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# Subglacial volcanic eruptions

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# Ice–volcano interaction of the 1996 Gjálp subglacial eruption, Vatnajökull, Iceland

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**Volcanic eruptions under glaciers can cause dangerous floods and lahars[1–3] and create hyaloclastite (fragmented glassy rock) mountains[4–8]. But processes such as the rate of heat transfer between ice and magma, edifice formation, and the response of the surrounding glacier are poorly understood, because of the lack of data. Here we present observations from the fissure eruption at Vatnajökull ice cap, Iceland, in October 1996. In the 13 days of the eruption 3 km$^3$ of ice were melted and the erupted magma fragmented into glass forming a hyaloclastite ridge 6–7 km long and 200–300 m high under 500–750 m of ice. Meltwater of temperatures of 15–20 °C flowed along a narrow channel at the glacier bed into the Grímsvötn subglacial lake for five weeks, before draining in a sudden flood, or jökulhlaup. Subsidence and crevassing of the ice cap occurred over the eruptive fissure and the meltwater path, whereas elsewhere the glacier surface remained intact, suggesting that subglacial eruptions do not trigger widespread basal sliding in warm-based glaciers.**

Despite the lack of direct observations of subglacial eruptions, qualitative models of the formation of hyaloclastite mountains within glaciers were presented in the 1940s and 1950s[5–7] and have been advanced by more recent work[8–10]. These models are based on geological mapping and involve melting of the overlying ice and the emergence of a mountain within a water-filled depression confined by the surrounding glacier. The lowermost parts of the mountains are often made of pillow lavas, whereas the upper parts are hyaloclastites. Subaerially erupted lavas may cap the mountain, forming a table mountain or tuya. Mountains considered to have been formed in this way are found in Antarctica[4], British Columbia[5] and Iceland[6–8]. Magma fragmentation into pyroclastic glass and its subsequent alteration into consolidated hyaloclastite was observed during the formation of a marine table mountain in the Surtsey eruption off the south coast of Iceland in 1963–67 (refs 11–13).

The eruption in 1996 took place within Vatnajökull, which is temperate (warm based) and the largest ice cap in Europe. On 30 September, at about 22:00 GMT, the beginning of the eruption was marked by the onset of continuous seismic tremor[14]. The subglacial eruptive site was located by aeroplanes the following day, midway between the Bárdarbunga and Grímsvötn volcanoes (Fig. 1). Heavily crevassed ice cauldrons had formed overnight; no signs of unusual melting were observed before the eruption. A shallow linear depression had developed on the ice surface from the eruptive site towards the Grímsvötn lake, and new fractures on the ice shelf covering the lake indicated increased water level. From the first day of the eruption we monitored the eruptive activity and melting of ice, by mapping changes on the ice cap surface, and by monitoring the accumulation of melt water in the Grímsvötn lake (Figs 2 and 3).

On 13 October when the eruption ended, 3 $km^3$ of ice had melted, with a further 1.2 $km^3$ being melted in the following three months (Fig. 3). Ice cauldron formation was fastest during the first four days of the eruption when melting rate averaged 0.4–0.6 $km^3$ per day. At

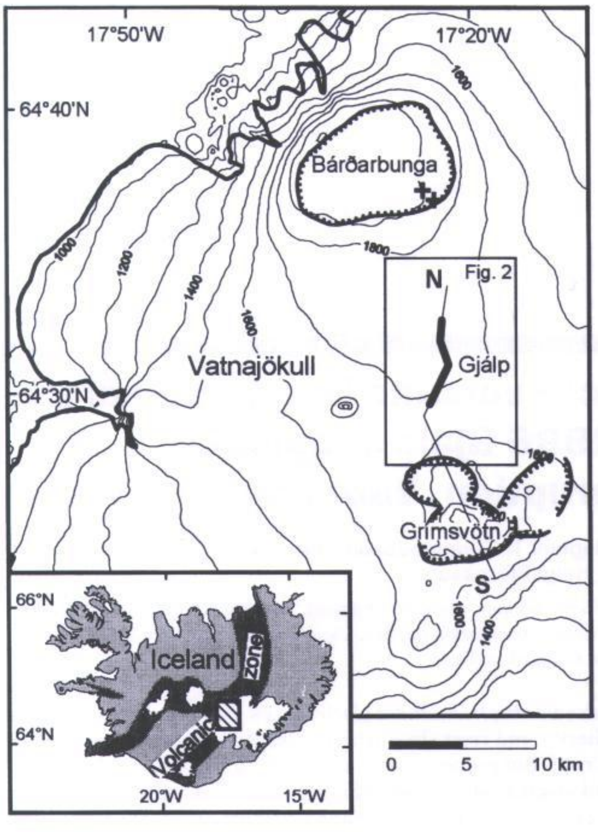


**Figure 1** Map of the northwest part of Vatnajökull showing the location of the 1996 eruptive fissure (solid line, in box), midway between the subglacial calderas of the Bárdarbunga and Grímsvötn volcanoes[22,26]. The contours show ice surface topography (in metres). The eruptive site is well known from accurate mapping and radio-echo soundings that have revealed the bedrock topography[22]. The Bárdarbunga and Grímsvötn volcanoes are among the most active in Iceland with frequent eruptions and associated jökulhlaups documented in the historical record. A subglacial lake is sustained by geothermal activity within the Grímsvötn caldera[22,27]. Jökulhlaups originate from the lake, usually at intervals of a few years, but eruptions may disrupt the periodicity of the jökulhlaups[22,23,28]. The most recent eruptions in the Grímsvötn area occurred in 1934, 1938, 1983 and 1984 (ref. 23). The crosses in the southeast corner of Bárdarbunga are small ice cauldrons considered formed by a minor subglacial eruption at the same time as the main eruption. The locations of the maps and sections in Fig. 2 are shown. Inset shows ice caps and the volcanic zones of Iceland.

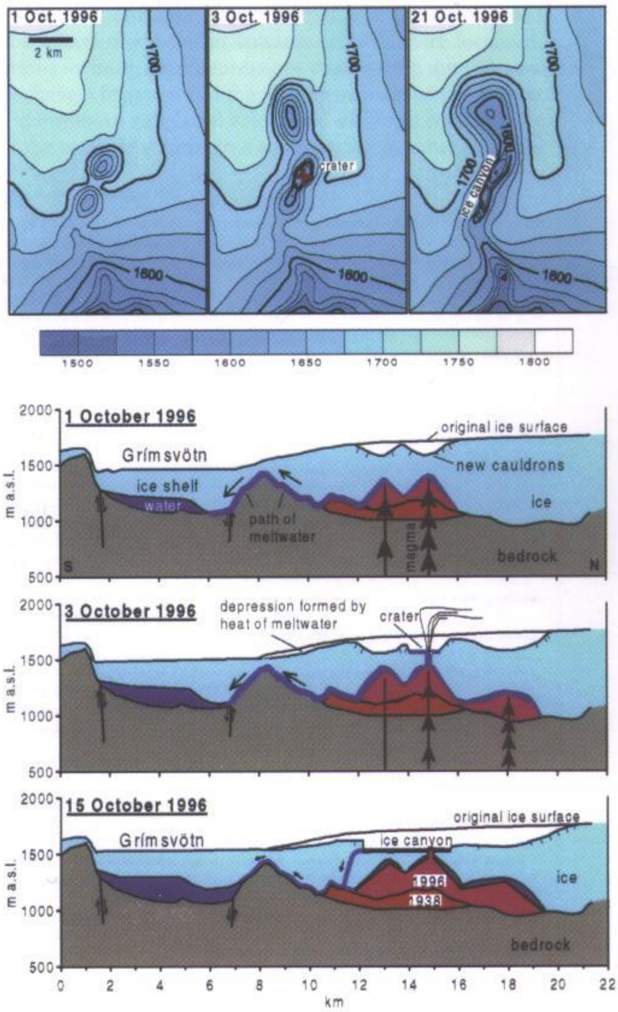


**Figure 2** Top, development of ice cauldrons at the eruption site. The cauldrons over the initially 4-km-long eruptive fissure had reached a depth of about 100 m in the first 15 hours. On 2 October the fissure reached a length of 6–7 km as it grew towards the north. Cauldron development was most rapid during the eruption (30 September to 13 October) but continued after the eruption ended. Bottom, sections showing the ice surface and bedrock topography, and, schematically, the formation of a subglacial hyaloclastite ridge at the eruption site, the subglacial meltwater flow path that is mostly determined by ice surface slope, and water accumulation in the Grímsvötn lake (m.a.s.l., metres above sea level). During the latter part of the eruption, a 3.5-km-long canyon developed in the ice surface above the southern part of the volcanic fissure. Water flowed along the canyon, disappearing down into the glacier at its southern end. The material erupted in 1996 lies partly on top of material from an eruption in 1938.

one location the eruption melted its way through 500-m-thick ice in 30 hours. Other parts of the fissure, where initial ice thickness was 550–750 m, remained subglacial. The maximum elevation of the eruption column was 9 km on 3 October, but typically its elevation was 4–5 km and aerially dispersed tephra amounts to no more than 1–2% of the total volume erupted. The visible subaerial eruption was thus always a minor part of the activity. The volcanic fissure has been given the name Gjálp (after a giantess in Nordic mythology) at the suggestion of G. Larsen.

Formation of ice cauldrons on the surface of ice caps during subglacial eruptions is the response to melting at their base[15] as eruptive products give off thermal and kinetic energy to the environment. If the eruption breaks the ice cover, some of the energy creates an eruption plume. The kinetic energy of eruptive products from typical subaerial fissure eruptions is minuscule when compared with the thermal energy. The total energy of such an eruption is therefore about equal to its thermal energy[16,17]. Heat released to the environment from eruptive products, $Q_e$, is

$$Q_e = M_e(C_e \Delta T_e + L_e) \qquad (1)$$

where $M_e$ is mass of erupted materials, $C_e$ is their heat capacity, $\Delta T_e$ is their drop in temperature and $L_e$ is their latent heat of fusion. Magma erupted subglacially is cooled rapidly and may solidify in a mostly crystalline state as pillow lavas, or, if subjected to lower confining pressure, may fragment and solidify as pyroclastic glass, releasing no latent heat[18], $L_e = 0$. The very fast rate of heat transfer (Fig. 3b) in the Vatnajökull eruption cannot be explained by cooling of pillow lavas, as the solidification time of pillows[19] is at least an order of magnitude too long to account for the observed energy flux per unit area. Thus, quenching and fragmentation of the magma must have been the dominant form of activity. The magma erupted is basaltic andesite with an eruption temperature of $1{,}090 \pm 50\,°C$ and density of $2{,}700\,kg\,m^{-3}$ (N. Óskarsson and K. Grönvold, personal communication), and the mean heat capacity of volcanic glasses when quenched from melting temperature to 0 °C is about $1.10 \pm 0.05\,kJ\,kg^{-1}\,K^{-1}$ (ref. 20). If all heat from eruptive products melts ice that is at the melting point, then

$$Q_e = M_{ice}L_{ice} + M_w C_w T_w \qquad (2)$$

where $M_{ice}$ and $M_w$ are the masses of ice and meltwater respectively, $T_w$ is the temperature of the meltwater, $L_{ice} = 335\,kJ\,kg^{-1}$ is the latent heat of ice melting, and $C_w = 4.2\,kJ\,kg^{-1}\,K^{-1}$ is the heat capacity of water near 0 °C. The amount of erupted magma can be estimated from equations (1) and (2) if the volume of melted ice is known. The volume of ice melted during the eruption and the following six weeks was $4.0\,km^3$, requiring $1.1 \times 10^{12}$ kg of magma if $\Delta T_e = 1{,}000\,°C$, $T_w = 0\,°C$ and efficiency was 100%. This equals $0.4\,km^3$ of magma, making the 1996 Vatnajökull eruption the fourth largest in Iceland this century. The dry density of the pile of volcanic glass is likely to be $1{,}500$–$1{,}600\,kg\,m^{-3}$, similar to that of the Surtsey hyaloclastite[21]. The volume of the eruptive products under the ice cap is then $0.7$–$0.75\,km^3$. In June 1997 the top of the new hyaloclastite mountain was observed within the ice canyon at the site of the subaerial crater in October. The exposed part was a 300-m-long ridge made of partly altered pyroclastic glass. Isolated radio-echo point soundings made in June 1997 suggest that the volcanic edifice is a 6–7-km-long ridge, similar to that shown in Fig. 2.

Highly turbulent convection of a mixture of meltwater and quenched ash fragments is called for to obtain the observed high rate of heat transfer. This mixture had a mean temperature of at least 15–20 °C, the temperature required to account for the volume of ice melted over the subglacial path of the meltwater. Such heat in the meltwater greatly increases the rate of enlargement of subglacial tunnels[22] and must be a key factor in the swiftness and destructive power of volcanic jökulhlaups. The meltwater drained towards Grímsvötn, apparently at the same rate as it was formed (Fig. 3a), precluding any significant water accumulation at the eruption site. Therefore the eruptive products mostly piled up against ice walls, containing the emerging volcano, and the walls remained steep as melting was compensated for by rapid ice flow towards the eruptive site. Thus, hyaloclastite mountains formed within ice caps may be steeper and not as laterally extensive as hyaloclastite mountains or islands of similar volume formed in water.

The product of the Vatnajökull eruption is a subglacial hyaloclastite ridge, partly located on top of a ridge formed in a similar eruption in 1938 (refs 22, 23; Fig. 2), increasing its maximum relief from 200 m to 500 m. Figure 4 shows schematically how the

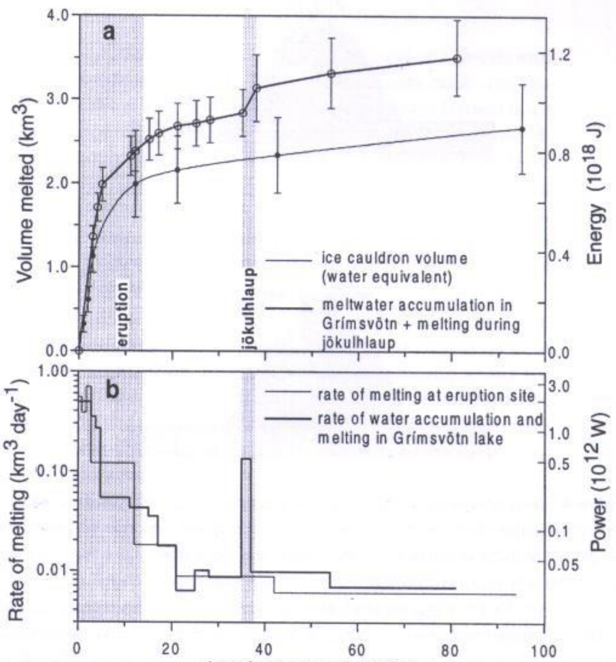


**Figure 3** Ice melting and heat transfer rate from 1 October 1996 until 3 January 1997. **a**, Melting by the eruption could be measured by two independent methods: first, from the water equivalent volume of ice cauldrons above the eruptive fissure (obtained by radar altimetric measurements from an aeroplane equipped with a navigation unit); second, from the volume of meltwater in the Grímsvötn lake (by measuring the elevation of the lake's ice shelf with differential Global Positioning System measurements (DGPS) and knowledge of bedrock topography). Volume of the ice cauldrons is lower because it does not include the volume of the volcanic pile nor that of water melted at Grímsvötn by heat of the meltwater from the eruption site. To obtain total melting by the eruption, the volume of water stored in the porous volcanic pile, estimated as $0.3\,km^3$, should be added to the accumulation curve for Grímsvötn. During the eruption $3.0\,km^3$ of ice was melted and in the following three months, $0.8\,km^3$ melted because of cooling of erupted material and the release of heat as pyroclastic glass was altered into consolidated hyaloclastite. During the jökulhlaup on 4–7 November a further $0.4\,km^3$ was melted by the remaining heat of the water, mainly at the subglacial outlet from Grímsvötn. The level of Grímsvötn fell by 175 m in the jökulhlaup when a total of $3.5\,km^3$ of water was drained from the lake. **b**, Rate of ice melting in water equivalent volume based on the same data as in **a**. In order to obtain the rate of melting at the eruption site from the ice cauldron data, it is assumed that volume of the volcanic pile increased from 0 to $0.4\,km^3$ in the first three days and from 0.4 to $0.7\,km^3$ in the next 10 days.

eruption penetrated the ice cap by melting a narrow chimney through the ice. The subaerial eruption was fed through this chimney in the ice, 200–300 m wide and 50–100 m high, for 12 days without significant widening of the chimney. This gives insight into the conditions needed to form table mountains within ice caps. Lava caps on table mountains in Iceland[6–8] and British Columbia[5] indicate that openings melted in the overlying ice by eruptions must have had diameters of a few kilometres. This suggests that only prolonged or repeated eruptions at the same place can create the conditions required for table mountain formation within glaciers. Initial eruption of pillow lavas may also be important. Gradual heat release from a pile of pillow lava may lead to slower but more widespread melting, causing widening of any chimney formed through the ice cap in early stages of an eruption. Further volcanic activity at the 1996 eruption site might lead to the formation of a table mountain that would persist as a nunatak in the glacier.

It has been suggested that subglacial volcanism may play a role in the dynamics of West Antarctic ice streams by supplying water to their base[24]. At Vatnajökull no rapid basal sliding on a regional scale was associated with the eruption. The formation of ice cauldrons over the eruptive fissure, and their subsequent widening, is the ice cap's response to sudden removal of mass from its base. Intense surface crevassing is similar to that observed in glacier surges but ice flow is probably dominated by internal deformation, not basal sliding as in surges[25]. The crevassed subsidence structure at the eruption site widened from 4 to 7 km in the three months after the eruption, reaching a width of 8 km in June 1997. Otherwise the surface of Vatnajökull remained intact with no signs of increased ice flow such as subsidence or crevassing. The effects of meltwater are also localized: it drains away either continuously, as at the 1996 eruption site, or in a sudden jökulhlaup as observed when the meltwater escaped from the Grímsvötn lake on 4–7 November. The removal of meltwater coupled with the formation of deep depressions may create steep gradients in basal water pressure, towards the depressions, and decrease water pressure at the base of the glacier surrounding the eruption site. This would impede basal sliding. Moreover, the mass removal by melting and drainage will decrease ice flow along the flowline downstream from the eruption site, as the glacier will act to heal itself by ice flow into the depression. Overall, our data suggest that at least for warm-based glaciers, the effects of subglacial volcanic eruptions are localized, with eruptions forming deep depressions and causing jökulhlaups. Significant changes in extent and shape of an ice sheet would require extensive, voluminous subglacial volcanism, melting a considerable fraction of the total ice volume in a short period of time. □

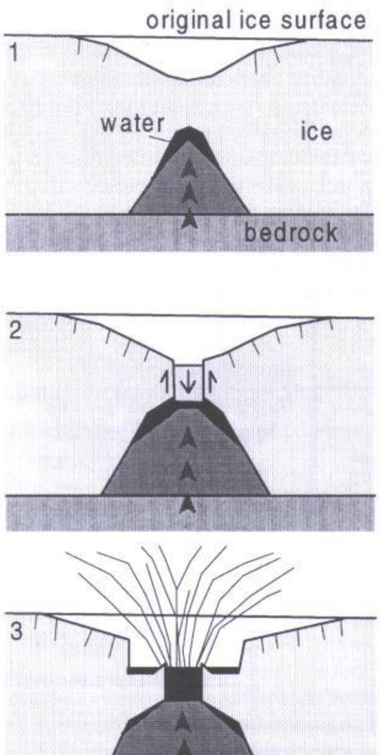


**Figure 4** Schematic sections showing how the eruption penetrated the ice cap, through a narrow chimney through the ice. As the eruption melted its way through the glacier and the volcanic pile grew higher, the thickness of the overlying ice decreased until a block, 200–300 m in diameter, collapsed onto the underlying erupting vent. When this piston of ice had melted completely, the subaerial eruption commenced. Any melting of the chimney walls was apparently compensated by ice flow, as the width of the chimney remained 200–300 m despite continuous eruption for almost two weeks.

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