Reviewed research article

Subglacial and intraglacial volcanic formations in Iceland

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Abstract — *Landforms created in eruptions within glaciers are conspicuous features of the volcanic zones in Iceland and eruptions occur frequently under present-day glaciers. The subglacially and intraglacially created landforms include volcanic structures like tuyas, tindars, móberg sheets, and a variety of proximal sedimentary beds. These landforms constitute a prominent part of the Móberg Formation, a term used for rocks generated during the Brunhes geomagnetic epoch to the end of the Pleistocene (0.78–0.01 Ma). Subglacial and intraglacial rocks of the Móberg Formation cover about 11,200 km^2 of the presently ice free areas. These rocks are predominantly basaltic and the main units of the volcanoes are pillow lava, hyaloclastite tuffs, flow-foot breccias, cap lavas and minor intrusions. Recent eruptions within glaciers have generated tindars and mounds, lead to the formation of widespread basaltic tephra layers, and caused major jökulhlaups. No intraglacial tuya-forming eruptions have been observed. Much of the basaltic glass formed in subglacial eruptions during the Pleistocene has been altered to palagonite, forming consolidated edifices resistant to glacier erosion. Data from recent submarine and subglacial eruptions (Surtsey 1963–1967, Gjálp 1996) indicate that palagonitization and consolidation takes place during the first years after eruption driven by mild hydrothermal activity in the interior parts of the edifices. On the outer slopes of the volcanoes the alteration of the hyaloclastites is dominantly diagenetic. The height of tuyas and tindar in Iceland indicates that they were formed within a glacier that was considerably less than 1 km thick and probably smaller than the Weichselian ice sheet at its maximum. A possible explanation for this might be that tuya-forming eruptions in Iceland were linked to increased magma generation caused by declining pressure in the mantle under a decreasing ice sheet.*

INTRODUCTION

Interaction of water and magma has a major effect on the style of volcanic activity and the morphology of volcanic landforms. At high water pressures effusive activity dominates leading to pillow lava formation, while at lower pressures magma fragmentation and explosive activity are most common (e.g. Wohletz, 1986; Stroncik and Schmincke, 2002; White *et al*., 2000; Chapman *et al*., 2000). Eruptions under ice share the same characteristics in terms of style of volcanic activity while ice confinement and changes in water level due to drainage of meltwater are among features that distinguish subglacial volcanism from submarine and subaqueous eruptions (e.g. Moore and Calk, 1991; Smellie, 2000, 2006; Gudmundsson *et al*., 2004).

Volcanic activity within glaciers has been and still is very common in Iceland (e.g. Kjartansson 1960; Saemundsson, 1980; Gudmundsson, 2005). The terms subglacial and intraglacial are often used to classify the volcanic eruptions and the resulting landforms. Strictly speaking, the term subglacial only applies to processes occurring under ice cover without direct contact with the atmosphere. The term intraglacial is more general since it also applies to eruptions and volcanoes that have broken through the glacier. About 20% of the active volcanic zones is at present ice covered, including many of the most active central volcanoes such as Grímsvötn and Katla (Figures 1 and 7. More than 50% of historical eruptions have taken place within glaciers (Larsen, 2002), mostly in the western part of Vatnajökull (Grímsvötn, Bárðarbunga) and Mýrdalsjökull (Katla).

Subglacial and intraglacial volcanic rocks are especially prominent in the Móberg (Palagonite) Formation of Iceland, which is a chronostratigraphical unit (Kjartansson 1960; Einarsson 1994), and comprises all strata formed during the Brunhes geomagnetic epoch to the end of the Pleistocene (0.01–0.78 Ma). The rocks, which include pillow lavas, hyaloclastites (Table 1) and cap lavas, have sometimes been called the Móberg Formation *sensu stricto* (Kjartansson 1960). The Móberg Formation covers about 11,200 km^2 (Figure 1) in the presently ice-free parts of the volcanic zones. In most areas it is composed of subglacial and intraglacial mountains including tindars and tuyas (Table 1) and the more complicated polygenetic edifices of active central volcanoes. Much of the low-standing areas within the volcanic zones do not belong to the Móberg formation since they are largely covered with subaerially-erupted Holocene lavas (e.g. Jóhannesson and Saemundsson, 1998).

Table 1. Terminology – *Fræðiheiti*

Term	**Definition**
flow-foot breccia	sediments that are deposited on the advancing frontal slope of lava which flows into water (Jones, 1969); synonym with “foreset breccia” (Jakobsson, 1978) and “lava-fed delta” (Skilling, 2002)
hyaloclastite	volcaniclastic deposits formed by explosive magma-water fragmentation and non-explosive granulation of glassy lava rims; used for both unconsolidated and consolidated deposits (Fisher and Schmincke, 1984)
móberg	consolidated, mafic to intermediate, hyaloclastite (cf. Kjartansson, 1943)
móberg sheet	a flat layer of hyaloclastite with isolated pillows and pillow fragments, and usually columnar jointed basalt at the base (Walker and Blake, 1966; Loughlin, 2002)
palagonite	altered, mafic to intermediate, volcanic glass (cf. Peacock, 1926a); the first stable alteration product of mafic to intermediate, volcanic glass (cf. Stroncik and Schmincke, 2002)
passage zone	the zone of transition between a subaerial lava and its subaqueous flow-foot breccia (Jones, 1969)
tindar	a linear, constructional, often serrated, ridge, made up of hyaloclastite and/or pillow lava, occasionally with cap lava (Jones 1969); synonym with “móberg ridge” (Kjartansson, 1960) and “hyaloclastite ridge” (Chapman *et al.*, 2000)
tuya	a subrectangular or circular, constructional, flat-topped mountain, made up of hyaloclastites and/or pillow lava, usually with cap lava (Mathews, 1947); synonym with “stapi” (Kjartansson, 1943) and “table mountain” (Bemmelen and Rutten, 1955)

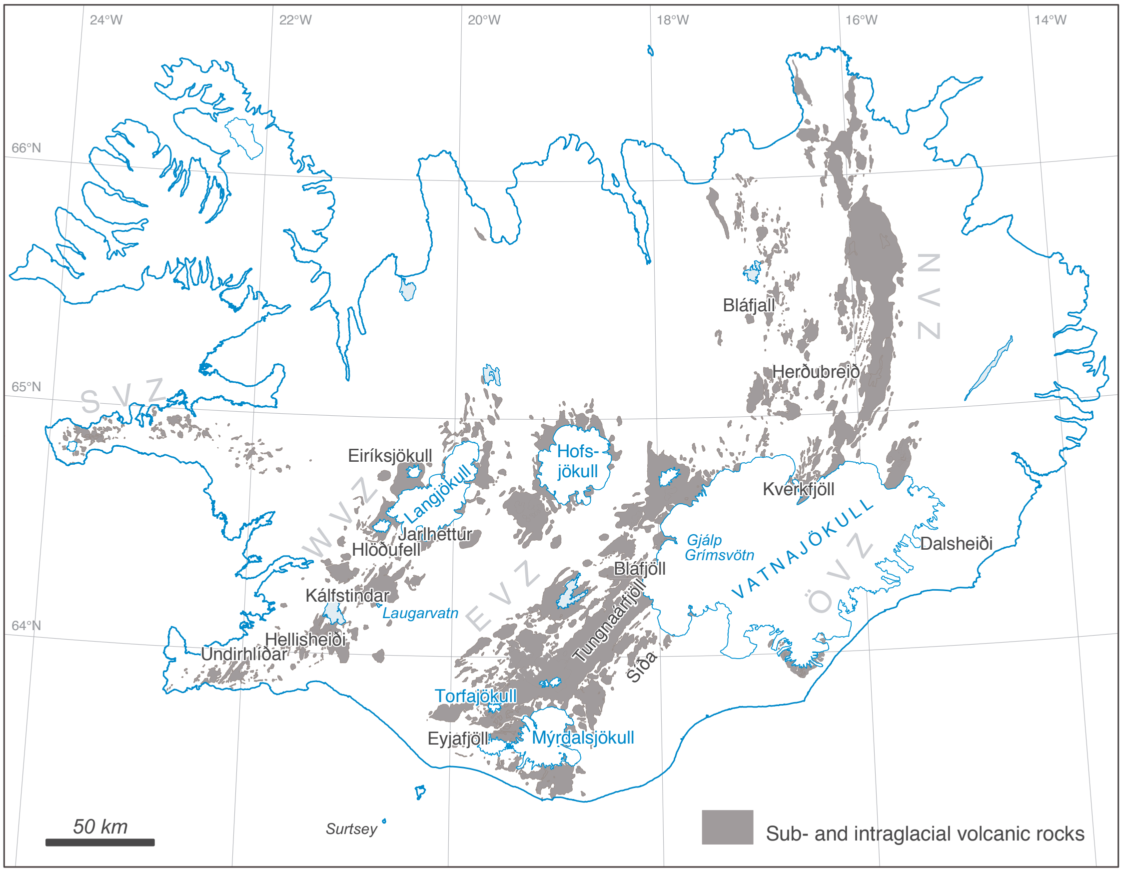

Figure 1. Exposures of subglacial and intraglacial volcanic rocks of Late Pleistocene age (0.01–0.78 Ma) in Iceland. Modified from Jóhannesson and Saemundsson (1998). WVZ: Western Volcanic Zone, EVZ: Eastern Volcanic Zone, NVZ: Northern Volcanic Zone, SVZ: Snæfellsnes Volcanic Zone, ÖVZ: Öræfajökull Volcanic Zone. – *Útbreiðsla jarðmyndana sem orðið hafa til við gos undir jöklum á síðjökultíma. WVZ: Vesturgosbelti, EVZ: Austurgosbelti, NVZ: Norðurgosbelti, SVZ: Snæfellsnesgosbelti, ÖVZ: Öræfajökulsgosbelti.*

In this paper the main characteristics of subglacial and subaqueously-formed volcanics found in Iceland are reviewed. A brief history of the development of ideas is presented, then the basic units of a subglacial/subaqueous volcanic edifice are defined, the most common landforms described, the process of palagonitization and consolidation is discussed, and recent eruptions are reviewed including the constraints they put on the rates of volcanic processes in subglacial volcanic activity.

HISTORICAL OVERVIEW

The subglacial Pleistocene origin of the Móberg formation was discovered by Pjeturss (1900, 1904). Peacock (1926a, 1926b) demonstrated that the Móberg deposits were created by volcanic activity under glaciers and in lakes, that basaltic glass is the principal component of the hyaloclastite, and that a large part of the glass has altered into palagonite (Table 1). After studying the 1934 eruption in Grímsvötn and hyaloclastite formations in south Iceland Noe-Nygaard (1940) discussed palagonitization and was the first to sketch the possible evolution of a subglacial eruption. Early attempts of explaining the existence of the steep-sided and flat-topped tuyas had lead to two hypotheses: (1) They are remnants of pre-existing highlands that had been subjected to regional subsidence (Reck, 1922); (2) they are volcano-tectonic horsts (Sonder, 1938). Kjartansson (1943) distinguished between "móberg" ridges and tuyas and suggested a third alternative hypothesis, (3) that the ridges and tuyas might be formed in subglacial and intraglacial eruptions.

Independently, Mathews (1947) had reached the same conclusion on the formation of tuyas in British Columbia. Bemmelen and Rutten (1955) came to similar results for the subglacial mountains in north Iceland. Einarsson (1960) studied subglacial mountains at Hellisheiði in SW-Iceland and was the first to identify the existence of a basal unit of pillow lavas that forms during the first eruptive phase in a subglacial eruption. Jones (1969, 1970) studied the hyaloclastites around Laugarvatn in SW-Iceland (Figure 1) and defined the stages of a tuya forming eruption and how they relate to individual eruptive units. Later work on the subglacial and intraglacial volcanics of the Móberg formation has included application of sedimentology, petrology and geophysics and lead to deeper understanding of the formation of hyaloclastites and pillow lavas and their relation to the overlying ice sheet (e.g. Bergh and Sigvaldason, 1991; Smellie and Skilling, 1994; Werner and Schmincke, 1999; Schopka *et al.*, 2006; Höskuldsson *et al.*, 2006). Work on subglacial silicic rocks has revealed important differences when compared to basalts in style and formation morphology (Furnes *et al.*, 1980, Tuffen *et al.*, 2001, 2002).

Eruptions in recent decades have greatly advanced the understanding of subglacial and subaqueous eruptions. The formation of Surtsey off the south coast of Iceland in 1963–1967 (Figure 1) offered important insight into the phreatomagmatic volcanism, island building, formation of flow-foot breccia and post-eruption palagonitization (e.g. Thorarinsson *et al.*, 1964; Thorarinsson, 1967; Jakobsson, 1978; Jakobsson and Moore, 1986). In a similar way, the Gjálp eruption in Vatnajökull in 1996 demonstrated how a tindar is formed within a large glacier, including response of the ice to the eruption, the drainage of meltwater and subsequent cooling and evolution of the subglacial ridge (e.g. Gudmundsson *et al.*, 1997; 2004; Björnsson *et al.*, 2001; Gudmundsson, 2005; Jarosch *et al.*, 2008).

BASIC UNITS

The basic units formed in basaltic eruptions under glaciers are pillow lava, hyaloclastite, irregular intrusions and cap lava flows. A complete subglacial and intraglacial formation contains all units, with the pillow lavas at the base, overlain by hyaloclastites and capped with subaerial lavas. Irregular intrusions occur mainly within the hyaloclastites. Tuyas typically consist of all units while most tindars lack cap lava (Figure 2).

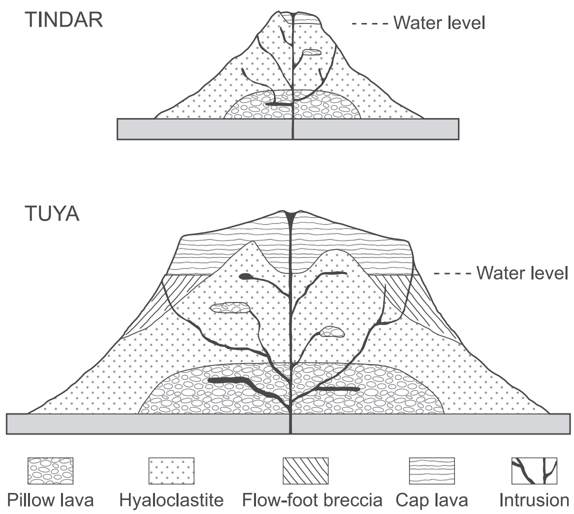

Figure 2. Simplified cross-sections of a tindar and a tuya. Based on observations of subglacial and intraglacial mountains in the Western Volcanic Zone (Jakobsson and Johnson, 2008). – *Einfölduð þversnið af móbergshrygg og móbergsstapa. Byggt á mælingum í Vesturgosbeltinu.*

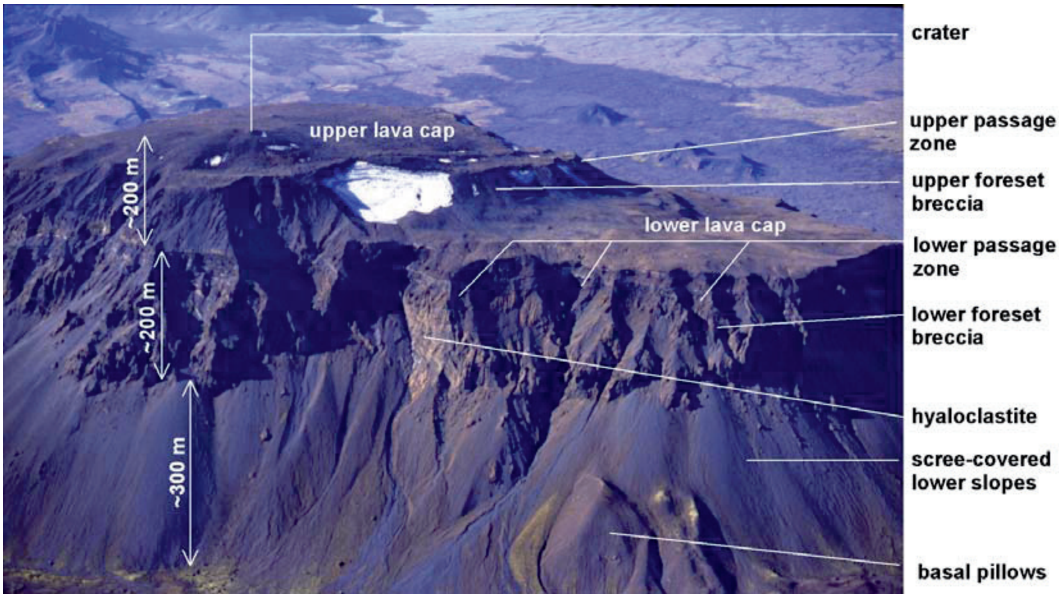

Figure 3. The tuya Hlöðufell in the Western Volcanic Zone, aerial view from the northeast (Figure 1). The summit rises 700 m above the surrounding Holocene lava fields. The two lava caps indicate a rise in the level of the intraglacial lake during the formation of the mountain. The lower lava cap is 50–100 m thick and the elevation difference between the passage zones is about 150 m. – *Móbergsstapinn Hlöðufell í Vesturgosbeltinu, séður úr norðaustri. Hraunlögin tvö í kolli fjallsins gefa til kynna að vatnsborð í jöklinum hefur hækkað meðan á gosi stóð*. Photo/*Ljósm.* O. Sigurðsson.

Pillow lava commonly represent the basal unit of subglacial mountains. Pillow lavas form where the water pressure has been sufficiently high to prevent efficient mixing of magma and water and hence fragmentation of magma by thermal granulation or fuel-coolant interaction (e.g. Zimanowski and Buettner, 2003). Pillow lavas may locally be absent (e.g. Schopka *et al.*, 2006; Jakobsson and Johnson, 2008) but a thickness of up to 280 m has been reported in the Western Volcanic Zone (Jones, 1970). In places elongated ridges of pillow lava and water-sorted hyaloclastite, extending outwards from the edifice are observed, as at Hlöðufell, SW-Iceland (Figure 3). Such ridges may have formed when pillow lava flowed along subglacial drainage channels.

In the south central highlands flat, sheet-like formations of pillow lava occur (Figure 3), up to 3 km in width and 15 km in length (Vilmundardóttir *et al.*, 2000). Individual pillow tubes or sacks are commonly 0.5–1.0 m in width and their porosity increases with increasing height of the edifice (Jones, 1968, 1970).

Hyaloclastites typically overlie the basal pillow lavas. Hyaloclastite (Table 1) is formed by fragmentation of the magma as a consequence of mixing with external water. A distinction can be made between coarse hyaloclastite where rock fragments are common and finer-grained surtseyan tephra. The lower part of the hyaloclastite unit is in some volcanoes coarse-grained and poorly stratified, while the top part is often fine grained and stratified. In other volcanoes, particularly tuyas (Werner and Schmincke, 1999), several successions of coarse- and fine grained hyaloclastite can be observed. An apparent total thickness of 300 m for the hyaloclastite unit has been observed (Jones, 1969).

In the island of Surtsey the subaqueous part is fine grained and stratified while the coarser unit is

Figure 4. Typical tindar morphology southwest of Hlöðufell, SW-Iceland (Figure 1). Aerial view from the southeast. – *Dæmigert landslag móbergshryggja suðvestur af Hlöðufelli í Vesturgosbeltinu, séð úr suðaustri.*

found below 20 m depth below sea level, reaching down to the pre-eruption sea bottom at 130 m (Jakobsson, 1978; Jakobsson and Moore, 1982). Apparently, the coarser hyaloclastite is formed subaqueously or subglacially, and at less efficient explosive activity than typical for the highly energetic surtseyan style of eruption. Helgafell, adjacent to Undirhlíðar (Figure 1), is an example of a tindar made of only hyaloclastite, lacking a lava cap and apparently also basal pillow lava (Schopka *et al.*, 2006).

For basalts it is common that 80–90% of the hyaloclastite is glass, 5–12% phenocrysts and 2–8% rock fragments. The basaltic glass is unstable and alters easily to palagonite with the formation of secondary minerals leading to consolidation of the hyaloclastite.

Intrusions are common although they are volumetrically a minor part of the subglacial and intraglacial edifices. These intrusions are mainly dykes <1 m thick and minor irregular intrusions are common (e.g. Jones, 1970; Werner *et al.*, 1996; Schopka *et al.*, 2006). Some intrusions form as layers or lobes of pillow lava within the hyaloclastite (Saemundsson, 1967; Werner and Schmincke, 1999; Jakobsson and Johnson, 2008).

Lava, erupted subaerially, forms a gently sloping lava cap covering many tuyas and some tindars. They are formed when the vent emerges above the water level in an intraglacial lake. When water no longer has access to the vent, explosive activity stops. The eruption turns effusive and lava starts to flow. It advances by building a lava delta, a flow-foot breccia (Table 1), composed of pillows, pillow fragments and coarse-grained hyaloclastite, formed by fragmentation of batches of the magma as it enters the water (e.g. Jones, 1968, 1969; Werner and Schmincke, 1999; Skilling, 2002). These cap lavas can reach considerable thickness, with up to 350 m observed in the Western Volcanic Zone (Jakobsson and Johnson, 2008). Thin lava caps are found on some tindars.

EDIFICE MORPHOLOGY

Kjartansson (1943, 1960) divided the "móberg" mountains into two morphological types: "móberg" ridges and tuyas. Research carried out since the publication of this early work fits with this main classification, renaming the "móberg" ridges as tindars (Table 1). However, a third type needs to be added, the móberg sheets made of hyaloclastite and/or pillow lava, and columnar jointed basalt.

Tindars and tuyas

A characteristic of the tindars, especially those that have suffered little erosion, is that they form a row of peaks at semi-regular intervals (Figure 4). This corresponds to Holocene volcanic fissures where the activity is quickly concentrated in a row of craters although a continuous eruption took place at the beginning of eruption (cf. Wylie *et al.*, 1999). The length of each ridge, or tindar, is usually greater than double its width. In the Western Volcanic Zone (WVZ) the longest tindars are up to 9 km in length (Figure 5).

The majority of the largest tindars have basal pillow lavas with Kálfstindar in SW-Iceland (Figure 1) being a good example (Jones, 1970). In many of the smaller tindars, the basal pillow lavas appear to be absent, as in Helgafell, SW-Iceland (Schopka *et al.*, 2006). Some tindars are exclusively made of pillow lava, such as Undirhlíðar in SW-Iceland (Jónsson, 1978; Schopka *et al.*, 2006) and several of the ridges north of Kverkfjöll in central Iceland (Höskuldsson *et al.*, 2006). Apparently, in these cases the eruption stopped before magma fragmentation commenced. Several of the tindars in the WVZ, such as Jarlhettur and their southwest continuation, Brekknafjöll (Figure 1), have thin lava caps. Tindars are the dominant landform in parts of the Eastern Volcanic Zone (EVZ), especially in Tungnaárfjöll between the rivers Tungnaá and Skaftá (Figure 5). The longest of these tindars, Skuggafjöll, is 44 km long and its maximum width is 3.8 km (Vilmundardóttir *et al.*, 2000). In the Northern Volcanic Zone, especially its southern part, many large tindars occur (Helgason, 1989, 1990; Vilmundardóttir, 1997).

Tuyas (Figures 2 and 3) do usually only have a single central crater. However, since they mostly occur in the rift zones, the eruptions presumably started along a fissure. The largest tuya is Eiríksjökull ($\sim$48 km^3) in the WVZ (Figure 1), it has a basal area of 77 km^2 and its maximum relative height is about 1 km. Eiríksjökull is the largest monogenetic volcanic unit identified so far in Iceland (Jakobsson and Johnson, 2008). Two well-studied tuyas in the NVZ are Bláfjall (Schiellerup, 1995) and Herðubreið (Werner *et al.*, 1996). Werner *et al.* (1996) showed that Herðubreið actually consists of two eruptive units, the lower unit formed in lacustrine environment with little evidence for direct ice contact, while the upper, main part formed within a thick glacier. It is likely that the tuya-forming eruptions were long-lived and lasted considerably longer than the fissure eruptions that formed the ridges. Not all tuyas have a lava cap. In the WVZ tuyas can be found where the eruption apparently stopped at the pillow lava stage (Jakobsson and Johnson, 2008).

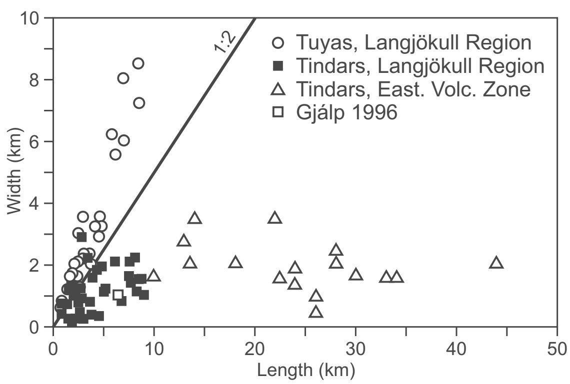

Figure 5. Width and length of tindars and tuyas in the Western Volcanic Zone and Eastern Volcanic Zone (Figure 1). The tindar formed in the 1996 Gjálp eruption is also shown. Most tuyas have length/width ratios <2 while most tindars have substantially higher length/width ratios. – *Lengd og breidd móbergshryggja og móbergstapa í Vesturgosbeltinu og Austurgosbeltinu. Lengd og breidd hryggjarins sem myndaðist í Gjálpargosinu 1996 eru einnig sýnd.*

Figure 5 shows the length and width of relatively well-preserved subglacial and intraglacial volcanoes in Iceland. It includes 72 tindars and tuyas in the WVZ (Jakobsson and Johnson, 2008) and 17 tindars in the EVZ (Vilmundardóttir *et al.*, 2000). The graph shows that tuyas generally form a group where the length/width ratio is less than 2, while all but the smallest tindars have length/width ratios greater than $\sim$2.

Móberg sheets

Layers of hyaloclastite and pillow lava, often with columnar jointed basalt at the base, have been described that apparently flowed considerable distances under ice or water. Dalsheiði in southeast Iceland which presumably is of Early Pleistocene age, is about 150 m thick, 2 km wide and 22 km long and its original length may have been 34 km (Walker and Blake, 1966). Hyaloclastite with isolated pillows and pillow fragments is the main component of Dalsheiði while the basal part is columnar jointed basalt overlain by an entablature that grades into the overlying hyaloclastite. Walker and Blake (1966) suggested that this layer was formed by lava flowing under a valley glacier. Similar formations occur elsewhere in southeast Iceland (Walker and Blake, 1966).

In the Síða district (Figure 1) and Fljótshverfi to the northeast of Síða, extensive móberg sheets of Early Pleistocene age occur (Bergh and Sigvaldason, 1991). These móberg sheets have similar stratigraphy as Dalsheiði and are commonly emplaced on layers of tillite (Noe-Nygaard, 1940). Bergh and Sigvaldason (1991) suggested that these sheets had formed in large volume eruptions under a glacier within the volcanic zone to the west. These lava flows had then advanced towards the southeast and onto the marine shelf. Smellie (2008) has, however, shown that the alteration minerals found in the Síða and Fljótshverfi sheets point to deposition in freshwater, suggesting a fully subglacial origin and deposition. Similar sheets but of much lower volume occur in the lower slopes of Eyjafjöll in south Iceland (Figure 1) thought to have formed under a Pleistocene glacier (Loughlin, 2002). In the southern central highlands of the EVZ, low aspect ratio sheet-like formations made exclusively of pillow lava are common (Vilmundardóttir *et al.*, 2000). An example of these formations is Bláfjöll near the western margin of Vatnajökull (Figure 6). It is elongated NE-SW, about 10 km long, 3 km wide, 0.2–0.3 thick and has a volume of several km^3. These low aspect-ratio high-volume pillow lava formations may have formed by high discharge fissure eruptions under thick ice.

SEDIMENTARY REWORKING OF HYALOCLASTITE

Fluvial transport of unconsolidated hyaloclastite and glacial, fluvial and eolian erosion have displaced, modified and removed considerable parts of the hyaloclastite formations (Bemmelen and Rutten, 1955; Jones, 1969; Werner and Schmincke, 1999). Detailed studies of the sedimentary formations have been few. It is clear, however, that sediments of this type are found in the lower slopes and around most hyaloclastite mountains. Debris flows are very common and are probably an integral part of the buildup of an evolving subglacial volcano (e.g. Smellie and Skilling, 1994; Werner and Schmincke, 1999; Schopka *et al.*, 2006). In places, large deltas of hyaloclastite at the base of the volcanoes occur, as seen in the southeast slopes of Kálfstindar (Jones, 1969).

Fluvial processes have in many cases transported parts of the unconsolidated hyaloclastite with jökulhlaups and glacial rivers out onto the outwash plains and the marine shelf of Iceland. The subglacial mountains are also eroded by glaciers after their formation. The older formations are often heavily eroded while the more recently formed ones are better preserved. Detailed volumetric estimates of total erosion are lacking at present although it is likely that up to 50% of many of the older edifices have been removed while some of the most youthful looking mountains have suffered little erosion.

RECENT SUBGLACIAL ERUPTIONS

Studies of tephrochronology of soil sections and outlet glaciers of Vatnajökull, and historical records have revealed that over 50% of eruptions since the settlement of Iceland before 900 AD occurred within glaciers (Larsen, 2002). The majority of eruptions

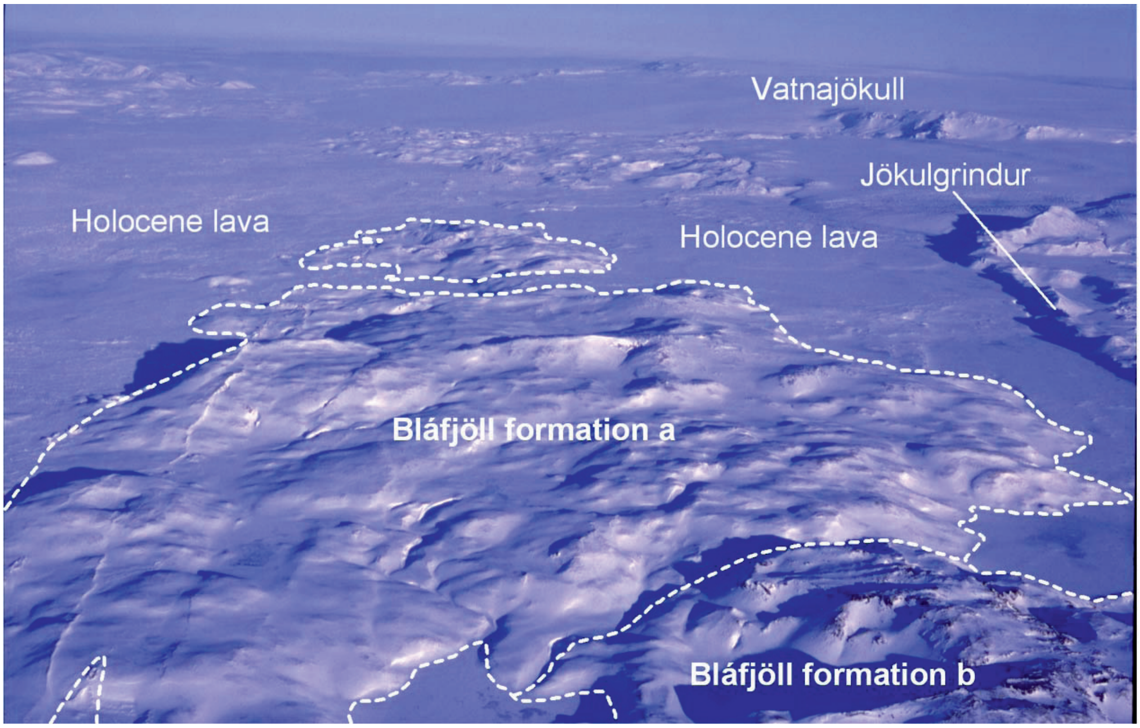

Figure 6. Aerial photo of pillow lava sheets in Bláfjöll in the south central highlands (Figure 1). The formation Bláfjöll a is up to 10 km long and 3 km wide while its relative height is about 0.2 km. In stark contrast is the tindar of Jökulgrindur, also composed of pillow lava (based on Vilmundardóttir *et al.*, 2000). – *Vetrarmynd af bólstrabergsbreiðum er mynda Bláfjöll í Austurgosbeltinu. Til samanburðar eru Jökulgrindur sem einnig eru úr bólstrabergi.* Photo/*Ljósm.* O. Sigurðsson.

have occurred in Vatnajökull where the Grímsvötn central volcano dominates the record with about one eruption every 10 years (Figure 7). However, activity in the Vatnajökull region is periodic, with high activity intervals of 60–80 years alternating with low activity intervals of similar length (Larsen *et al.*, 1998). Many of these eruptions have caused considerable melting and jökulhlaups, notably the eruptions in Katla (e.g. Larsen, 2000), Grímsvötn (Thorarinsson, 1974; Gudmundsson *et al.*, 1997) and Öræfajökull (Thorarinsson, 1958).

Several eruptions occurred within glaciers during the first four decades of the 20th century while 1939–1996 was the most quiet volcanic period in Icelandic glaciers since the latter part of the 16th century (Larsen *et al.*, 1998), with only one confirmed minor eruption occurring in Grímsvötn in 1983 (e.g. Gudmundsson, 2005). This quiet period came to an end with the Gjálp (Figure 1) eruption in October 1996, followed by eruptions in Grímsvötn in 1998 and 2004, and a possible minor eruption in Katla in 1999 (Gudmundsson *et al.*, 1997, 2004; Gudmundsson *et al.*, 2007).

Gjálp 1996

The eruption occurred under initially 600–750 m thick ice when a 6 km long fissure erupted between the subglacial central volcanoes of Grímsvötn and Bárðarbunga (Gudmundsson *et al.*, 2004). The eruption lasted 13 days. It melted its way through the 600 m

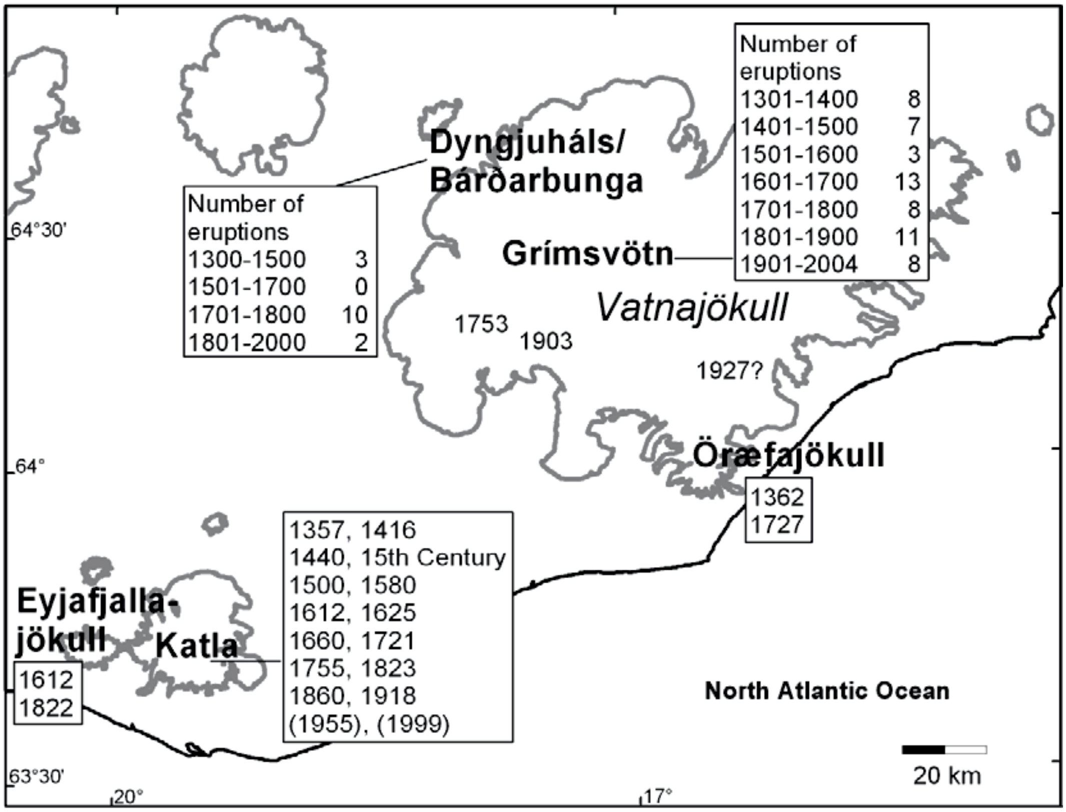

Figure 7. Known eruption sites in Icelandic glaciers since about 1300 AD. For Bárðarbunga and Grímsvötn the number of confirmed eruptions over intervals of 100 or 200 years is given while the eruption years are shown for other volcanoes (based on Thorarinsson, 1974; Larsen *et al.*, 1998 and Thordarson and Larsen, 2007). – *Eldgos sem kunnugt er um að hafi orðið í jökli hér á landi síðan um 1300. Byggt á ýmsum heimildum.*

thick ice in 31 hours, leading to a phreatomagmatic eruption that dispersed tephra over north Iceland in the first 24 hours. Subsequent tephra fall was confined to Vatnajökull. The subaerial eruption, however, was always a minor part of the activity since no more than 2–4% of the erupted material reached the surface of the glacier. The major part was always subglacial (Figure 8). Ice melting was very fast with the meltwater accumulating in the Grímsvötn caldera lake. The meltwater was released in a major 2-day long jökulhlaup about 3 weeks after the end of the eruption (e.g. Gudmundsson *et al.*, 1997, 2004; Björnsson *et al.*, 2001; Snorrason *et al.*, 2002).

The Gjálp eruption was the first of its kind to be monitored in any detail from repeated flights where airborne radar profiling played an important role in quantifying glacier response and ice melting (Gudmundsson *et al.*, 2004). To date, it is also the only fissure eruption under thick glacier forming a subglacial tindar to be observed. The eruption itself and the post-eruption evolution of the glacier at the eruption site has therefore provided important new data on volcano-ice interaction and post-eruption edifice evolution. The reader is referred to publications of various aspects of Gjálp for details (e.g. Gudmundsson *et al.*, 1997, 2002, 2004; Björnsson *et al.*, 2001; Jarosch *et al.*, 2008) and only a brief summary is given here.

The eruption confirmed that extremely fast heat transfer rate occurs in a vigorous subglacial eruption. Melting occurred under well-defined ice cauldrons over the erupting fissure and along the path of the meltwater, down-glacier of the eruption site. The average heat flux during the first three days of the eruption was 5–6x10^5 W m^{-2} indicating magma fragmentation as the dominant style of activity (Gudmundsson *et al.*, 2004), an observation supported by the low density of the edifice, estimated from gravity surveying (Gudmundsson *et al.*, 2002). A tindar, 6 km long and up to 500 m high was formed in the eruption which produced 0.45 km^3 DRE of basaltic andesite. No indications could be found that meltwater ponded at the eruption site. On the contrary, continuous subglacial drainage prevailed. However, supraglacial meltwater flow occurred within a 3.5 km long ice canyon above the subglacial edifice, but this meltwater entered the bed and flowed subglacially into Grímsvötn. The temperature of the meltwater as it flowed away from the eruption site was about 20°C. This high temperature, obtained by efficient mixing of the pyroclastic material with water and ice above the erupting fissure, should be an important factor in widening subglacial tunnels and a cause for the rapid rise in discharge of volcanogenic jökulhlaups (e.g. Jóhannesson, 2002; Björnsson, 1992). Water pressure at an erupting vent under a subsiding ice cauldron should be substantially less than suggested by considering only the static load of the ice and water. Shear stresses in the deforming ice carry part of the load. This reduced pressure may be an important factor in inducing magma fragmentation in subglacial eruptions and is one of the factors that makes subglacial eruptions different from submarine or subaqueous eruptions.

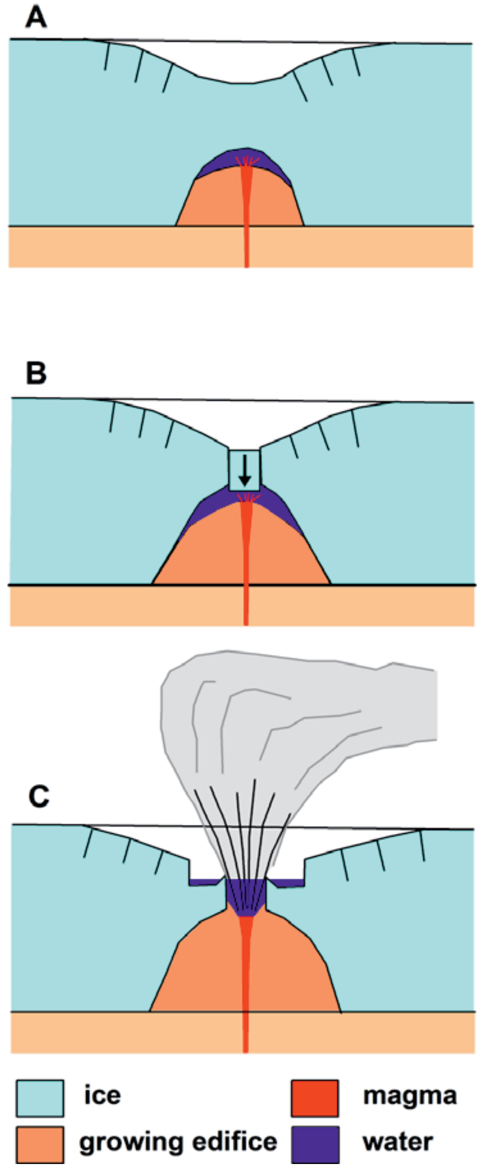

Figure 8. Cross section of the evolution of Gjálp in Vatnajökull, its emergence through the ice cover and formation of a supraglacial ice canyon. In A the eruption was fully subglacial, while in B a central piston of ice is collapsing onto the subglacial crater. C shows the subaerial crater stage, when the crater was a 100–200 m wide chimney through the ice. The eruption stopped after 13 days at stage C. Melting by water created the vertical cliffs bordering the ice canyon (after Gudmundsson *et al.*, 1997). – *Þversnið er sýna myndun Gjálpar í Vatnajökli í eldgosinu 1996. Gosið stóð aðeins í 13 daga; það var í byrjun undir jökli en braut sér síðan leið í gegn um jökulinn. Sigketill myndaðist kringum gossprunguna.*

Grímsvötn eruptions

Most Grímsvötn eruptions are relatively small (0.01–0.1 km^3 DRE, Thorarinsson, 1974; Gudmundsson and Björnsson, 1993; Gudmundsson, 2005) and large-scale fallout of tephra seldom occurs outside Vatnajökull. Most eruptions occur within the Grímsvötn caldera, along the southern caldera fault, where ice thickness is usually 50–200 m. Some of these eruptions seem to break through the ice cover almost instantly when magma reaches the glacier bed. This seems to have been the case with the 1934 and 1998 eruptions (Gudmundsson, 2005) while the 2004 eruption may have taken about an hour to melt its way through 150–200 m of ice in the southwest corner of the caldera (Figure 9). These eruptions melt several hundred meter wide openings in the ice cover where a phreatomagmatic crater develops. Some craters can develop on the surrounding ice and in many cases they do not survive as landforms at the glacier bed. These formations may be classified as irregular small tindars. Repeated eruptions built a succession of such formations. Retreating ice walls and collapse of the unstable craters leads to turbidity currents and accumulation of sediments at the bottom of the Grímsvötn lake.

Melting of ice in a typical caldera eruption in Grímsvötn is usually limited. For example, in each of the 1998 and 2004 eruptions, about 0.1 km^3 of ice were melted. Neither of these eruptions caused major jökulhlaups. Some of the eruptions in Grímsvötn (1934, 2004, several early 20th century and 19th century eruptions) are triggered by the unloading effects on the caldera floor of a falling lake level during draining of the lake in jökulhlaups (Thorarinsson, 1953; Gudmundsson, 2005).

Katla eruptions

The last visible eruption in Katla (October 1918) was a major event. A catastrophic jökulhlaup with a peak discharge of 250–300 thousand m^3/s took place on the sandur plain southeast of the volcano (Tómasson, 1996). The erupted products seem to have been only tephra, the majority of which was transported away from the craters with the meltwater. This tephra formed a 3 km long peninsula at the southern coast and deposited several meters on the sandur between the glacier and the coast. This is in sharp contrast with the Gjálp eruption where most of the erupted material remained at the eruption site. The removal of the erupted material at Katla in 1918 is probably the result of a combination of factors such as a very high discharge eruption leading to extremely large volumes of meltwater being generated, the relatively steep flow path and a sloping glacier bed at the eruption site.

In June 1955 and July 1999 ice cauldrons suddenly formed within the Katla caldera, in both cases leading to sudden jökulhlaups of moderate size (of order 2000 m^3/s). The cauldron formed in 1999 was about 50 m deep and 1.5 km wide. It is possible that cauldron formation in 1955 and 1999 was caused by minor subglacial eruptions. However, this is difficult to prove and unusual changes in geothermal activity cannot be ruled out as an alternative explanation (Gudmundsson *et al.*, 2007).

EDIFICE PALAGONITIZATION AND CONSOLIDATION

In Surtsey a network of minor intrusions formed, especially during the late stages of the eruption, leading to the onset of hydrothermal convection within the edifice (Jakobsson, 1978; Jakobsson and Moore, 1986). Research at Surtsey showed that the rate of the alteration was mainly controlled by temperature and the tephra had altered into consolidated palagonite tuff within 2–3 years when subjected to temperatures of 80–100°C.

The post-eruption evolution of Gjálp seems to conform to this (Jarosch *et al.* 2008). The depression in the ice surface was up to 60 km^2 in area, 8 km wide and 200 m deep at the end of the eruption. Despite inflow of ice and positive surface mass balance the influx of new ice was offset by basal melting. Hence, the depression volume changed little in the first five years after the eruption. Reduction in depression volume only started after 2001. Estimates of temperature of the edifice suggests that it was a liquid dominated geothermal system in the first few years, with a mean temperature of 240°C at the end of the eruption, dropping to 40°C in 2001. An apparent reduction in

Figure 9. The eruption in the southwest corner of the Grímsvötn caldera in Vatnajökull, on November 2, 2004. – *Eldgosið í suðvesturhorni Grímsvatnaöskjunnar 2. nóvember 2004.*

permeability suggests that palagonitization and consolidation of the edifice probably occurred in the first 1–2 years. No ice flow over the edifice, giving rise to glacial erosion has occurred in 1996–2006. Thus, it has been effectively shielded from erosion since ice flow has only acted to fill in the ice depression surrounding the edifice.

Field observations and petrological studies generally indicate that the hyaloclastites of the edifices of the subglacial volcanoes are altered and consolidated within short-lived geothermal systems. The alteration of the hyaloclastites of the outer slopes of the volcanoes is, however, dominantly a low temperature diagenetic process and probably develops over thousands of years (Jakobsson, 1978; Fisher and Schmincke, 1984).

Silicic volcanic glass is much more resistant to alteration than basaltic glass and Pleistocene silicic hyaloclastites are therefore commonly unconsolidated (Tuffen *et al.*, 2001, 2002; McGarvie *et al.*, 2006). These formations are, however, often altered by high temperature hydrothermal activity. Silicic formations are common in the Torfajökull area, south central Iceland and Kerlingarfjöll, SW of Hofsjökull, in central Iceland (Figure 1). Outside these two silicic centers, subglacial silicic rocks are a minor component of the volcanic zones (Jóhannesson and Saemundsson, 1998).

CONDITIONS DURING FORMATION OF SUBGLACIAL AND INTRAGLACIAL MOUNTAINS

The height of the lava cap on tindars and tuyas provides clues to the thickness of the ice cap during the time of eruption. The elevation of the passage zone (Table 1) determines the level of the englacial lake at the time of formation (e.g. Jones, 1969; Smellie, 2006). Walker (1965) constructed a profile of ice thickness across northern central Iceland extending to the northern coast and used the maximum elevation of the tuyas to define ice thickness. During the Gjálp eruption in 1996, the water level at the crater was 150–200 m lower than the original ice surface, much lower than the roughly 9/10 of the ice thickness required to float the ice. Thus, caution is needed when constructing palaeo-ice thickness on the basis of tuya height and considerable uncertainty exists in such reconstructions. A striking feature of many tuyas is the semi-constant level of the passage zone, indicating stable lake level for prolonged periods during the tuya-forming eruptions. However, observations in the WVZ and in Antarctica indicate variations and multiple levels in subglacial and intraglacial mountains (e.g. Jones, 1969; Smellie, 2006). At Hlöðufell (Figure 3), two lava caps occur, separated by hyaloclastite and flow-foot breccias, indicating a substantial rise in lake level during construction of these tuyas.

Passage zone height at hyaloclastite mountains in the WVZ is variable but ranges between 50 and 550 m above the surroundings. These mountains are considered to have formed during the last 2–3 glaciations (Jakobsson *et al.*, unpubl. results). Apparently the extent and thickness of the Icelandic ice sheet has varied considerably during glaciations. During the last glacial maximum 25 ka BP, the ice sheet covered all of Iceland and a large part of the shelf area around the island (Norddahl and Pétursson, 2005). The thickness of ice sheets is dependent on their size (e.g. Paterson, 1994). During the last glacial maximum the WVZ was covered by the ice sheet and located 120–150 km away from its margin. The corresponding ice thickness is 1.0–1.5 km. This is two to three times the maximum passage zone elevation observed in WVZ, suggesting that most of the hyaloclastite mountains were formed in eruptions when the ice sheet was considerably less extensive than during the last glacial maximum. The implications of this have not been explored in detail. However, models of the relationship between lithospheric loading by ice and partial melting in the mantle indicate a strong pulse in magma generation under Iceland at the end of the last glaciation (Jull and McKenzie, 1996). Studies of volcanic production rates indicate that such a pulse did occur at the beginning of the Holocene, especially through the formation of large lava shields (e.g. Maclennan *et al.*, 2002). Holocene lava shields and intraglacial tuyas occur in the same areas of the volcanic zones and a similar geographical association applies to Holocene volcanic fissures and tindars (Jakobsson and Johnson, 2008). If tuya eruptions are caused by de-loading effects under a decreasing ice sheet, it would provide an explanation for their generally larger volume in comparison with the tindars; magma supply would have been greater and more stable than at other periods.

CONCLUDING REMARKS

Although understanding of the main processes responsible for the formation of intraglacial mountains has existed for considerable time, and the eruptions in Surtsey and Gjálp in particular have thrown new light on important aspects of the behaviour of subglacial, intraglacial and subaqueous eruptions, many aspects remain unclear. Only a small fraction of the Móberg formation has been mapped in detail. The conditions leading to the formation of móberg sheets, whether mainly composed of pillow lava or hyaloclastite, have not been explained in a satisfactory manner. Climate change may lead to a drastic reduction in size and extent of Icelandic glaciers over the next 1–2 centuries. However, subglacial and intraglacial volcanic eruptions will remain a major part of Icelandic volcanism for some time to come.

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ÁGRIP

Jarðmyndanir sem orðið hafa til við eldgos undir jökli á Íslandi

Jarðmyndanir sem orðið hafa til við eldgos undir jökli eru algengar og áberandi hér á landi. Sá hluti þeirra sem myndast hefur á síðjökultíma, þ.e. fyrir 0,78 til 0,01 milljón árum, hefur verið nefndur Móbergsmyndunin (í þrengri merkingu þess orðs). Þessar jarðmyndanir þekja um 11.200 ferkílómetra. Um er

að ræða þrenns konar megin gosmyndanir, móbergshryggi, móbergsstapa og móbergslög. Grunneiningar þessara gosmyndana eru bólstraberg, gjóska og/eða móberg, óregluleg innskot og hraun. Auk þess eru setlög algeng, einkum við jaðra gosmyndananna. Töluverð reynsla hefur fengist af eldgosum undir jökli á seinni tímum, einkum í Vatnajökli. Mjög hröð kólnun gosefnanna skýrir myndun gjósku og bólstrabergs og ákafa bráðnun íss umhverfis gosstöðvar. Goshryggir hafa hlaðist upp, gjóska hefur dreifst víða og meiri háttar jökulhlaup orðið. Stapar hafa ekki myndast í þeim gosum sem upplýsingar eru um í Vatnajökli eða Kötlu. Meginþáttur gjósku er basaltgler. Það er yfirleitt ummyndað í palagónít í kjarna gosmyndunarinnar og móberg hefur myndast. Gögn frá Surtseyjargosinu 1963–1967 og Gjálpargosinu 1996 benda til þess að þetta sé afleiðing staðbundinnar jarðhitavirkni sem varir í nokkur ár eða áratugi eftir að gosi líkur. Jaðrar gosmyndananna eru yfirleitt minna ummyndaðir. Móbergshryggir og móbergsstapar geta gefið hugmynd um þykkt jökulsins þegar þeir mynduðust. Flest móbergsfjöllin eru mynduð í jökli sem var vel undir einum kílómetra að þykkt, mun þynnri en áætluð þykkt ísaldarjökulsins í hámarki á helstu móbergssvæðunum. Hugsanleg skýring á þessu gæti verið að stapparnir hafi einkum myndast í gosum í ísaldarjökli sem fór minnkandi. Við þær aðstæður lækkar þrýstingur í möttlinum undir landinu og meiri kvika verður til vegna hlutbráðnunar en þegar jökull er stöðugur eða stækkandi.

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