long iron stake at one locality. The annual velocity obtained was 12.1 ± 1.5 m year\(^{-1}\), not significantly different from the summer velocity in 2002 of 9.7 ± 1.5 m year\(^{-1}\). This suggests that the summer velocities are representative of the annual velocities.

The annual surface velocity measurements play an important role in the estimation of the ice transported into the system (m\(_{in}\)). The analytical, parallel-sided slab model of a glacier gives a ratio between the vertically averaged velocity of the glacier, \(v_\bar{v}\) and the surface velocity \(v_s\) as \(v_\bar{v}/v_s = 0.8\) for a Glen nonlinearity of \(n = 3\) (Paterson, 2001). This ratio is only valid for a parallel-sided slab geometry. The geometry of the inflow region north of Gjálp is more complex (Fig. 5) and the ratio \(v_\bar{v}/v_s\) is not known. Therefore a numerical model of the inflow region was created to evaluate the vertical velocity distribution and to estimate \(v_\bar{v}/v_s\).

3.5. Ice inflow model: m

Bedrock data from radio-echo soundings are available for most parts of Vatnajökull (Bjornsson et al., 1992), and the Gjálp area was remeasured in 1997, 1998 and 2000 to acquire the shape of the edifice formed by the eruption (Gudmundsson et al., 2002a). This, in combination with the surface maps and the annual surface velocity measurements, makes it possible to create a 2D, finite element model of the inflow area north of Gjálp. Using the Icetools software (Jarosch and Gudmundsson, 2007; Jarosch, 2008), the 2005 surface along with the 2005 surface velocities were used to estimate the \(v_\bar{v}/v_s\) ratio at the location of the inflow cross-section (Fig. 3b). The model computes flow velocities along an approximately north–south trending inflow line, which is displayed in Fig. 3c.

The flow field computed with the model is shown in Fig. 5, using \(n = 3\) in the Glen rheology, \(\dot{e}_{ij} = A \tau^{n-1} \sigma_{ij}'\) (Glen, 1955;
Nye, 1957), and estimating the rate factor $A$ with the surface velocity data. Here $\dot{\varepsilon}_{ij}$ denotes the strain rates and $\sigma'_{ij}$ the deviatoric stresses.

The vertical velocity distribution at the inflow cross-section is estimated to be $\bar{v}/v_s=0.8$, the same value as for a parallel-sided slab model. The rather smooth bedrock topography on the 2 km long section north of Gjálp (Fig. 5) causes this local agreement between the numerical model and the much simpler model of a parallel-sided slab. Other regions within the inflow area have quite different $\bar{v}/v_s$ ratios. With the numerically-estimated ratio it is possible to use the annual surface velocities to calculate $m_{in}$ for each year through the cross-section defined in Fig. 3b.

4. Results

4.1. Thermal power as a function of time

By subtracting the annual surface maps from the pre-eruption surface we obtain the volume of the Gjálp depression as a function of time (Fig. 6). The change with time, $\dot{V}$, is the derivative of this function. The heat output, $Q_{\text{heat}}$, was estimated using Eqs. (1) and (2) for the whole study period. A detailed plot of $Q_{\text{heat}}$ for the first 100 days after the eruption is shown in Fig. 7 and the long term evolution in Fig. 8.

The evolution of the heat output from Gjálp from the end of the eruption can be divided into four episodes (Figs. 7 and 8): (I) The eruption (13 days), (II) end of eruption until June 1997, (III) June 1997–June 2001, and (IV) the period since June 2001. During the eruption (I) the heat output dropped from an initial value of $2\times10^{12} \text{ W}$ to $7\times10^{10} \text{ W}$ (Gudmundsson et al., 2004). A period of rapid drop in heat output followed until June 1997 (II).

In order to construct a volume curve that could be fitted to all data points, the period June 1997 to June 2001 (III) was split into two, two-year intervals and linear regression lines used to identify the average change in volume for these two intervals (Fig. 6). A similar regression line was used for the average volume change during the four year period from June 2001 to June 2005. Until June 2001 changes in the surface depression volume were very small. The volume increased slightly from June 1997 to June 1999 and thereafter decreased slightly until June 2001 (Figs. 6 and 8). From June 2001 onward the depression is closing at an almost constant rate. This approach leads to an average heat output of $17\pm2\times10^8 \text{ W}$ for the period between June 1997 and June 1999. From June 1999 to June
2001 an average heat output of $7\pm 2\times 10^8$ W is obtained. Between June 2001 and June 2005 (IV) no significant heat output was measured. However, since the uncertainty is of the order of $2\times 10^8$ W heat output of $1-2\times 10^8$ W is possible during that period.

4.2. Energy budget of edifice

The energy balance of the Gjálp eruption can be investigated by considering the following two aspects:

- The total volume of the erupted material can be used to calculate the total eruption energy as

\[
E_{\text{tot}} = \int_{T_0}^{T_{er}} m \cdot c_m \cdot dT
\]  

(3)

with $m$ being the mass and $c_m$ the specific heat content of the erupted material, $T_{er}$ the eruption temperature and $T_0$ the final temperature after cooling, i.e. that of the ice ($\sim 0$ °C). By ignoring latent heat of crystallization it is assumed that crystalline material constitutes a minor part of the edifice, an assumption based on the rapid heat transfer (Gudmundsson et al., 2002a), direct observations of the top in 1997, and gravity modeling yielding very low bulk density of the edifice (Gudmundsson et al., 2002a).

- The presented $Q_{\text{heat}}$ data can be integrated over time to derive energy released during that time period. The total energy released during the study period is defined as $E_{\text{rel}} = E_{\text{er}} + E_{\text{post}}$, where the energy released during the eruption ($t_{er}$) is

\[
E_{\text{er}} = \int_0^{t_{er}} Q_{\text{heat}} \cdot dt
\]  

(4)

and the energy released from the end of the eruption until the end of the study period ($t_{\text{end}}$) is

\[
E_{\text{post}} = \int_{t_{er}}^{t_{\text{end}}} Q_{\text{heat}} \cdot dt
\]  

(5)

ignoring the insignificant energy loss to the atmosphere.

The first estimation of total energy is based on the mass of the erupted material, whereas the second method is based on our record of ice melting and calorimetry.

The total volume of erupted material during the Gjálp eruption was $0.8\pm 0.1$ km$^3$ with an average porosity of 45% (Gudmundsson et al., 2002a). This yields the mass of erupted magma $m_m = 1.2\pm 0.2 \times 10^{12}$ kg, using a density of volcanic glass as $2750$ kg m$^{-3}$ (Oddsson, 1982). The energy stored in the erupted magma can be calculated as $E_m = m_m \cdot c_m \cdot \Delta T$ (acc. to Eq. (3)). With $\Delta T= 1090 \pm 50$ °C being the temperature difference between the initial eruption temperature and $0$ °C (Gudmundsson et al., 1997), and $c_m = 1100\pm 50$ J kg$^{-1}$ K$^{-1}$ denoting a temperature-averaged value of the specific heat capacity of the magma (Bacon, 1977), the total energy stored is estimated as $E_m = 1.4 - 1.5 \times 10^{18}$ J.

Initial volatile content of the Gjálp magma is not known, but only $0.5-1.0\%$ H$_2$O is likely to have been present in sufficient quantities to be relevant for energy considerations. Taking a plausible value of $0.5-1.0\%$ H$_2$O of total mass for basaltic andesite (Wallace and Anderson, Jr., 2000), the maximum energy of volatiles may have been $\sim 3\%$ of $E_m$. Thus, including the volatile contribution, the total thermal energy of the eruption is estimated to have been $E_{\text{tot}} = 1.5 \pm 0.3 \times 10^{18}$ J.

$E_{\text{tot}}$ can be compared with $E_{\text{rel}}$, the energy released from the Gjálp edifice over the study period 1996–2005 ($E_{\text{rel}} = 1.4 \pm 0.2 \times 10^{18}$ J, see Fig. 9). This comparison indicates that only a small fraction of the initial energy could have remained by the end of 2005. During the eruption itself, $E_{\text{er}} = 1.0 - 1.1 \times 10^{18}$ J were released, a remarkable two thirds of $E_{\text{tot}}$. Calculations of the rate of heat loss from magma fragmentation into hyaloclastite in subaqueous or subglacial environments suggest that 60–80% of heat should be lost before burial within the growing volcanic pile (Gudmundsson, 2003). Our results conform with these predictions. The energy released gradually since the eruption is $E_{\text{post}} = 0.5 \pm 0.1 \times 10^{18}$ J (Fig. 9).

4.3. Temperature of edifice

Since the energy released from Gjálp ($E_{\text{rel}}(t)$) is known and was used to melt ice, it is possible to estimate an average temperature within the edifice after the eruption. Assuming that after the eruption the pore space is occupied by pore fluid in thermal equilibrium with the rock matrix, there are two end-member cases: that the pore space is occupied with liquid water, or that it is filled with steam. The average temperature, $T_{\text{av}}$ at a given time $t$ can be estimated as

\[
T_{\text{av},1}(t) = \frac{E_{\text{tot}} - E_{\text{rel}}(t)}{m_{\text{ridge}} \cdot c_m + m_w \cdot c_w}, \quad \text{(liquid water)}
\]

(6a)
$T_{av,v}(t) = \frac{E_{tot} - E_{rel}(t) - m_vh_v}{m_{ridge}c_m}$, \hspace{1em} (vapor) \hspace{1em} (6b)

In Eq. (6a), $m_w$ denotes the mass of the water within the pore space and $c_w=4200$ J kg$^{-1}$ K$^{-1}$ the specific heat capacity of water. For the density of the pore water, which was assumed to be at the pressure boiling point, 900 kg m$^{-3}$ was used. The saturation temperature within the central part of the Gjálp edifice is shown in Fig. 10. The north and south parts of the edifice (Fig. 1) are buried under several hundred metres of ice and subjected to 3–5.5 MPa pressure. The saturation temperatures in these buried parts are therefore in the range 230–270 °C.

For Eq. (6b) $m_v$ can be estimated from the density of saturated steam which ranges from 0.6 kg m$^{-3}$ at 0.1 MPa, to 20 kg m$^{-3}$ at 4 MPa (Linstrom and Mallard, 2005) with 10 kg m$^{-3}$ being a suitable mean value. Saturated steam enthalpy ($h_v$) is only weakly pressure dependent and can to a good accuracy be taken as $2.8 \times 10^6$ J kg$^{-1}$. Since 1/8 of the erupted material was transported away from the eruption site by meltwater draining into Grímsvötn (Gudmundsson et al., 2002a), $m_{ridge} = 1.1 \pm 0.2 \times 10^{12}$ kg is the mass of the Gjálp ridge after the eruption.

The volcanic glass in Gjálp was highly vesicular (Steinthorsson et al., 2000). This contributes to the high porosity of 45%. Parts of the pore space are expected to stem from closed minor vesicles in tephra grains, not accessible to pore fluids. It is therefore sensible to assume a somewhat lower effective porosity, here cautiously estimated as 40%. For liquid water as pore fluid the resulting mass is $m_w = 2.5 \pm 0.5 \times 10^{11}$ kg and the total heat capacity of the ridge can be estimated as $C_{ridge, l} = m_{ridge}c_m + m_wc_w = 2.2 \pm 0.4 \times 10^{15}$ J K$^{-1}$. If saturated steam is taken as a pore fluid the total steam mass is $m_v = 3 \times 10^9$ kg. The total energy stored in the steam would then be $m_vh_v = 8 \times 10^{15}$ J. This, however, is less than 2% of the total heat remaining in the edifice at the end of the eruption, about 4% of the estimated heat left in June 1997 (Fig. 9) and much less than the uncertainties in the energy budget. The steam term can therefore be ignored in Eq. (6b) and only the heat stored within the rock matrix is considered. The total heat capacity of the ridge in this case is simply $C_{ridge,v} = m_{ridge}c_m = 1.2 \pm 0.2 \times 10^{15}$ J K$^{-1}$.

Using Eq. (6a), Eq. (6b) and the given parameter values, we obtain $T_{av} \approx 240$ °C for a liquid-filled edifice and $T_{av} \approx 460$ °C for a steam-filled edifice at the end of the eruption. Furthermore, the change in average temperature with time within the edifice for these two end-member cases can be estimated using Eqs. (6a) and (6b) and the energy release history (Fig. 9). It is assumed that porosity remains constant and that the effective pore space is always saturated with water or steam. The resulting cooling history is displayed in Fig. 11.

In Fig. 11a, both cases are displayed for the first three years. The steam dominated case becomes increasingly unlikely with time. The top of the edifice was submerged in water in late summer of 1997 (Gudmundsson et al., 2004) indicating close to hydrostatic pressures at that time, with most of the edifice being subjected to pressures of 2–5 MPa, requiring edifice temperatures of 210–270 °C for saturated steam to exist, higher than the predicted temperature according to Eq. (6b). Moreover, the permeability of an unconsolidated pile of hyaloclastites is very high (see below) and rapid convective heat transfer of steam would quickly have lead to reduced temperature and ingress of liquid water. It should also be borne in mind that throughout the
eruption the vent was submerged in water (Gudmundsson et al., 2004) and the phreatomagmatic activity observed requires access of liquid water to the magma (Wohletz, 1986; Zimanowski, 1998). An unconsolidated subglacial edifice with pores dominated by superheated steam must therefore be considered unlikely. We therefore favor the liquid dominated case as being more realistic in describing the temperatures within the edifice, except perhaps during a short initial period.

The estimated average temperature, assuming 100% liquid pore water, decreased from initially ~240 °C (compared to ~460 °C for 100% steam) right after the eruption to ~110 °C (~200 °C for steam) in June 1997. Temperatures of 60–70 °C were measured at 0.5 m depth in the exposed part of the ridge at this time (Gudmundsson et al., 2002a, 2004). The ambient air temperature was ~0 °C and it is likely that below 0.8–1.0 m depth the exposed top 40 m of the edifice was at the boiling point at atmospheric pressure is observed above the groundwater table (~30 m wide) in June 1997. Temperatures of 60–70 °C were measured at 0.5 m depth in the exposed part of the ridge at this time (Gudmundsson et al., 2002a, 2004). The ambient air temperature was ~0 °C and it is likely that below 0.8–1.0 m depth the exposed top 40 m of the edifice was at the boiling point of water (~95 °C at 1500 m elevation); such a steam region at atmospheric pressure is observed above the groundwater table (sea level) on the island of Surtsey (Stefansson et al., 1985). At Gjálp in June 1997 the groundwater table is expected to have been at about 40 m below the summit, level with a 20–30 m wide pool of water dammed between the edifice and surrounding ice.

After the decline in temperature from 110 °C to 40 °C between 1997 and 2001, little change is estimated in 2001–2005 since the energy release during that period was negligible. The values obtained are based on several simplifying assumptions and have considerable uncertainty. Errors of ±40 °C in average values obtained are based on several simplifying assumptions and have considerable uncertainty. Errors of ±40 °C in average temperature have been estimated. The results are shown in Fig. 12 together with the heat flux history of Gjálp which was calculated from the heat output, $Q_{\text{heat}}$ (Fig. 8) and an area $A_{\text{gjalp}}$, where the heat transport occurs, or $q_{\text{heat}} = Q_{\text{heat}} / A_{\text{gjalp}}$. The area $A_{\text{gjalp}}$ was assumed to be either 100% or 10% of the total area of the Gjálp edifice, thus yielding a region of possible heat flux values.

As can be seen in Fig. 12, the only permeability value matching the heat flux record of Gjálp after the end of the eruption is the one for the altered Surtsey hyaloclastite ($k_3$). For comparison, heat flux values assuming conduction as the only heat transport mechanism ($q_{\text{cond}}$) are also shown in Fig. 12. An average thickness of $d = 200$ m for the conducting layer (c.p.

## 5. Discussion

### 5.1. Permeability of edifice

Heat transport by advection in a porous medium is strongly dependent on the permeability, regardless of whether one or two phase flow is considered. The permeability is related to the porosity as well as fracturing of a medium but consolidation and alteration act to reduce permeability (Ingebritsen and Sanford, 1999). We estimate possible permeability values for the Gjálp edifice in order to explain the cooling record presented. In the absence of a drill core and permeability measurements, we attempt to explain the cooling record by a simple heat transport model.

If liquid-phase buoyancy driven flow through a porous medium is assumed to be the main heat transport mechanism within the edifice, possible heat flux values can be calculated. The volumetric flow rate per unit area $q_w$ (Darcy velocity) (Ingebritsen and Sanford, 1999) was estimated using

$$q_w = \frac{k \rho_0 g \alpha_w (T_{\text{core}} - T_{\text{surf}})}{\mu_w},$$

with $k$ being the permeability of the rock, $\rho_0$ the density of water at $T_{\text{surf}}$, the surface temperature, $T_{\text{core}}$ the core temperature in the center of the Gjálp ridge, $g$ gravitational acceleration, and $\alpha_w$ and $\mu_w$ respectively the coefficient of thermal expansion and the dynamic viscosity of water. Heat flux values for given parameters can easily be obtained by using the enthalpy difference $\Delta h = h(T_{\text{core}}) - h(T_{\text{surf}})$ of the water as $q_{\text{heat}} = \rho_0 g \alpha_w \Delta h$.

The core temperature of the Gjálp edifice ($T_{\text{core}}$) was assumed to be 1.5 times the estimated average temperature $T_{\text{ave}}$ given in Fig. 11 and the surface of the edifice was assumed to be at ~0 °C.

The range in published permeability values for hyaloclastite is quite large. In order to take this into account three different permeability values are used to estimate possible heat flux values. For the high permeability case unaltered and unconsolidated hyaloclastite from Surtsey is used $k_1 = 1.2 \times 10^{-10}$ m²; for recent but moderately altered and consolidated hyaloclastite, a value estimated for 60–100 m depth b.s.l. in Surtsey is used, $k_2 = 4.1 \times 10^{-13}$ m² (Stefansson et al., 1985); and measurements of samples taken at 2000–3000 m depth b.s.l. (Dannowski, 2002) from the HSDP2 drillhole in Hawaii (De Paolo et al., 2001) give results for highly altered and compacted hyaloclastite, $k_3 = 1 \times 10^{-15}$ m². Using these permeability values together with Eq. (7) and the assumed temperature conditions, a set of heat flux values can be predicted for each point in time where average temperature has been estimated. The results are shown in Fig. 12 together with the heat flux history of Gjálp which was calculated from the heat output, $Q_{\text{heat}}$ (Fig. 8) and an area $A_{\text{gjalp}}$, where the heat transport occurs, or $q_{\text{heat}} = Q_{\text{heat}} / A_{\text{gjalp}}$. The area $A_{\text{gjalp}}$ was assumed to be either 100% or 10% of the total area of the Gjálp edifice, thus yielding a region of possible heat flux values.

### Fig. 12. Heat flux estimates based on three different permeability values, estimated temperature conditions and assuming liquid-phase buoyancy driven flow (see text and Fig. 11). The Gjálp heat flux record is displayed in red. For comparison, heat flux values assuming conduction as the only heat transport mechanism ($q_{\text{cond}}$) are also shown in Fig. 12. An average thickness of $d = 200$ m for the conducting layer (c.p.

- $k_1 = 1.2 \times 10^{-10}$ m²
- $k_2 = 4.1 \times 10^{-13}$ m²
- $k_3 = 1 \times 10^{-15}$ m²
Fig. 10a) is used to estimate \( q_{\text{cond}} = \lambda (1.5 * T_{av}/d) \) together with a thermal conductivity \( \lambda \approx 1.1 \text{ W m}^{-1} \text{ K}^{-1} \), a value for porous basalts similar to Bjálp (Robertson and Peck, 1974). Heat conduction has clearly played a minor role in transporting heat out of the edifice. For the period before January 1997 thermal conditions inside Bjálp may have been controlled by two phase flow and very efficient heat transport, which cannot be explained by this simple model. However after January 1997 this simple model for heat transport within the Bjálp edifice matches the recorded heat flux changes quite well.

On the basis of our heat output record and the derived temperature history we propose the following scenario: A short period of very high heat fluxes right after the eruption was driven by two phase convection that lasted roughly until the end of 1996 or the first months of 1997. High permeability of unconsolidated hyaloclastite was probably important to sustain this initial heat transfer. If we assume that the heat release was evenly distributed over the surface area of Bjálp, the heat fluxes calculated with Eq. (7) are not significantly different from the values derived from the heat output record (Fig. 12). The gradual reduction in heat flux occurs as a consequence of lowering of the core temperature. After June 2001 no significant heat flux is measured. However, when the error margins are considered, this result is not very robust and the existence of the small cauldron in the center of Bjálp (Fig. 2) after 2001 indicates that some thermal energy is still being released.

Several simplifying assumptions have been made to derive the cooling history presented. Alternatives are possible such as very low permeabilities being responsible for the lowered heat flux and that temperatures within parts of the ridge are still considerably higher than suggested here. It seems reasonable, however, to accept the simple model presented as the most plausible. This model provides a logical explanation for the large drop in heat flux which is the result of the nonlinear relationship between buoyant heat flux and temperature difference, since both volume expansivity and viscosity are temperature dependent (Eq. (6a)). Thus, a drop in base temperature from \( \sim 90 \text{ °C} \) to \( \sim 60 \text{ °C} \) (corresponding to average temperatures of \( 60 \text{ °C} \) to \( 40 \text{ °C} \)) leads to a drop in convective heat flux from about \( 100 \text{ W m}^{-2} \) to \( \sim 30 \text{ W m}^{-2} \).

5.2. Implications for development of hyaloclastite mountains

The cooling record indicates that the Bjálp edifice sustained large scale geothermal activity over a period of \( \sim 5 \) years. After this period no significant heat output was detected, except the small, localized activity at the top of the ridge. The question arises whether this rather short period of geothermal activity was long enough for the process of palagonitization to consolidate the Bjálp edifice. The observed thermal history cannot be explained assuming permeability values of loose tephra (\( \sim 10^{-10} \text{ m}^2 \)). That would result in much higher heat flux values than observed, given the same edifice temperatures, and faster cooling. The lower values of permeability inferred in our model could be caused by consolidation of the material and therefore one could argue that the thermal history indicates considerable consolidation and hence palagonitization. However, in the absence of samples from the ridge confirming compaction and alteration, the extent to which palagonitization has occurred at Bjálp remains speculative.

A key point in the argument for palagonitization at Bjálp is the edifice temperature of \( \sim 200 \text{ °C} \) at the end of the eruption, with the temperature remaining above \( 100 \text{ °C} \) for about 1 year (Fig. 11). These temperatures are in agreement with the results of a simple cooling model for grain size populations like those generated by fragmentation in Bjálp and settling through a water column within a subglacial lake (Gudmundsson, 2003). Available data on palagonitization rate from Surtsey (Jakobsson, 1978) and experimental results (Fumes, 1975) indicate that consolidation should have occurred within 1–2 years at the temperatures suggested by our data. For englacial eruptions of long duration (months to years) similar or higher edifice temperatures are to be expected. Thus, it is possible that palagonitization of the hyaloclastite may be well advanced at the end of such a long-lived eruption.

Another aspect of the preservation potential for the ridge is the diverted local ice flow field. Fig. 3c shows clearly that the local ice flow field is still diverted towards the edifice with no overflow of ice occurring in the first 10 years after the eruption. Ice flow is required for glacial erosion to take place (e.g. Benn and Evans, 1998). Thus, the process of glacial erosion of the edifice has not yet been effective as confirmed by the observations (Fig. 4). This type of shielding of a hyaloclastite edifice in its early post-eruption development may play an important role in the preservation of hyaloclastite ridges formed in subglacial eruptions. Once ice flow starts, erosion of the edifice may occur, but the rates of such erosion are poorly known.

It may be instructive to compare the post-eruption development of Bjálp with that of the Surtsey eruption. The island of Surtsey was formed over a four year period in repeated volcanic activity. The initial phase was subaqueous, followed by several months of surteyan activity turning into effusive activity generating a lava cap and lava delta (Thorarinsson, 1967). A geothermal area exists on Surtsey but it has been characterized by a very slow decrease in average temperature and low heat flux values. Borehole measurements done in 1982, 15 years after the end of the island’s formation, show that the average temperatures in the core of the island were still above \( 100 \text{ °C} \) while the average heat flux in Surtsey was \( \sim 6 \text{ W m}^{-2} \) (Stefansson et al., 1985). Heat transfer in Surtsey was dominated by hydrothermal convection, both above and below sea level. In contrast, average temperatures in Bjálp decreased to values well below \( 100 \text{ °C} \) over the first five years after the eruption. Moreover, the observed heat flux at Bjálp was about two orders of magnitude higher than found in Surtsey. As in Surtsey, slower cooling rates and a longer duration of thermal anomaly are observed for the ignimbrite in the Valley of Ten Thousand Smokes formed in the Katmai eruption of 1912 (Hogeweg et al., 2005). At Katmai, cooling rates after 1912 have been governed by deposit permeability and the availability of groundwater. The island of Surtsey is only partly
submerged in water and the remaining part is subaerial. In contrast to these two places, Gjalp is completely covered with glacier ice, resulting in a practically inexhaustible supply of groundwater that submerges the edifice. This leads to highly efficient heat transfer from hot rocks to the surroundings and faster cooling times than can be expected for an oceanic island like Surtsey.

6. Conclusion

We have derived a 10 year record of the heat output and cooling history for the subglacial hyaloclastite ridge formed in the Gjalp eruption in 1996. The main conclusions are:

1. The heat output history of Gjalp can be divided into four episodes: (I) The eruption (13 days), (II) end of eruption until June 1997, (III) June 1997–June 2001, and (IV) the period since June 2001. During episode (I) heat output dropped from $>2 \times 10^{12}$ W to $7 \times 10^{10}$ W and further decreased to $3 \times 10^{9}$ W by the end of episode (II). An average value of $1.2 \times 10^{10}$ W characterized episode (III) and no significant heat output was measured in episode (IV).

2. The total eruptive energy was $1.5 \pm 0.3 \times 10^{18}$ J, estimated from the volume of erupted material. A remarkable two thirds of the total energy was released during the eruption itself and by June 2005, only a small fraction remained within the edifice.

3. The heat remaining in the edifice at several points in time has been determined and its average temperature estimated. The temperature dropped from $\sim 240$ °C at the end of the eruption to $\sim 130$ °C after three months and $\sim 110$ °C after nine months. In mid 1999 an average temperature of $\sim 60$ °C is estimated, and $\sim 40$ °C by mid 2001, with little cooling occurring since.

4. Using a liquid-phase buoyancy driven convection model and the derived edifice temperatures, it is found that the cooling history is consistent with permeability values of order $10^{-12}–10^{-13}$ m$^2$, similar to that estimated for consolidated hyaloclastite in the island of Surtsey, but inconsistent with a pile of loose tephra. This may indicate that the edifice consolidated to dense hyaloclastite in the first one or two years.

5. No traversing ice flow over the edifice was observed in the surface velocity data record, indicating that the surface depression closure still dominates the local ice flow field.

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